SILURIAN LANDS AND SEAS

Paleogeography Outside of Laurentia

edited by
Ed Landing
and
Markes E. Johnson

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…om Linnaeus' field notes made at the Upper Silurian section at Kyllaj, Gotland (compare with cover illus- *…rom M. Åsberg and W. T. Stern's translation "Linnaeus's Öland and Gotland Journey 1741" (1973, pub- …he Linnean Society of London by Academic Press).*

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Cover illustration: Coastal outcrops along the Baltic Sea in southeastern Gotland, southern Sweden. These sea stacks eroded in the Sundre beds (Upper Silurian, upper Ludlow) are similar to those described by Linnaeus during his 1741 visit to Kyllaj, northeastern Gotland (see title page).

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PREFACE

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This volume is the second installment of proceedings from an international symposium on the Silurian System, which was held at the University of Rochester in Rochester, New York, on August 4–9, 1996. Convened under the primary sponsorship of the Subcommission on Silurian Stratigraphy, (or SSS; a subdivision of the International Commission on Stratigraphy under the International Union of Geological Sciences), the Rochester conference attracted 75 participants from sixteen countries. These participants represented all of the continents where Silurian strata are extensively exposed. The first volume of the proceedings appeared in 1998, and included 21 articles on "Silurian cycles," most of which were originally presented as posters at the Rochester conference (Landing and Johnson, 1998).

The centerpiece of the 1996 meeting was the presentation of seventeen keynote lectures in a symposium titled "Silurian Lands and Shelf Margins." These lectures were designed to address the stratigraphic architecture and basic paleogeography of the principal continents affected by highstands in sea level during the Silurian. Fundamental to the organization of the sessions was the concept that good paleogeography cannot be accomplished outside a solid framework of biostratigraphy. Furthermore, global paleogeography cannot be undertaken without a reasonable system of global correlation.

Members of a special task force under the SSS laid the groundwork for the 1996 symposium by preparing a uniform graptolite zonal scheme that was especially designed for global paleogeographic studies (Koren' et al., 1996). The parent Commission on Stratigraphy sponsors subcommissions that specialize in each of the geological systems and periods. Thus, the 1996 Rochester conference was intended to demonstrate that the stratigraphic subcommissions are more than debate clubs where endless wrangling over series and epoch boundaries take place. Biostratigraphy, together with other means of stratigraphic correlation, has a highly practical application to paleogeographic reconstructions. The work of any stratigraphic subcommission is not finished until the hard field work and intellectual labor represented by such correlation schemes are put to the test of paleogeographic synthesis. Thus, the 1996 Rochester conference marked the first time that a subcommission marshaled its efforts to attempt something approaching global paleogeographic coverage.

This volume entails a collection of ten papers on stratigraphic correlations and paleogeography for the Silurian continents or microcontinents of Avalonia; Baltica; western, central, and southern Europe (Perunica); Siberia (two papers); Kazakhstan; China (with separate sections on distinct microcontinents), and eastern and western Gondwana. The latter mega-continent is represented by separate papers on Australia, India, and North Africa. South America is represented by the original abstract provided for the conference. The seven keynote addresses on Laurentia, or "ancestral" North America, are being developed for inclusion in a future volume. None of the keynote speakers were able to deliver manuscripts at the time of the conference; hence the present collection represents the continuation of considerable research after the Rochester conference, most of which culminated with a review process in 1998 and final submission of revised manuscripts in 1999. One of the original research teams that made a presentation in Rochester was unable to meet their commitment for a text, and a new team was constituted in 1999 to deliver a manuscript on the Baltica paleocontinent (northern Europe).

Any mention of the Silurian System naturally brings to mind the extraordinary efforts by Roderick I. Murchison (1792–1871), who established the system based on a type district in southern Wales and the Welsh Borderland (Murchison, 1839). The First International Symposium on the Silurian System was held in 1989 at the University of Keele, Staffordshire, U.K., to celebrate the 150th anniversary of that important reference work. The Second International Symposium on the Silurian System (1996) in Rochester, New York, celebrated both the life and career of James Hall (1811–1898). The meeting marked the 150th anniversary of the submission to the governor of New York State of Volume One of Hall's encyclopedic *Palaeontology of New York.* Hall's *Palaeontology* subsequently expanded year by year until it reached a monumental size of thirteen quarto volumes, with thousands of pages of taxonomic descriptions and hundreds of illustrated plates. It was fitting that the papers presented at Rochester on Silurian cycles were published on the centennial of Hall's death (Landing and Johnson, 1998).

Both Hall and Murchison are rightfully identified with major labors that were undertaken in separate regions. However, both Hall and Murchison made their field areas on both sides of the Atlantic famous for their fossil content, and both used the fossils in biostratigraphic correlation. It is not generally recognized, however, that Hall carried out extensive studies beyond New York State and that Murchison did the same outside the British Isles.

Although the Silurian System is associated by name with Wales and the Welsh Borderland, Silurian strata are well represented in Scandinavia and the Baltic States, where Murchison eventually conducted extensive field studies. Outside of the British Isles, it may be said that Scandinavia has the longest history of studies on Silurian strata and fossils. The first tabulation of fossils from the "Upper Silurian" strata of Scandinavia and the Baltic States (Murchison, 1845, table 2) preceded comparable work in England and America. Of course, Murchison's "Lower Silurian" is now the Ordovician System. In that work, Murchison (1845, p. 11) revealed that the Rev. William Bilton had repeatedly urged him to visit Norway. Bilton was an English clergyman who, in 1840, authored a fishing guide titled *Two Summers in Norway*. He was also an amateur geologist who enthusiastically practiced his creed that "the hammer often accompanies the fishing rod." It was Bilton (1840, v. 2, p. 180) who made the first printed reference to the Silurian System of Norway:

> "The numerous fossils contained in these deposits prove that they belong to the ancient group of rocks, lately named and admirably described by Mr. Murchison, as Silurian. But it remains still to be shown whether the order of succession observed in South Wales obtains equally in the Norwegian beds. . . . It is hoped that ere long, the Author of the Silurian System will himself visit this interesting locality, and decide the question by an examination of the rocks and their contents on the spot."

Well before he first traveled to Norway in 1844, however, Murchison was aware of the presence of Silurian strata on the Swedish island of Gotland in the Baltic Sea. The earliest descriptions of fossils from Gotland were made by the great Swedish systemicist, Carolus Linnaeus (1707–1778), who toured the islands of Öland and Gotland in 1741 and kept a journal of his observations. A facsimile of a page from his original Öland and Gotland journal, together with a full English-language translation and commentary by Åsberg and Stearn (1973), was published by the Linnean Society of London. The following passage records the naturalist's discovery of fossil corals on Gotland (Åsberg and Stearn translation, 1973, p. 122):

"'Coral shore' [is what] I call that which lies to the eastern side of Kappelshamn, which was very broad and covered with white and grey stones; this greatly surprised us, since each one stone was nothing but a coral, called *Madreporae*, so that anyone who wants exquisite corals for collections of natural history specimens need try no other place but this; every man in the world could probably get a cartload of his own of corals here. . . . The *Madreporae* lying close to the water were clean and clear, adorned with stars like the backs of playing cards or the cells of a honeycomb."

The village of Kappelshamn is located in a deep bay at the northern end of the island of Gotland. Most of the shoreline in the bay exposes strata belonging to the Högklint Beds, regarded as Lower Wenlock in position. Farther along on his journey, Linnaeus visited a raised area of limestone sea stacks at Kyllaj, located some 12 km southeast of Kappelshamnsviken. These rocks belong to the Slite Beds, also Wenlock in position but laying above the Högklint Beds. Linnaeus wrote the following description in his journal (Åsberg and Stearn translation, 1973, p. 1136):

> "We called the 'Stone Giants,' as had the learned bishop G. Wallin, that which we saw by the sea near to Kyllej: between the customs officer's house and the lime kiln of Kyllej there was a slope towards the sea where stood many high and thick limestone rocks 4–6 fathoms [24–36 feet] high, arranged in a row like the ruins of churches or castles, of which those standing at a lower level of the slope were taller than those higher up, so that the heads were all at the same level. From a distance they looked like statues, horses, torsos and I do not know what kind of ghosts. Evidently, this had been formerly a limestone mountain, the roots of which had been ground, cut and formed by the heaving waves of the sea, till it finally left those stones in their present form."

Linnaeus' sketch of the Kyllej coast was rendered in the margin of his journal (see half-title page). Linnaeus observed that similar sea stacks occur along the shore as far as the village of Slite. In fact, similar sea stacks also occur on the southeastern end of Gotland, where they are eroded in the Sundre Beds (upper Ludlow). A photo of the Sundre sea stacks is printed on the front cover of this volume, in tribute to the long history of Silurian studies in Scandinavia.

The 103 years that lapsed between the visit of Linnaeus to Gotland and the arrival of Murchison in Norway bracketed a dramatic change in the sophistication of knowledge involving geological history. During their much-publicized travels in Russia, Murchison et al. (1845) saw no more "Upper" Silurian strata after he left the territory of Estonia. The paper by Baarli et al. (this volume) brings together the most complete analysis of Silurian strata on the paleocontinent of Baltica, and includes extensive areas of northwestern Russia. The editors are grateful to her for taking on the coordination of this large work as a substitute for previously promised materials.

All the authors who contributed to this volume are experts on the Silurian geology and paleontology of their assigned regions. Except for North America, which will be treated separately, the present volume is the most complete and current treatment of Silurian paleogeography of its kind and should provide a standard reference work for many years to come.

ACKNOWLEDGMENTS

We wish to thank the many reviewers of the manuscripts submitted as part of this volume, the authors for their patience shown while this volume was in preparation, and Jeanne C. Finley for the final editing of this volume.

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ABBREVIATIONS

In many of the figures used in this publication, biostratigraphic zones and intervals named for conodonts and graptolites are abbreviated, and only the trivial (species) name is recorded. As detailed by the North American Commission on Stratigraphic Nomenclature (1983, article 54.1), trivial names are not unique, and properly require the genus name in a binomial or trinomial combination. For this reason, we supply the current (2002) binomial names for all of the biostratigraphic zones used in this publication. The listing of the trivial (species) names used is alphabetical. The binomial names used for graptolites correspond to Melchin et alii's (1998, appendix 1) usage.

GRAPTOLITES

acuminatus; Parakidograptus acuminatus
argenteus; Monograptus argenteus
bohemicus tenuis-kozlowskii; Bohemograptus bohemicus tenuis-Nucucullograptus kozlowskii
bouceki-transgrediens; Monograptus bouceki-M.? transgrediens
branikensis-lochovensis; Monograptus branikensis-M. lochovensis
centrifugus-murchisoni; Cyrtograptus centrifugus-C. murchisoni
convolutus; Monograptus convolutus
cyphus; Parakidograptus (Coronograptus) cyphus
formosus; Monograptus (Formosograptus) formosus
fritschi linearis; Saetograptus fritschi linearis
griestoniensis-crenulata; Monoclimacis griestonensis-M. crenulata
guerichi; Spirograptus guerichi
lapworthi-insectus; Cyrtograptus lapworthi-C. insectus
leintwardinensis; Saetograptus leintwardinensis
leudensis; Colonograptus ludensis
lundgreni; Cyrtograptus lundgreni
nilssoni; Neodiversograptus nilssoni
parultimus-ultimus; Monograptus parultimus-M. ultimus
parvus-nassa; Pristiograptus parvus-Gothograptus nassa
praedeubelli-deubelli; Colonograptus praedeubelli-C. deubelli
progenitor; Lobograptus progenitor
riccartonensis-belophorus; Monograptus riccartonensis-M. belophorus
rigidus-perneri; Cyrtograptus rigidus-C. perneri
scanicus; Lobograptus scanicus
sedgwickii; Stimulograptus sedgwickii
spiralis; Oktavites spiralis
triangulatus-pectinatus; Demirastrites triangulatus-D. pectinatus
turriculatus-crispus; Spirograptus turriculatus-Monograptus crispus
ultimus; Monograptus? ultimus
vesiculosus; Cystograptus vesiculosus

CONODONTS

amorphognathoides; Pterospathodus amorphognathoides
bohemicus; Ozarkodina bohemicus
celloni; Pterospathodus celloni
crispa; Ozarkodina crispa
eosteinhornensis; Ozarkodina eosteinhornensis
kentuckyensis; Distamodus kentuckyensis
obesus; Spathognathodus obesus
parahassi-guizhouensis; Spathognathodus parahassi-S. guizhouensis
ploeckensis; Ancoradella ploeckensis
ranuliformis; Kockelella ranuliformis
remscheidensis-eosteinhornensis; Ozarkodina remscheidensis-O. eosteinhornensis
sagitta rhenana; Ozarkodina sagitta rhenana
sagita sagita; Ozarkodina sagitta sagitta
siluricus; Polygnathoides siluricus
snajdri; Ozarkodina snajdri
stauros; Kockelella stauros
tenuis-staurognathoides; Pterospathodus? tenuis-Distomodus staurognathoides
variabilis; Kockella variabilis

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PART I: THE BALTIC AND AVALON

Silurian Stratigraphy and Paleogeography of Baltica

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ABSTRACT — The Silurian continent of Baltica was a small- to-medium-sized craton (ca. 7.8 million km^2). Silurian outcrops cover only 1% of the area of this former continent, although the extent of subsurface Silurian strata is appreciable. Stratigraphic patterns influenced by tectonic and eustatic cycles are documented from 25 lithologic and bathymetric profiles drawn from four principal areas in the Central Scandinavian, East Baltic, Dniester, and Timan–Pechora Depressions. Baltica was bounded on its west flank by the developing Scandian orogen, which resulted from the closure of the Iapetus Ocean. On its southwest flank, Baltica underwent a collision with the microcontinent of Avalonia. The remaining continental margin was passive. Base maps for central Scandinavia, the combined East Baltic region and Dniester River area of Podolia, and the Timan–Pechora region of northwestern Russia allow reconstruction of the Silurian across approximately 35% of Baltica and its shelf margins. Paleogeographic maps are provided for Baltica through the Llandovery, Wenlock, and Ludlow Series, but exclude the youngest Pridoli Series due to insufficient or unavailable data. Primary control of continental and marine sedimentation was exerted by tectonic factors, but the secondary imprint of eustasy is preserved in the stratigraphic record of Baltica's mid-shelf sectors.

Introduction

The earliest general purview of Scandinavia with adjacent areas as part of a continent coincided with the development of the plate tectonic theory (Wilson, 1966). The first use of the name "Baltica" was probably on a map by McKerrow and Ziegler (1972) that depicts a triangular-shaped paleocontinent bounded on three sides by the Proto-Atlantic (Iapetus), Rheic, and Pleionic Oceans. A subsequent treatment by Ziegler et al. (1977) adopted the same orientation for Baltica in the context of a global reconstruction of Silurian continents in which the plate boundaries were more rigorously defined. Baltica was one of about ten Silurian continents and microcontinents. It was an order of magnitude smaller than the supercontinent of Gondwana, but approximately two-thirds the size of neighboring Laurentia, the next-largest continent.

Silurian Baltica was a small- to medium-sized continent approximately 7.8 million km^2 in area. Its Precambrian basement exposures have received various regional names. The Fennoscandian Shield takes its name from Finland and Scandinavia. Another region of highly metamorphosed Precambrian rock exposure on this continent is called the Sarmantian Shield, from the classical name for the region between the Vistula and Volga Rivers in Poland and adjacent areas of the Ukraine and Russia. Precambrian crystalline basement rocks crop out widely in southern Norway, Sweden, and Finland, as well as in the Ukraine. A single name, the Fennosarmatian Craton, is preferred by Nikishin et al. (1996). "Baltica," however, has persisted as the name for the lands and marine shelves of a Precambrian–Silurian continent inclusive of and peripheral to the present-day Baltic Sea in northern Europe.

Cocks and Fortey (1998) discussed the margins of Baltica. We follow their interpretations, with the exception of the margin on the far northeast. Cocks and Fortey (1998) included parts of Taymyr and the islands of Pai-Khoi and Severnaya Zemlya in Baltica. Zonenshain et al. (1990) referred these areas to one separate paleoplate. As long as the paleogeographic position of these areas is controversial, we choose to exclude them from Baltica.

Earlier attempts at regional syntheses of the Silurian paleogeography of Baltica were focused mainly on the eastern Baltic states of Estonia, Latvia, and Lithuania and the adjacent Swedish island of Gotland (Kaljo and Jürgenson, 1977; Kaljo and Rubel, 1982; Bassett et al., 1989). In addition, two maps by Einasto et al. (1986) for the early and late Wenlock show a broader paleogeographic reconstruction that reaches from the East Baltic region to Podolia in the southern Ukraine. Less emphasis has been placed on the Oslo Region of southern Norway (Worsley et al., 1983).

The goals of this report are to: 1) provide an updated tectonic review of the borderlands that defined Baltica as a distinct Silurian paleocontinent, 2) summarize new data on the stratigraphy and paleoecology of the Central Scandinavian and Timan-Petchora Basins and allow comparison with the East Baltic Basin, and 3) critically assess the influence of eustasy on cratonic marine deposits. Our treatment does not include the Peri-Caspian Basin to the southeast, which is still poorly understood from subsurface borings. During Silurian time, the microcontinent of Avalonia probably was conjoined with Baltica. By design, our coverage of Baltica excludes the paleogeography of Avalonia, which is treated separately by Cocks et al. (this volume).

Location of Baltica and Tectonic Constraints

By the start of the Silurian, Baltica reached low equatorial latitudes and had approximately the same orientation it has today, after drifting northward from more southerly latitudes and rotating counterclockwise through the Ordovician (Torsvik et al., 1996). On Baltica's present southwestern margin, the Tornquist Sea between Baltica and eastern Avalonia closed toward the end of the Ordovician, and eastern Avalonia may have docked with Baltica in the latest Ordovician (Meissner et al., 1994). Later dates and alternative models for the final closure that spans the Silurian to Early Devonian were proposed by Oliver et al. (1993), Tanner and Meissner (1996), and McCann (1998), among others. The combined continents, referred to as Balonia by Torsvik et al. (1993), eventually collided with the Laurentian continent as the Iapetus Ocean closed in a scissors motion between these continents. The main collision between Baltica and Laurentia occurred along the margins of Greenland–Scotland and western Norway, and initiated the Scandian phase of the Caledonian orogen during the Early Silurian. The Caledonian deformation front is traced into the western Barents Sea and east of Spitsbergen (Gee and Page, 1994). A growing database indicates this was not a simple continent-to-continent collision, but probably involved Early and Late Ordovician orogenic events (Andersen et al., 1998), the main Silurian event, and a conclusion in the Early Devonian.

To the north in the eastern Barents Sea area, there appears to have been a broad depositional basin situated between the Timan Ridge and the northern Ural orogen, which is called the Timan–Petchora Basin. According to Torsvik et al. (1995), this area was separated from an inverted Siberian continent by a narrow, closing ocean. The eastern margin of Baltica was passive, as seen in a belt from the Timan–Petchora Basin to Novaya Zemlya in the northeast (Nikishin et al., 1996; Malyshev, 2000). The exact position of the margin along the Urals is not known, but hydrothermal vent faunas from the south-central Urals prove the presence of an ocean, and its margin must lie west of these faunas (Cocks, 2000). The Lower Paleozoic is known only from a few deep wells in the Peri-Caspian Basin to the southeast, but that too was part of the passive margin during the Silurian (Nikishin et al., 1996). The Dniester Depression in the southern Ukraine, however, is well explored (Koren' et al., 1989). Both the Peri-Caspian and Dniester Depressions are the results of post-rift subsidence (Nikishin et al., 1996).

Thus the Early Silurian was a very active tectonic interval along the western and southwestern margins of Baltica, and this is well indicated by the sedimentary record. There are five major basins where the Silurian is preserved: the Central Scandinavian, East Baltic, Timan–Pechora, Dniester, and Peri-Caspian Depressions (Fig. 1). A Middle Silurian tongue that extends across the Baltic platform into the Moscow Basin (Einasto et al., 1986) constitutes the only other area with preserved Silurian rocks with Baltic character. This bifurcation of Baltica by a southeasterly-trending basin lends support to the retention of two different names for the continent's cratonic shields.

In this paper, four basins are reconstructed on the basis of their more widespread outcrop, superior record, and accessibility of data. The Central Scandinavian and East Baltic Basins are close geographically, but are isolated on the opposite side of the Fennoscandian Shield from the Timan–Pechora Basin. Similarly, the Dniester and Peri-Caspian Depressions are isolated from all the other basins on the southern flanks of the Sarmantian Shield (Fig. 1).

Methodology

The paleogeographic approach of this report starts with stratigraphic columns that relate lithology to changes in bathymetry and community composition through time. We recognize benthic assemblage zones 0–6 of Boucot (1975), which are widely used in the analysis of global Silurian fossil associations or communities (Boucot and Lawson, 1999). According to this system, 0 represents land; BA 1 corresponds to near-shore, shoal, or lagoonal facies; BA 2 is the fore-shoal to inner shallow shelf; BA 3 and BA 4 relate to mid-shelf positions; and BA 5 and BA

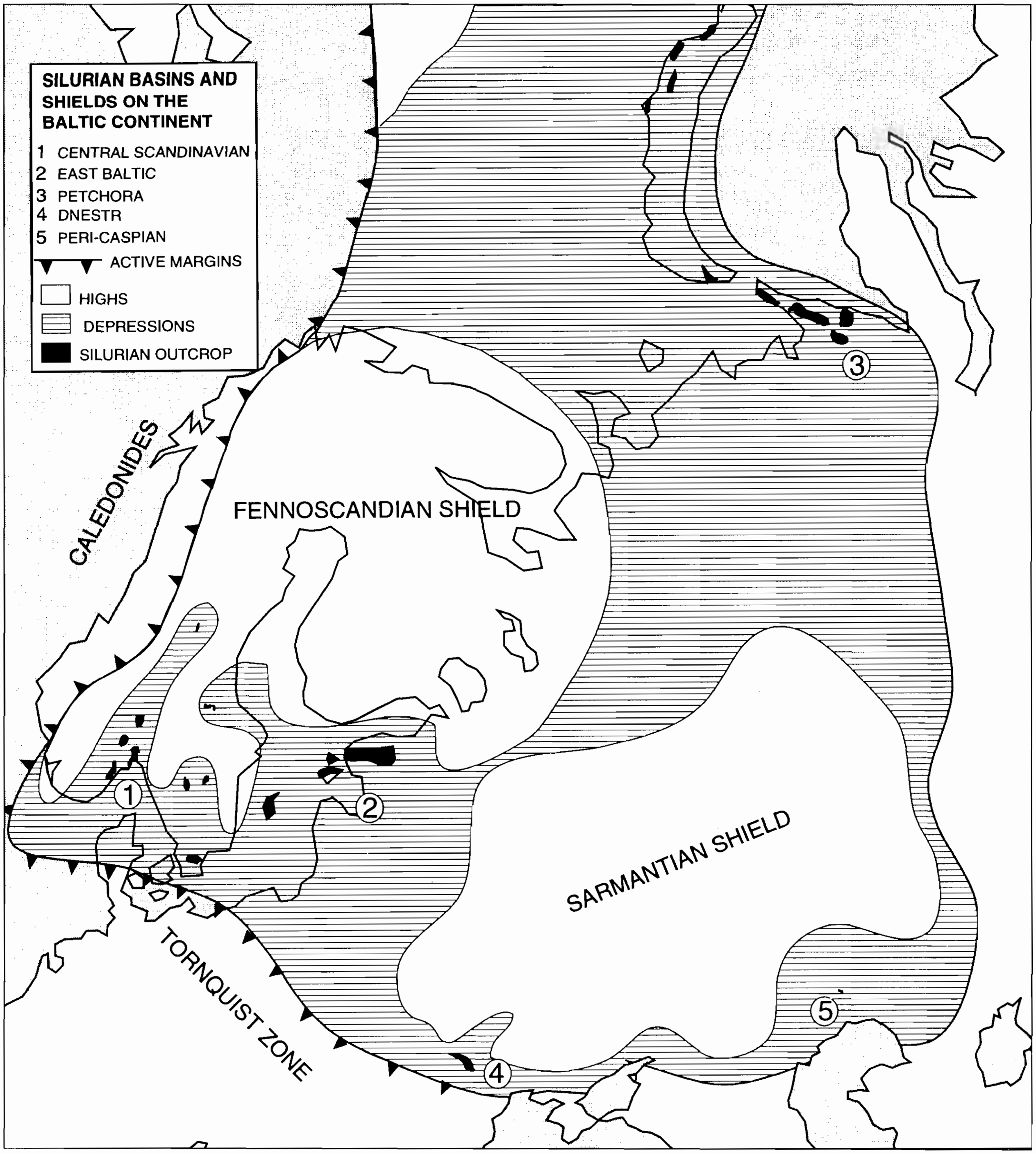

FIGURE 1 — Silurian Baltica showing principal shield areas and basins.

6 equate with the deeper parts of the outer shelf. The bathymetric model assigns normal (fair-weather) wave base to the boundary between BA 2 and BA 3, and generally fixes maximum storm base in the lower middle shelf in BA 4. Application of this model to Baltica is based on two lines of evidence that are predicated on physical and biological grounds. Proximality trend analysis assumes that the sedimentological fabric of the marine shelf responds to distance from shore and to water depth, as related to the effects of storms. As tested in the Oslo

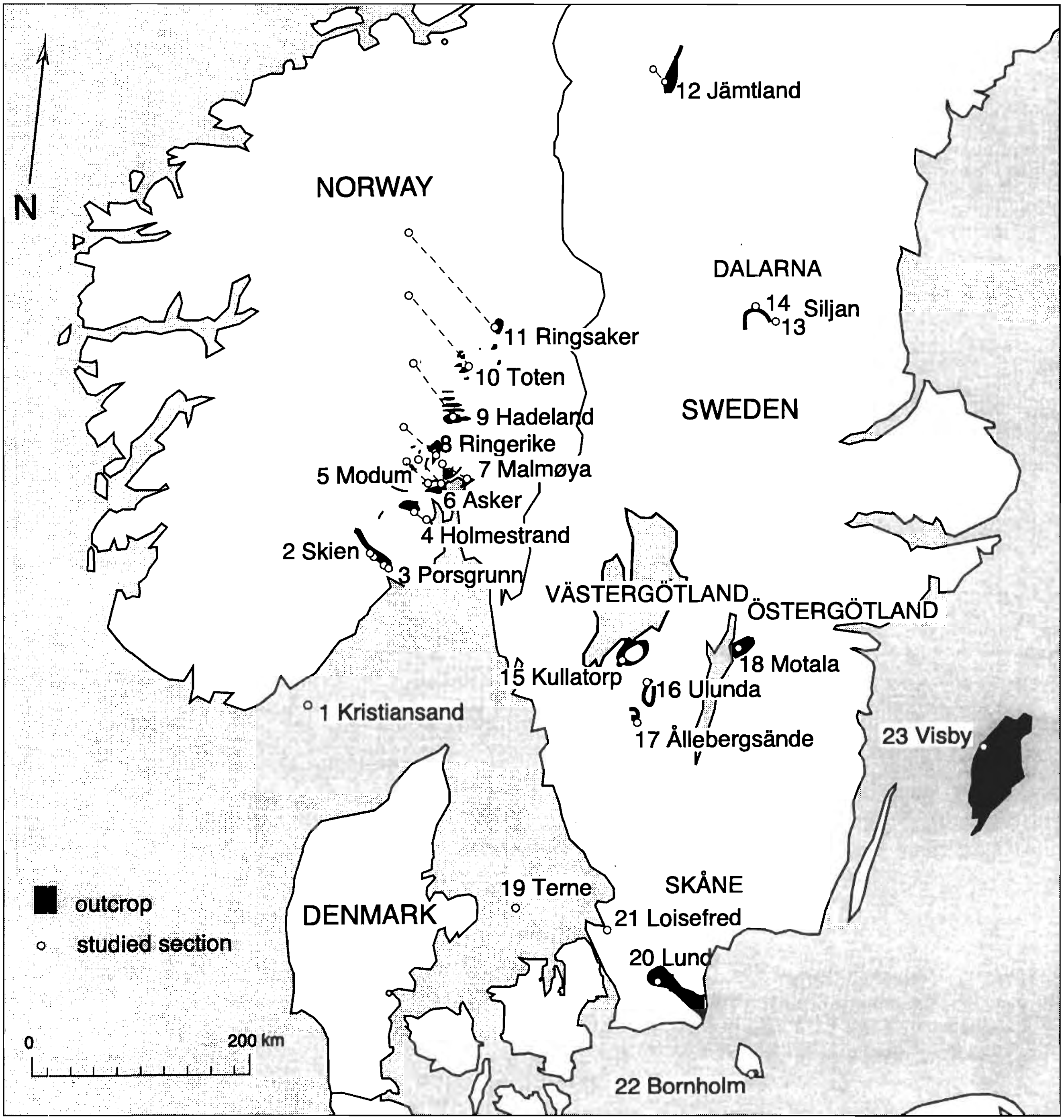

FIGURE 2 — Silurian outcrops (in black) of the Central Scandinavian Basin.

Region (Baarli, 1988), proximality trends are found to be coordinated with a wide range of well-defined marine paleocommunities (Baarli et al., 1999). Similarly, a study of shelly tempestites was made in southern Norway (Johnson, 1989), in which *Pentamerus oblongus* is interpreted to have lived in benthic environments in a range of water depths affected by storm turbulence. Persistent wave energy at the shallower end of the spectrum (BA 2) frequently scattered and fragmented shells of this Silurian brachiopod and generally prevented maximum growth. As a result of sediment deposition and limited physical agitation at the deeper end of the bathymetric spectrum (BA 3 to upper BA 4), populations of large-shelled *Pentamerus oblongus* were more typically suffocated and preserved in life position. It is our working premise that similar conditions prevailed throughout the flooded shelves of Baltica.

Paleogeographic maps are formulated through inter-regional comparison of correlated horizons at specific localities with relatively complete stratigraphic sections. This report provides only limited space for reporting the sampling through fully illustrated profiles. Many more auxiliary sections were utilized in the construction of the paleogeographic maps, the locations of which are indicated by open dots as opposed to filled dots for illustrated sections (Figs. 2, 7, and 11). The maximum number of time-successive maps generated for any one region is eight. For the most part, the designated time planes from the Llandovery Series correspond to the *Coronograptus cyphus, Stimulograptus sedgwickii, Spirograptus turriculatus,* and *Cyrtograptus lapworthi* Zones. Those adopted from the Wenlock Series correlate with the *Monograptus belophorus* and *Cyrtograptus lundgreni* Zones. Maps correlated with the base and top of the Ludlow Series correspond to the Neodiversograptus nilssoni and Monograptus formosus Zones. Our reference to these zones conforms to the generalized graptolite zonal sequence for the Silurian of Koren' et al. (1996).

Central Scandinavian Basin

The shallow Early Silurian basins in the Baltoscandinavian region (Fig. 2) are located along an old Middle Proterozoic rift system that stretches from the East European Craton through Baltoscandia to Greenland (Van Balen and Heeremans, 1998). Along this system lay the East Baltic Depression, with intracratonic basins that extend into the Sea of Bothnia, and the Central Scandinavian Basin with narrow basins like the Colonus Trough in Skåne and the Oslo graben. Closure of the Tornquist Sea between eastern Avalonia and Baltica probably occurred by subduction of the ocean floor beneath the Avalonian plate, while the southwestern edge of Baltica was passive (Mona Lisa Working Group, 1997). During Late Silurian time, increased deposition rates are reflected by up to 2.6 km of strata in the Danish Basin (Michelsen and Nielsen, 1991) and synsedimentary tectonism. These indicate development of a foredeep as the eastern part of the Avalonian plate thrust over the southern edge of Baltica (Mona Lisa Working Group, 1997). The source for these shaly sequences may have been mountains that rose in the south at the onset of the collision between Baltica and eastern Avalonia (Erlström et al., 1997). The Upper Silurian sandstones that succeed those shales may be related to the renewal of Caledonian thrusting in the south.

The western margin of Baltica extended at least 400 km west of the present thrust front of the orogen (Stephens and Gee, 1989). The edges of the continent, as shown in Fig. 1, are not the original ones, but are margins defined by later tectonic events. A Late Ordovician phase of orogenic activity may have led to uplift along this margin. Later erosion of these uplands provided a westerly source for Early Silurian siliciclastics now in the Oslo graben and the developing Jämtland Basin. During the Early Silurian, there also seems to have been an easterly source of sediments for parts of the Oslo Region (Braithwaite et al., 1995). Manten (1971, p. 25–27) and Bassett et al. (1989) inferred exposed land northwest of Gotland from the early Wenlock onward. According to them, this accounts for the lack of Wenlock and Ludlow sections in central Sweden and the intermittent influx of siliciclastic sediment into western Gotland. In their analysis of sedimentary rocks from the intracratonic basin preserved in the Sea of Bothnia, Van Balen and Heeremans (1998) proposed that the bulls eye-shaped Bothnian Basin was formerly twice as wide as today. If so, the exposed area must have been much narrower than suggested by Bassett et al. (1989, figs. 121–122). Thus the Early Silurian Baltoscandinavian areas featured an uneven topography with several small intracratonic basins or shallow depressions.

In connection with the onset of the Scandian phase of the Caledonian orogeny, southeasterly-directed nappe displacement led to development of foreland basins (Baarli, 1990a) and later deposition of continental sediments conformably above the marine deposits. The effect of this orogeny was first manifested in the late Llandovery to the northwest in Jämtland, and reached Skåne in southern Sweden by latest Ludlow to early Pridoli time (Bassett, 1985). Allochthonous and parautochthonous Silurian deposits are found within the Scandinavian Caledonides and along the thrust front (Stephens and Gee, 1989). Of these, only the parautochthonous deposits are considered herein, along with the autochthonous deposits. The parautochtohnous deposits have been subject to palinspatic reconstruction (Fig. 2), following Cocks and Worsley (1993).

Oslo Region and Jämtland — Detailed stratigraphic columns are provided from ten section localities in southern Norway and Sweden (Figs. 3–5). The primary sources of information on the Oslo Region are from the classic treatment by Kiær (1908) and the overview of Worsley et al. (1983). The latter's contribution revised the region's stratigraphic nomenclature into a form compatible with modern usage. The Llandovery succession of the Oslo Region was updated by Baarli and Johnson (1988), Worsley (1989), Baarli (1990b), and Johnson et al. (1991),

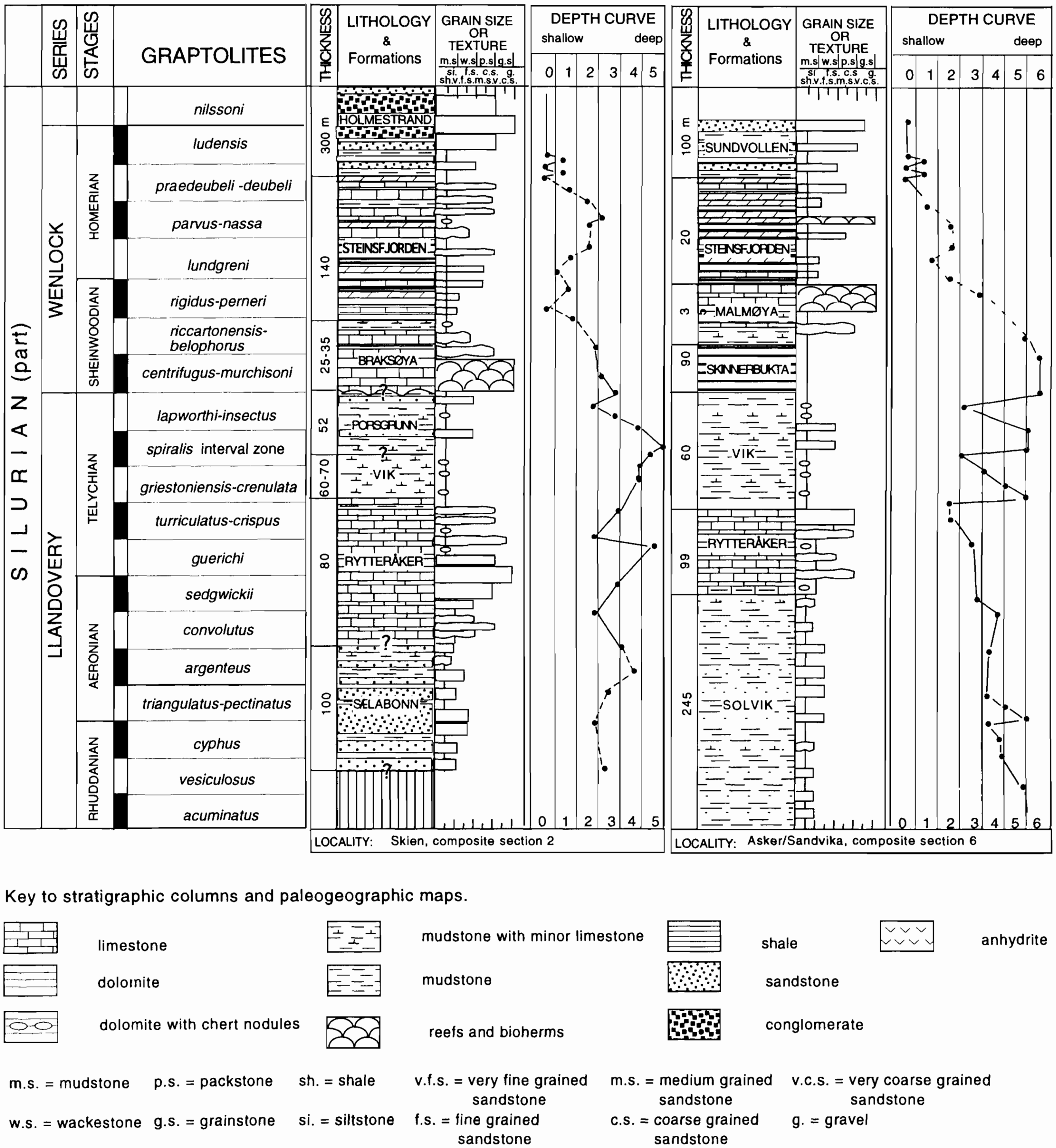

FIGURE 3 — Silurian stratigraphy of localities 2, 6, 7, and 8 in the Central Scandinavian Basin (see Fig. 2). Lithologic key applies to all stratigraphic sections in this report.

among others. Cocks and Worsley (1993) further refined the Telychian succession and expanded the stratigraphic and paleoecologic overview of the region through the lower Wenlock. Heath and Owen (1991) and Braithwaite et al. (1995) revised the Hadeland section (Fig. 4). The youngest strata in the Silurian of the Oslo Region are not well investigated, with the exception of a few local studies by Turner and Whitaker (1976), Olaussen (1985), and

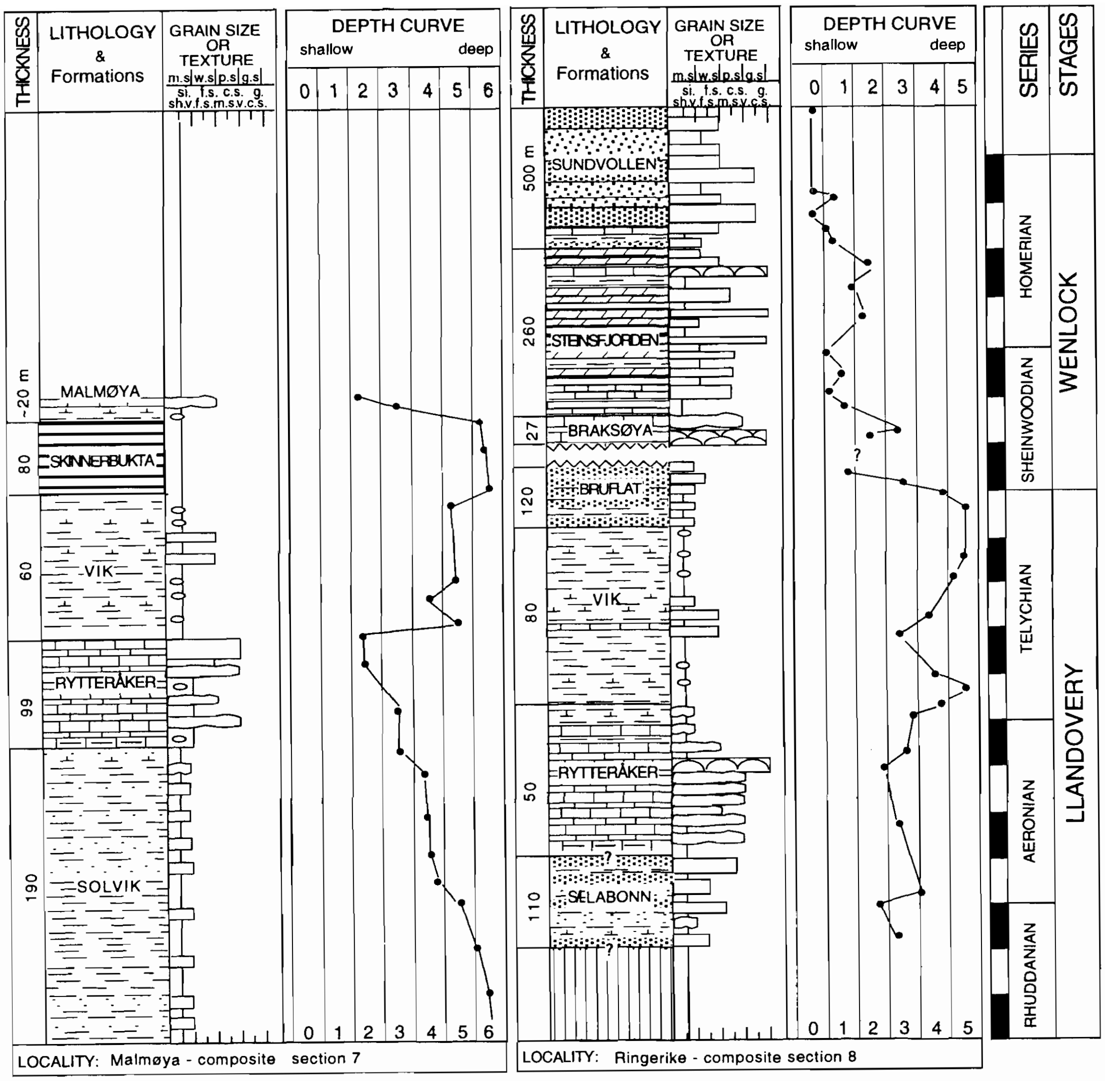

FIGURE 3 *continued.*

Dam and Andreasen (1990). The latter demonstrated a western source for the youngest Silurian marine and continental deposits of the central Oslo Region, as opposed to the eastern source for the central area proposed in earlier studies (e.g., Turner and Whitaker, 1976; Worsley et al., 1983). An overview of Silurian benthic marine communities of the Oslo Region was conducted by Baarli et al. (1999).

The locations of all localities are marked in Fig. 2, and are distinguished as primary and auxiliary sections in Figure 6a–h. Sources other than those mentioned above for the auxiliary sections can be briefly summarized. Smelror et al. (1997) described an offshore stratigraphic core (location 1) drilled outside Kristiansand with Rhuddanian and Aeronian deposits. Locations 3 and 4 (Fig. 2) entail general information from Kiær (1908), Worsley et al. (1983), and Cocks and Worsley (1993) that covers the Silurian marine sequence to possible Ludlow continental deposits. The lower Llandovery to the *Monograptus crispus* Zone was described by Baarli (1988) for location 5, at Modum (Fig. 2). Location 11 (Fig. 2) at Ringsaker is not well known, but is included in general reports treating the entire Oslo Region. These reports include Möller (1989) on the upper Aeronian Rytteråker Formation, while Cocks and Worsley (1993) provide additional information on the succeeding Telychian.

The Jämtland section (Fig. 4) from west-central Sweden clearly is not traditionally part of the Oslo Region, but includes parautochthonous nappes along the Caledonian mountain range much further north. When sections in the Oslo Region are palinspatically restored (Fig. 2), they lie along a trend that includes Jämtland. In this

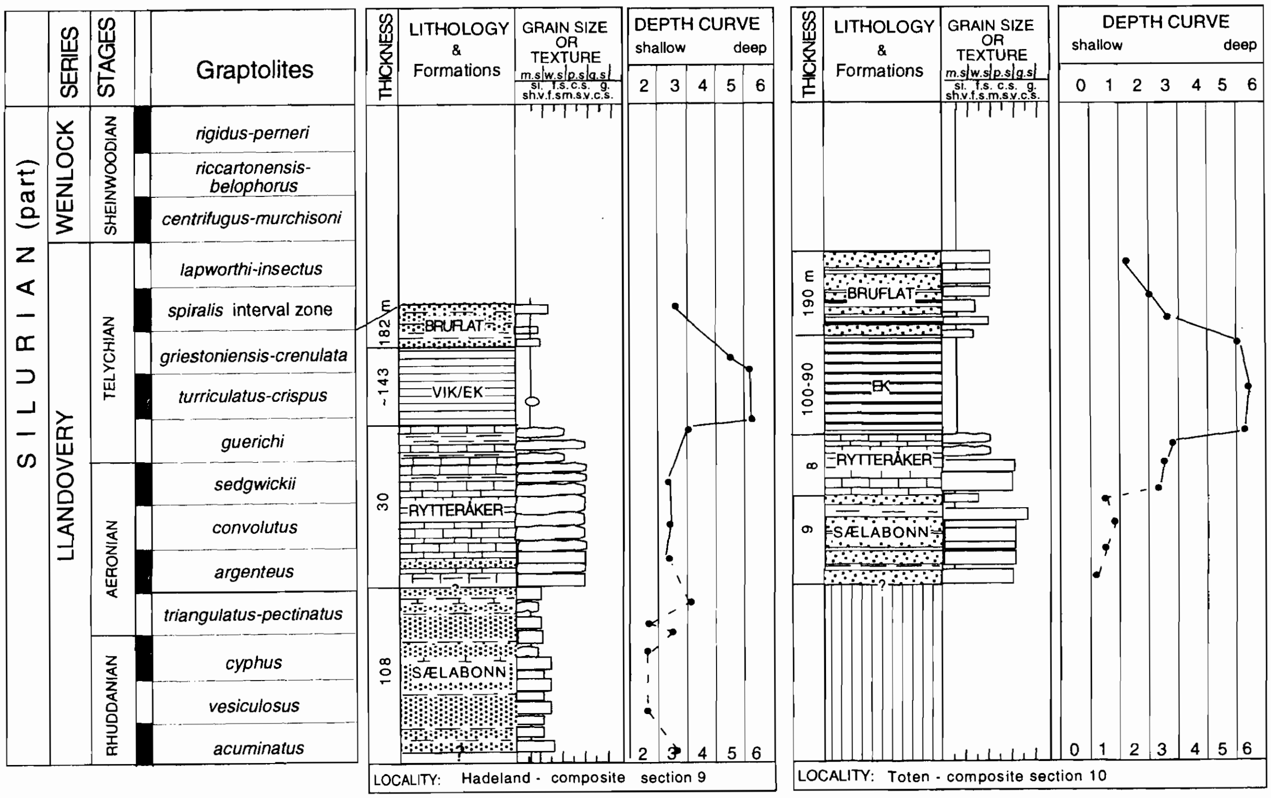

FIGURE 4 — Silurian stratigraphy of localities 9, 10, 12, and 15 in the Central Scandinavian Basin (see Fig. 2).

report, Jämtland has a position in relation to the Caledonian front that is similar to the Ringsaker, Toten, and Hadeland areas of Norway. Jämtland experienced greater post-depositional folding and metamorphism than the other areas. This, in addition to an extensive cover of till, makes stratigraphic data difficult to compile. The Upper Ordovician and Ordovician–Silurian boundary sections are described by Cherns and Karis (1995). Additional information is from Thorslund and Jaanusson (1960), Bassett et al. (1982), Strömberg (1986), and Grahn (1998). There are fossiliferous Silurian shallow-water sediments in several other places in the Caledonides of Norway and Sweden, such as the Ofoten–Troms area of northern Norway. These were deposited in basins developed off the Baltic platform and emplaced onto the platform during the final Caledonian thrust phase (Andresen and Steltenpohl, 1994).

Central to southern Sweden and Denmark — The stratigraphic information from central to southern Sweden and Denmark (Figs. 4 and 5) is much less detailed than that from the Oslo Region. The chitinozoan biostratigraphy of the Llandovery to middle Wenlock of mainland Sweden was documented by Grahn (1998). However, our estimations of sea-level changes interpreted from the limited lithological and paleontological information are subject to emendation. There are four main areas with Silurian exposures in central and southern Sweden. The first, in the Skåne area of southern Sweden, is included with the Silurian sections from the nearby Danish island of Bornholm. The Silurian sections of Västergötland and Östergötland in the area near Lakes Vättern and Vänern (Fig. 2) lie to the north. Farther north are the Silurian sequences in Dalarna Province of central Sweden. Gotland is treated as part of the East Baltic Basin (described below), but is included on the paleogeographic maps (Figs. 6a–h).

With the exception of Gotland, the areas where Silurian strata are best known from southern Scandinavia are in Skåne and Bornholm. There are several review articles on Skåne, but those by Regnéll (1960), Laufeld et al. (1975), and Larsson (1979) are the primary sources. Grahn (1996, 1998) has reviewed the biostratigraphy based on chitinozoans. No single section reveals the entire Silurian sequence, but a combination of cores and outcrops allow

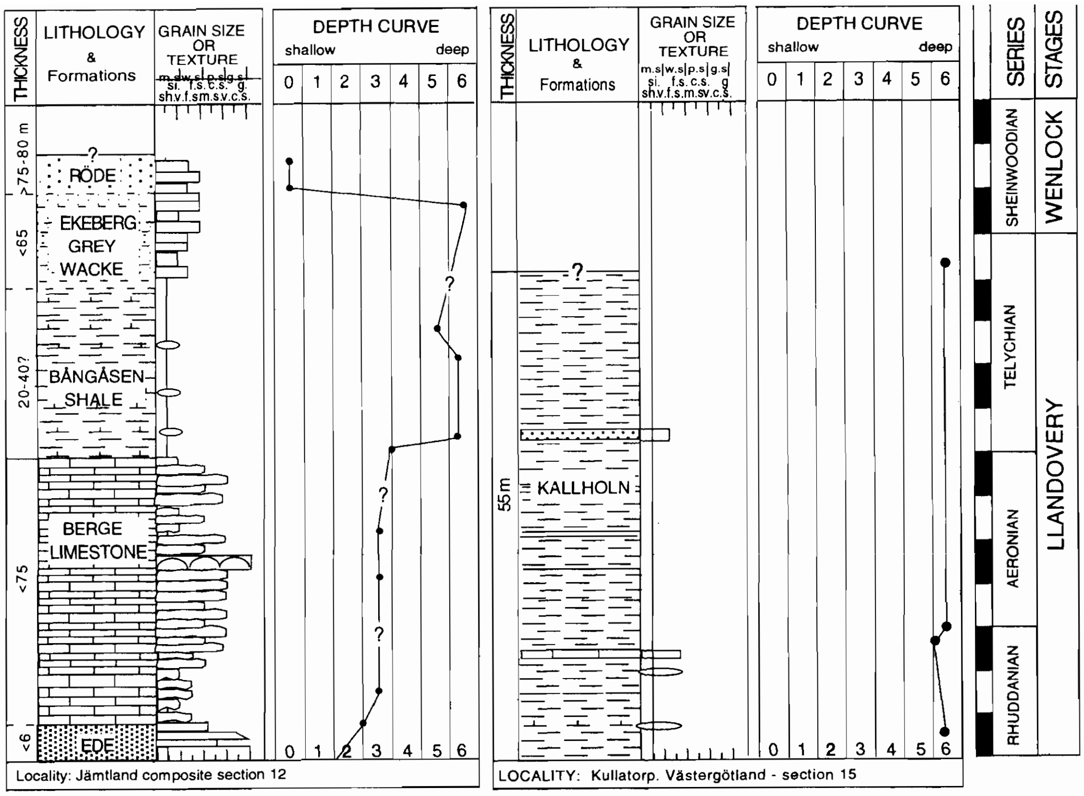

FIGURE 4 *continued.*

construction of a composite section from the Lund area (Fig. 2, location 20) based on studies by Bergström et al. (1999), Laufeld et al. (1975), and Jeppsson and Laufeld (1986). Information from other wells, such as the long Lovisenfred core (Fig. 2, location 21) is summarized by Bergström et al. (1999). The Silurian sequence on Bornholm (Fig. 2, location 22) is described in detail by Bjerreskov (1975, 1986) and Bjerreskov and Jørgensen (1983). More or less continuous sections extend into the Wenlock, while the middle Wenlock *Cyrtograptus lundgreni* Zone is represented by boulders on the southeastern coast of Bornholm. The Terne-1 well (Fig. 2, location 19) in the Danish waters of the Kattegat between Jylland and Skåne was described by Michelsen and Nielsen (1991).

The Lower Silurian of Västergötland (Fig. 2, locations 15–17), Östergötland (Fig. 2, location 18), and the Siljan area in Dalarna (Fig. 2, locations 13–14) was described in studies on the Ordovician–Silurian boundary and Silurian bentonites (Bergström and Bergström, 1996; Bergström et al., 1998), while the biostratigraphy was described by Grahn (1998). Few other comprehensive recent studies are available for this region, with the exception of the description of a core at Kinnekulle (Fig. 2, location 15) in Västergötland (Waern et al., 1948). Additional data from locations 13–18 (Fig. 2) are available in Thorslund and Jaanusson (1960) and Ramsköld (1994).

East Baltic and Dniester Basins

The base map for Silurian outcrops, sections, and core localities on Gotland, Sweden; the Baltic States; Belarus; and the Ukraine is shown in Fig. 7. The East Baltic Basin featured a fairly large, shallow, intracratonic sea on the western part of the East European Platform. The basin developed on a Precambrian basement with very low relief and had a gentle dip toward the present southeast. This basin covered the Baltic states and northern Poland in the south, and the island of Gotland, Sweden, to the north and west. The southeastern edge of the basin is a continuation of the Tornquist–Tesseyre lineament (Fig. 7). Increased subsidence on the southwest margin of the East Baltic Basin began in the Caradoc and the subsidence rate increased through the Ordovician–Silurian (Poprawa et

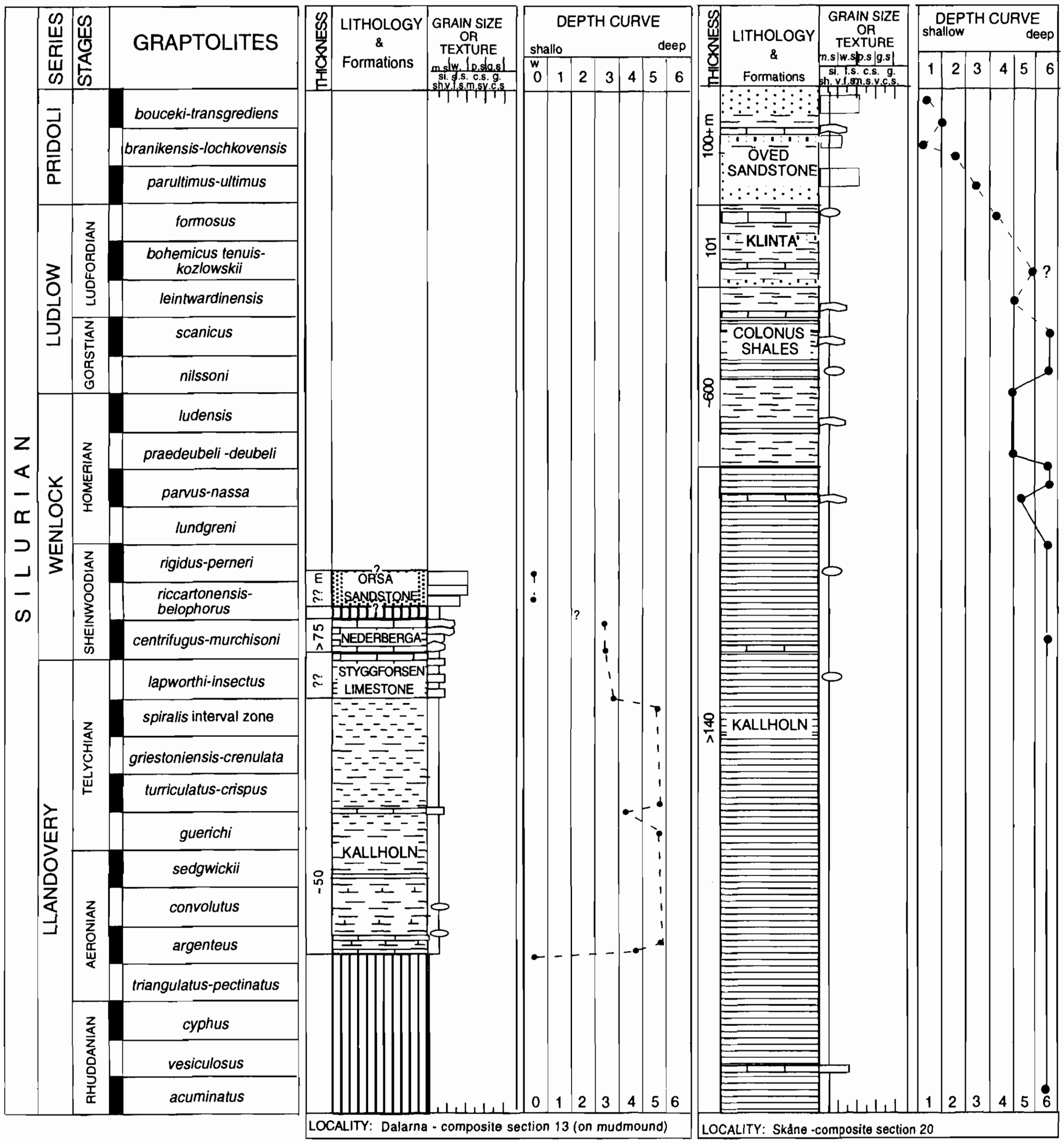

FIGURE 5 — Silurian stratigraphy of localities 13 and 20 in the Central Scandinavian Basin (see Fig. 2).

al., 1999). The resulting accumulation of Silurian rocks ranges from a few hundred meters in the northeast to more than 3.0 km in the southeast, and was mainly deposited during the Ludlow and Pridoli (Poprawa et al., 1999). This was probably coeval with closure of the Tornquist Sea and the thrusting of eastern Avalonia over the edge of Baltica (Mona Lisa Working Group, 1997). The overthrusts may have been comprised of accretionary wedges derived from the North German–Polish Caledonides (Poprowa et al., 1999). Oliver et al. (1993) proposed that the initial continent-to-continent collision occurred in southwest Poland during the middle Silurian with a later transpressional closure, although most other models stipulate earlier contact further west. The East Baltic Basin changed from a passive platform to a foreland basin with lithospheric flexure shown by progressive movement of the depositional towards the east during the Silurian (Poprawa et al., 1999). This foreland basin was a sediment-starved "bay" with hemipelagic deposits formed in present western Latvia that extended to western Lithuania, through Kalingrad, and into northern Poland (Nestor, 1990a).

The Dniester Basin developed as a rift-related, pericontinental depression in Volynia, Podolia, and Moldavia. After a Late Proterozoic rift phase involving massive flood basalts, this basin commenced subsidence that lasted from the Cambrian through the Silurian (Nikishin et al., 1996). The basin roughly parallels the southwest margin of the Baltic continent, has a steep slope towards the Tornquist-Tesseyre lineament (Fig. 8), and features narrow facies belts. A carbonate-dominated shelf was developed during the Silurian, and features a shoaling-up succession (Koren' et al., 1989). Abundant bentonites deposited from the late Wenlock to the end of the Silurian suggest a nearby active subduction zone, possibly an arc in the Rheic Ocean to the southeast of Baltica (Huff et al., 2000).

GOTLAND — A composite section for western Gotland is shown in Fig. 9. The island of Gotland features excellent Silurian exposures, and the area has been intensely studied since the mid-eighteenth century. The lithostratigraphic scheme is somewhat in flux, because only some units are formally defined as formations; the rest are referred to as "Beds." The biostratigraphic correlation follows Jeppsson (1994, 1997) and Kaljo et al. (1998). Our composite section comes from the area around Visby on the west coast of Gotland (Fig. 9). The basal Llandovery is not exposed, but is known from several cores. Detailed information for this part of the sequence is lacking, except for discussion of the biostratigraphic horizon at the base of the Lower Visby Beds in the File Hadar and Visby cores (Thorslund, 1968). Facies and lithostratigraphy for exposed parts of the Lower and Upper Visby Beds and the succeeding Høgklint Formation were defined and described by Riding and Watts (1991), although they defined the upper parts of the Högklint and the succeeding Tofta Formation as the Kopparsvik Formation. This name was rejected by later workers (e.g., Jeppsson, 1997; Kaljo, 1998). The Upper Visby Beds and Høgklint Formation were previously detailed in the Vattenfallet Project (Jaanusson, 1979; Bassett 1979), and all major fossil groups were described and related to lithostratigraphy. The Slite Group was studied by Jacobsson (1997) and Calner and Säll (1999). The latter authors also described the basal beds of the Halla Formation, where they included the Mulde Beds as a member within this formation. The Fröyel Formation, between the Slite Group and the Halla Formation, was defined and separated from the upper Slite Group by Calner (1999). The Hemse Group was treated by Sandström (1998) with emphasis on the reefal facies. The transition between the Hemse Group and Eke Formation was reviewed by Cherns (1983), and rocky-shore conditions within the Eke Formation were discussed by Cherns (1982). The Hamra and Sundre Beds were treated in Long (1993). Where no recent study was available, the lithostratigraphy follows Laufeld (1974), who provided a description of each rock unit.

Not reflected in our paleogeographic maps for this region is the arrival of subaerial siliciclastic deposits from the Caledonian front. None of these are preserved on Gotland, but they occur in submarine outcrops south of the island and are at least partly of Pridoli age (Jeppsson et al., 1994). Subaerial siliciclastic deposits subsequently reached central Poland and adjacent regions by the early Emsian (Bassett et al., 1989).

BALTIC STATES — Estonia features coastal Silurian outcrops on its islands and an extensive network of core data from the mainland. The Ohesaare core from the island of Saaremaa (Fig. 9) shows most of the Silurian (Nestor, 1990b). An auxiliary section from Panga Cliff (Fig. 7, locality 4) is Wenlock (Rubel and Einasto, 1990). Additional sections in the Llandovery (Fig. 7, localities 5–8) come from the Eikla core on Saaremaa, as well as from the Varbla, Kirikukula, and Ikla cores from mainland Estonia (Johnson et al., 1991). The biostratigraphy of Saaremaa was revised by Loydell et al. (1998) for the Lower Silurian and by Jeppsson et al. (1994) for the rest of the Silurian. The subsurface Silurian throughout Latvia and Lithuania is well investigated, but there are no outcrops. Two cores from Latvia (Fig. 7, localities 9 and 10) are from Kolka (Männil, 1977; Märss, 1986) and Ventspils (Märss, 1986; Bassett et al., 1989). Both span the upper Llandovery through Pridoli.

In Lithuania, Musteikis (1993) provided a detailed description of Silurian communities through the Pilviskiai core (Fig. 7, locality 13), which spans the upper Llandovery–Pridoli. We relied on Musteikis and Kaminskas (1996) and Musteikis and Paskevicius (1999) for information on sections through the Silurian in western, central, and eastern Lithuania (Fig. 7, localities 11, 12, and 14).

BELARUS — Our data from southwestern Belarus comes from a core at Rataichitsy (Fig. 7, location 15). Available information is available on communities from the deeper parts of the shelf margin (Pushkin and Modzalevskaya, 1999). The Belarus succession ranges from the upper Llandovery through the rest of the Silurian.

DNIESTER BASIN — On the basis of excellent exposures on the Dniester River and its tributaries in the Skala area (Fig. 7, locality 16), a composite stratigraphic section (Fig. 10) is adapted from Tsegelnjuk et al. (1983), Gritsenko et

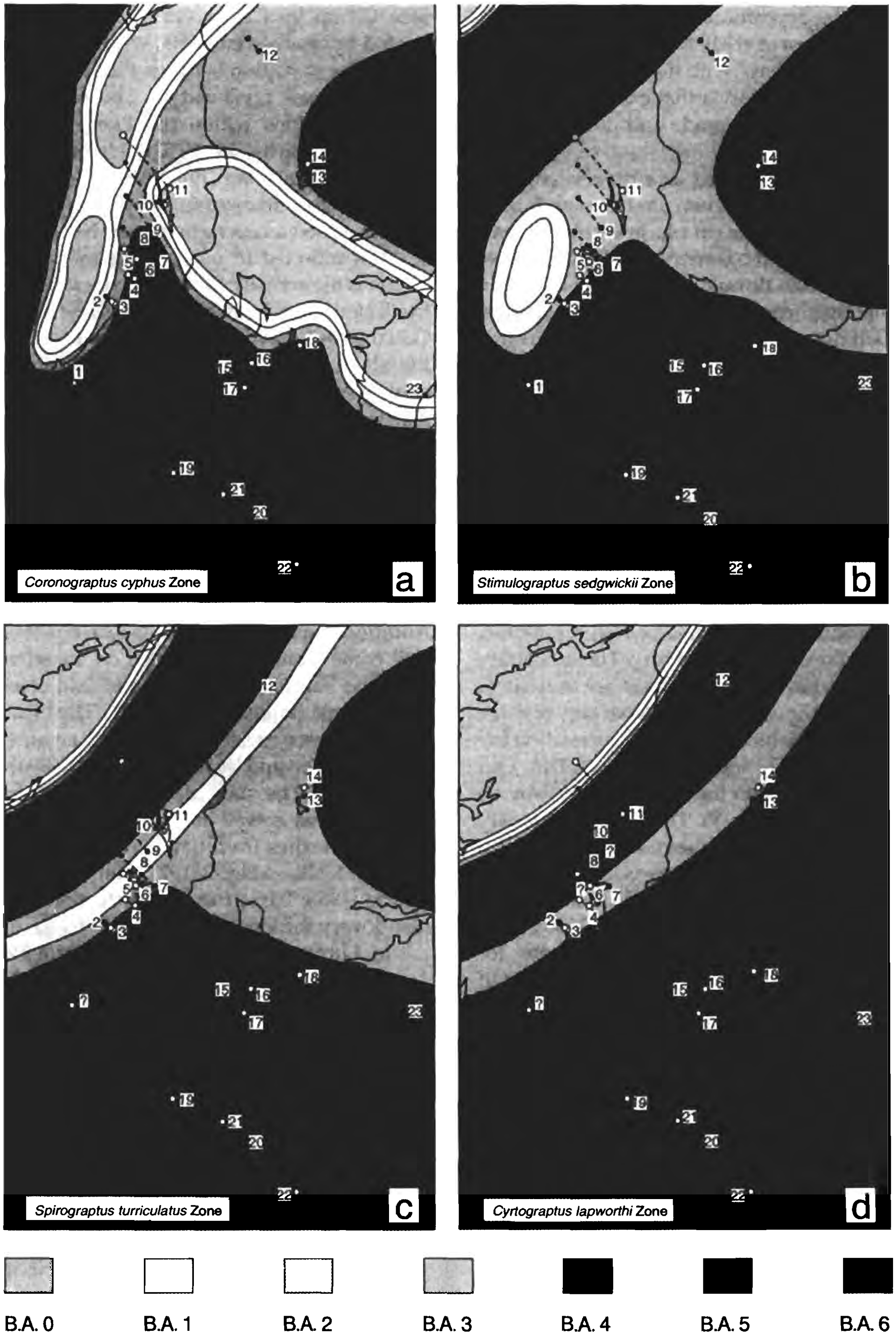

FIGURE 6 — Paleogeographic maps for eight Silurian time intervals: a—d, Llandovery; e and f, Wenlock; g and h, Ludlow. Benthic Assemblage Zone numbers 0–6 (Boucot, 1975) apply to all paleogeographic reconstructions in this report.

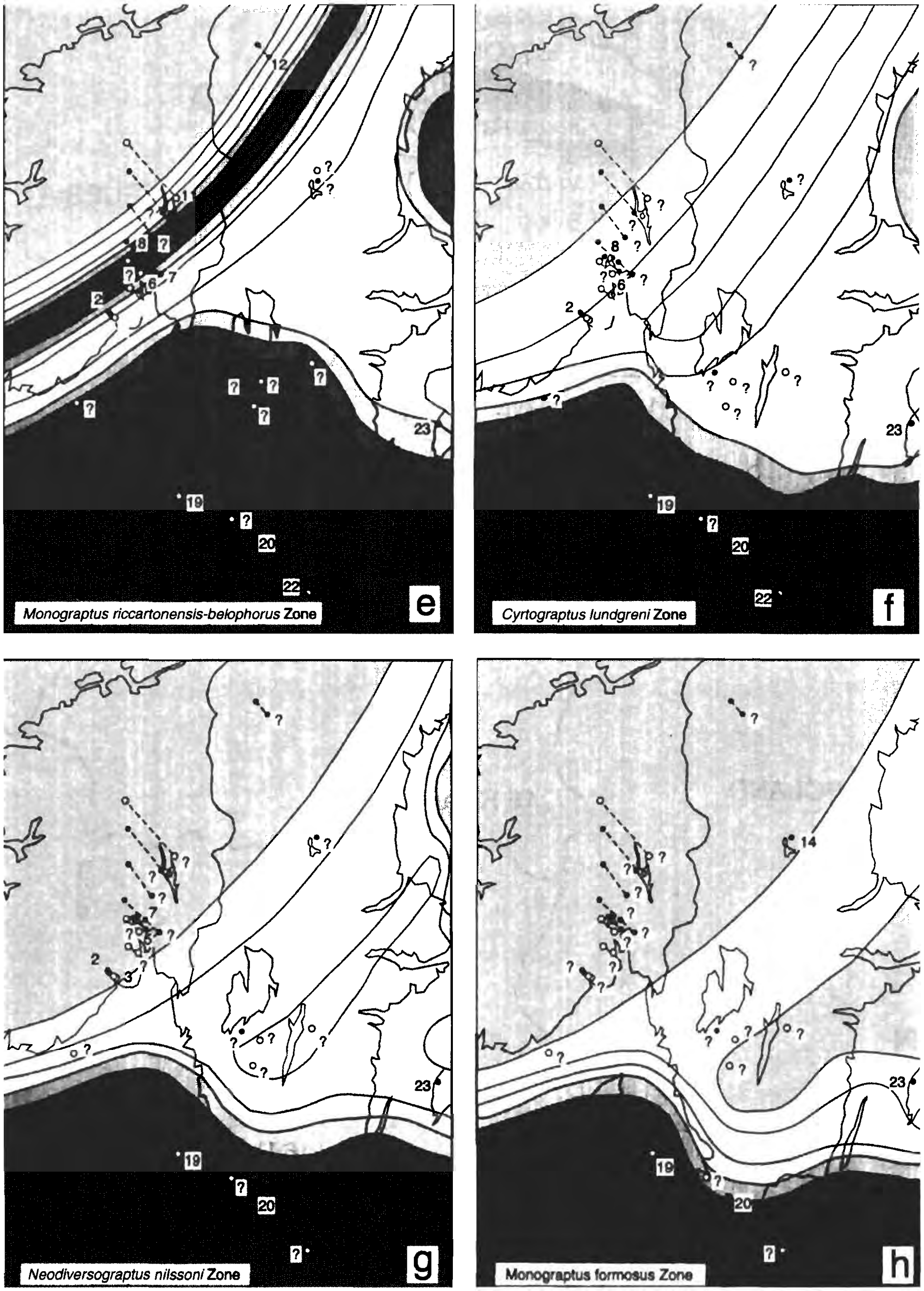

FIGURE 6 *continued.*

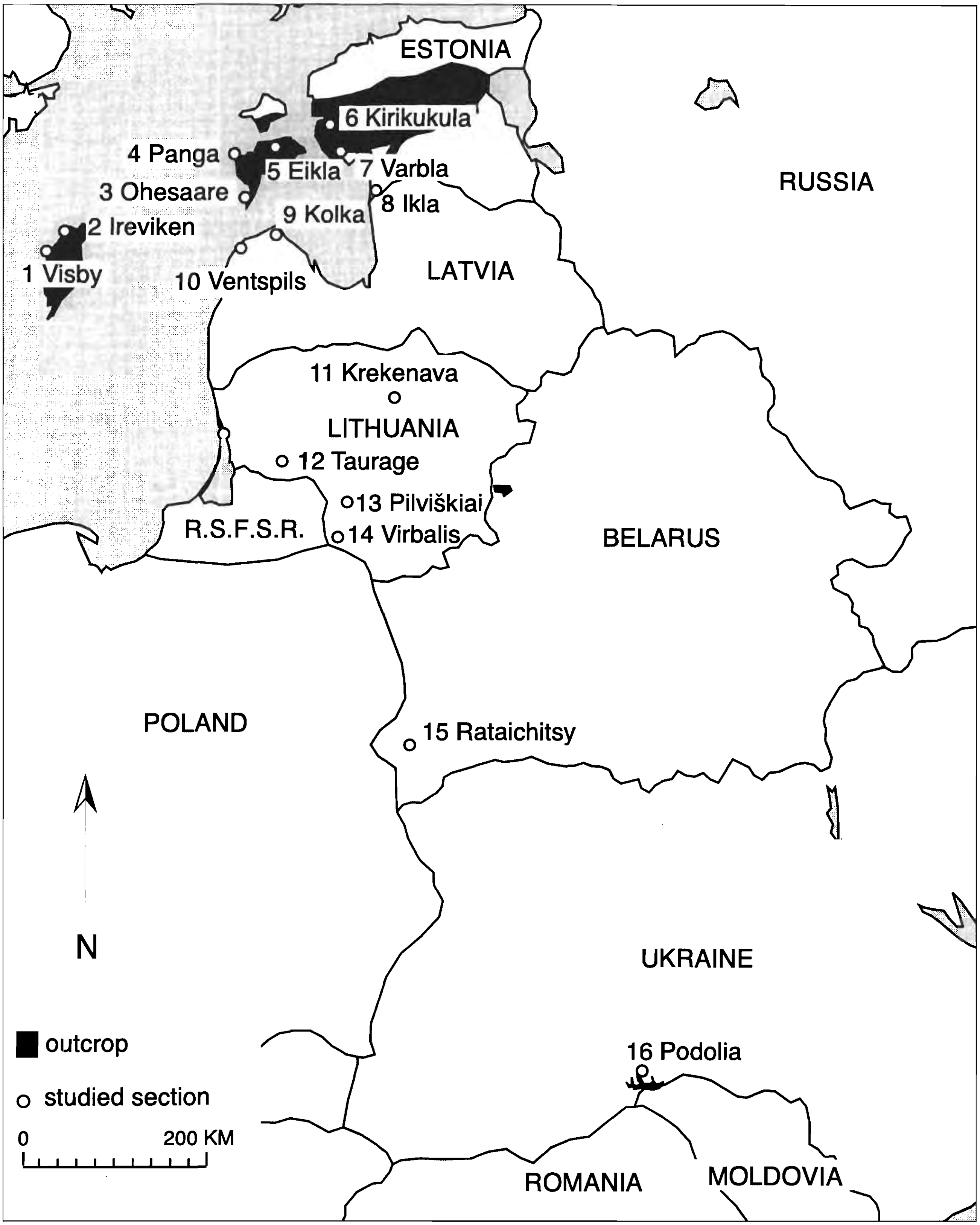

FIGURE 7 — Silurian outcrops of the East Baltic Basin, Belarus, and the Dniester Basin of Podolia.

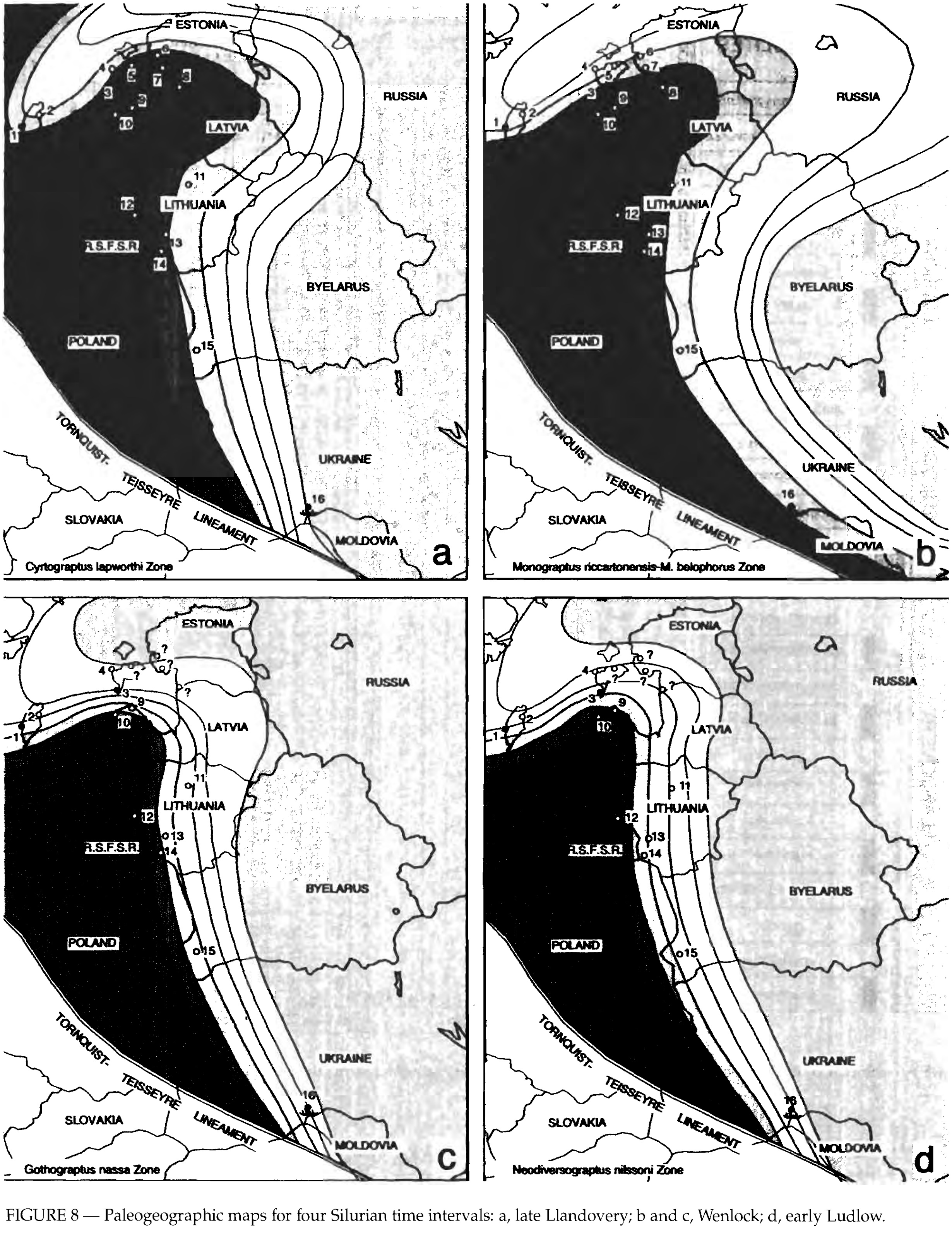

FIGURE 8 — Paleogeographic maps for four Silurian time intervals: a, late Llandovery; b and c, Wenlock; d, early Ludlow.

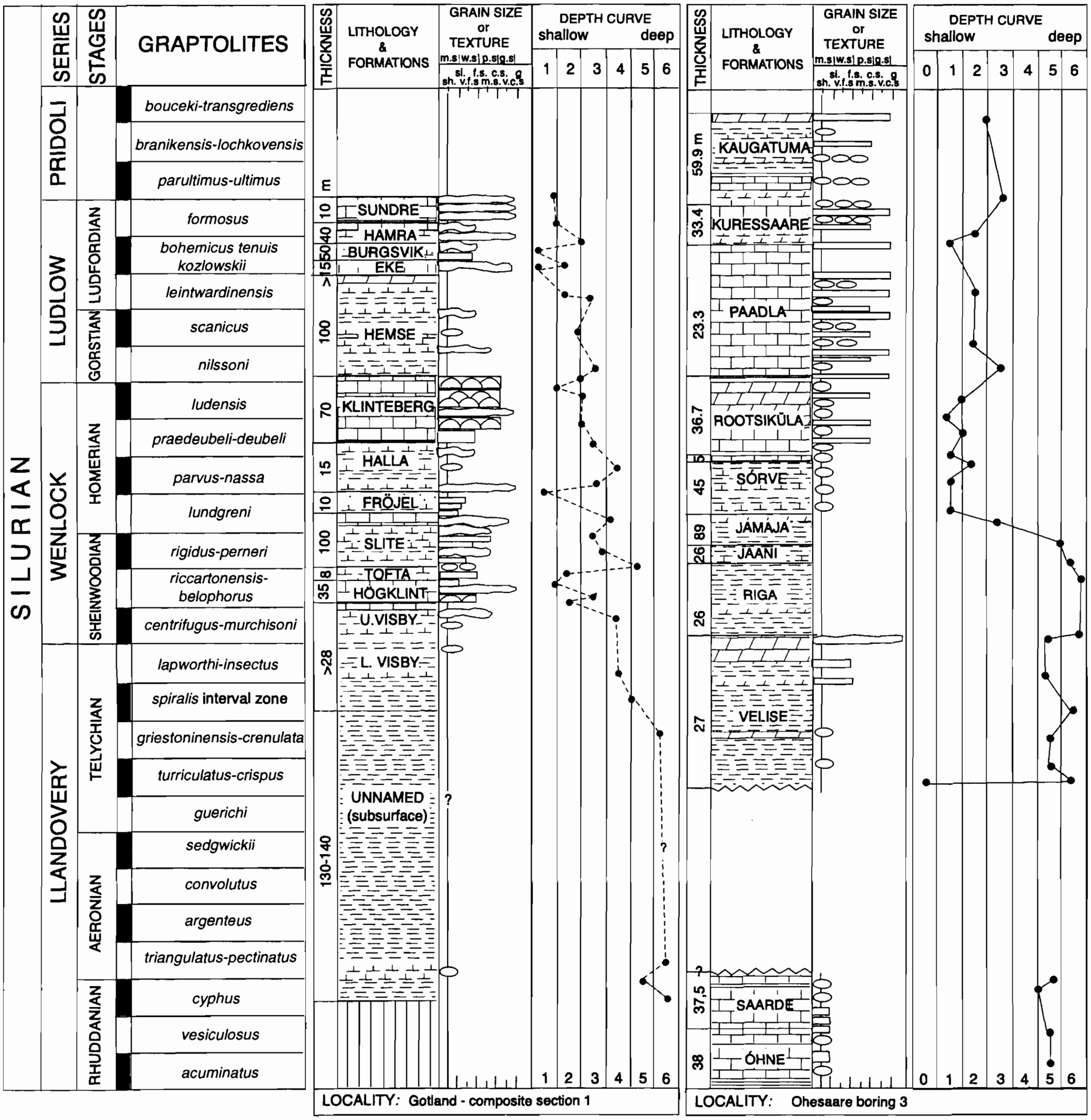

FIGURE 9 — Silurian stratigraphy of Gotland, Sweden, and the Ohesaaer boring in Estonia (see Fig. 7).

al. (1983), Abushik et al. (1985), and Koren' et al (1989). Benthic assemblage zones are recognized through this sequence according to the interpretations of Gritsenko et al. (1999). Correlation of the Lower Silurian, based on conodonts, follows Jeppsson (1997).

Timan–Pechora Basin

The Novaya Zemlya terrane, possibly together with the North Tamyr–Zevernaya terrane, appears to have been accreted to the Petchora–Barentia Basin on the Baltic continent during the Cambrian. The collision may have ter-

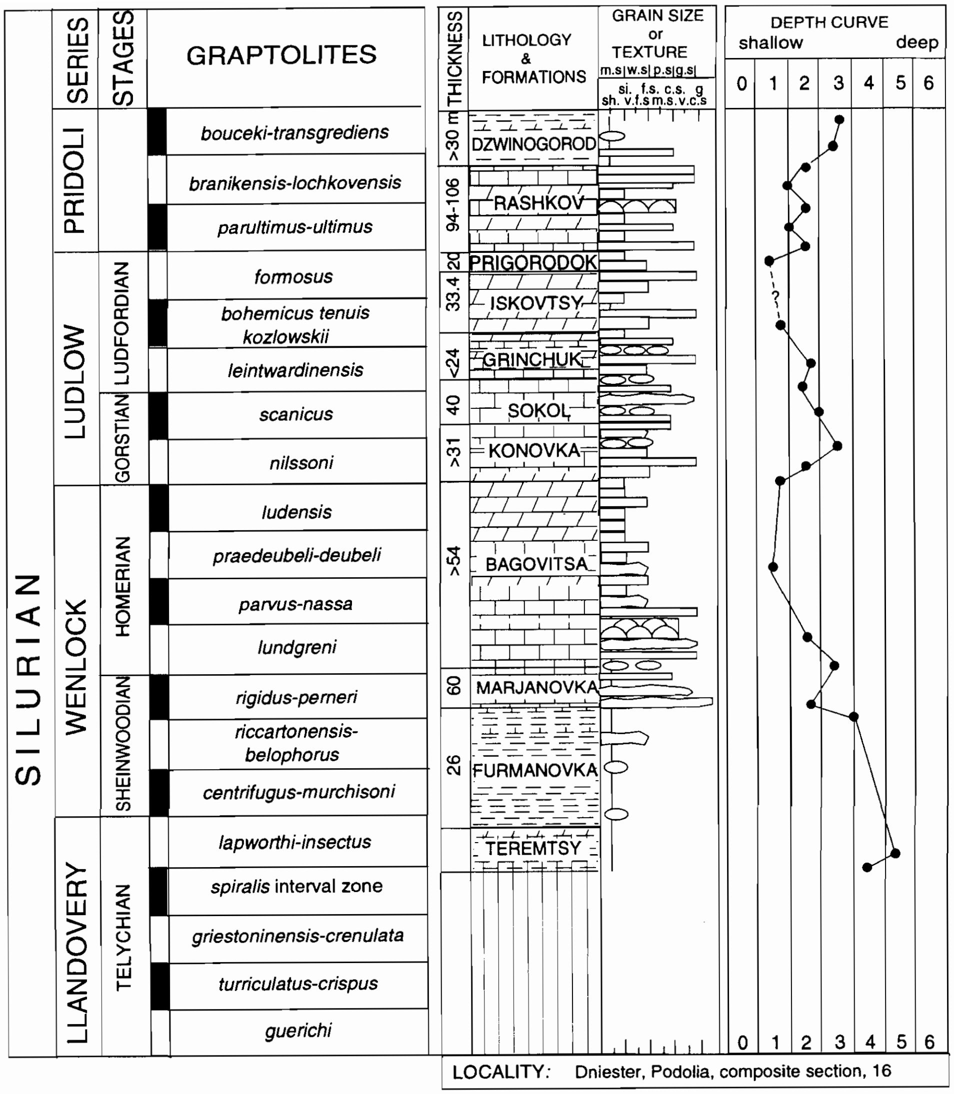

FIGURE 10 — Composite Silurian section for the Dniester Basin, Podolia (see Fig. 7).

minated during the Late Cambrian–Early Ordovician (Nikishin et al., 1996). The opening of the Uralian Ocean to the east also began in the Late Cambrian–Early Ordovician (Malyshev, 2000). A shallow marine carbonate platform developed on a generally passive margin during the Late Ordovician and Silurian. Weak extensional tectonics in the Silurian led to pull-apart subbasins in the central area and resulted in irregular isopachs (Martirosyan et al., 1998; Malyshev, 2000). There was a general subsidence towards the east, with thin sedimentary sequences towards the Timan Ridge and very thick sequences towards the Uralian Ocean in the east (Malyshev, 2000). Possible loading with deposition of sediments derived from the Caledonian front led to a deepening of the basin in the north toward the end of the Silurian (Nikishin et al., 1996). Simultaneously, renewed passive rifting was initiated in the Late Silurian (Malyshev, 2000).

A base map for this region of northeastern Russia is shown in Fig. 11. Detailed stratigraphic columns are provided from twelve localities in an area that extends from the polar areas of Novaya Zemlya southward to the subpolar eastern side of the Ural Mountains (Figs. 12–14). This entire region has been extensively studied, and provides a wealth of stratigraphic information generally unavailable to non-Russian geologists. An updated stratigraphic scheme for the region was published by Antsygin et al. (1993). A biostratigraphic correlation of major strati-

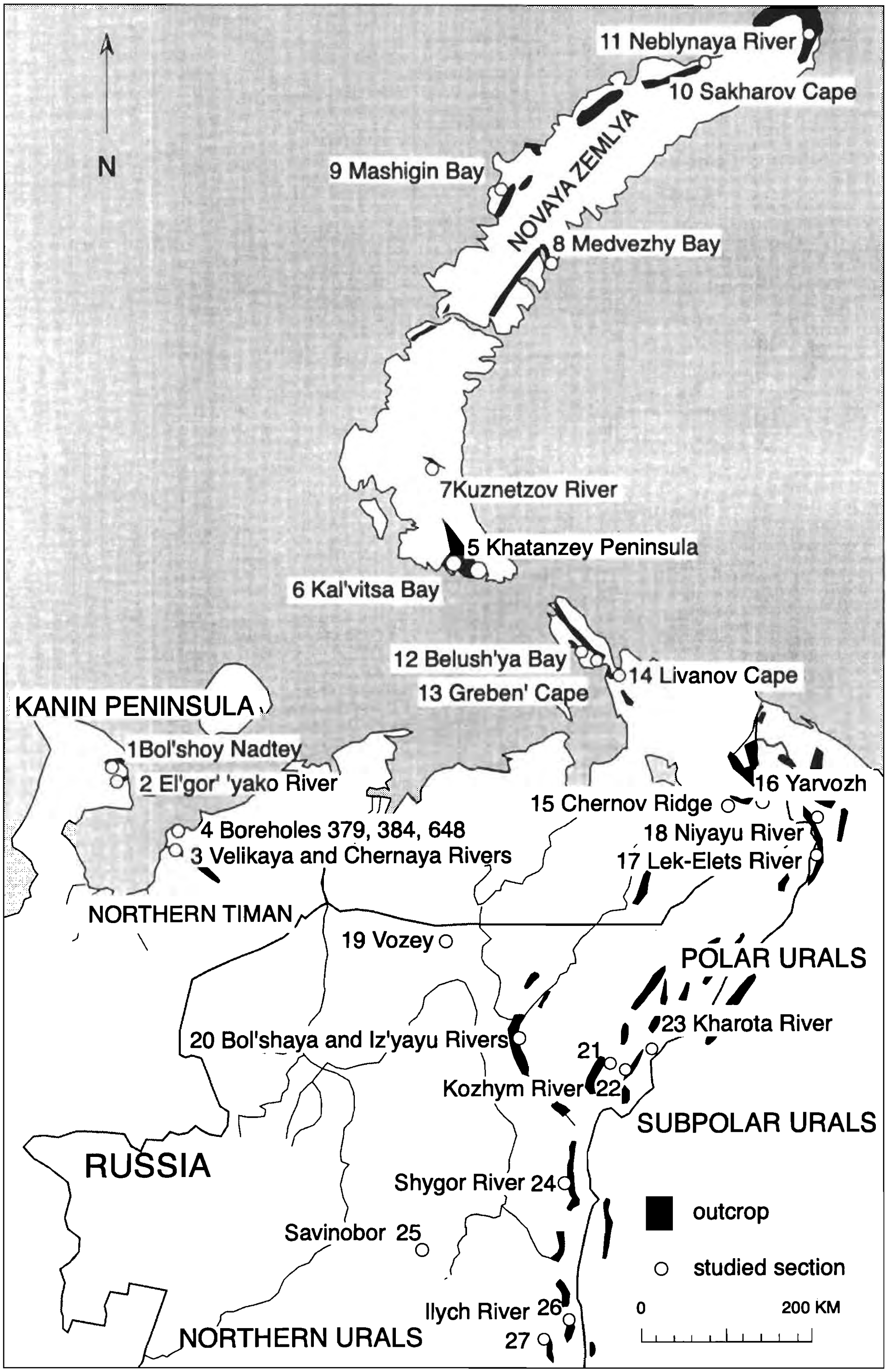

FIGURE 11 — Silurian outcrop (in black) of the Timan–Pechora Basin of northwest Russia.

graphic boundaries within the Timan–northern Ural regions into the East Baltic Basin is given by Antoshkina et al. (2000). Excellent reviews of Silurian community data from Vaygach and Novaya Zemlya, and the Polar to Subpolar Urals, are found in Nekhorosheva and Patrunov (1999) and Sapelnikov et al. (1999), respectively.

The Kanin Peninsula (Fig. 12, locations 1, 2) and the northern Timan area (Fig. 12, localities 3, 4) do not have extensive outcrops. Apart from the overview articles cited above, there are only stratigraphic notes in Gercen et al. (1975) and Valiukevicius et al. (1983). Novaya Zemlya (Figs. 11 and 12, localities 5–11) and the Vaygach Peninsula (Fig. 11, localities 12–14) have been more thoroughly investigated (Blinkov, 1981; Bondarev et al.,1985; Cherkesova, 1970; Tsyganko et al., 2000). The lithostratigraphy of the northern Urals was treated by Antoshkina et al. (1989), and a field guide treats the Subpolar Urals in detail (Antoshkina et al., 2000). In compiling sections from this area (Figs. 13 and 14, localities 17, 18, 26, and 27), the published reports were supplemented by unpublished data collected by A.I. Antoshkina as part of her reef facies studies (Antoshkina 1988, 1994). Locality 21 (Fig. 11) was discussed by Tsyganko and Chermnykh (1983, 1987). Most of the sections in the Timan–Pechora Basin feature shallow-water carbonate deposits with limited possibilities for precise correlation with the graptolite zonation. Koren' (1964), however, studied the graptolites from slope deposits (Fig. 11, locality 23) and correlated them with faunas from elsewhere in the Polar Urals. A set of four paleogeographic maps for the late Llandovery to late Ludlow is compiled in Figs. 15a–d.

Record of Coordinated Sea-Level Changes on Baltica

The evidence of Silurian eustasy has been reviewed by Johnson (1996) and Johnson et al. (1998). Eight eustatic highstands are identified through the Silurian. Four major events occurred in the Llandovery Epoch, and they correlate with: 1) the approximate boundary between the Rhuddanian and Aeronian Ages, 2) the early part of the late Aeronian Age *Stimulograptus sedgwickii* graptolite Chron, 3) the earliest part of the Telychian, and 4) the latest Telychian prior to or roughly equivalent with the Llandovery–Wenlock boundary. A fifth event is a minor highstand equated with the *Monograptus riccartonensis* Chron within the Sheinwoodian of the Wenlock. Two major events (6 and 7) are recognized in the Ludlow. One is in the earliest Gorstian and corresponds to the *Neodiversograptus nilssoni* Chron, and the other is within the late Ludfordian and approximately coeval with the *Bohemogratpus bohemicus tenuis-Neocucullograptus kozlowskii* Chron. The Pridoli had a single eustatic highstand.

Loydell (1998) suggested modifications to this scheme in the lower part of the Silurian. In particular, he proposed a minor highstand correlated with the *Monograptus argenteus* Zone in the middle Aeronian Stage. He also distinguished separate highstands in the *Spirograptus guerichi* and *S. turriculatus* Zones in the lower Telychian Stage, and proposed an additional minor event correlated with the *Gothograptus nassa* Zone in the upper Wenlock. Data from the graphic logs and attendant paleogeographic maps assembled in this report offer the opportunity to test the applicability of the various schemes.

Llandovery eustasy — The Central Scandinavian Depression was subject to extensive tectonic movements during the Silurian with loading of its southern margin that resulted from docking of eastern Avalonia and the advancing Caledonian front to the northwest. The paleogeographic model in Figs. 6a–h reveals, however, that eustatic changes were superimposed on the dominant tectonic regime. We provide sea-level curves for ten stratigraphic sections in this region (Figs. 3–5).

The close of the Ordovician featured a strong regression, followed by a transgression, that began in the latest Ordovician *Glyptograptus persculptus* Chron. A continuous transition occurs between the Ordovician and Silurian along the southern margin of the Baltic Platform from Bornholm to Skåne, and probably further north in the Kinnekulle section in Västergötland (Figs. 4, 6 a–d). There are stratigraphic gaps, however, and the boundary interval is missing immediately to the east at localities 16–18 (Fig. 2) of Bergström and Bergström (1996). In the Oslo Region, the Asker–Sandvika and Malmøya sections (Fig. 3) include continuous boundary sequences (Baarli, 1988). A gradual transgression west and northward at localities 1 through 5 and 8 through 10, through the *Coronograptus cyphus* Chron (Fig. 6a) to Jämtland (Fig. 4), allowed preservation of a continuous section across the boundary. This is attributed to uplift and instability towards the west in the Late Ordovician, with erosion and deposition in a shallow basin that developed through the latest Ordovician and Llandovery.

Four out of ten bathymetic curves (Figs. 3–5) record the initial Silurian highstand that is coeval with the latest *Coronograptus cyphus* Chron (Johnson et al., 1998). Bergström and Bergström (1996) demonstrated that the Motola Formation, which represents a brief shallowing before the *C. cyphus* Chron deepening, can be traced throughout Västergotland and Östergotland. The Skien curve (Fig. 3, locality 2) picks up the subsequent *Monograptus argenteus* Zone event of Loydell (1998). Also at Kallholn (Fig. 5, locality 13) in Dalarna Province, limestone is deposited on a mudmound and overlain by shales with graptolites of the *M. argenteus* Zone. Two out

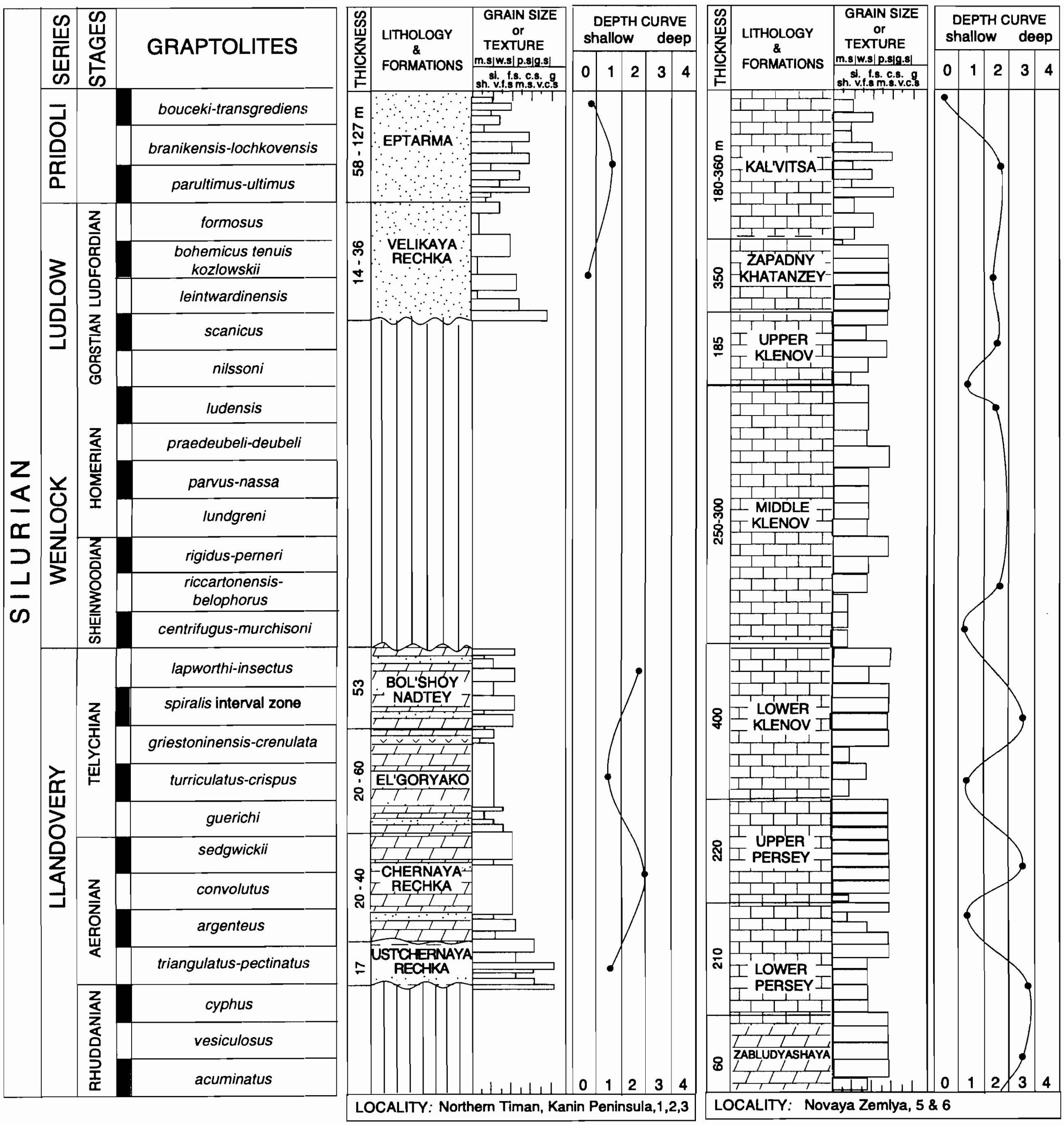

FIGURE 12 — Silurian stratigraphy of localities 1–3, 5, 6, 10–13, 15, and 16 in the Timan–Petchora Basin (see Fig. 11).

of ten curves specifically register the highstand coeval with the base of the *Stimulograptus sedgwickii* Chron (Johnson et al. 1998), but the paleogeographic map in Fig. 6b shows a substantial drowning of paleotopography after the start of the Aeronian. The poor record of the later Aeronian highstand in the Oslo Region and Jämtland is attributed to the position and southeast advance of a peripheral bulge and developing foreland basin in connection with the approaching Caldonian front (Baarli, 1990a). The Swedish localities, during this time, all feature deep-water graptolite shales, where few details are known and minor sea-level changes may go unnoticed.

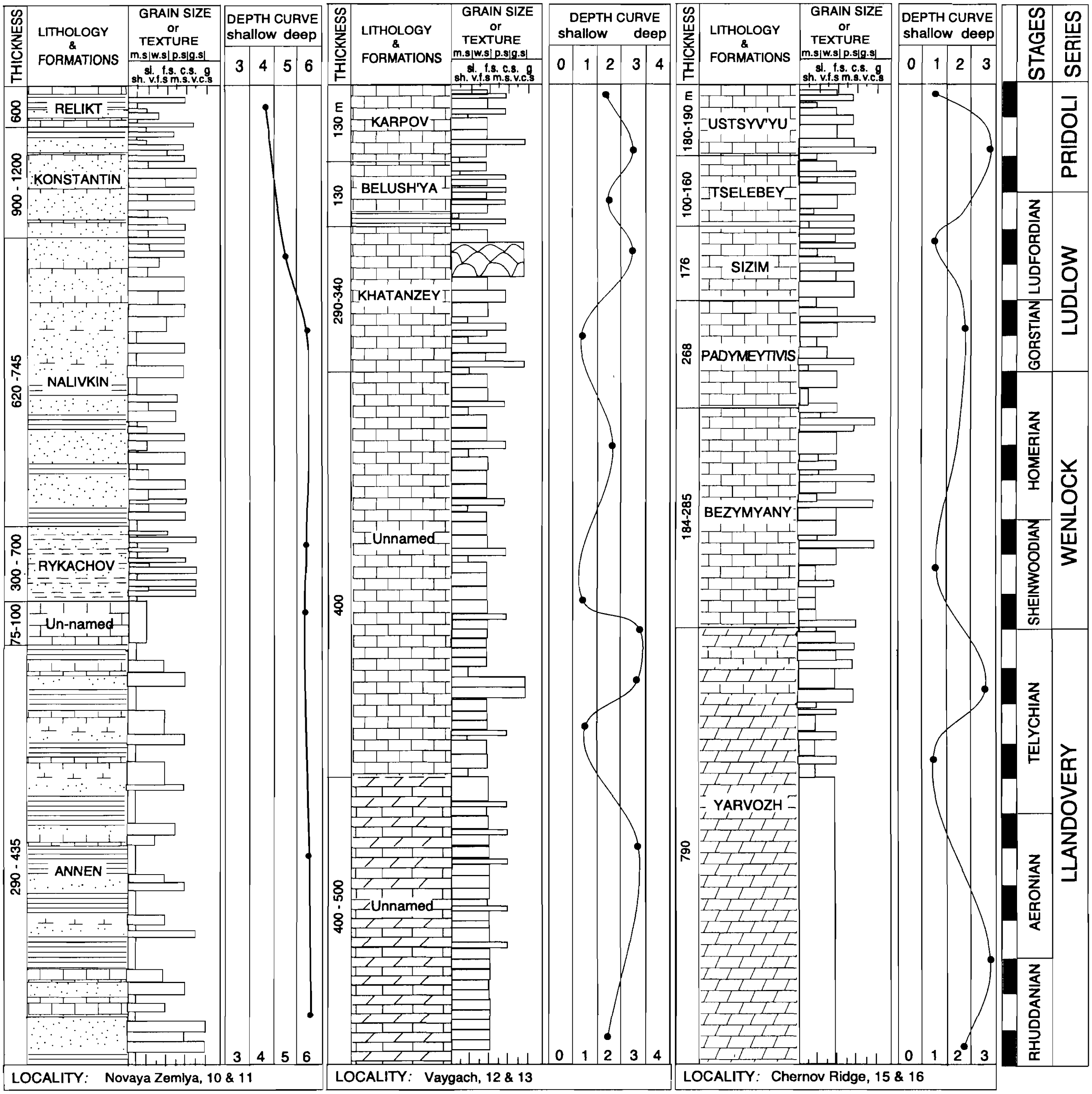

FIGURE 12 *continued.*

At Osmundsberget (Fig. 2, locality 14), the section commences with shales in the *S. sedgwickii* Zone that overlie a thin conglomeratic limestone (Thorslund and Jaanusson, 1960) as the mound was submerged. The Jämtland section is very poorly described and has little stratigraphic control, but the general deepening of the basin continued through the *S. sedgwickii* Zone.

A sea-level highstand equated with the basal Telychian *Spirograptus guerichii* Zone is well documented in the Skien and Ringerike sections (Fig. 3, localities 2 and 8) and the Hadeland, Toten, and Jämtland sections (Fig. 4, localities 9, 10 and 12). The absence of a record of this event in the Asker–Sandvika and Malmøya sections (Fig. 3, localities 7 and 8) may be due to these areas being sub-

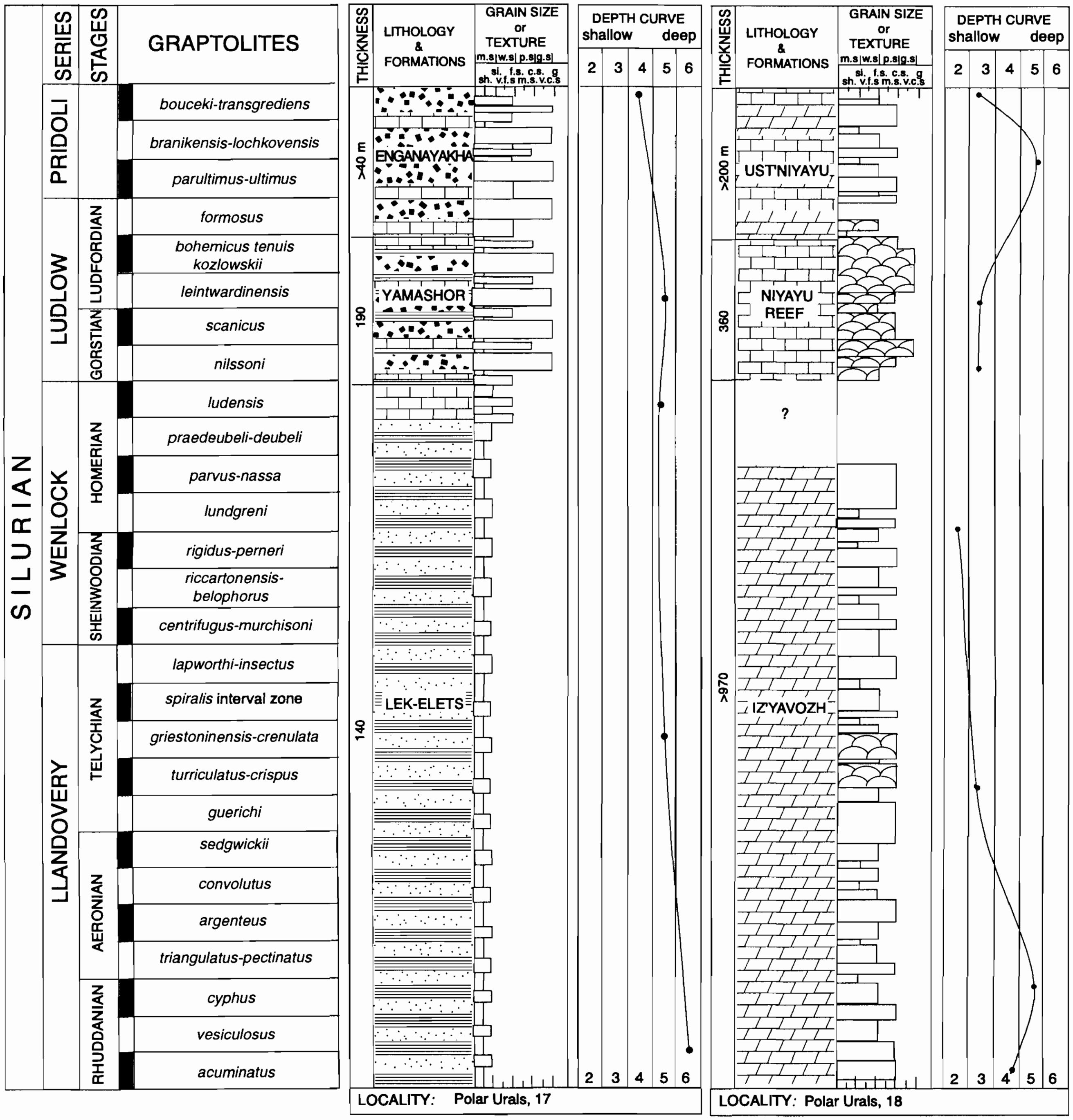

FIGURE 13 — Silurian stratigraphy of localities 17, 18, and 21–24 in the Timan-Petchora Basin (see Fig. 11).

jected to uplift with passage of a peripheral bulge that locally negated a rise in sealevel. The same bulge passed through Ringerike, and thus the eustatic fall in sea level was probably accentuated. Hadeland and Jämtland were on the margin of the approaching fordeep during the *S. guerichii* Chron, while Toten was already in the foredeep and, therefore, experienced a continuous deepening. The Swedish sections (Figs. 4 and 5) are in deep-water graptolite shales that reveal little about bathymetric changes during the time interval.

Changes in sea level during the *Spirograptus turriculatus* Chron are better expressed. A sea-level low followed

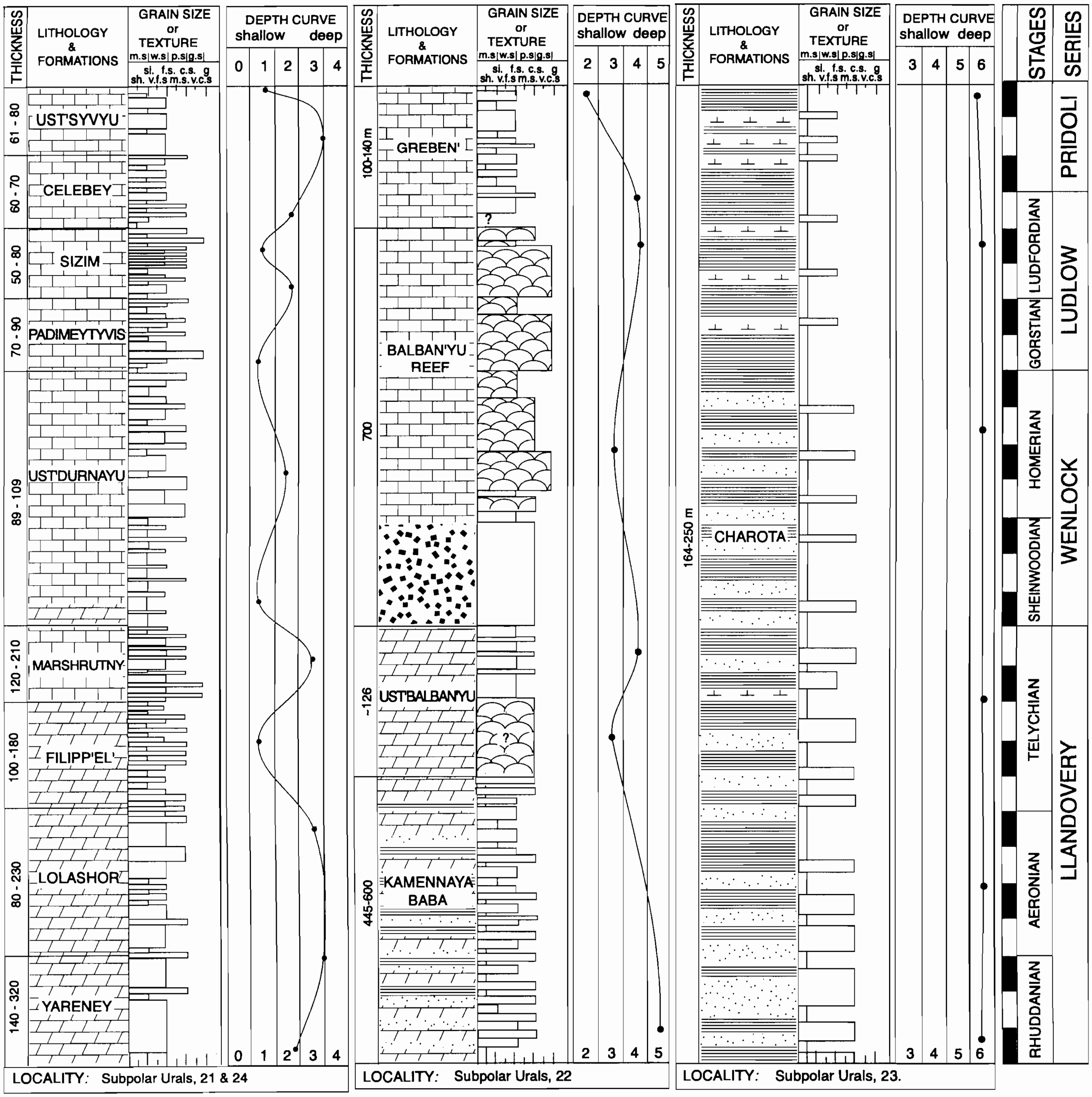

FIGURE 13 *continued.*

by renewed deepening is seen in all areas of the Oslo Region, with the exception of Hadeland, Toten, and Jämtland (Figs. 3 and 4). For this time interval, these areas exhibit deep-water graptolite shales. In Dalarna (Fig. 5), a limestone-rich interval in the Kallholn Formation overlain by shales containing *S. turriculatus* (Grahn, 1998) may record the same deepening.

The highstand at or near the Llandovery–Wenlock boundary is well represented in five out of seven sections that span this interval in the Central Scandinavian Depression (Figs. 3–5). The Skåne section (Fig. 5, locality 20) reflects long-term graptolite shale deposition before, during, and after this interval, but water depth was too great to register small-scale changes. Section 13 in

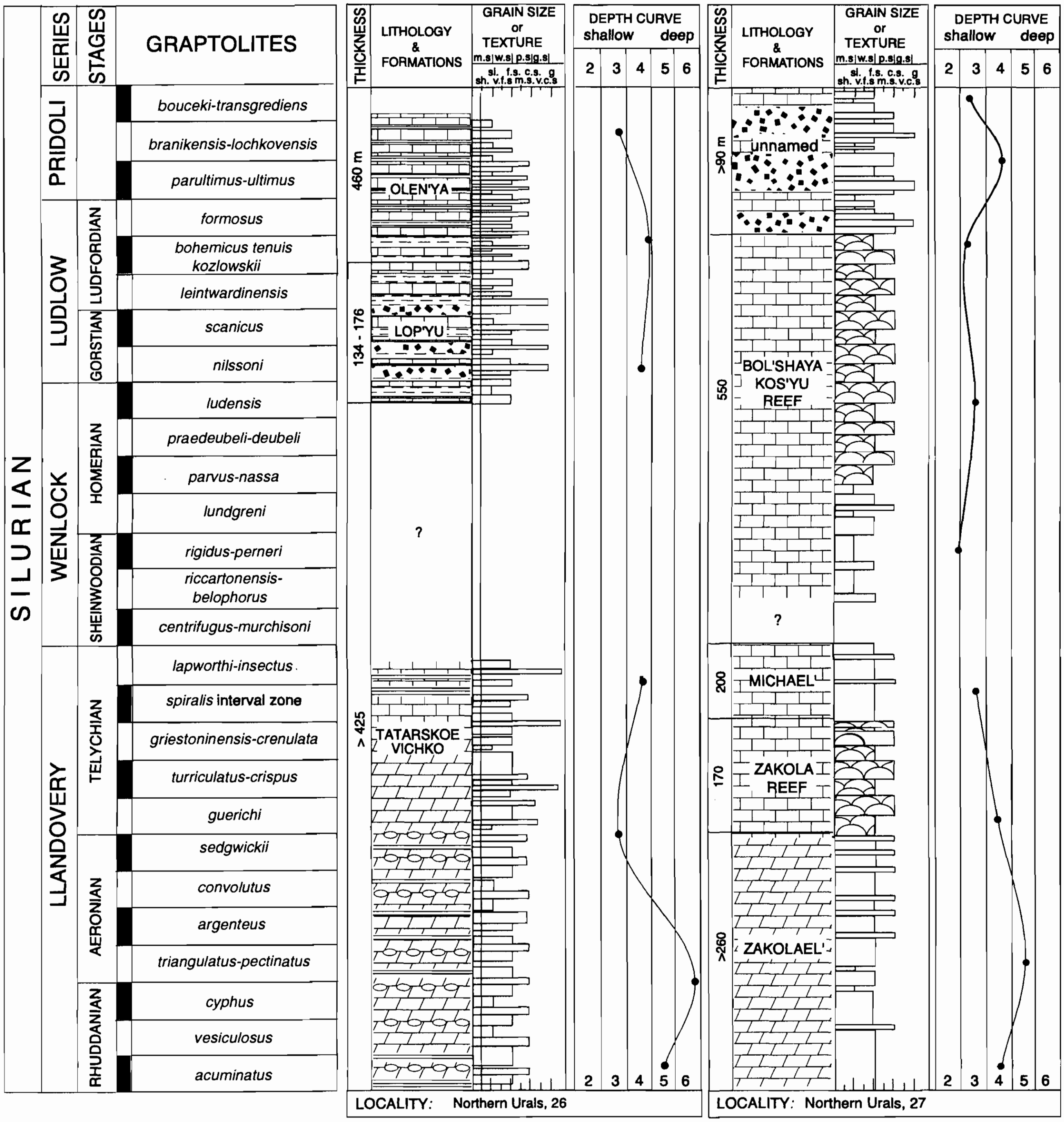

FIGURE 14 — Silurian stratigraphy of localities 26 and 27 in the Tilmen-Petchora Basin (see Fig. 11).

Dalarna (Fig. 5) exhibits a shallowing at this time that demonstrates advance of the peripheral bulge across central Sweden.

In the East Baltic and Dniester Basins, the initial Llandovery transgression first reached the islands west of the Estonian mainland during the Rhuddanian. Locality 13 in western Lithuania was first onlapped in the late Aeronian to early Telychian. Locality 15 in Belarus was flooded in the middle Telychian. Podolia, which is farthest east, was drowned during the late Llandovery (Fig. 7).

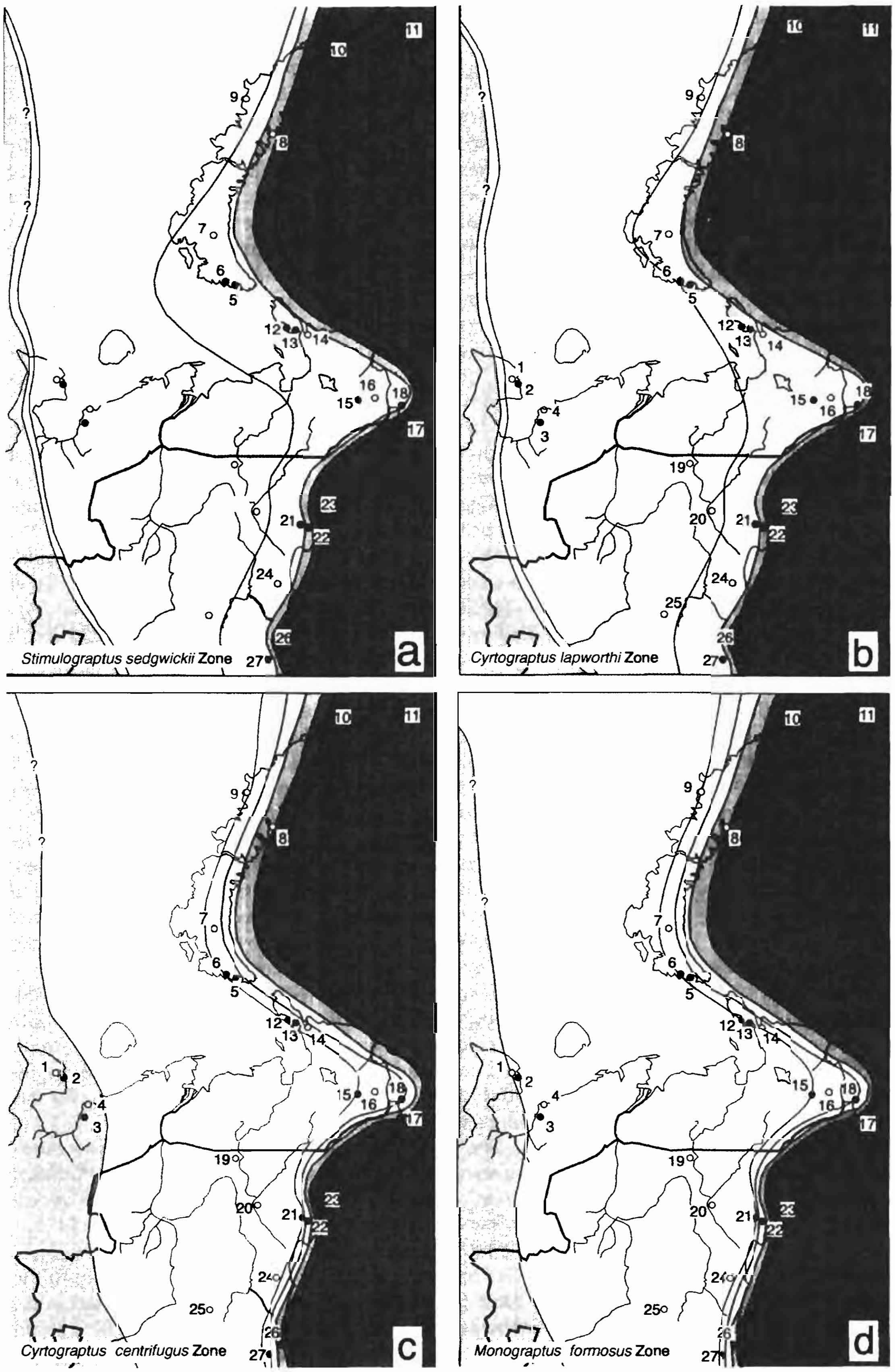

FIGURE 15 — Paleogeographic maps for four Silurian intervals in the Timan–Petchora Basin: a and b, Llandovery; c, early Wenlock; d, late Ludlow.

Johnson et al. (1991) examined the Estonian Llandovery for sea-level changes. There is general agreement among auxiliary sections 5–8 and the Ohesaare sequence (Fig. 9) for a Rhuddanian–Aeronian highstand. Bergström and Bergström (1996) documented the shallowing of the *Coronograptus cyphus* Zone.

The latter zone is followed by the same deepening into the *Demirastrites triangulatus* Zone in the subsurface of Gotland. There is a major regional transgression in the *Spirograptus turriculatus* Zone (Johnson et al. (1991) that is picked up in all Estonian sections (Fig. 7). A minor shallowing immediately followed in the Ohesaare section (Figs. 8a, 9). The transgression of the Podolian section occurred simultaneously with a small deepening in the *Oktavites spiralis* Chron that is known in the Ohesaare section (Figs. 9, 10). This is a minor eustatic high picked up by the more detailed curve of Loydell (1998), as opposed to that of Johnson et al. (1998). The same small highstand is also expressed in the Asker–Sandvika and Skien districts in the Oslo Region (Fig. 3), but in no other sections there. A major transgression is found in the Ohesaare section at the Llandovery–Wenlock boundary.

In contrast to central Scandinavia, the Timan–Pechora Depression is generally considered to have developed on the relatively passive northern margin of Baltica in the Timan–Urals region. The Pechora Basin, however, sits astride several zones of uplift or megaswells that strike at an oblique angle toward the pre-Uralian foredeep on the outer shelf. On the Chernaya River of the Kanin Peninsula in northern Timan, for example (Fig. 12, localities 1-3), middle Aeronian carbonates of the Chernaya Rechka Formation rest on the eroded surface of the terrestrial Ustchernaya Rechka Formation, which is provisionally assigned to the basal Aeronian. This area is situated on the Paj–Khoj uplift, and probably experienced a delayed effect of the initial Llandovery marine transgression.

Elsewhere, rather abrupt lithological changes occur at the Ordovician–Silurian boundary. On the Kozhym River in the Subpolar Urals, for example (Fig. 13, locality 18), Upper Ordovician, light-gray, dolomitized limestones with the brachiopod *Holorhynchus* and the tabulate coral *Eocatenipora* are overlain by black dolomitized limestones with crinoid debris, rare *Mesofavosites* corals, and interbeds dominated by radiolarians and sponge spicules. At deep-water localities, the outer shelf margin with BA 6 communities is a poor place to detect sea-level changes, as shown by locality 23 in the Subpolar Urals (Fig.13).

We provide bathymetric curves for twelve stratigraphic sections in this latter region (Figs. 12–14). Five of the nine bathymetic curves that cover the interval record the initial Silurian highstand that is roughly coeval with the lowest *Coronograptus cyphus* Chron (Johnson et al., 1998). None of the Russian sections pick up the subsequent *Monograptus argenteus* event of Loydell (1998). Four out of twelve curves specifically register the highstand coeval with the lowest *Stimulograptus sedgwickii* Chron (Johnson et al., 1998). It is noteworthy, however, that the *S. sedgwickii* Chron featured a substantial expansion of the mid-shelf region at the expense of more landward environments (Fig. 15a). None of the Russian sections picks up either of the two highstands attributed to the early Telychian. The interval corresponds to conversion of a sloping shelf margin to a rimmed shelf with extensive reef growth that kept up with rising sea level. These conditions are well represented at locality 22 in the Subpolar Urals (Fig. 13) and locality 27 in the northern Urals (Fig. 14).

The Llandovery–Wenlock boundary is difficult to pinpoint in the Timan–Pechora Depression. Three out of our twelve sections lack this interval. Only three sections are suggestive of the highstand proposed for this problematic interval (Figs. 12 and 13).

WENLOCK EUSTASY — The middle to late Sheinwoodian eustatic highstands of Johnson (1996) and Johnson et al. (1998) are recorded only in the Ringerike district (Fig. 3) of southern Norway. Malmøya and Asker (Fig. 3, localities 6 and 7) preserve this part of the stratigraphic column, but these areas underwent transition from foredeep to shallow-water facies as the Caledonian front approached. On Gotland and in Estonia, the maximum Wenlock highstand is noted in the Wenlock. The clearest sea-level pattern recorded in central Scandinavia is a lowstand near the top of the *Cyrtograptus lundgreni* Zone, followed by a pronounced deepening approximately in the *Pristiograptus parvus-Gothograptus nassa* Zone in the Skien, Asker, Ringerike, Skåne, and Gotland areas. The same trend is more weakly represented in the Ohesaare area of Estonia (Figs. 3, 5, and 9). Lloydell (1998) reported the same event from the Ulst core in Latvia. This eustatic trend was not registered by Johnson et al. (1998), although the more detailed curve based on graptolite data constructed by Loydell (1998) does show this event. Based on a study of rocky shoreline submergence in Gotland, Calner (1999) made a convincing argument for this being a eustatic event.

The stratigraphy of the Oslo Region records extensive terrestrial deposition that began in the late Wenlock. The only place with marine stratigraphic sections in central Scandinavia for this interval is in Skåne. Laufeld et al. (1975) described the fauna at the transition from the Kallholn Formation (formerly the *Cyrtograptus* Shale) to the *Colonus* Shale as evidence of an eustatic lowstand in the *"Monograptus" ludensis* Zone. This is followed by a recurrence of deeper graptolite shales at the base of the Ludlow. The same eustatic lowstand, followed by a deepening at the base of the Ludlow, occurs in the Gotland, Ohesaare, and Podolia areas. This pattern agrees with the curve of Johnson et al. (1998).

None of the Russian sections in the Timan–Pechora Depression reflects the minor highstand assigned to the middle to late Sheinwoodian by Johnson (1996) and Johnson et al. (1998). Only the Russian sections (Fig. 13, localities 21 and 24) pick up the highstand advocated by Loydell (1998) in the *Pristiograptus parvus-Gothograptus nassa* Zone in the middle Homerian, although three of the twelve available sections lack the interval. This interval is also a time of renewed reef growth on the rim of the Uralian foredeep (Fig. 13, locality 22; Fig. 14, locality 27).

LUDLOW EUSTASY — The early Ludlow highstand of Johnson (1996) appears both in Gotland and Estonia in the *Neodiversograptus nilssoni* Zone (Fig. 9), as well as in the Kolka core in Latvia (Fig. 7, locality 9). The early Ludlow highstand peaks slightly later in Podolia (Fig. 10). The late Ludlow highstand, approximately coeval with the *Bohemograptus bohemicus tenuis–Neocucullograptus kozlowskii* Chron, is picked up in the Hamra Beds of Gotland (Fig. 9). A possible eustatic highstand recorded in the Klinta Formation of Skåne is also considered Ludfordian (Jeppson and Laufeld, 1986). In general, the Klinta Formation was deposited in shallower waters than the underlying *Colonus* Shale, but the middle Bjär Member shows a recurrence of shale similar to the *Colonus* Shale. We tentatively interpret the Bjär Member as evidence of a brief deepening that corresponds to the middle Ludfordian eustatic maximum advocated by Johnson (1996) and Johnson et al. (1998).

None of the Russian sections in the Timan–Pechora Depression records the prominent highstand correlated with the *Neodiversograptus nilssoni* Zone at the base of the Ludlow (Johnson, 1996; Johnson et al., 1998) that is evident in the East Baltic Basin. Again, this interval in the Russian quarter of Baltica is dominated by extensive reef growth on the rim of the Uralian foredeep (Fig. 13, localities 18 and 22; Fig. 14, locality 27). It is surmised that vigorous reef growth kept pace with the global rise in sea level at that time. Conversely, the Ludfordian highstand that is considered approximately coeval with the *Bohemogratpus bohemicus tenuis–Neocucullograptus kozlowskii* Chron (Johnson, 1996; Johnson et al., 1998), appears in five of twelve Russian profiles that span the interval (Figs. 12–14). Eight out of twelve Russian successions indicate a drop in sea level through the end of the Ludlow, and this is well illustrated by the paleogeographic map for the *Monograptus formosus* Zone (Fig. 15d).

PRIDOLI EUSTASY — Information on Pridolian sea-levels is sparse. Except for Skåne in southern Sweden, none of the successions in central Scandinavia reach into the Pridoli. The Skåne area records a possible middle Pridolian highstand (Fig. 5, locality 20). Our sections from Estonia and Podolia suggest either a continuous regression or a continuous transgression across the Ludlow–Pridoli boundary. The most extensive data on the Pridoli comes from the Timan–Pechora Depression of Russia. Only two of twelve sections fail to cross the Ludlow–Pridoli boundary. Five of ten sections that preserve this interval register a single transgressive–regressive cycle with a middle Pridoli highstand in sea level (Figs. 12–14, localities 1–3, 12, 13, 15, 16, 21, 24, 27).

INFLUENCE OF WENLOCK–LUDLOW TECTONICS

From the Wenlock on, eustatic changes are particularly difficult to discern in the Central Scandinavian Depression. It is likely that movement on the Caledonian front slowed down or halted, while the pronounced foredeep filled to sea level with mostly siliciclastic material after the beginning of Wenlock. The last indication of a foredeep in the Oslo Region (Fig. 6e) is suggested by the sudden occurrence of graptolite shales in the *Cyrtograptus centrifugus–C. murchisoni* Zone that continued into the next graptolite zone (Fig. 3, localties 6, 7). There is no information on marine strata from central Sweden between the Oslo Region and Skåne after that time.

The combined thickness of the *Colonus* Shale and subsequent Öved–Ramsåsa Beds are estimated to be on the order of 800 m in Skåne. This indicates rapid basin subsidence that started near the end of the *Monograptus predeubeli–M. deubeli* Chron and continued until the middle Ludlow. Simultaneously, thick terrestrial deposits accumulated over the entire Oslo Region. We envision that coeval, near-shore terrestrial deposits to deep-water shales in Norway and Sweden were deposited, with progradation of sediments supplied from the Caledonian front to the northwest.

Cores from the Danish Basin indicate that syntectonic deposition occurred on fault blocks during this time, and there is much variation in sedimentary rock thickness over short distances (Michelsen and Nielsen, 1991). Evidently there was strong tectonic activity during the Late Silurian in this region. Upper Silurian volcaniclastic sediments, as seen interbedded with shales in the Terne-1 well, are common. Basalts are found in the deep well at Nøvling on Jylland west of location 19. Volcanic tuff is reported from Bornholm, where associated graptolites indicate the *Cyrtograptus lundgreni* Zone (Bjerreskov and Jørgensen, 1983). The source for these andesitic tuffs is interpreted to lie to the south. Combined faulting and volcanic activity may indicate renewed activity, with thrusting of Avalonia over the margin of Baltica to the south. It is possible that an uplifted source to the south or southwest was more important in deposition in the Skåne area than the approach of the Caledonian front from the northwest. Cut-outs of sections and lack of data in central Sweden may be due to crustal flexure in response to the rapidly subsiding basin in the south, or to the approaching foreland basin to the northwest. Through the Wenlock, there were several episodes of siliciclastic influx from the east into Gotland. Long (1993) suggested that the influx of detrital material, including metamorphic rock fragments during the *Cyrtogratpus lundgreni* Chron, implies a source from the Precambrian shield of eastern Sweden. Whether or not uplift caused by a peripheral bulge could account for this kind of crustal erosion is debatable.

CONCLUSIONS

Approximately 7.8 million km^2 in total area, the paleocontinent of Baltica retains only a miniscule record of its former marine shelves and cratonic depressions in the remaining exposures of fossil-rich Silurian strata. Combined with a more extensive record of Silurian strata in the subsurface, however, it is possible to convert paleontologic and lithologic information from key sequences into reliable bathymetric profiles. Backed by a wealth of auxiliary information, 25 bathymetric logs from the Central Scandinavian, East Baltic, Dniester River, and Timan–Pechora regions of northern Europe; Belarus; the Ukraine; and northwestern Russia allow the reconstruction of a series of paleogeographic maps that cover approximately 35% of Silurian Baltica.

Maps for various time slices through the Llandovery, Wenlock, and Ludlow Series show that sedimentation in the Central Scandinavian Depression and adjacent East Baltic Basin was strongly influenced by the Scandian orogeny to the west and the docking of the microcontinent Avalonia on the southwest. Compilation and interpolation of map data for these sectors yield a more comprehensive model for tectonic influences on the evolution of marine shelves and epicontinental basins. The major tectonic signature of the region during the Early Silurian (Llandovery) is a narrow foreland basin preceded by a peripheral bulge that advanced southeast across Scandinavia. During later Wenlock and Ludlow time, central Scandinavia was deluged with siliciclastics that prograded into shallow water. Simultaneously, the downwarping of the cratonic margin to the south from the impingement of Avalonia brought about relatively deepwater conditions that long persisted in Denmark and Skåne. A tongue of graptolitic shales prograded as far as Gotland in the Llandovery and remained somewhat longer in western Estonia through the earliest Wenlock, but accommodation space for thick accumulations of shale never developed to the same degree as in southern Scandinavia.

By contrast, the Timan–Petchora Depression in northwest Russia developed on a comparatively passive cratonic shelf. Two distinct phases of extensive reef growth occurred behind a rimmed platform margin during the late Llandovery and again from the middle Wenlock through earliest late Ludlow. These episodes are roughly coeval with the record of at least two highstands in sea level that are often recorded elsewhere in the upper Llandovery, and with a highstand commonly recorded at the Wenlock–Ludlow transition.

Against this regional background, the record of Silurian eustasy is surprisingly robust. Four out of eight eustatic peaks proposed by Johnson (1996) and Johnson et al. (1998) are registered with some regularity in Baltica. All shelf regions were affected by the initial rise in sea level following the retreat of Late Ordovician glaciers in the North African and South American sectors of Gondwana. The first prominent highstand in sea level at or near the Rhuddanian–Aeronian boundary is well represented in fully half of the profiles illustrated in this report that span the interval. A rapid rise in sea level that is particularly well exhibited in central Scandinavia and the East Baltic Basin is correlated with the early Telychian *Spirograptus turriculatus* Chron. We surmise that this event was not recorded in the Timan–Petchora region due to vigorous reef growth that kept up with rising sea level. A subsequent highstand at or near the Llandovery–Wenlock boundary is well recognized along an axis that extends from Skien, southern Norway, to Jämtland in west-central Sweden, but is less well known in northwestern Russia. It is important to note that the late Wenlock eustatic draw-down is registered in more than half the profiles for that interval illustrated in this report. It is notable that half the cited Russian profiles show the late Ludlow sea-level excursion, as do the profiles for Skåne and Gotland. Among the additional Silurian eustatic peaks proposed by Loydell (1998), the most promising for Baltica is that associated with the *Pristiograptus parvus-Gothograptus nassa* Zone in the Homerian Stage of the upper Wenlock.

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THE SILURIAN OF AVALONIA

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ABSTRACT — Seventeen sections—six in Wales, five in England, two in Ireland, two in eastern Canada, and two in Belgium — give lithological and thickness data for representative sequences on the Silurian continent of Avalonia, together with depth curves constructed from analyses of the faunal assemblages. Avalonia was an independent continent from the Early Ordovician, when it detached from Gondwana, until the Early Silurian, when it collided with Baltica and subsequently with Laurentia. The relatively small size of Avalonia and the relatively intense tectonic activity it was subjected to led to swiftly changing lithologies and benthic communities through space and time. The Llandovery, Wenlock, and Ludlow Series all have type areas in or near the Welsh Basin in Avalonia.

INTRODUCTION

Avalonia was originally part of Gondwana (Figs. 1, 2), and adjacent to the northern margin of Amazonia. In the Early Ordovician, perhaps in the Llanvirn, it rifted off from Gondwana (Cocks et al., 1997; Prigmore et al., 1997; Cocks, 2000; Woodcock, 2000) to leave a widening Rheic Ocean to its south. During the Early Silurian, Avalonia collided with Baltica, and by the Late Silurian this combined continent moved further across a fast-vanishing Iapetus Ocean to collide and merge with Laurentia during the Caledonian orogeny. This collision led to continental Old Red Sandstone deposition over much of the continent by Pridoli time. Thus Avalonia probably existed as a separate continent only during the Ordovician. At no time was Avalonia faunally unique. In the Cambrian and Early Ordovician, it had faunas similar to the southern continent (including the polar regions) of Gondwana, while from the Middle Ordovician most of its faunas became increasingly similar to those of Baltica and Laurentia (Cocks and Fortey, 1982, 1990). The faunal evidence for the northward movement of Avalonia across the Iapetus Ocean during the Ordovician has been amply confirmed by paleomagnetic data (Torsvik et al., 1996). Siliciclastic sedimentary rocks are dominant in Avalonia through much of the Early Paleozoic, although a few shell hash limestones occur locally. The earliest substantial warm-water carbonates appeared in the Wenlock. The faunal and climatic changes are reviewed later in this paper.

Avalonia is comprised of a number of now-separate regions which form much of New England (south to Cape Cod), U.S.A., the Maritime Provinces of eastern Canada, much of southern Ireland, all of Wales, England, northernmost France, Belgium, the Netherlands, part of northern Germany, and probably southern Portugal (Cocks et al., 1997). However, to the south of what is today Belgium, there are no relatively complete, well-dated, or fossiliferous sections in Avalonia, and this latter region is not considered further herein. In a similar way, only two sections are provided for North America: New World Island (Newfoundland) and Arisaig (Nova Scotia), even though Avalonia extended as far as Massachusetts.

CORRELATION PROGRAMS

The columns which follow are a small selection from the many available. These sections may be found in reports by Berry and Boucot (1970) for the United States, Norford et al. (1997) for Canada, and Cocks et al. (1992) for the British Isles, including Ireland. The seventeen columns which follow have been selected for their relative stratigraphic completeness and for the variation across the paleocontinent which they demonstrate. Perhaps more columns from North America might have been included to achieve a more even geographical balance across the terrane. For example, sections in southern New Brunswick, western Nova Scotia, and the Avalon Plat-

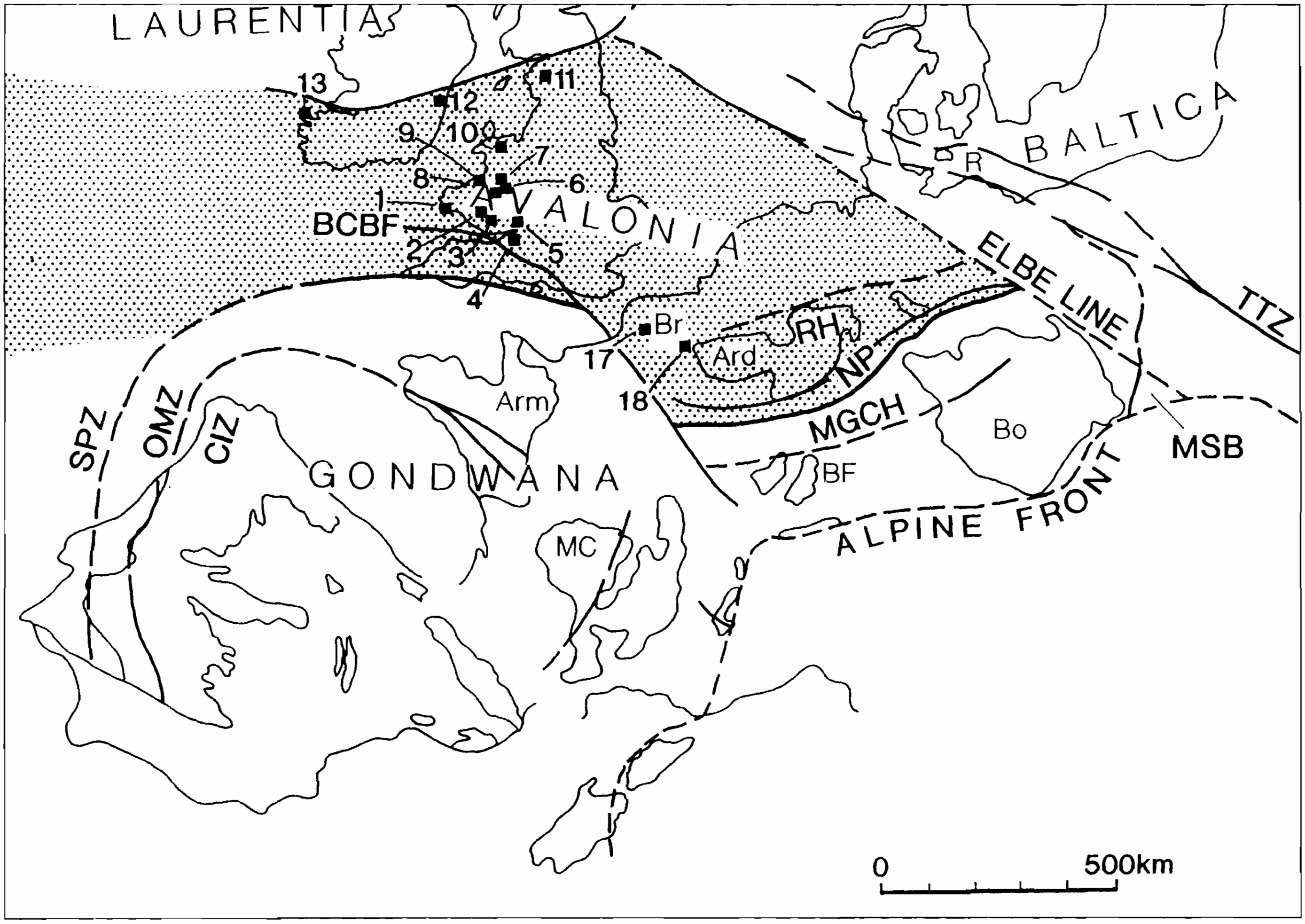

FIGURE 1 — Map of terranes assigned to Avalonia in Europe (after Cocks et al., 1997). Note that Avalonia is now considered to have extended northwestward to the Tornquist–Tesseyre Zone (TTZ). Numbers indicate locations of Silurian sections discussed in text.

form represented in the offshore Grand Banks all contain Silurian rocks. However, substantial lithological, faunal, and community data were not available from these areas at the time of our compilation. For the locations of the columns, see Figs. 1 and 2. The Benthic Assemblage (BA) zones of Boucot (1975), which range from BA 1 (shallow) to BA 5 (deep), are used where possible. "BA 6" is used for areas too deep for the appearance of abundant and diverse benthic organisms in the Silurian, but from which graptolites and other planktic or nektic fauna elements have been recovered. "BA 0" is used for non-marine assemblages.

Marloes and Skomer — This column (Fig. 3, left) is based on well-exposed rocks along the southwestern Welsh coastline at Marloes Bay and nearby Skomer Island. The Skomer Volcanic Group was long thought to be Ordovician until the discovery of the Aeronian brachiopod *Eocoelia hemisphaerica* in interbedded shales (Ziegler et al., 1969). The unconformable and overlying Coralliferous Group of late Telychian age (and possibly also early Wenlock age) includes BA 4 *Costistricklandia* Community assemblages with *C. lirata* that shallow up into a BA 2 *Eocoelia* Community. The succeeding Gray Sandstone Group includes brachiopods of generalized Wenlock age and bivalve-dominated assemblages that represent a shallowing-up sequence (Walmsley and Bassett, 1976; Hurst et al., 1978). The overlying Red Cliff and Sandy Haven Formations are in terrestrial Old Red Sandstone facies and represent continental emergence. These formations are the lowest within the Milford Haven Group (Allen and Williams, 1978), which continues upwards into the Devonian. Correlation within these higher rocks is locally quite uncertain, although comparable rocks 80 km east near Llandeilo yield Late Silurian spores.

Llandovery — The formal, international base of the

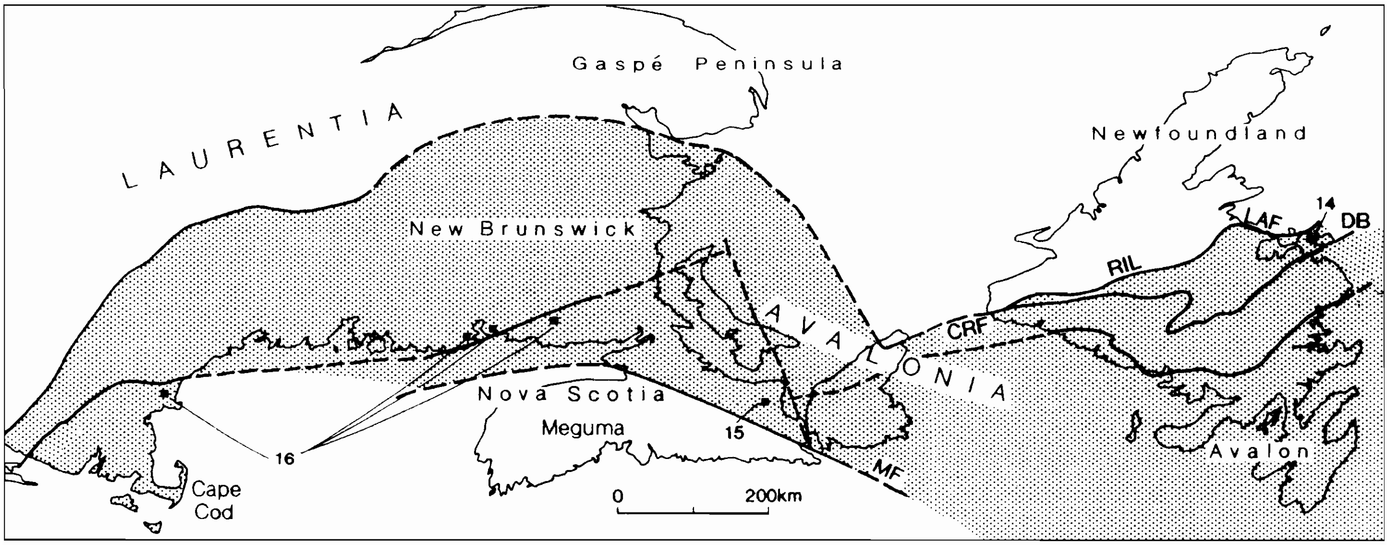

FIGURE 2 — Map of terranes assigned to Avalonia in eastern North America (after Cocks et al., 1997). Numbers indicate locations of Silurian sections discussed in text.

Llandovery Series (and that of its lowest Rhuddanian Stage) is at Dob's Linn, in the Southern Uplands of Scotland (Cocks et al., 1992). This local succession was deposited in an ocean a long distance offshore from Avalonia and far from any source of coarse siliciclastic sediment. However, the basal stratotypes of the higher two Llandovery Stages, the Aeronian and Telychian, are in the Llandovery district in the county of Dyfed, south-central Wales, an area made famous by the original detailed work of Murchison (1839). The area has been revised several times, most recently by Cocks et al. (1984), who formalized the stratigraphic nomenclature (Fig. 3, right) and recorded large numbers of brachiopods, graptolites, and other fossils from the area, including enough chitinozoans to base a zonal scheme on them. The base of the Aeronian is within the sandy mudstones of the Trefawr Formation in the north of the area and at the base of the *Monograptus triangulatus* Zone. The base of the Telychian Stage lies within the thin-bedded sandstones and mudstones of the Wormwood Formation to the south of Llandovery, and coincides with the lowest appearance of *Eocoelia curtisi* and *Stricklandia laevis.* Most of the Llandovery succession around Llandovery is represented by middle- to outer-shelf siliciclastics. The succeeding Upper Silurian is best exposed in the Sawdde Gorge, 10 km south-southwest of Llandovery, where the Wenlock is a generally regressive sequence from deep to shallow shelf (Hurst et al., 1978), followed by deep-shelf environments in the basal Ludlow. A second regressive sequence shoals to non-marine prior to the end of the Pridoli. The Wenlock of the Sawdde Gorge has three units (Hurst et al., 1978): 1) the Sawdde Siltstone Formation, 325 m of siltstones and shales with rotational slumps in the upper half; 2) the Sawdde Sandstone Formation, which includes interbedded sandstones, siltstones, and shales with an in-situ BA 4 *Isorthis* Community; and 3) the shallower-water Ffinnant Sandstone Formation, which shows a regressive sequence from the BA 4 *Isorthis* Community through the BA 3 *Homoeospira* Community and into a BA 2 *Salopina* Community. The Wenlock succession culminates in a 10 m sandstone, interpreted as a barrier bar, overlain by 6 m of siltstones with a BA 1 bivalve-*Lingula* Community interpreted as a restricted lagoonal assemblage. The uppermost bed of the formation is a 1 m ironstone oolite with a BA 3–4 *Gypidula* Community that heralds the basal Ludlow transgression within the Tresglen Beds and their BA 5 *Dicoelosia* Community. The overlying Ludlow and Pridoli beds are variable in character (Potter and Price, 1965). They consist of shelf deposits (the Tresglen and Black Cock Formations) overlain by coarser-grained sandstone deltaic deposits (Trichrûg Beds). These are in turn overlain by the Cwm Clíd and Roman Camp Beds, which carry mid-shelf faunas, are progressively regressive, and terminate upwards with BA 2 *Protochonetes-Salopina* Community faunas at the top of the Roman Camp Beds. After a relatively brief period of non-deposition, the Long Quarry Beds were deposited in a shallow-marine basin with a BA 1 *Lingula*-bivalve Community, and this unit is followed by the non-marine Raglan Mudstones with a typical Old Red Sandstone, non-marine facies.

Usk — The Usk Inlier in southeast Wales only has exposures of middle Wenlock and later beds (Fig. 4, left), although unpublished borehole data reveal the underly-

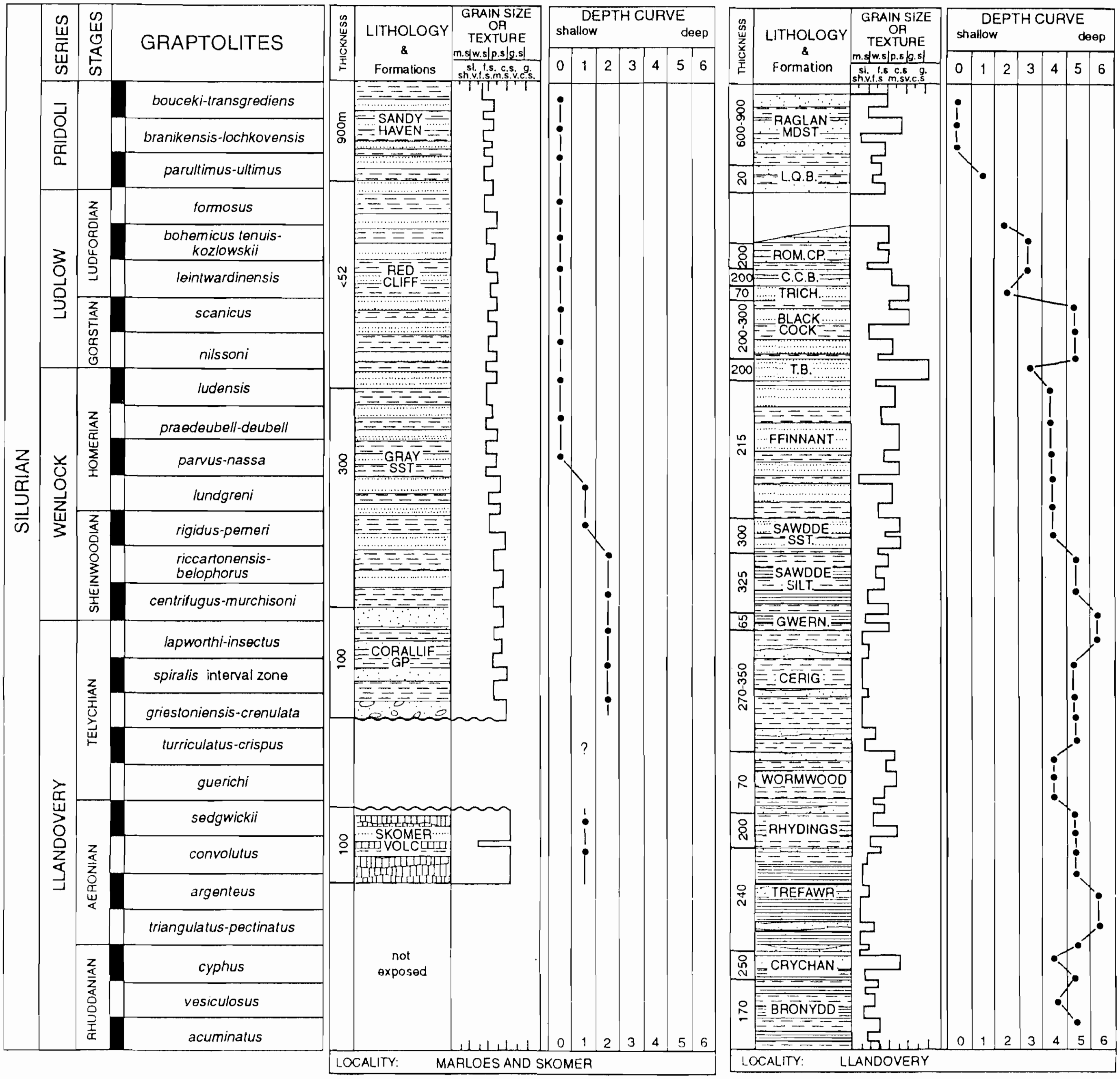

FIGURE 3 — *left*, Locality 1, Marloes and Skomer Island, southwest Wales. For location see Fig. 1. Corallif Gp, Coralliferous Group; *right*, Locality 2, Llandovery area, Powys, south central Wales. Raglan Mdst. Is Raglan Mudstone Formation; LQB; Long Quarry Beds; Rom CP, Roman Camp Beds; CCB, Cwm Clyd Bed; Trich, Trichrug Beds; TB, Tresglen Beds; Gwern, Gwernfelen Formation.

ing Wenlock and Llandovery (Butler et al., 1997). The rich shelly succession in the Upper Silurian was described in detail by Walmsley (1959). Bassett (1974) recorded *Monograptus flemingii flemingii* near the exposed axis of the anticline. The latter species indicates a post-*Monograptus riccartonensis* Zone-age for that part of the Coalbrookdale Mudstone. The British Geological Survey (Squirrel and Downing, 1969) confirmed Walmsley's (1959) work, but employed the stratigraphical nomenclature from the type Ludlow area, which we follow herein.

TORTWORTH–TITES POINT — These two outcrop areas are geographically separated; however, they are probably

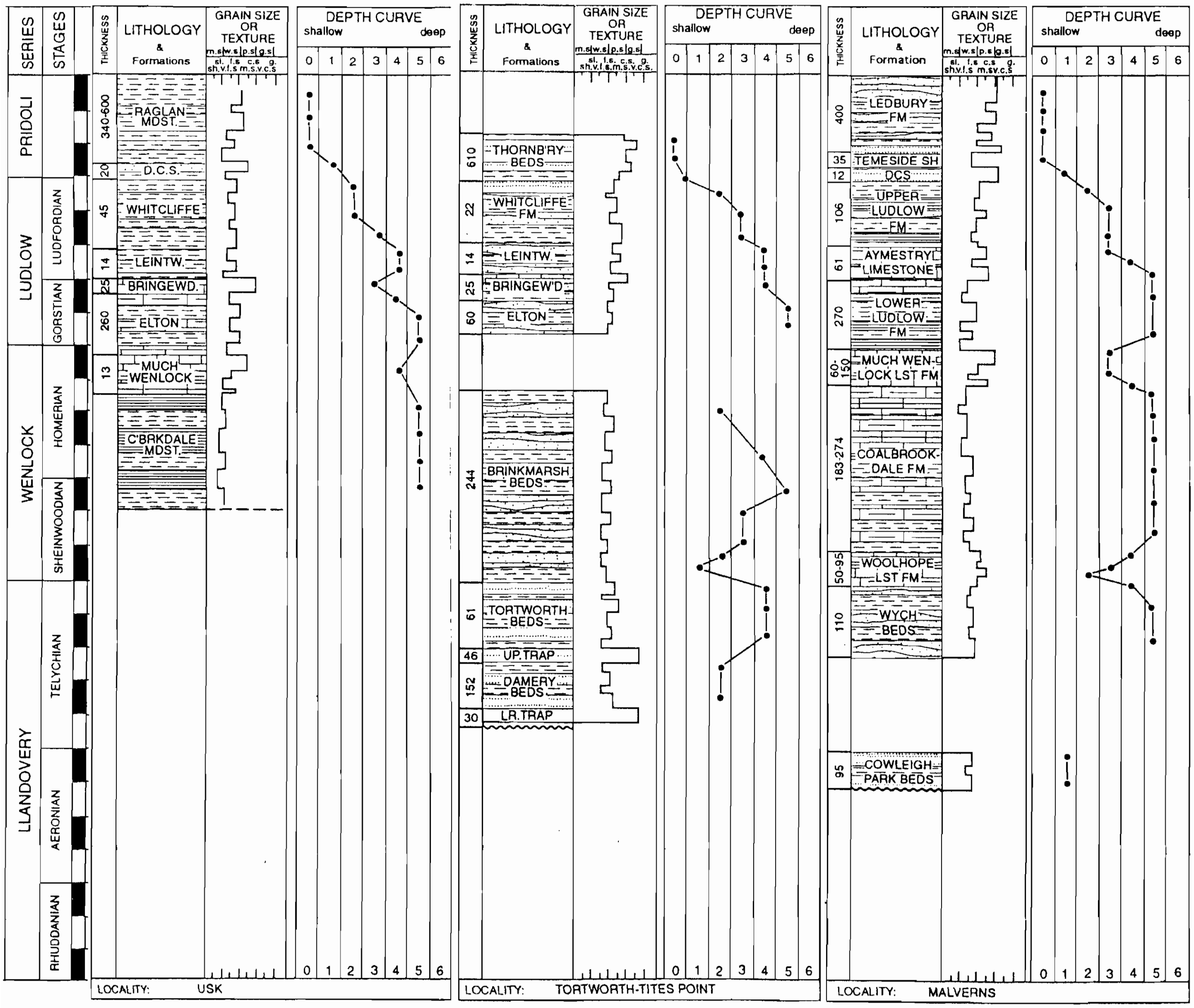

FIGURE 4 — *left*, Locality 3, Usk Inlier, Gwent, southeast Wales; DCS, Downton Castle Sandstone Formation; Leintw, Leintwardine Formation; C'brkdale Mudst, Coalbrookdale Mudstone Formation. *center*, Locality 4, Tortworth and Tites Point, Avon, western England. For location, see Fig. 1. Leintw, Leintwardine Formation; Up Trap, Upper Trap (andesite); Lr Trap, Lower Trap (andesite). *right*, Locality 5, Malvern Hills, Hereford and Worcester, western England. DCS, Downton Castle Sandstone.

connected at depth. The Tortworth Inlier north of Bristol includes Telychian to Homerian horizons, and the Tites Point and associated areas in Gloucestershire include rocks of early Ludlow to early Pridoli age (Fig. 4, center). Tortworth was investigated by Curtis (e.g. 1972), and the Tites Point area was investigated by Cave and White (1971); there is a further summary in Cave (1977). The unconformity below the upper Llandovery Damery Beds, within which are the Lower Trap volcanics, has been proven by exposures in temporary excavations. The Damery Beds have BA 2 *Eocoelia* Community assemblages, and the Tortworth Beds have BA 4 *Stricklandia* Community assemblages, although redeposited shell horizons rich in *Eocoelia* lie within them (Ziegler et al., 1968b). Ludlow rocks are present in the northern Tortworth Inlier, but the Wenlock is succeeded unconformably further south by the Upper Old Red Sandstone of Late Devonian age. At Tites Point itself, the Ludlow Bone Bed indicates a conformable passage into the Pridoli, but the overlying non-marine Thornbury Beds of

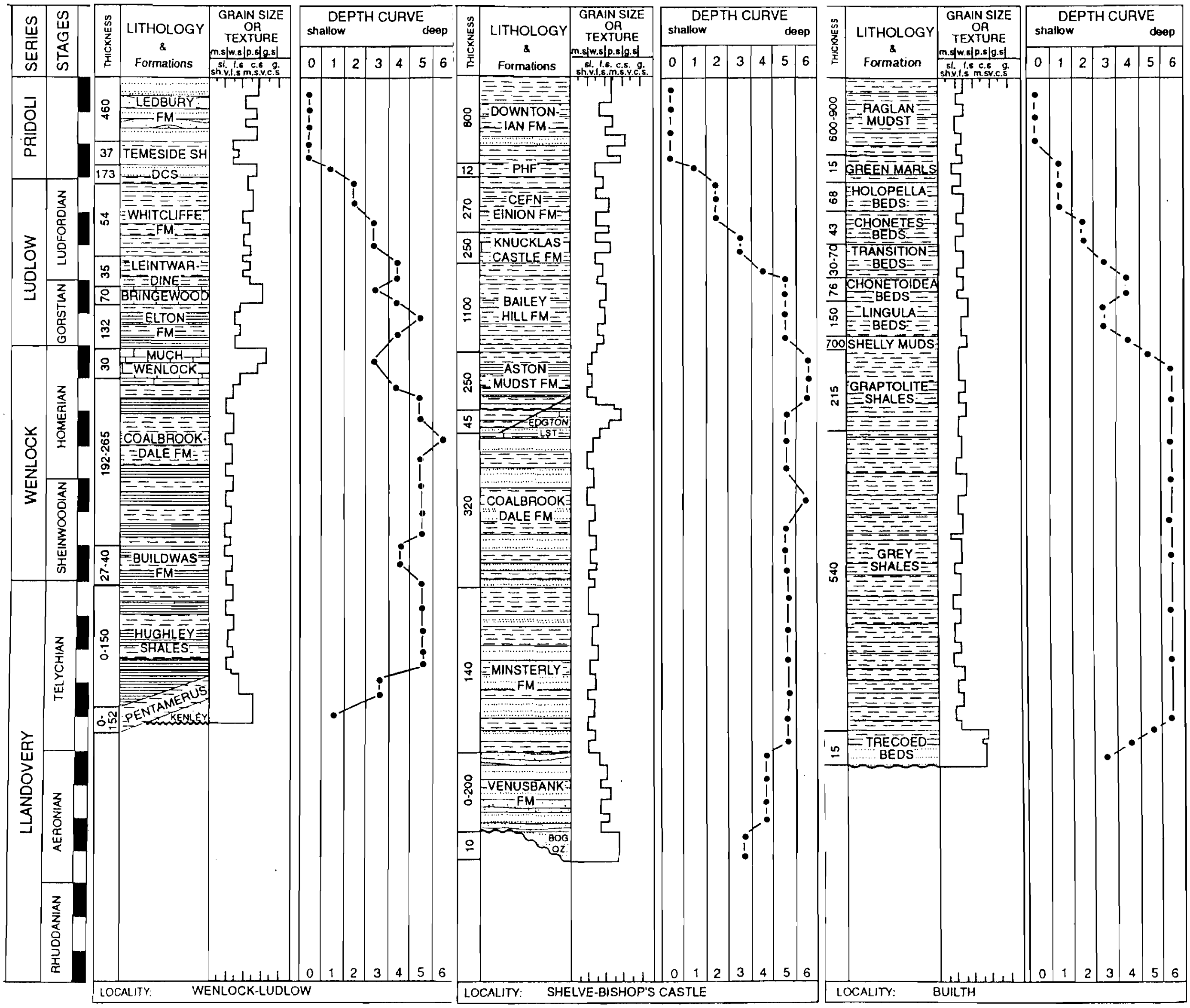

FIGURE 5 — *left*, Locality 6, Wenlock and Ludlow area, Shropshire, western England. DCS, Downton Castle Sandstone. *center*, Locality 7, Shelve and Bishop's Castle area, Shropshire, western England. For location, see Fig. 1. PHF, *Platyschisma helicites* Formation; Bog Qz, Bog Quartzite. *right*, Locality 8, Builth Wells area, Powys, eastern Wales.

Pridoli age are overlain unconformably by the Upper Old Red Sandstone.

MALVERN HILLS — The Precambrian rocks of this area mark a line of minimal subsidence through the Early Paleozoic. At the start of the Silurian, the Malverns formed part of a landmass that extended across much of the English Midlands (Bassett et al., 1992). The sea did not cover the area until the latest Aeronian, when the conglomerates, sandstones, and siltstones of the Cowleigh Park Beds with a BA1 *Lingula* Community were deposited (Ziegler et al., 1968b). The overlying rocks (Fig. 5, right) are latest Telychian; they show a gradually deepening sequence (from BA3 to BA5) of *Pentameroides*, *Stricklandia* and *Costistricklandia*, and *Clorinda* Community assemblages. As the assemblages above this early Telychian hiatus are deeper-water than those below, it has been suggested (Ziegler et al., 1968b, p. 776) that the sea may have been present in the Malvern area during this time, but that no sediments were deposited. The progressive Llandovery deepening continued through the early to middle Wenlock Woolhope Limestone and Coalbrookdale Formations (Hurst et al., 1978). A sequence deposited under shallower water is present in the upper Wenlock; the shallower (BA 2, BA 3) faunal assemblages present in the Malverns occur in the locally developed Much Wenlock Limestone Formation (in contrast to the Shropshire region further north). The Ludlow starts with a sharp return to deeper (BA4, BA5) environments (Hurst et al., 1978), but there was a gradual shallowing to non-marine red mudstones and sandstones of the Ledbury Formation after the Gorstian (Phipps and Reeve, 1967; White et al., 1984). Overlying non-marine, Old Red Sandstone facies of Pridoli and later age are described in Mitchell et al. (1961).

WENLOCK–LUDLOW — These two adjacent areas in Shropshire include the type areas for the second and third series of the Silurian System. They lie to the east of the major Church Stretton tectonic lineament, where the Silurian is not strongly deformed (in contrast to the cleaved rocks of similar sedimentary facies to the west of the lineament). The Llandovery rests unconformably on Caradocian and older rocks, and the deposits contain a BA 1 *Lingula* Community in the Kenley Grit (Fig. 5, left). BA 2 *Eocoelia* and BA 3 *Pentamerus* Communities appear in the succeeding siltstones and mudstones of the *Pentamerus* Beds, and a BA 5 *Clorinda* Community appears in the Hughley Shales (Ziegler et al., 1968a, 1968b). As elsewhere in western England, the Llandovery progressively deepens through time. The conformably overlying Buildwas and Coalbrookdale formations (Wenlock) carry mostly deeper-water, BA 5 community faunas, with sporadic development of BA 4 shelly and BA 6 graptolite-dominated faunas. Gradual shallowing through BA 4 shelly assemblages, with an increasing proportion of carbonate beds within the dominant shales, precedes the Much Wenlock Limestone Formation, which generally has BA 3 shelly assemblages and bioherms (Bassett et al., 1975; Hurst et al., 1978). There is a dispute whether the top of the Much Wenlock Limestone is diachronous, but we follow Hurst (1975) in believing it to be diachronous. In the Much Wenlock area, the Much Wenlock Limestone is conformably overlain by the siltstones and fine sandstones of the Elton Formation, which can be traced southwards to the type Ludlow area (Holland et al., 1963; White and Lawson, 1978). Above the Much Wenlock Limestone, the basal siltstones of the Elton Formation contain BA 2–3 assemblages. The sudden deepening that is characteristic of the base of the Ludlow in many places (McKerrow, 1979) occurs 5 to 10 m above the limestone. Hurst (1975) gave cogent reasons why this widespread deepening event is more likely to be synchronous than diachronous and local in its switch from carbonate to siliciclastic sedimentation. The Elton Formation includes BA 3 to BA 5 shelly faunas, and is succeeded by the Bringewood Limestone, with horizons dominated by the large pentamerid *Kirkidium* (BA 3). The overlying Whitcliffe Formation has faunas that decrease in diversity upwards and culminate in the remanié Ludlow Bone Bed at the base of the Downton Castle Sandstone. The latter carries sparse BA 1 ostracode and rare brachiopod faunas (Bassett et al., 1982). Higher strata of the Downton Castle Sandstone and Temeside Shales represent non-marine and deltaic environments (Allen and Tarlo, 1963), followed by the Old Red Sandstone facies of the fish-bearing Ledbury Formation (White, 1950). More details are given on Wenlock–Pridoli communities in this region by Boucot and Lawson (1999).

SHELVE–BISHOP'S CASTLE — West of the Church Stretton tectonic lineament, the folded rocks of the Precambrian–Caradocian Shelve Inlier are unconformably overlain by middle Llandovery (Aeronian) rocks. These rocks include the Bog Quartzite and the Venusbank Formation (Fig. 5, center), which yield diverse BA 3 rocky bottom and BA 4 *Stricklandia* Communities, respectively (Ziegler et al., 1968b). Bridges (1975) described transgression over a hard substrate in this area. Above these lower two units, the Minsterley Formation yields chiefly BA 5 *Clorinda* Community assemblages in facies that range from outer-shelf siltstones to thin turbidites. The Coalbrookdale Formation (Wenlock) overlies the Minsterley Formation and is relatively sparsely fossiliferous, with deeper-water outer-shelf benthic assemblages interspersed with thin turbidites and occasional horizons rich in graptolites. The lenticular Edgton Limestone is a deeper-water limestone with a high proportion of shale interbeds and relatively few shells. The Aston Mudstone is also outer-shelf and poor in fossils, and has been lithologically correlated with the lower Ludlow Elton Formation in the type Ludlow area. The higher beds shown in this column are closely comparable to the Upper Silurian of the Knighton area (Holland, 1959), and a very similar lithological classification is followed, apart from the recognition of the Cefn Einion Formation for the highest Ludlow unit (Cocks et al., 1992). Three small igneous intrusions near Bishop's Castle apparently record local Silurian volcanism (Sanderson and Cave, 1980). The Downtonian Formation is usually divided into the lower Yellow Downtonian (ca. 10 m thick), the middle Green Downtonian (20–100 m) and the upper Red Downtonian, which is over 700 m thick and extends up to include a non-marine, fish-bearing, Old Red Sandstone facies of Early Devonian age.

BUILTH — Ziegler et al. (1968b) reviewed the Llandovery faunas of this area and reported a BA 3–4 *Pentamerus* and *Stricklandia* assemblage from the Trecoed Beds (Fig. 5, right). This assemblage includes *Stricklandia laevis*, an early Telychian species. The Trecoed Beds are unconformable on the Middle Ordovician. The overlying Wenlock is represented by a continuous graptolitic sequence, and most of the Wenlock graptolite zones are present up to the *Cyrtograptus lundgreni* Zone (Elles, 1900; Wood, 1900; Hurst et al., 1978). There is no evidence of the basal Wenlock unconformity suggested by Jones (1947). Water depths became shallower in the latest Wenlock and through the Ludlow, where the basinal siltstones have yielded a variety of shelly faunas (Straw, 1937). These earlier records have been translated into BA numbers on our chart based on the communities described by Calef and Hancock (1974). The overlying Pridolian Raglan Mudstone is a non-marine Old Red Sandstone facies.

RHAYADER–KNIGHTON — Rhayader is on the eastern margin of the central Wales Basin. During all of the Llandovery and early Wenlock, the Silurian of this region was deposited well below the depth range of shelly faunas. The graptolites recovered represent almost all the zones known from this substantial time interval (Fig. 6, left). At various periods (e.g., during deposition of the late Aeronian Caban Conglomerate), very substantial olistostromes filled canyons at the basin edge, and most of the other deposits represent some kind of turbidite (Kelling and Woodlands, 1969). BA 5 assemblages appear in the Homerian. Later on, the basin was not as deep, and various shelly assemblages are known in the Ludlow, particularly from the Knighton area (Holland, 1959), where the succession is lithologically similar to the sequence northeast near Bishop's Castle. Above the Ludlow, the Pridoli again represents the non-marine regime of the Old Red Sandstone facies.

DENBIGH AND CONWAY — The Silurian rocks of this area have been remapped by the British Geological Survey (Warren et al., 1984). The Llandovery consists of a relatively narrow outcrop belt of mudstones and siltstones, and nearly all of the Llandovery graptolite zones are known. The Wenlock and Ludlow have much more substantial outcrops and include many slumps which carry shelly faunas, in addition to the graptolites shown on the column (Fig. 6, center). No deposits later than very early Ludfordian are known below the unconformity with the Carboniferous.

ENGLISH LAKE DISTRICT — A great deal of new work has been carried out in the Ordovician and Silurian of this area by the British Geological Survey and others. The transition from the type Ashgillian into the Silurian occurs within deep-water mudstones (Fig. 6, right), and the whole of the Llandovery, Wenlock, and Ludlow is represented by BA 6 graptolite assemblages (Kneller et al., 1994). The mudstones were succeeded by coarse-grained turbidites in the late Homerian (Rickards, 1978, p. 140), when the rates of subsidence increased. Increased subsidence suggests development of a foreland basin (Kneller, 1991; King, 1994), perhaps related to the subduction of English continental crust (then part of Avalonia) below Laurentian Scotland (Leggett et al., 1983; King, 1994). At the top of the Silurian, the Scout Hill Flags show a slight shallowing, and have a sparse Pridoli fauna of ostracodes and other forms. The Scout Hill Flags are shown here as a separate formation above the Kirkby Moor Flags, as suggested by most authors (see Cocks et al., 1992), but others prefer to treat these beds as the Helm Member within the Kirkby Moor Flags.

BALBRIGGAN, IRELAND — The Silurian of this inlier north of Dublin (Fig. 1, locality 6) is entirely graptolitic (Fig. 7, left) and is very similar to that of the Lake District, although sand-grade turbidites occur in the Telychian as well as the Homerian. The youngest exposed fossiliferous beds are Homerian, though the coarse turbidites of the Skerries Formation may extend up into the Ludlow (Rickards et al., 1973). There has been some debate about which areas of modern central Ireland formed part of Avalonia (e.g., Harper and Parkes, 1989). The Silurian deposits in central Ireland consist mainly of deeper-water graptolite shales and turbidites, which are of little help in resolving the problems. However, there is little doubt that eastern inliers such as Balbriggan lay on the Avalonian side of the Iapetus Ocean.

DINGLE — The oldest beds exposed in this area (Fig. 1, locality 13) are mainly siltstones of the Ferriters Cove Formation (Fig. 7, center), which has Wenlock fossils (Bassett et al., 1976; Holland, 1988). The lower beds contain a *Dolerorthis rustica* Association, which can be interpreted as a mid-shelf environment (BA 3), and the majority of the formation contains brachiopods and corals indicative of shallower, storm-influenced environments (Watkins, 1978). The overlying Clogher Head Formation is dominantly volcanic, but contains some rhynchonellides characteristic of BA 2 (Holland, 1988). It is overlain successively by the unfossiliferous red and grey sands and tuffs of the Mill Cove Formation and by the green and gray siltstones of the Drom Point Formation. The latter formation contains the rhynchonellide *Sphaerirhynchia wilsoni*, which probably indicates BA 2 (Watkins, 1978). The Ludlow is represented by Croaghmarhin Formation siltstones, which shows a return to deeper-water environments, with BA 3 *Rhipidium* and coral faunas in the lower part of the formation and BA 4–5 brachiopods (*Dayia*, *Isorthis*, *Shaleria*) at higher levels. The topmost marine beds show some shallowing and have rhynchonellides typical of BA 2. The Croaghmarhin Formation is overlain, apparently conformably, by the nonmarine red sandstones of the Dingle Group, which are probably late Ludlow or Pridoli based, on spores (Holland, 1987, 1988).

NEW WORLD ISLAND, NEWFOUNDLAND — It has long been known that the Avalon Peninsula in eastern Newfoundland contains *Paradoxides* and other Cambrian trilo-

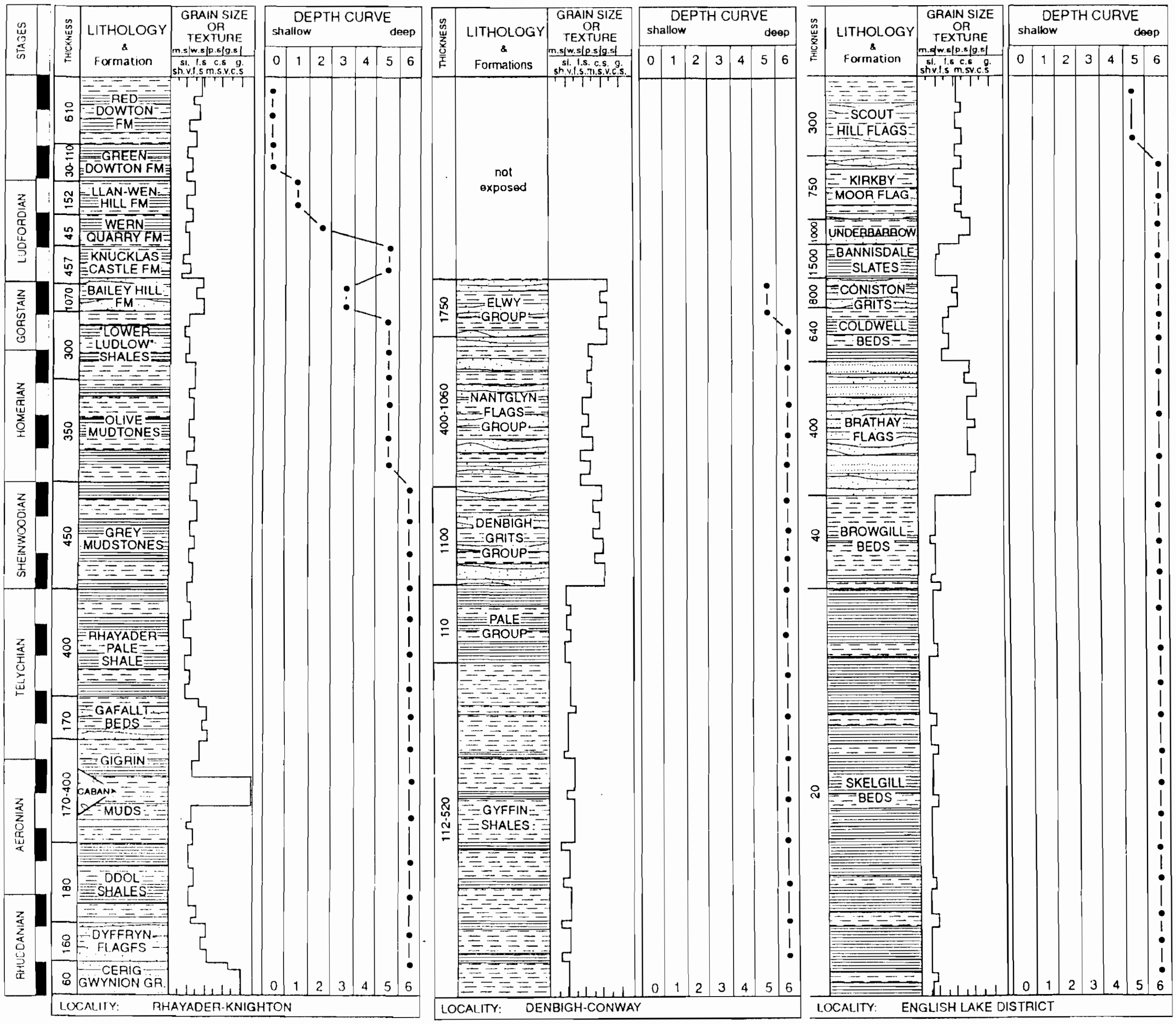

FIGURE 6 — *left,* Locality 9, Rhayader area, east-central Wales. *center,* Locality 10, Denbigh and Conway area, North Wales. For location, see Fig. 1. *right,* Locality 11, the Lake District, northwest England.

bites which show strong affinities with southern Britain (then part of Gondwana). Similar faunas occur also in Nova Scotia, coastal New Brunswick, and New England (Conway Morris and Rushton, 1988; Nowlan and Neuman, 1995), where they are also interpreted as lying on Avalonian (Gondwanan) crust (Rast, 1980; Rankin et al., 1989, p. 79; van Staal et al., 1996). Cambrian faunas similar to these also occur in the southern Appalachians, but they are geographically widespread and occur in beds which were probably accreted from Gondwana to Laurentia in the Alleghanian orogeny and were never part of Avalonia.

The "Central Mobile Belt" of Newfoundland (and its equivalents in other parts of the northern Appalachians) is now known to consist of many small terranes, some of which collided with Laurentia in the Ordovician Taconic orogeny. However, paleomagnetic data from Early Ordovician rocks show that the terranes now lying east of the Red Indian Line (in both Newfoundland and New Brunswick) were then at high southern latitudes, as was Avalonia, and substantially different from the Early Ordovician equatorial latitudes of contemporary Lauren-

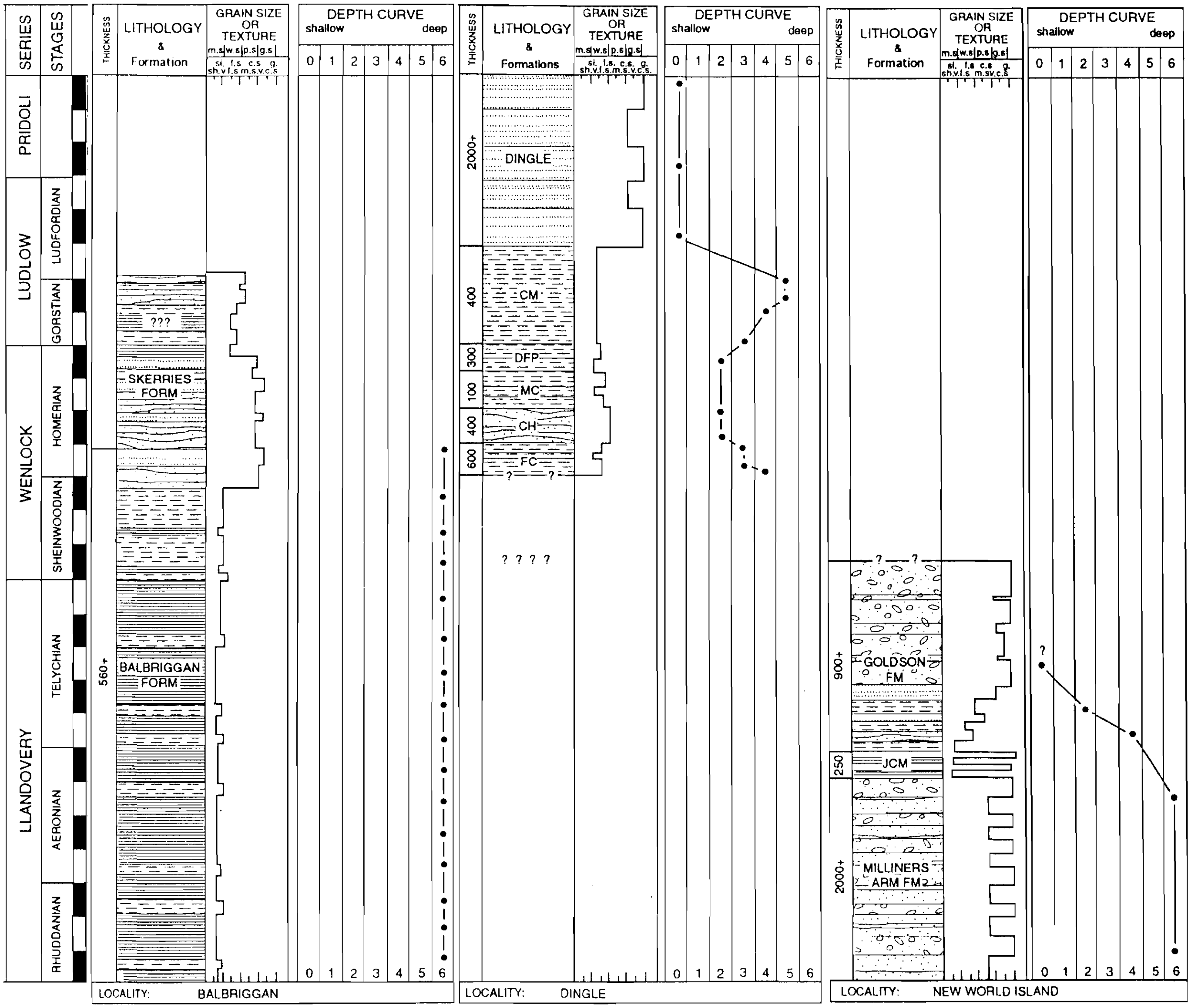

FIGURE 7 — *left,* Locality 12, Balbriggan, eastern Ireland. *center,* Locality 13, Dingle Peninsula, western Ireland. For location, see Fig. 1. CM, Croaghmarhin Formation; DFF, Drom Point Formation; MC, Mill Cove Formation; CH, Clogher Head Formation; FC, Ferriters Cove Formation. *right,* Locality 14, New World Island, north-central Newfoundland. For location see Fig. 2. JCM, Joey's Cove Mudstone Formation.

tia (van der Voo et al., 1991; van Staal and van Roo, 1995; Cocks et al., 1997). These terranes, which appear to have migrated north from high southern latitudes during the Ordovician, include the Summerford and Chanceport Groups of New World Island. Although the Iapetus Ocean substantially narrowed in the Silurian, these Silurian sequences are now thought to have developed on Avalonia, and not on Laurentia, as previously suggested (McKerrow and Cocks, 1977). The presence of the Llandeilian warm water Cobbs Arm Limestone, which yields brachiopods of Laurentian affinity on New World Island, suggests at least intermediate latitudes. However, the paleomagnetic data (van der Voo et al., 1991) shows that these terranes were distinct from Laurentia until the convergence of Avalonia and Laurentia in the Early Silurian.

There are several Silurian sequences on the north coast of Newfoundland between the Red Indian Line and Avalon *sensu strictu.* We illustrate them with the section in New World Island (Fig. 2) exposed northwards from Cobbs Arm (Fig. 7, right), where coarse deep-water turbidites in the upper Milliners Arm Formation continue from the Ashgill into the Llandovery (Arnott et al., 1985; Williams et al., 1993). Redeposited Aeronian brachiopods occur locally in the Milliners Arm Formation (McKerrow and Cocks, 1978). The succeeding Joey's Cove Melange contains blocks of various Ordovician rocks (including the Cobbs Arm Limestone). Above this, the Goldson Formation consists of sandstones with *Stricklandia lens progressa* in growth position, which are overlain by sandstones with *Eocoelia curtisi.* This indicates a progressive shallowing from BA 4 to BA 2 in the Telychian (Arnott et al., 1985).

ARISAIG, NOVA SCOTIA — Although the Silurian occurs over much of the Avalonian northern Appalachians (including Nova Scotia, southern New Brunswick, coastal Maine and New Hampshire, eastern Massachusetts, and Rhode Island), most of these sequences are much disturbed by faults and Acadian deformation. Only at Arisaig on the north coast of Nova Scotia (Fig. 2) is there a complete, largely continuous sequence from the lower Llandovery to the Pridoli (Fig. 8, left). The majority of the brachiopod-dominated assemblages are assigned to BA2, but there are some relatively thin intervals that represent deeper-water facies in the Aeronian, basal Wenlock, and basal Ludlow (Boucot et al., 1974; Watkins and Boucot, 1975; McKerrow, 1979; Hurst and Pickerill, 1986). The higher part of the succession includes thin, nonmarine facies at the end of the Wenlock, and shows a general shallowing from BA 2 through BA 1 to BA 0 in the Pridoli and basal Lochkovian.

NEW BRUNSWICK AND NEW ENGLAND — Silurian shelf sandstones and shales occur in coastal New Brunswick, coastal Maine, and eastern Massachusetts (Fig. 2) in regions which are underlain by rocks with Cambrian and Ordovician faunas of clear Avalonian affinities. Inland, the boundaries of Avalonia with Laurentia have been revised to include the Bronson Hill anticlinorium and Popelogan arc rocks within Avalonia (Cocks et al., 1997). The Silurian sedimentary rocks in these inland areas contain greywackes of the Merrimack Trough, the Fredericton Trough, and the Matapedia Basin (McKerrow and Ziegler, 1971; Osberg et al., 1989; Williams, 1995). There are also several volcanic sequences. Conditions were not favorable for benthic faunas in most of these inland areas of the northern Appalachians. However, scattered collections of Telychian brachiopods in northern New Brunswick suggest shallower communities around the Matapedia Basin, with deeper ones (up to BA 4) towards the center (McKerrow and Ziegler, 1971). In the modern Atlantic coastal areas, most of the shelf successions are tectonically disturbed (van Staal, 1987, 1994). As a result, there are no continuous measured sections through several formations, and no section has been good enough to include as a separate column in this paper. Watkins and Boucot (1975) indicated that the variety of shelf assemblages present in the Llandovery decreases upwards, so that BA 2 assemblages with *Protochonetes* and *Salopina* become dominant in the Pridoli.

SOUTHEAST ENGLAND — Correlation columns cannot be shown for this region, as the known sections are too incomplete and no formal formation names exist in the area. However, Silurian rocks have been encountered in many boreholes in southeast England (Cocks et al., 1992; Woodcock and Pharoah, 1993). The four facies belts that have been recognized probably represent a transition from an anoxic basin slope or outer shelf in the northeast to a storm-dominated inner-shelf area to the southwest. Most of the Silurian rocks in the boreholes have been dated by microfossils (Molyneux, 1991), though some samples contain macrofossils. The boreholes have penetrated upper Llandovery to Pridoli rocks, and clearly represent the northeastern margin of Avalonia—a margin which extends southeastward to north of the Brabant Massif, Belgium.

BRABANT MASSIF, BELGIUM — The Brabant Massif (Fig. 1), part of the larger Anglo-Brabant Fold belt, was folded, cleaved in most places, and faulted by suspected thrust faults during the Acadian orogeny, and is now metamorphosed from zeolite to greenschist facies. The Silurian of the Brabant Massif was reviewed by Verniers and Van Grootel (1991). The anticlinorial structure results in several Silurian outcrop areas in the south and a poorly known subcrop area in the north. At the southern margin, the Silurian is present in the east and west. In the subsurface, it is present in a large area of southern Flanders that continues under northern France to the Boulonnais area, and in a central rim in the northwest. In the northern part of the massif, three boreholes reach the Silurian in Belgium, and only one in the Dutch sector of the North Sea (Legrand, 1967; Verniers and Van Grootel, 1991). Graptolites are often the only macrofaunal elements present and have been studied from the Llandovery–lower Ludfordian, and most of the older-known graptolite zones are recognized. However, more work is necessary to locate several horizons in the graptolite zonation used in this report. Organic microfossils, such as acritarchs and chitinozoans, are used to date thick units in the succession that lack graptolites. The following chitinozoan zones (Verniers et al., 1995) have been demonstrated: *Belonechitina postrobusta*, *Spinachitina maennili*, *Eisenackitina dolioliformis*, *Angochitina longicollis*, *Margachitina margaritana*, *Cingulochitina cingulata*, *Sphaerochitina lycoperdoides,* and *Angochitina elongata* (Martin, 1969; Verniers, 1983; Van Grootel, 1990; Louwye et al., 1992). A local, more refined chitinozoan zonation was proposed for the Wenlock of the Mehaigne area (Verniers, 1999). Several parts

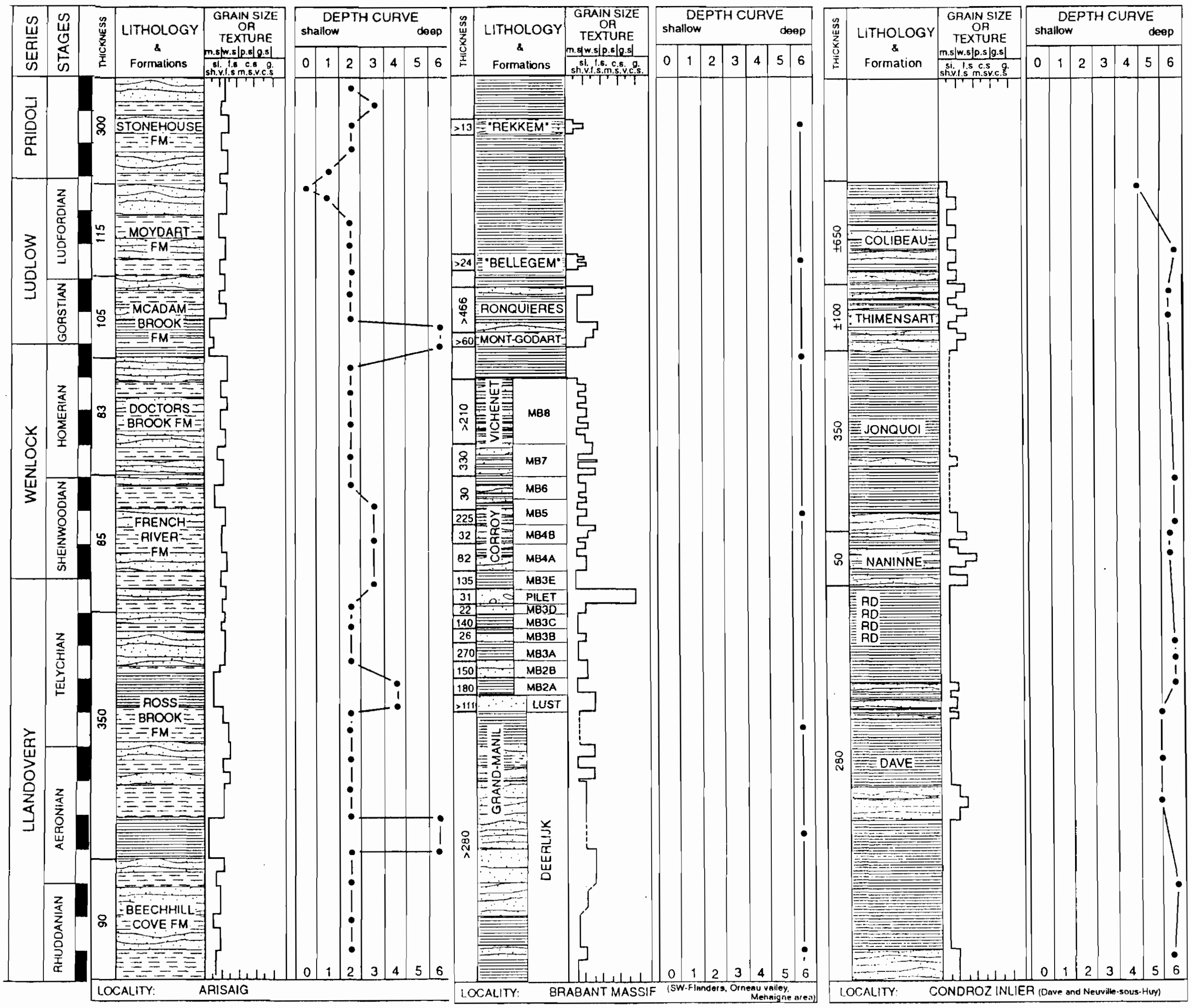

FIGURE 8 — *left*, Locality 15, Arisaig, Pictou County, Nova Scotia. For location, see Fig. 2. *center*, Locality 17, Brabant Massif, Belgium. For location, see Fig. 1. Numbered members discussed in text. *right*, Locality 18, Condroz inlier, Belgium. RD is purple-red intervals in the Dave Formation.

of the lithostratigraphy in outcrops and boreholes are still not known in detail, especially much of the Llandovery, Ludfordian, and Pridoli. The Deerlijk and Lust Formations are defined in the subsurface of southern Flanders. The first contains greenish and dark gray, compact or laminated soft shale and clayey sandstone, with alternating deposits either of distal turbidites or (laminated) hemipelagites. The rocks are often rich in graptolites, with rare orthocone nautiloids, sponge spicules, foraminiferans and benthic algae (Martin, 1974). The Lust Formation is similar to the Deerlijk, but contains up to 45% of shaly fine sandstone. In the outcrop area, the green, soft, Llandovery shales are grouped within the Grand-Manil Formation. As the overlying Wenlock, the Llandovery Corroy and Vichenet Formations are poorly studied.

In the Mehaigne area (Fig. 8, center), an informal lithostratigraphy shows a thick green shaly MB3 formation with several members at the top of the Telychian. A rapid increase in energy of the turbidites starting across the Llandovery–Wenlock boundary culminated at the top of both members during MB4A (*Monograptus riccartonensis* Zone) and MB4B (*Cyrtograptus rigidus* Zone). Within MB4A, the green color typical of the Llandovery and lower Wenlock units is replaced by the gray of middle Wenlock to Pridoli units. Graptolites are regularly found in both members, together with rare *Conularia* sp. In the base of the Ronquieres Formation, *Cardiola* sp., *Orthoceras* sp., and bioturbation are known. Thicker-bedded but less energetic turbidites occur in formations MB5–8. They are mostly fine silty slate, clayey siltstone, and (rarely) very fine-grained sandstone, with more sandstone in MB6 and very thick, slightly calcareous beds in MB8. The Mont Godart and Ronquières Formations in the Sennette Valley contain turbidites with a smaller proportion of intercalated, laminated hemipelagites (LHP). This facies forms a shallow–deep–shallow megacycle with some variations. Unit B, near the base of the Ronquières Formation, displays 30 m of LHP with six metabentonite levels in the deepest part of the megacycle, and the higher unit (K) shows another deepening event. The Mont Godart has now been included as a member of the Ronquieres Formation (Verniers et al., 2001). Ludfordian or Pridoli stratigraphic units have not been defined, and are only found in a few outcrops or boreholes as gray thin-bedded turbidites, often with burrowed horizons and traces of the archaeostracan *Ceratiocaris* sp. To produce the pre-Givetian cleavage, some authors have suggested continuing sedimentation in the Brabant Massif through the Silurian, with another 3 to 6 km of post-Gorstian rock. Using reworked palynomorphs in the Lower Devonian south of the massif, Steemans (1989) postulated that sedimentation continued in the Brabant Massif until the early Lochkovian. The Stratigraphic Commission of Belgium has recently redefined Silurian stratigraphy, and has now formalized many of the hitherto informal units in the Mahaigne area (Verniers et al., 2001).

CONDROZ INLIER, BELGIUM — This inlier, also called the Condroz Ridge or Sambre and Meuse Belt (or Strip), was gently folded but not cleaved in the Pridoli to early Lochkovian, and was later deformed by the Variscan orogeny. An overview of the literature is given in Michot (1954) and Martin (1969). The Condroz Inlier is a 69 km-long, 0.5–2.0 km-wide, narrow strip located south of the Sambre and the Meuse Rivers. It extends from east of Charleroi to southwest of Liège. Its location and deformation along the Variscan front between the Namur and Dinant synclinoria are the cause of the discontinuous sections. In contrast to the Brabant Massif, the Llandovery to Ludlow (Fig. 8, right) is much thinner, non-turbiditic, and richer in macrofossils (mostly graptolites, with rare brachiopods, trilobites, crinoids, ostracodes, conularids and *Orthoceras* sp.). The graptolite studies all date from before Maes et al. (1979). Most of the graptolite zones as defined by Rickards (1976) have been found, except the *Cystograptus vesiculosus, Coronograptus cyphus, Gothograptus nassa,* and *Monograptus ludensis* Zones, but a modern review is necessary. Acritarch studies have elucidated a few of the previously undated sections (Martin, 1969), and unpublished chitinozoan studies show the presence of the *Eisenackitina dolioliformis, Angochitina longicollis, Margachitina margaritana, Cingulochitina cingulata,* and *Angochitina elongata* Zones. The rocks are mostly shales, siltstones, and some fine-grained sandstones and calcareous shales. The dark gray or greenish Dave Formation contains silty shale with some sandstone and sandier parts in the *Monograptus convolutus* and the *Spirograptus turriculatus* Zones. Several purple-red intervals occur in the *Monoclimacis griestoniensis* and *Monoclimacis crenulata* Zones and in the lower Sheinwoodian. The Naninne Formation shows a rapid change to green or gray, laminated, locally calcareous silty shales, fine sandstones, and shales (Michot, 1954). The Jonquoi Formation marks the return to laminated or compact grayish-brown or green shales with some levels of calcareous nodules (Michot, 1954; Maes et al., 1979). The Thimensart Formation contains gray, laminated, fine-grained sandstone and olive-green shale, and the poorly studied and supposedly thick Colibeau Formation consists of dark shale with minor fine sandstone beds. In the Telychian part of the Dave Formation and in the Naninne Formation, eleven acid volcanic and volcano-sedimentary layers are present.

PATTERNS IN LATERAL AND VERTICAL LITHOSTRATIGRAPHY

In contrast to Laurentia, Baltica, and above all Gondwana, Avalonia was a relatively small paleocontinent. It also possessed active margins and, in addition, collided progressively with both Baltica and Laurentia during the Silurian and Early Devonian. Thus, with the possible exception of the Avalon Peninsula of Newfoundland and the Midlands Microcraton of England, the whole of Avalonia was subject to constant tectonic activity during the Silurian. The result of this is that the patterns to be seen in both lateral and vertical lithostratigraphy and biostratigraphy are essentially brief and relatively local in extent, in contrast to the widespread and relatively uniform distribution of rocks and biota on the larger paleocontinents.

During most of the Llandovery, the lithofacies consisted almost entirely of siliciclastic rather than calcareous rocks. These siliciclastics varied from the shallower-shelf lithologies around such land areas such as central England to outer-shelf and slope deposits, which include submarine canyon fills such as those found at Rhayader, Wales (Kelling and Woollands, 1969). Intra-terrane basin areas, such as west-central Wales, are known, where turbidites from 3.0 m down to a few millimeters thick filled tectonically activated depressions to form rocks of very substantial thickness. Woodcock (1990) has defined all the Silurian deposits as a single supergroup within the Lower Paleozoic succession of Wales. On the shallow shelves of the Welsh Borderland and adjacent areas in Wales and England, a transgression has been well known since pioneer work by A. C. Ramsay and others in the nineteenth century. The timing of the various pulses of this transgression has been documented by Ziegler et al. (1968b). The latter authors dated the shelly rocks that encircle the shelf in a wide arc from Meifod and the Breidden Hills in the north; through the Shelve-Longmynd Inlier and Wenlock Edge areas (Cocks, 1989); and southwards through the Malvern Hills, the Woolhope Inlier, and May Hill to the Tortworth Inlier near Bristol. Within this central arc lie the outcrops at Presteigne, Old Radnor, Builth Wells, Garth, and the type Llandovery area. Outside the arc are sporadic outcrops near Birmingham (Walsall, Great Barr, and Rubery) and Coventry, as well as Llandovery of the Lower Lemington borehole in Oxfordshire. These disparate outcrops have yielded a wealth of data which demonstrate ever-deepening water depths with time in any particular area as the Llandovery transgression progressed unconformably over Precambrian to Ordovician rocks. This unconformity surface was a seafloor with very variable topography. Bridges (1975) has documented such a varied topography in Shropshire and described the sedimentology of these shallow-water siliciclastics. Woodcock et al. (1996) analyzed the Late Ordovician and Early Silurian sequence stratigraphy of the Welsh Basin, and concluded that the depositional patterns reflect contemporary tectonic activity and not proposed global patterns of eustatic shallowing and deepening.

The oldest substantial deposits of Silurian carbonates in Avalonia are lower Wenlock and include the Nash Scar Limestone of east-central Wales. This is a thick development of massive, dark gray to white, crystalline, algal and bryozoan limestone that ranges in thickness from 24 m to 60 m (Bassett, 1974). In the Welsh Borderland, there are more carbonate rocks of the same age, such as the Woolhope Limestone of the May Hill, Woolhope, and Malvern areas, but these are not so massive, and represent relatively thin limestones with siliciclastic interbeds. At the end of the Wenlock, there was another development of massive limestones, in particular the Much Wenlock Limestone of Shropshire and the Dudley Limestone in central England. This limestone development is known as far east as the Ware borehole in Hertfordshire. Within the Much Wenlock Limestone, substantial biotherms are developed at intervals (Bassett et al., 1975). In the Ludlow, the Welsh Borderland rocks are chiefly represented by siltstones and shales, with sporadic local development of such carbonate rocks as the Aymestrey Limestone. These represent a variety of shelf sedimentary regimes, including storm deposits (Tyler and Woodcock, 1987). Energy conditions varied from high to low, and the inshore areas in particular were marked by intermittent sedimentation that included erosional gaps and hardgrounds. Tectonic controls are reflected by downslope slumps and submarine channels and by low-energy basinal environments (Cherns, 1988). At the top of the Ludlow, faunas dwindle in both abundance and diversity and indicate a progressive decrease in marine-influenced deposition within the Welsh Borderland. These are succeeded by the remanié deposits of the Ludlow Bone Beds and the overlying deltaic deposits of the Downtonian and lower Dittonian (Allen and Tarlo, 1963), which represent the first phases of the non-marine Old Red Sandstone rocks that continued into the Devonian. In the Welsh Basin, the Upper Silurian still includes turbidites of varying thickness and extent, although marine rocks later than Ludlow are unknown and extensive slumping is present, particularly in the Ludlow of North Wales (Warren et al., 1984). In the Lake District, the Llandovery is represented by thin turbidites and graptolitic black shales that total no more than 70 m. Much thicker turbidites were deposited in the Wenlock and Ludlow, with 3.7 km in the Ludlow of the Windermere area, before deposition ceased near the end of the Pridoli (Kneller et al., 1994).

Silurian igneous activity is marked by the Skomer Group volcanics in southwest Wales (Ziegler et al., 1969), which are middle Llandovery, and by equally substantial Wenlock volcanics in the East Mendips area southeast of Bristol (Hancock, 1982). There was volcanic activity in the Brabant Massif in the Llandovery, and from the Telychian to the Sheinwoodian in the Condroz Inlier. Throughout Wales, the Welsh Borderland, central England, the English Lake District, and Belgium, there is further evidence of volcanism shown by the many bentonites throughout the Llandovery, Wenlock, Ludlow, and Pridoli, of which many have been used for radiometric dating.

Throughout the Silurian, the sediment accumulation rate in the Brabant Massif increased drastically. The distal turbiditic sedimentation with minor pelagic intervals was moderate in the Llandovery, with a rather thick upper Telychian (about 624 m), and slowed in the shallower, lower Sheinwoodian, with fossils indicating BA 4 to BA 6 assemblages that were transported by turbidites into the basin. The sedimentary rocks change from green and clay-rich to gray and more quartz-rich in the *Monograptus riccartonensis* Zone. Sudden increases of sedimentation rate and depth started in the *Cyrtograptus rigidus* Chron, with thick but still mostly clay/silt-dominated turbidites, and continued up to the *Gothograptus nassa* Zone (*lycoperdoides* Zone). Above a poorly known, upper Homerian interval, a thick shallow–deep–shallow cycle is present at Ronquières near the Wenlock–Ludlow boundary to the top of the Gorstian. The supposedly very thick Ludfordian and Pridoli turbidites there are thin-bedded, often bioturbated, and very rich in leiospherid acritarchs.

Deposition in the Condroz Inlier was almost all in the graptolitic shale environment (BA 6), except at the top of the Colibeau Formation where such brachiopods as *Delthyris elevatus*, *Plethorhyncha percostata*, and *Strophеodonta simulans* indicate BA 3 to BA 5. Coarser-grained, laminated, and slightly calcareous intervals, such as in the *Monograptus convolutus*, *Spirograptus turriculatus*, and *Coronograptus cyphus* Zones in the Dave, Naninne, and Thimensart Formations, point to shallower but still deep-shelf environments. However, the lateral variability in lithofacies and faunas has not been studied enough for firm conclusions.

Thus the relatively coherent picture in the eastern part of Avalonia is that of a stable Midlands Microcraton with fringing and variably developed basins that surrounded it in the Welsh Borderland, English Lake District, eastern England, and Belgium. Southwards, thick Mesozoic cover and Variscan tectonics hide the true situation during Silurian times, but some authors have postulated a now-vanished land mass in that area. In western Avalonia, the picture is less coherent, and the land areas from which the Silurian sediments of Nova Scotia, Newfoundland, and elsewhere were derived are not well constrained.

Patterns in Lateral and Vertical Biofacies

Most of the Silurian rocks on the paleocontinent of Avalonia were deposited in water depths too deep to support shelly benthic communities. The substantial exceptions to this were the rocks deposited to the east and south of the Welsh Basin. It was here that Ziegler et al. (1968a) defined what has become the classic sequence of *Lingula, Eocoelia, Pentamerus, Stricklandia,* and *Clorinda* Communities, which were taken by Boucot (1975) as the blueprints for the definition of his benthic assemblages BA 1 to BA 5. Analysis of the distribution of these communities allowed the chronology of the Llandovery transgressions of the Welsh Borderland to be documented in detail (Ziegler et al., 1968b). Comparable Wenlock and Ludlow communities were defined by Calef and Hancock (1974), and their distributions were charted in the Wenlock by Hurst et al. (1978). The distribution of benthic animal communities in the Ludlow was complicated by the gradual diminution in open-ocean access to the Welsh Borderland as time progressed, but Watkins (1979) documented their diachronous and sometimes lithofacies-specific distribution in key sections in Shropshire and Herefordshire. There has been much discussion on the relationship of these communities to water depth and to such other factors as the composition and nature of the substrate, temperature, salinity, and exposure to bottom currents. However, the marked parallelism seen in many Silurian sequences from different terranes (McKerrow, 1979, and data in this report) strongly suggest that sea-level changes affected the distribution of the communities, and their relationship to water depth seems to be very strong. These sea-level changes were in turn caused by variable combinations of eustasy and local tectonics.

Climatic Indicators and Climatic Variations

Avalonia was characterized by the presence of cold faunas related to northwestern Gondwana in the Early Ordovician (in contrast to the progressive change towards warmer-water equatorial faunas in the eastern part of Gondwana at the same time [Cocks and Fortey, 1988]). This changed to a progressive climatic warming as Gondwana moved over the South Pole (and the northwestern part moved towards the equator) and Avalonia moved even faster to the north throughout the later Ordovician and Early Silurian (Cocks and Fortey, 1982). These latitudinal changes are also recorded in the paleomagnetic data (Torsvik et al., 1996), and are reflected in the changing sedimentary facies. During the Early Silurian, there were few carbonates in Avalonia. However, by the early Wenlock, a few massive carbonates were deposited, for example at Old Radnor, Wales, and subtropical climates are indicated by the sedimentology. In contrast, most of the environments present in Belgium were in water too deep to record the contemporary climate, and Silurian carbonates are largely absent from the North American part of the paleocontinent. The change in climate in Avalonia was probably due entirely to its northward migration toward lower latitudes. However, there is also a suggestion that warm climates were generally more prevalent in the Silurian than in the Ordovician (Scotese and McKerrow 1990, fig. 22).

Chronological Summary

During the Silurian, Avalonia collided sequentially with Baltica to the northeast and Laurentia to the northwest. The timing is reflected in sediment distribution, for example in the English Lake District (McCaffrey and

Kneller, 1996), where the thin Upper Ordovician to lower Llandovery is characterized by a progressive rise in the proportion of hemipelagic material derived from rocks to the south. From the late Llandovery on, petrographic and geochemical data indicate sediment supply from a recycled orogenic source and probably reflect the collision with Baltica. No conclusive evidence is recorded of Laurentian input until the late Wenlock. By the end of the Ludlow, an abrupt reversal in provenance records basin inversion and occlusion of most of the Iapetus Ocean. This pattern has been proposed for the whole paleocontinent by Soper and Woodcock (1990). That the paleocontinent was tectonically unstable is demonstrated by the extensive slumping to be seen in some areas, for example in the Ludlow of North Wales (Warren et al., 1984). Island arcs on the margins of Avalonia are better known in Ordovician rather than Silurian times. However, there is some Silurian volcanism, as seen in the Llandovery of Pembrokeshire (Ziegler et al., 1969) and Tortworth (Curtis, 1972); the Wenlock of the East Mendips area near Bristol (Hancock, 1982); and the many bentonite horizons in the Llandovery, Wenlock, and Ludlow of Wales, the Welsh Borderland and the English Lake District. The Tortworth volcanics may be part of a more extensive sequence of mainly intermediate-composition tuffs that are seen in boreholes and seismic profiles. The geochemistry of these tuffs suggests intra-plate rifting rather than a subduction zone (Woodcock, 2000). Rankin et al. (1989, p. 74) mentioned "Ordovician- to Devonian-aged alkalic plutonic rocks plus associated volcanic and volcaniclastic rocks" as forming part of the Avalonian rocks of New England, but there is as yet no consensus as to the precise pattern of igneous activity in the area. In particular, the volcanics of the Bronson Hill and Popelogan arcs at the margin of Silurian Avalonia represent island arcs accreted to Avalonia in the Caradocian (Cocks et al., 1997). The subsequent Silurian volcanics and plutons in the northern Appalachians are products of these arc-continent collisions and the later collision of Avalonia and Laurentia. The latter collision was accompanied by as yet unconstrained amounts of large-scale lateral movement along the sutures. Rankin et al. (1989, fig. 28) showed Silurian volcanicastics in the Boston area. In a comparable way, the lateral and vertical extent of the sedimentary basins in which the Nova Scotia Silurian rocks (including Arisaig, Locality 25), New Brunswick, and Newfoundland (including New World Island, Locality 14) were deposited are not well known.

As Avalonia moved northwards, its brachiopods, trilobites, and other taxa with pelagic larval stages became progressively more like those of Laurentia and Baltica (Cocks and Fortey, 1990, figs. 4, 5). By the late Llandovery, the only faunal distinctions between Avalonia and the northern continents (Laurentia and Baltica) was in the distribution of ostracodes (Berdan, 1990) that had no pelagic spat and in fish distribution (McKerrow and Cocks, 1976). By the late Llandovery, the ostracodes in Avalonia and Baltica became similar (Berdan, 1990), but they remained distinct from Laurentia until the Devonian. These differences appear to be related to local tectonic developments. The collision between Avalonia and Laurentia appears to have produced an extensive, elongate foreland basin that acted as a barrier to ostracode migration until the Middle Devonian (Berdan, 1990). This foreland basin is reflected in the thick Wenlock turbidite sequences of central Ireland, and in the even thicker Ludlow of northern England (Kneller, 1991; King, 1994; McCaffrey and Kneller, 1996).

As Avalonia approached and collided with Baltica, Tornquist's Sea was subducted below England. Unlike the margins of Laurentia, a long foreland basin did not develop on the southern margins of Baltica, and thus the migration of ostracodes across the suture was unaffected. In southern Poland, the Silurian and Devonian sediments appear to have been deposited in a succession of small basins (Ksiazkiewicz et al., 1977, pp. 180–183, 260–266, 346). The exact timing of the collision between Avalonia and Baltica is uncertain, but it had to be before the late Llandovery, when the ostracodes crossed the suture; part of it might have coincided with the early Ashgillian Shelvian orogeny in western England (Toghill, 1992).

The collision of Gondwana with Avalonia (by then a part of Laurussia) did not occur until the Devonian. Final closure of the Rheic Ocean did not occur until the Emsian, when the Baltic–British ostracodes reached Morocco (Berdan, 1990).

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Part II: Western Gondwana and Related Continents

Silurian Glacial Paleogeography in South America

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ABSTRACT — In the northern intra-cratonic basins of South America, continental sedimentation succeeded the deposition of shallow marine strata in response to sea-level falls brought about by the Hirnantian glaciation in north-central Africa. Meltwater from the Hirnantian ice caps in Central Africa carried coarse siliciclastics into the Accra (Africa), Parnaíba, and Paraná Basins. Alternating glacial and interglacial conditions persisted until the early Wenlock, when migration of intermittent ice caps from Africa towards central South America caused the deposition of diamictites in continental and marine environments. Evidence of Silurian glaciations is found in the Amazon and Paraná Basins of Brazil, as well as in the Andean basins of Argentina, Bolivia, and Peru (Caputo, 1998). Although no direct evidence of glaciation is currently known in the Silurian of Paraguay, conglomeratic units of this age may be related to glaciation in nearby regions. In the Amazon Basin, three glacial episodes of Silurian age (early Aeronian, latest Aeronian–early Telychian, and late Telychian) led to deposition of diamictites and associated fluvial sandstones, which replaced littoral and shoreface sediments in the Nhamundá Formation. Subglacial deposits in this area consist of quartzose subgraywackes, which often display deformational features induced by ice-push and -shear.

In the Paraná Basin, a shale interval immediately overlying basal diamictites of the Vila Maria Formation probably records a transgressive peak synchronous with the latest Aeronian–early Telychian deglaciation. Therefore, this interval can be correlated with the Vargas Peña Shale in Paraguay. On the other hand, the basal Vila Maria diamictites are probably of the same Hirnantian age as those in the Cape Basin (Pakhuis tillite) and in Argentina (Don Braulio Formation).

In Bolivia, the Cancañiri Formation has yielded fossils from a few sites in the Cochabamba area that contain glacial sediments. The age of the Cancañiri Formation has been regarded as latest Llandoverian–eariest Wenlockian based on chitinozoans and acritarchs. Therefore, two glaciations may be recorded in the Andean area: Ashgillian (Argentina) and latest Llandoverian–earliest Wenlockian (Peru, Bolivia, and Argentina).

Ice movements during the Silurian were probably from upland shield areas down to basinal margins in Brazil and southeastern Andean areas, and from the Pampean and Arequipa massifs to the Andean basins.

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SILURIAN STRATIGRAPHY AND PALEOGEOGRAPHY OF THE NORTHERN AFRICAN MARGIN OF GONDWANA

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ABSTRACT — The Silurian of the northern African margin of Gondwana is geographically widespread and typically exhibits very uniform facies, considerable thickness, and dominantly graptolitic faunas. It is often a petroleum source-rock. The Silurian of each northern African country is described. Twenty-six representative sections are figured, and over 100 localities have been considered in this synthesis. The paleogeography of the northern African Silurian is figured in nine successive maps for the *Akidograptus ascensus-Parakidograptus acuminatus, Cystograptus vesiculosus-Coronograptus cyphus, Coronograptus gregarius-"Monograptus" convolutus,* and *"M." sedgwickii* and equivalent Zones; the *"M." guerichi–"M." turriculatus–"M." crispus–"M." griestoniensis* Zones; the lower, upper, and uppermost Wenlock; and the Pridoli. Four conclusions on the Silurian of northern African Gondwana are emphasized: 1) Several paleogeographic zones can be distinguished. 2) The Late Ordovician glaciation was a very complex event. The effect of the Taconic orogeny during this glaciation has been inadequately considered in earlier syntheses. The Late Ordovician ice cap melted before the end of the Ordovician, and its local role in Silurian transgression is unclear because glacial rebound caused regression during deglaciation. The position of the poles is very hypothetical and cannot be used to explain the differences in facies and subsidence of the Ahaggar Basin. Glacial rebound followed by uplift of the Egypto–Sudanese High caused regression northward and westward in a scheme different from the classic transgression model. It must be noted that explanations that link the presumed extinction and re-radiation of many groups to the beginning and end of this glaciation are hypothetical. 3) An important event occurred in the late Wenlock. 4) The Pridoli is sometimes thick and well developed but incomplete at its base and eroded at the top as a result of later epirogenic movements.

INTRODUCTION

The Silurian of the northern African margin of Gondwana stretches more or less continuously for over 4,000 km from the Lybian–Egyptian Western Desert in the east to the Senegalese coast in the west. It also extends over 3,000 km from the Moroccan Rif in the north to the Bov Basin of Guinea and Guinea Bissau in the south. In this large area, the Silurian crops out under diverse climates (Mediterranean, desert, tropical) that partly control field conditions. The outcrop map (Fig. 1) shows only part of the total extent of the Silurian; and it must be complemented in the subsurface by cores. Sedimentological reconstructions must also take into consideration the size of the large subsurface basins. The northern boundary of the study area, practically the entire Mediterranean coast, does not seem to have been a structural feature during the Silurian.

The Silurian on the northern African margin of Gondwana has seven salient features. 1) It features very uniform facies, mainly shale with rare carbonates that pass upward into siltstone and sandstone with south and southeast transport directions. 2) It is very thick, especially the Lower Silurian in the south (i.e., 200–300 m of lower Llandovery in east Tassili N'Ajjer, with the *"Monograptus" segdwickii* Zone 200 m thick in the Illizi Basin, and more than 700 m of Ludlow–Pridoli in Tripolitania). 3) The faunas are dominantly graptolitic, with other macrofossil groups only locally present. 4) It includes diverse structural domains, from cratonic margins (i.e., Tassilis N'Ajjer) to thrust sheets (i.e. north Moroccan Rif), and this means regional differences in outcrop area and quality. Much of the north Gondwanian Silurian lies in cratonic basins (i.e., the Algerian north Saharan basins) and can be examined and correlated only from bore holes. 5) The Silurian includes good petroleum source-rocks. 6) The difficult working conditions in some regions mean that they have been only cursorily studied. Some sections have been examined only once, and there is no possibility of further observations. 7) A great deal of data

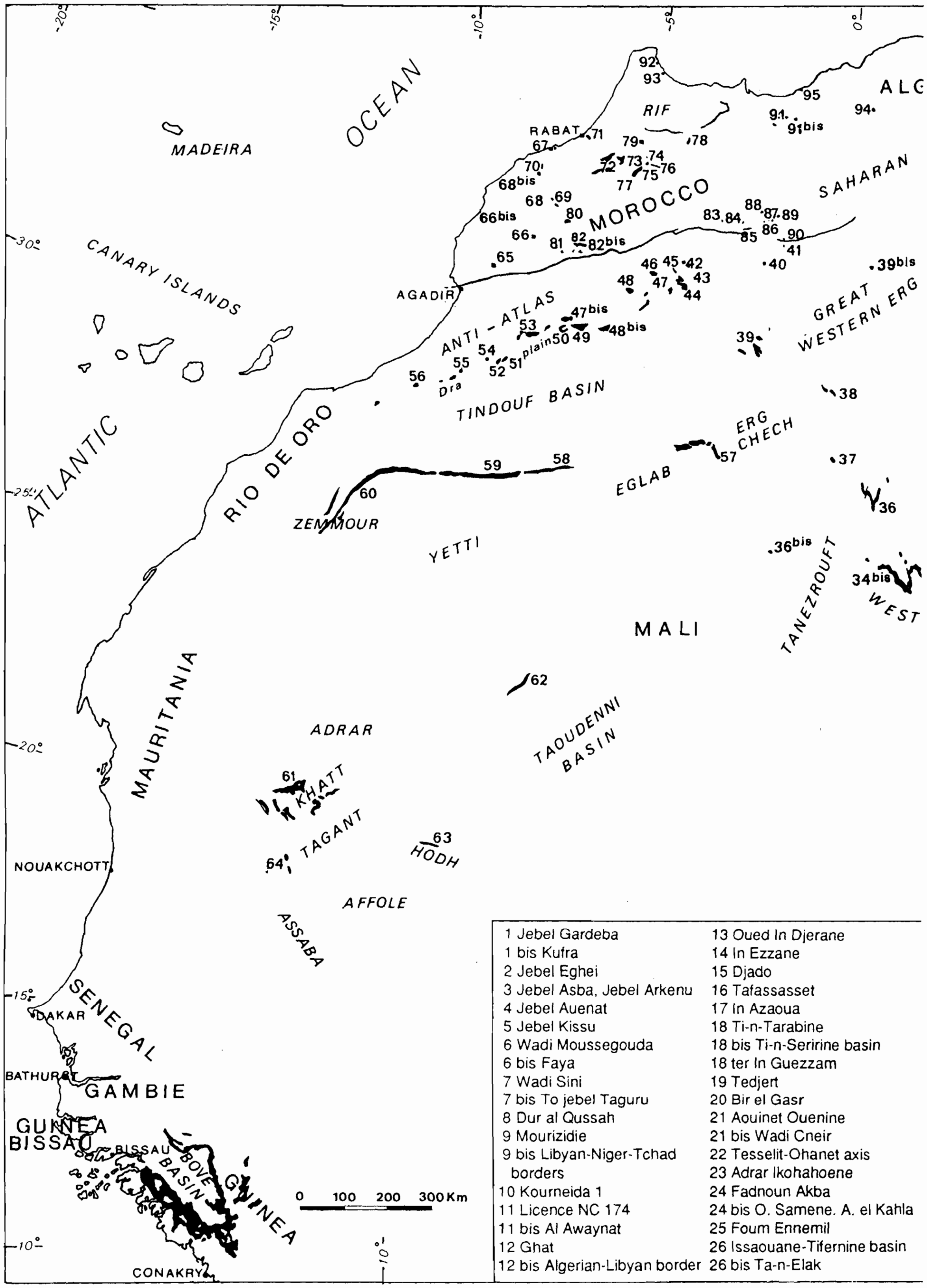
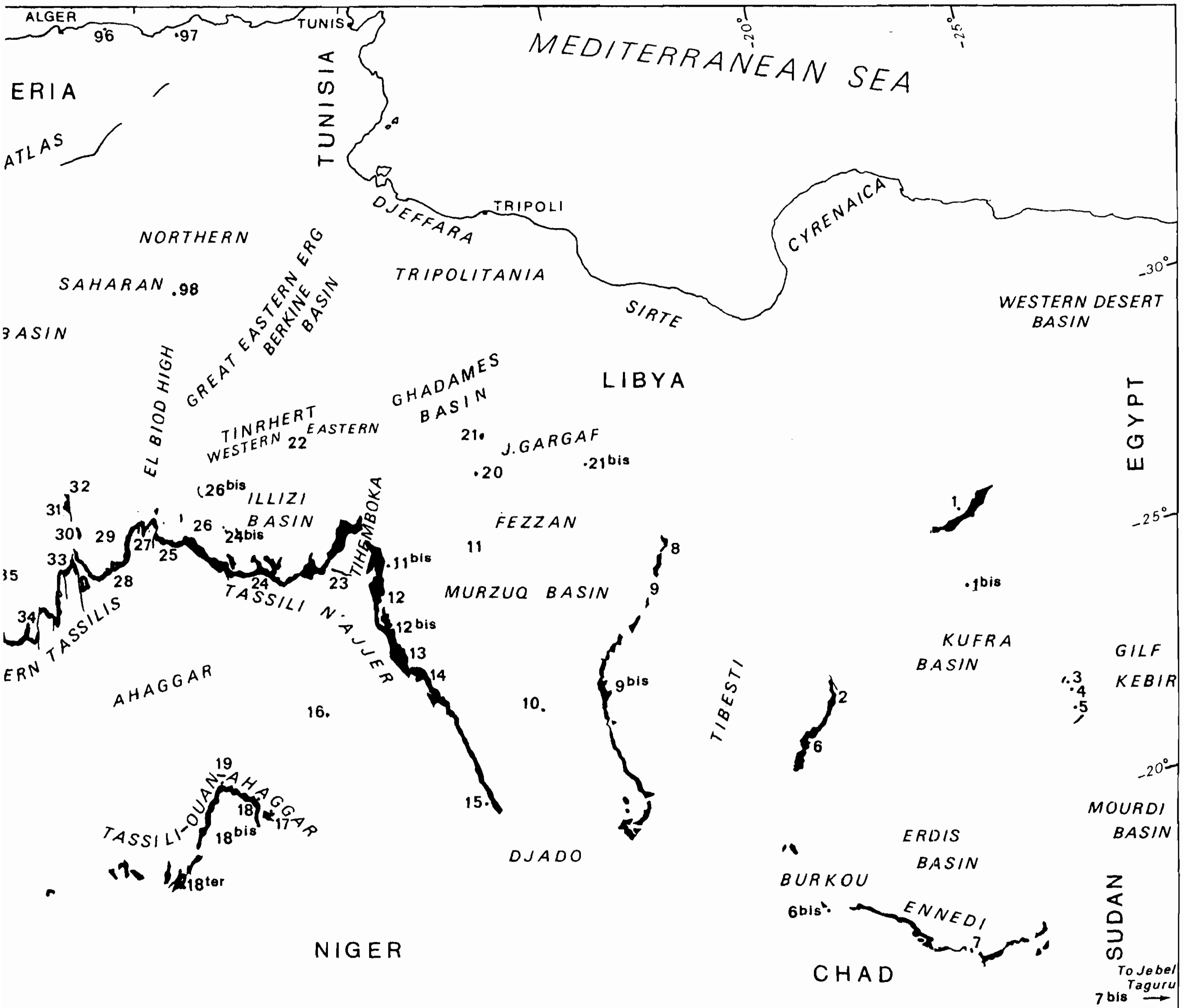


FIGURE 1 — Location map of North African Silurian outcrops (in black).



27 Amguid
28 Mouydir
29 Habadra basin
30 Jebel Idjerane
31 Jebel Azaz-Aïn Beida
32 Ers Oum el Lil
33 Jebel Settaf, Foum Belrem
34 Tassili of Tarit
34 bis Ouallene
35 Oued Djaret basin
36 Azzel Matti
36 bis Bir Ould Brini
37 Touat
38 Gourara
39 Ougarta Range
39 bis Oued Rharbi
40 Bechar
41 Ben Zireg
42 Erfoud. Hamar Laghad
43 Tafilalt
44 Taouz
45 Tizi Ambed
46 Tamaghrout
47 Maïder
47 bis Foum Zguid
48 Tazzarine
48 bis Mel'Alg
49 Iriqui
50 Jebel Amsaïlik
51 Jebel Addane
52 Oued Aguemamou
53 Akka
54 Foum el Hassane
55 Assa
56 Aïn Oui-n-Delouine
57 Bou Bernous
58 Aouinet bel Egrâ
59 Oued El Hamra
60 Gara Bouya Ali
61 Oued Chig
62 Mejahouda
63 Aratane
64 Tagant
65 Haut Atlas : Talma-kent etc...
66 Marrakech
66 bis Essaouira basin
67 Casablanca, Ben Sli-mane country
68 Rehamna
68 bis Doukkala
69 J. Krarou and Mechra ben Abbou outcrops
70 Oulad Abou Syncline
71 Bou Regreg
72 Kouribga-Oulmes an-ticlinorium
73 Kasbah Tadla-Azrou "anticlinorium"
74 Ito, Eguer-Iguiguema
75 Touchent
76 Azrou region
77 South-west nose of the Kasbah Tadla-Azrou "anticlinorium
78 Tazzeka
79 Meknès
80 Eastern Djebilet
81 Adrar n'Dgout
82 Aït Mallahl, Aït Mdioual
82 bis Skoura country
83 Sebbab-Kebir
84 Jebel Korima
85 Jebel Aziza
86 El Zroug and El Hazma
87 El Atchana
Tamlelt
88 Aïn el Orak
89 Guelb Renazi
Tam-lelt
90 Teniet el Ghenia
91 Oujda area
91 bis Ghar Rouban
92 Nord Tetouan
93 Sud Tetouan
94 Tifrit rise (Saïda region)
95 Traras massif
96 Grande Kabylie
97 Petite Kabylie (Beni-Affeur
98 Hassi Messaoud

Ph. L., 2001

FIGURE 1 *continued.*

is unpublished, and belongs to private companies. A small number of publications remain the sole source of information, and paleontological monographs are almost non-existent. For Senegal, Mauritania, Guinea, and Guinea Bissau, the key source is the important compilation by Deynoux et al. (1985), which is complemented by a few reports by Villeneuve (1984), Villeneuve et al. (1984, 1989), and Racheboeuf and Villeneuve (1992). The Silurian of Morocco has been synthesized by H. Hollard and S. Willefert (*in* Destombes et al., 1985). Since then, few papers have appeared on the Moroccan Silurian. In Algeria, work on the Silurian has advanced somewhat since earlier syntheses by Legrand (1981, 1985b), thanks to studies on graptolites (Legrand, 1999) and miospores and chitininozoans (Boumendjel, 1987; Boumendjel et al., 1988). The Silurian of southern Tunisia is known only from cores, but a study by Jaeger et al. (1975) has done much to give precision to what I had written previously in 1974 (Legrand, 1985a). In Libya, the new data chiefly involve sedimentology since the synthesis by Klitzsch (1981) and Massa's thesis (1988), but some papers have appeared on Silurian acritarch and chitinozoan biostratigraphy (Grignani et al., 1991).

After the Late Ordovician glaciation, the African northern margin of Gondwana lay possibly near the South Pole in the Early Silurian. However, some sedimentological features (i.e., iron ores, carbonate units) pose problems. No direct evidence is available for the South Pole position during the Middle and Late Silurian.

General Features in Correlation

Lithostratigraphic nomenclature — The quality of the existing stratigraphic nomenclature is variable, and depends on the countries. Stratigraphic nomenclature is poorly developed in Morocco, but well established in Mauritania, Senegal, and Algeria. In Libya, the use of the same names for surface and subsurface formations leads to confusion. In general, because of the regionally uniform sedimentation, traditional lithostratigraphy is of little use in geologic syntheses, and the same formation may be of very different ages in different areas. Only biostratigraphy can give clues to understanding the region.

Biostratigraphy — Graptolites are the key fossils for Silurian biostratigraphy in northern African Gondwana. Because of general affinities with Bohemian and Sardinian faunas, the faunal zones established in these three regions are similar. However, Llandovery and lower Wenlock graptolites from the southern Sahara are of low diversity and endemic, and the key species for interregional correlations are often absent. Consequently, a local zonation has been proposed for lower–middle Llandovery graptolites. For a long time it has not been easy to correlate into the classical zones, but a satisfactory solution may now be proposed (Fig. 2), as I have explained at the Silurian symposium at the University of Rochester in 1996 and subsequently (Legrand, 1999). The scarcity of graptolites poses the main problem. Of course, graptolites cannot be found in laterally equivalent continental deposits. Sometimes identifications made by paleontologists from different regions differ, and the zonations may change from one country to another and affect definitions, such as that of the Wenlock–Ludlow boundary. The standard graptolite zonation "imposed" on this report contradicts earlier recommendations (Legrand, 1996). Even though globally interpreted zones may help paleogeographic reconstructions, they hinder the correlations between detailed local stratigraphic columns and the establishment of a more precise stratigraphy. The global graptolite zones are not all locally equivalent. Near the Wenlock–Ludlow boundary they provide detailed correlations, but the *Cyrtograptus rigidus* and *C. perneri* Zones are not separated in the global zonation, even though the former has a major importance in Gondwanan stratigraphy (Legrand, 1981, 1985b, 1994). Acritarchs and chitinozoans are becoming more and more valuable as information on their vertical ranges increases in precision. Crinoids, trilobites, mollusks, and brachiopods are also found locally and may be excellent markers. Finally, trace fossils deserve particular study as possible time indicators in littoral rocks (Crimes, 1981). However, systematic studies of Silurian ichnofossils in northern Africa are rare (Seilacher, 1969, 1990, 1992, 2000).

Sequence stratigraphy — Deepening–shallowing successions in the Silurian of North Africa consist of vertical sequences of sandstone, siltstone, shale, limestone, shale, silty shale, and sandstone. The significance of limestones remains uncertain. In addition to a relationship to distance from shore, perhaps a peculiar position (shoals?) and decreasing anoxia may be recorded. However, limestones are locally found in regressive sequences. Climatic influence is often put forward as an explanation of limestone deposition, but there are calcareous nodule beds near the Ordovician–Silurian boundary. Detailed sequence stratigraphic analysis is difficult because of low facies diversity, discontinuous outcrop, and great local thicknesses. These analyses can demonstrate transgressions and regressions, but the local synchronicity of these events remain uncertain. They require biostratigraphic analysis before any interpretation is possible. Some grave errors have been made because of a lack of biostratigraphic data. In the absence of satisfactory sedimentary models, and with a scarcity of benthic faunas, it is hazardous to prepare sea-level curves. It is only for the well-known sections in the Algerian Sahara that tentative sea-

Standard zonation	South-Saharan zonation	Known faunal elements used for correlations
cyphus Zone	*Nd. fezzanensis* Zone	
vesiculosus Zone	*Nd. africanus* and "*Gl.*" *tariti* Zone	*Dimorphograptus physophora* (Tassili of Tarit) *Atavograptus atavus*
	Nd. praeafricanus and "*Gl.*" e.g. *tariti* horizon	
acuminatus Zone	*Nd. imperfectus* Zone	
	Nd. incommodus and "*Gl.*" *saharensis* Zone	
ascensus Zone	*Po.* (?) *kiliani* Zone	
persculptus Zone	*N. tilokensis* and *N. pretilokensis* or *N. pseudovenustus* Zone	? "*Glyptograptus*" *persculptus* Zygospiraella *Normalograptus pseudovenustus*
	N. aff. *gelidus* and *N.* aff. *arrikini* Interzone	
extraordinarius Zone	*N. gelidus* and *N. arrikini* Zone	

Ph. L., 1995

FIGURE 2 — Late Ordovician–Early Silurian graptolite zonation of southern Sahara. Tentative correlation with the British zonation proposed.

level curves are proposed.

ABSOLUTE AGES — No Silurian formation is suitable for radiometric dating. Bentonites are unknown.

REFERENCE SECTIONS — The zone-level biostratigraphic divisions imposed on contributions to this volume are, in my opinion, not suitable for some of the thick northern Gondwanan deposits. Owing to the absence of thickness data in many Gondwanan regions, this zonal scheme imposes apparent uniform thicknesses on all stratigraphic columns and disguises lateral changes in thickness between coeval parts of the Silurian. This latter condition obscures intra- and inter-regional differences in thicknesses of accumulated rock, as well as differences in basin tectonics. The nature, exact location, and relative completness of biostratigraphic information cannot be shown in the same way as in a traditionally constructed column because part of the paleontological data is also lacking when a "standard," "global" zonation is used. Without proper assessment of the biostratigraphic information, many stratigraphic gaps either escape observation (see "bagstratigraphy" of Jeppsson et al., 1994) or, conversely, may artificially come into existence as a result of insufficient sampling. In order to avoid these drawbacks, the shorter columns in the figures of this report show those portions of sections that lack age data and that are dated only by their position within biostratigraphically dated levels. On the figures, white spaces do not necessarily indicate a gap, but only that the graptolite zone has not been identified.

North Africa is divided into two parts that lie north and south of the major South Atlas border fault. The significance of this boundary is questionable in the Silurian, whereas it was important in the Ordovician (Legrand, 1974) and the Carboniferous–Permian. The regions south of the South Atlas fault, which have generally continuous sections, well-preserved faunas, and more obvious facies variations, are described first. Regions lying north of the fault, mainly in Morocco and characterized by discontinuous exposures, higher deformation, and distorted fossils, are described next. Good sections are rare, but are essential to the study of relationships to the European northern margin of Gondwana (see Kříž et al., this volume).

Twenty-six representative sections are figured. In fact, a hundred localities have been considered, as well as many local or regional syntheses that are largely unpublished. I have added some information on those areas that lack complete sections.

KUFRA BASIN, SOUTHEAST LIBYA

The Kufra Basin is a large Paleozoic intracratonic basin that lies in southeast Libya and northeast Chad (Erdis Basin) and extends into northwest Sudan (Mourdi Basin) and southwest Egypt. It is bounded by the Jebel Gardeba in the north, the Tibesti Massif in the west, the Burkou and Ennedi Mountains in the south, and the Gilf Kebir in the east. This basin developed on an "Infracambrian" rift, supposedly related to the Trans-African lineament (Schandelmeier, 1988).

JEBEL GARDEBA (OR JEBEL DALMA, FIG. 1, LOCALITY 1) —

Above Cambrian–Ordovician sandstone, the Tanezzuft Formation consists of shale interbedded with siltstone and sandstone beds. The thickness can reach up to 120 m (Klitzsch, 1981) or 130 m (E. Bellini *in* Bellini and Massa, 1980, p. 41, 47). However, these figures vary significantly and are questionable because only the upper Tanezzuft Shale is generally observed (E. Bellini *in* Bellini and Massa, 1980, p. 35) and the contact with the Cambrian–Ordovician sandstone is rarely exposed. So, according to Lüning et al. (1999, 2000a–b), only about 15 m of shales without a gap can be seen, and the true thickness is unknown. However, in one case, this thickness is 110 m (E. Bellini *in* Bellini and Massa, 1980, p. 47). These shales were deposited on a very shallow shelf. In the upper part, *Neodiplograptus fezzanensis* indicates the Early Silurian (Rhuddanian, lower *Coronograptus cyphus* Zone) (Lüning et al., 1999). The change to the conformably overlying Acacus Sandstone is transitional; the thickness of this latter formation is from 100–165 m (E. Bellini *in* Bellini and Massa, 1980, p. 41) and could reach 240 m. (Lüning et al., 1999). The Akakus Formation represents marine–deltaic deposition. Its age may be Early Silurian (Aeronian) to Middle? Silurian.

JEBEL EGHEI (I.E., JEBEL DAHONE, DOHONE, OR DUHUN; FIG. 1, LOCALITY 2) — Above the Cambrian–Ordovician sandstone, the 120 m-thick Tanezzuft Shale consists of shale interbedded with siltstone and sandstone beds with rare graptolites, and with trace fossils in the sandy beds (Vittimberga and Cardello, 1963; Klitzsch, 1965b, 1968, 1981). Here also, the lower Tanezzuft Shale is not seen, as noted by Vittimberga and Cardello (only 68 m of shale sampled) and E. Bellini (*in* Bellini and Massa, 1980, p. 41, 47), who records a thickness of only 30–50 m. E. Bellini (*in* Bellini and Massa, 1980, p. 49) records 100 m of shale in a composite section, as do Lüning et al. (1999). The Acacus Sandstone conformably overlies the Tanezzuft Shale; its thickness reaches 40–50 m (sometimes 100 m); trace fossils occur locally. The Acacus is unconformably overlain by the Tadrart Sandstone (Lower Devonian?) (E. Bellini *in* Bellini and Massa, 1980, p. 43; Klitzsch, 1981). Sedimentological data indicate a deltaic environment that ranges from shelf to shoreface (Turner, 1991).

WESTERN EDGE OF GILF KEBIR — From north to south, sections at the northern, central, and southern Jebel Asba, west of Jebel Arkenu, and at Jebel Auenat and Jebel Kissu have to be distinguished (Fig. 1, localities 3, 4, 5). In the northern Jebel Asba (Fig. 1, locality 3), the Tanezzuft Formation is 18–24 m thick (E. Bellini *in* Bellini and Massa, 1980, p. 40, 41, 48). However, the facies can be finer-grained than usual, and confusion with the Upper Ordovician is possible (Lüning et al., 1999). The occurrence of *Cruziana acacensis* shows the formation to be Lower Silurian (Llandovery?). The overlying Acacus Formation does not exceed 10 m. In the central Jebel Asba, the Tanezzuft Formation is 40 m thick (E. Bellini *in* Bellini and Massa, 1980, p. 40, 48) and the Acacus Formation is very thin (2.5 m). A core (KW-2) near the outcrops cuts the Tanezzuft Formation, which reaches 82 m at this point, and specimens of *Normalograptus medius* and a chitinozoan assemblage have been found (Grignani et al., 1991; Paris et al., 1995). The lower part of the Tanezzuft Formation might be Upper Ordovician.

In the southern Jebel Asba west of Jebel Arknu, vast outcrops of Tanezzuft Shale are found, but the thickness is minor (15? m), as is that of the Acacus Formation (25 m). At Jebel Auenat (Fig. 1, locality 4), a sequence of sandstone of variable thickness containing *Cruziana acacensis* rests more or less on the Precambrian (Klitzch, 1981). A similar succession (30–50 m) is found 90 km of Jebel Kissu (Fig. 1, locality 5) in northwest Sudan (Klitzsch, 1981). To the south and east, it is often impossible to distinguish between the Tanezzuft and Acacus Formations.

KUFRA AREA — Two exploration wells near Kufra (Fig. 1, locality 1) have shown that the shale of the Tanezzuft Formation can be replaced by a silty sequence with few shales (at least 54 m thick), but it is not possible to extrapolate this transition to the whole basin. The overlying Acacus Formation reaches more than 220 m. According to Grignani et al. (1991), the equivalent of the lowermost Tannezzuft Shale should be upper Ashgill, with the same palynological association as in the underlying Memouniat Formation.

EXPOSURES ALONG THE BORDERS OF CHAD — Three regions may be distinguished: the eastern margin of Tibesti (a continuation of Jebel Eghei), Burku, and Ennedi (Fig. 1). In the first area, two lithological columns and schemes were published by de Lestang (1968), but only one unified nomenclatural scheme is needed. The Dohone Trail Formation, with three members, overlies the lower sandstones. These members include an unusual, lower calcareous, sandy member; a middle shaly member; and an upper araneceous member; the total thickness is about 150 m. The age of the lower member is uncertain (Late Ordovician–Early Silurian?). De Lestang (1968) reported graptolites apparently from the middle shaly member at Wadi Moussegouda (Fig. 1, locality 6), and Paris et al. (1995) noted lower Rhuddanian silty shales in a shallow well at Moussegouda. In the Burku, the Bedo Formation and lower member of the Gouring Formation are found above the lower sandstones (de Lesrang, 1968). The Bedo Formation is composed of more or less silty shales and argillaceous siltstone; the thickness is variable but does not reach 100 m. Its Silurian age is, in part, questionable, as diplograptids have been found near Faya (Fig. 1, locality 6) (Klitzsch, 1968; de Lestang, 1968). In the Ennedi, the upward succession above the lower sandstones includes the Wadi Djoua and Bedo Formations and the lower member of the Gouring Formation (de Lestang, 1968). The first formation is arenaceous with volcanic material, but the age is uncertain (Late Ordovician or Early Silurian?). The shaly Bedo Formation is thin (2 m at Wadi Sini (Fig. 1, locality 7; Klitzsch, 1968, 1981). The lower part of the Gouring Formation is probably equivalent to the Acacus Formation, and reaches 250 m. In the eastern Ennedi and at Jebel Taguru (Fig. 1, locality 7), chaotic rocks at the top of the Acacus Sandstone have been considered to be glacially deposited during the Silurian(?) glaciation (Semtner and Klitzsch, 1994).

WESTERN DESERT — There is no Lower Paleozoic outcrop in this area, but Silurian has been found in bore holes. Within the Silurian, the proportion of sandstone is higher and shale is lower than in northwest Libya, and graptolites seem to be less common (Klitzsch, 1981). On the southern border, the lower Llandovery is present, but on the western border, the middle Llandovery (upper Aeronian) and upper Llandovery (upper Telychian) are indicated by acritarchs, chitinozoans, and sporomorphs (Hill et al., 1985; Paris, 1988).

KUFRA BASIN SUMMARY — The lithofacies become more continental from north to south or to the southeast, but the limited biostratigraphic data do not allow precise reconstruction of lateral depositional environments. In addition, nothing is known about Silurian deposits in the middle part of the Kufra Basin, except that they do not seem to thin southward (Lüning et al., 2000b). The relationship between pre-Caledonian paleogeography and the modern structure of the Kufra Basin is very hypothetical. The re-activation of very old (pre-Panafrican?) west–east-trending structures may have been neglected in syntheses of basin deposition. The silty Silurian of the bore holes of the Kufra area might be an example of this re-activation.

MURZUK BASIN, SOUTHWEST LIBYA AND NORTHERN NIGER

The Murzuk Basin is a large Paleozoic intracratonic basin which lies in southwest Libya, along the Algerian border, and in northern Niger. It is bounded by the Gargaf area to the north, the Tihemboka and western Tassili N'Ajjer on the west, and by the Tibesti Mountains to the east. It extends into Niger, where it is named the Djado Basin. The origin of this basin is uncertain.

EASTERN FLANK, DUR AL QUSSAH (OR DOR EL GUSSA) SECTIONS — In northern Dur al Qussah (Fig. 1, locality 8), the Mamuniyat Sandstone (=Memouniat Sandstone) is overlain by the Tanezzuft Shale, which is about 160 m thick. These shales are intercalated with beds of fine-grained sandstone. Middle or early? Llandovery graptolites occur approximately 100 m above the base of the Tanezzuft Shale (Klitzsch, 1966, 1981). D. Massa (*in* Bellini and Massa, 1980, p. 25) noted the lower Llandovery form *Neodiplograptus fezzanensis* (Desio). There is no stratigraphic break with the Acacus Formation that overlies the Tanezzuft Shale, but there is a gradual change from a shale to a dominantly sandy–silty lithology. The maximum thickness is 465 m (Jacqué, 1963; Klitzsch, 1963, 1966, 1981). In the uppermost part of a section in the central Dor el Qussah, an important, apparently Silurian flora has been found (Klitzsch et al., 1973; Boureau et al., 1978; Douglas and Lejal-Nicol, 1981). As a result of sub-Lower Devonian erosional cut-out, the thickness of the overlying Acacus Sandstone decreases southward, and the Lower Devonian Tadrat Sandstone lies on strata as low as the lower Tanezzuft Shale (Fig. 3; Klitzsch, 1981). In the Mourizidie area (Fig. 1, locality 9), Silurian shales with some silty calcareous concretions (often exposed in the uplift zones in the Sahara) directly overlie the Serie Pharusian (Jacqué, 1963). Near the Libyian–Niger–Chad borders (Fig. 1, locality 9), the Tanezzuft Shale again thickens (100–125 m) and has rare *Normalograptus*, but the Acacus Formation, rich in *Harlania*, is not as thick as in the north.

CENTRAL MURZUK BASIN — The Silurian in the central part of the Murzuk Basin is only known from bore holes. South of the 24th parallel, the Tanezzuft Shale, probably of Silurian age, has been encountered in a few cores (e.g., bore hole Kourneida 1 [KR 1], Fig. 1, locality 10; Meister et al., 1991). North of the 24th parallel in the northwest part of the Murzuk Basin, many exploratory wells have been drilled, and several oil-fields have been recently discovered. However, biostratigraphic data are rare or unpublished. The transition from the top of the Ordovician sandstone into the Tanezzuft Shale is not clear (Echikh and Sola, 2000). A well on the northern border of the Murzuq Basin (licence NC 174; Fig. 1, locality 11) has the terminal Ashgill graptolite *Normalograptus tilokensis* at the base of the Tanezzuft Shale, and *Coronograptus cyphus* Zone graptolites 6.0 m higher (Lüning et al., (2000a). This sequence is appreciably different of that observed on the western flank of the basin (see below). The thickness of the Tanezzuft Shale could reach 1,500 m (Echikh and Sola, 2000). Above the Tanezzuft Shale, the subsurface equivalent of the Acacus Formation is about 360 m thick (Echikh and Sola, 2000). An interesting result is the Wenlock–Ludlow age suggested for the Acacus Formation from palynological studies (Pierobon, 1991). In the northeast Murzuk Basin north of the 24th parallel, in and near the Durr al Qussah area, a thick sequence of very silty Tanezzuft Shale was encountered in a water well. So there is a narrow northeast–southwest-trending basin that paral-

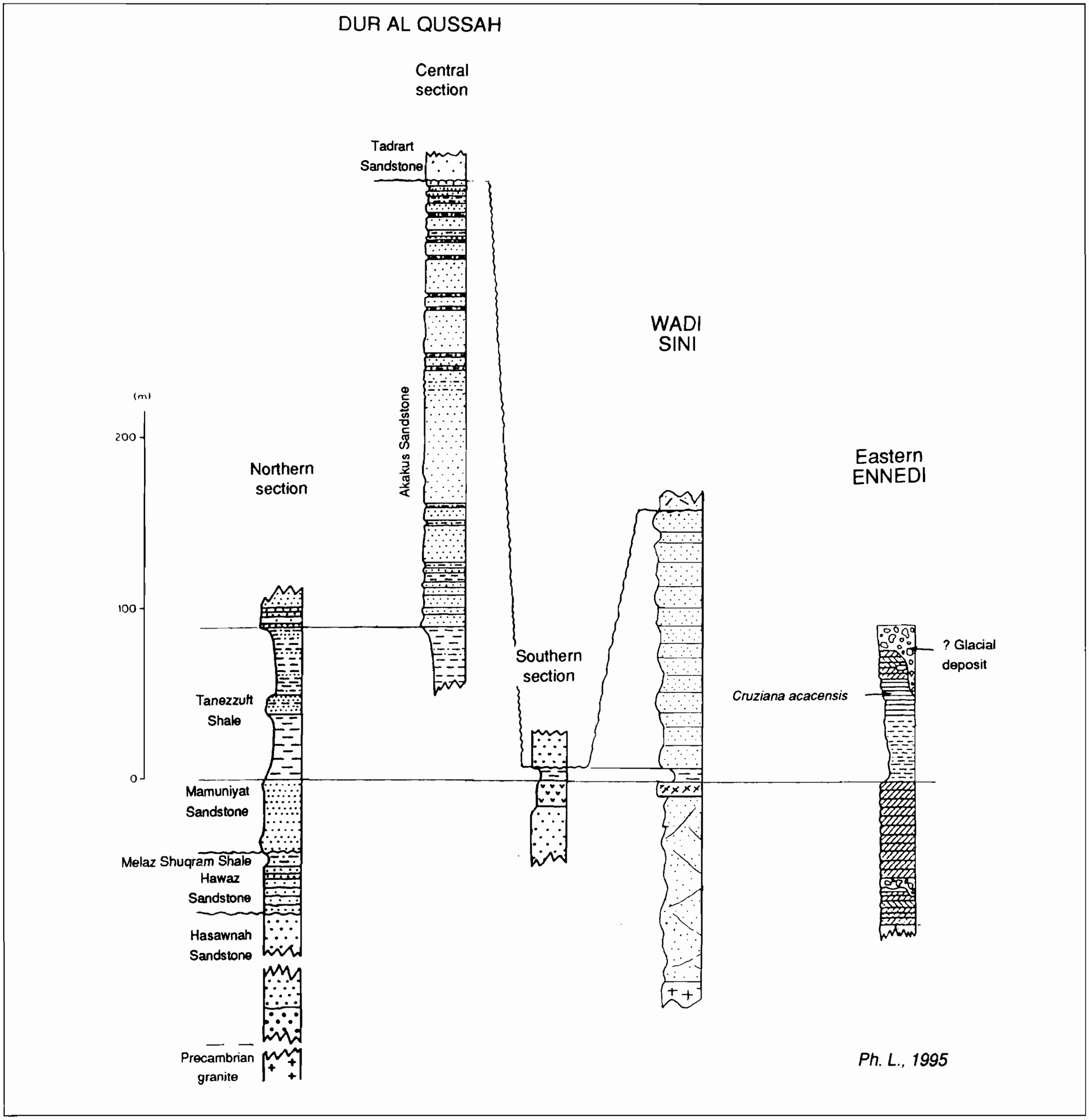


FIGURE 3 — Representative Lower Paleozoic sections from Dor el Gussa (Libya), Wadi Sini (Chad), and eastern Ennedi (Sudan). According to Klitzsch (1981) and Semtner and Klitzsch (1994).

lels the outcrops in this area.

SOUTHERN MURZUK BASIN — In the Djado area (Fig. 1, locality 15), the argillaceous–arenaceous Chirfa Formation, ca. 100 m thick, had previously been assumed to be Silurian (Beuf et al., 1971). However, it consists of Ashgillian (Rawtheyan) glacio-marine rocks in its lower part (Legrand, 1985c, 1993, 1995b, 1999). Consequently, the first phase of the glaciation is pre-Rawtheyan in this area. Above the Chirfa Formation in some sections, shales and siltstones possibly of Late Ordovician or Early Silurian age can be found, whereas in other sections, unconformably overlying Lower Devonian(?) sandstone occurs.

CENTRAL PART OF WESTERN FLANK OF MURZUK BASIN — In the Oued In Djerane sections (Fig. 1, locality 13; Figure 4; Legrand, 1976, 1985a, 1985b, 1986, 1988, 1999, 2000), sedimentation was continuous from the latest Ordovician (Hirnantian) into the Early Silurian (Rhuddanian). As a result, this is a key area for the stratigraphy of the Algerian Sahara and the biostratigraphy of the latest part of the Late Ordovician glaciation. Unhappily, the uppermost Ordovician and lowest Silurian graptolites are endemic species, and equivalency with the British zonation was difficult to establish (Fig. 2). The base of the Rhuddanian is now regarded to be at the level of the *Pseudorthograptus? kiliani* Zone, as proposed at the Silurian symposium in Rochester, NY (Legrand, 1996), and practically, for the purpose of constructing the paleogeographic map, at the base of this zone (Legrand, 1999, 2000). This presumed base of the Rhuddanian is characterized by the return to shale sedimentation in the *P? kiliani* Zone (=lower submember of the middle member of the Oued In Djerane Formation, 35 m.). Overlying strata include carbonates at the top of the lower *Normalograptus tilokensis-N. pseudovenustus* Zone. Above this interval, siltstones become more and more frequent, and a regression occurred at the end of the *Neodiplograptus imperfectus* Chron (=upper submember of the middle member of the Oued In Djerane Formation, 70 m). Subsequent to additional marine deposits of the lower *Neodiplograptus africanus-"Glyptograptus" tariti* Zone (ca. *Atavograptus atavus* Zone) and of the *Neodiplograptus fezzanensis* Zone (ca. *Coronograptus cyphus* Zone), sandy sedimentation resumed and indicate shore-to-lagoonal environments (upper member, 100 m). Southwards toward In Ezzane (Fig. 1, locality 14), all these deposits become sandier and thinner. Between In Ezzane and Djado, graptolites, erroneously determined but probably related to *N. fezzanensis*, occur in the sandier facies 80 m above the top of the Sandstone Group of the Inner Tassili (BRP-IFP, 1960; Legrand 1999). On the Algerian side of the Algerian–Libyan border (Fig. 1, locality 12; Fig. 5; Legrand, 1985), evidence for transgression is quite spectacular (Legrand, 1985c, 1999, 2000). However, it appears to be coeval with the regression of the upper *Neodiplograptus imperfectus* Zone in the Oued In Djerane section. Above the lower part, the section consists of shales and silty shales. A progressively weak regression preceded the appearance of *Normalograptus? libycus*, which indicates approximate correlation with the base of the *"Monograptus" triangulatus* Zone. On the Libyan side of the border (Fig. 1; Klitzsch, 1965a; Massa and Jaeger, 1971), the transgression has been considered to be earliest Llandovery — a doubtful age, in my opinion, as I would refer it to the *Cystograptus vesiculosus* Zone. The lower part of these sections consists of silty shales with some beds of

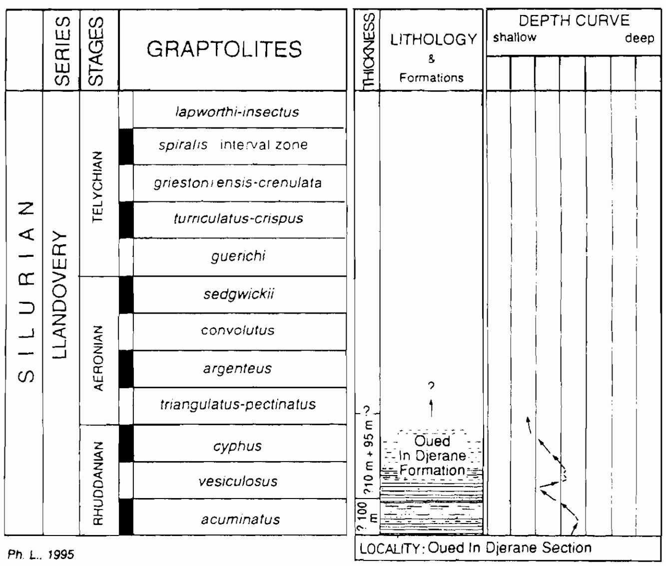


FIGURE 4 — Oued In Djerane section (Algeria); lithology, formations, and depth curve. Correlations between local and British graptolite zonation approximate. After Legrand (1976, 1986, 1988, 1999, 2000).

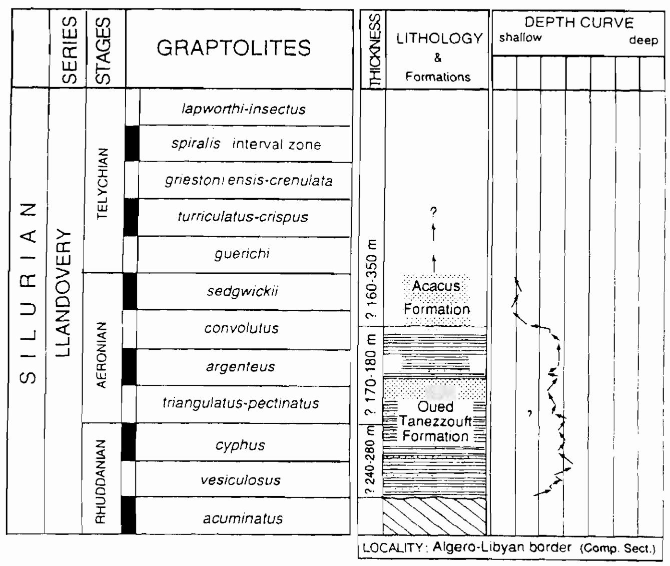


FIGURE 5 — Algerian–Libyan border; composite section, lithology formations, and depth curve. Correlations between local and British graptolite zonation approximate. Lower part after Legrand (1982, unpublished; 1985c, 1999, 2000); upper part after Collomb (1958, unpublished), Klitzsch (1965), and Massa and Jaeger (1971).

siltstone or sandstone of Rhuddanian age (*Coronograptus cyphus* Chron s.l.) overlain by Aeronian shales (*Coronograptus gregarius* Zone). According to Massa and Jaeger (1971), a disconformity, not recognizable in the field, separates the Rhuddanian (Iyadhar Formation) from the Aeronian (Tannezouft Formation), but this seems doubtful. Higher strata indicate that sandy sedimentation recommenced, followed by more shaly deposits (*"Monograptus" convolutus* Zone?). Progressively, these shaly deposits are transitional upward into the shore–deltaic Acacus Formation a little below the *"Monograptus" segdwickii* Zone. A facies analysis of this section has been carried out, and turbidites and tempestites have been recognized (Carneiro de Castro et al., 1991). In both areas, considerable thicknesses characterize the lower and middle Llandovery.

Northwestern flank of Murzuk Basin — In the western Al Awaynat section (Fig. 1, locality 11; Fig. 6), onlap and deposition of the Oued Tannezzuft Shale occurred only by the middle Llandovery. Above this, global regression led to the sandstone deposition of the Acacus Formation, which is especially rich in ichnofossils (Desio, 1936; Freulon, 1964; Klitzsch, 1981). The base of the Acacus Formation could be a little younger here than in the south.

Murzuk Basin synthesis — Poor biostratigraphic data on the eastern flank and in the central part of the basin limit any attempt at synthesis. On the western flank, biostratigraphic data, although inadequate, illustrate a more complicated story than the simple development of a prograding deltaic sequence. A large depression occurs between the Algerian–Libyan border and In Ezzane and, perhaps, in the Djado area, where the first melting of the Late Ordovician icecap began earlier than elsewhere in northern Africa. At least in the northern part of the Oued In Djerane area, the hypothesis of a "forced transgression" related to glacial rebound and/or other tectonic activity must be put forward to explain why the shoreline regressed from south to north (Legrand, 1999).

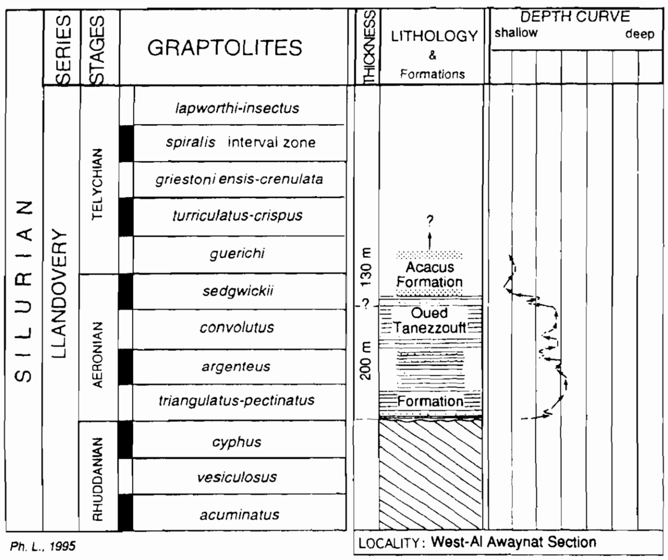


FIGURE 6 — West-Al Awaynat section (Libya); lithology, formations, and depth curve. Correlations between local and British graptolite zonation approximate. After Collomb (1958, unpublished) and Massa and Jaeger (1971).

Tassili Ouan Ahaggar, Southern Algeria and Northern Niger

Before discussing the Silurian of the western and the northern parts of North Africa, it is best to summarize what is known about the Silurian of the southern Ahaggar highlands (Fig. 1). The name "Tassili Ouan Ahaggar" as used herein includes all the Paleozoic outcrops along this more than 700 m-long border region.

Tassili of Tafassasset — These outcrops link the western flank of the Murzuk Basin and the Tassili Ouan Ahaggar (Fig. 1, locality 16; Gariel and Lacaze, 1965). An argillaceous–arenaceous sequence that is 30 m thick yields *Pseudorthograptus? kiliani* near its top (Attar and Saadallah, 1982; Legrand, 1985c). As a consequence, its age could be latest Ordovician–earliest Silurian (Legrand, 1999).

In Azaoua area — The outcrops of the Tedjert Formation in this area are poor and discontinuous (Kilian, 1928; Legrand, 1999; Fig. 1, locality 17). The observable sequence is less than 150 m thick and predominantly silty. The lower part is upper Ashgillian, and the upper beds are lower Llandovery (Legrand, 2001). Above the Tedjert Formation at localities to the east and south of In Azaoua is the Efeimazerta Formation. It consists of 10–15 m of argillaceous and ferrugineous rock; its age is uncertain, but is likely pre-Devonian (Beuf et al., 1971).

Oued Ti-n-Tarabine — The base of the Tedjert Formation is not visible, and its precise age is unknown in this area (Fig. 1, locality 18; Fig. 7). The exposed sequence is 60 m thick and silty rather than argillaceous. Many calcareous sandy lenses and microconglomeratics with fragments of inarticulate brachiopods and small phosphatic pebbles have been observed. The succession of *Neodiplograptus africanus africanus*, *N. fezzanensis* (Legrand, 1978, 1979), and *Normalograptus? libycus*, allow correlation of these outcrops with the uppermost *Cystograptus vesiculosus*, the *Coronograptus cyphus*, and lowest *Coronograptus gregarius* Zones (Legrand, 1999).

Tedjert section — This is a rare locality where the contact between the Sandstone Group of the Inner Tassili

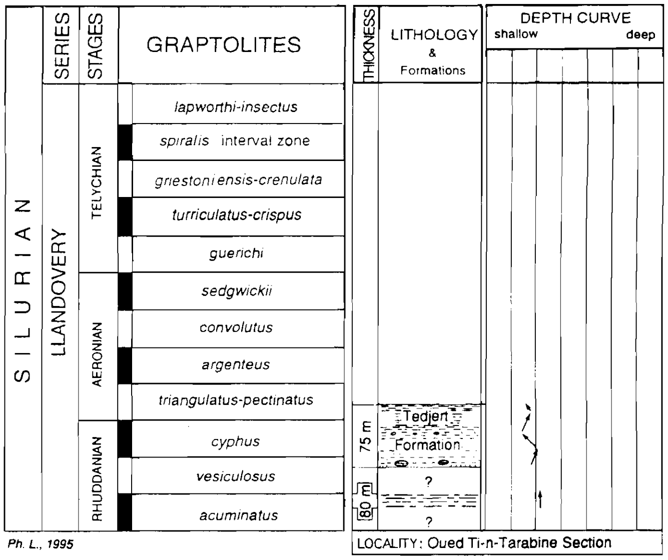


FIGURE 7 — Oued Ti-n-Tarabine Section (Algeria); lithology, formations, and depth curve. Correlations between local and British graptolite zonation approximate. After Collin et al. (1960, unpublished) and Legrand (1977 unpublished, 1999).

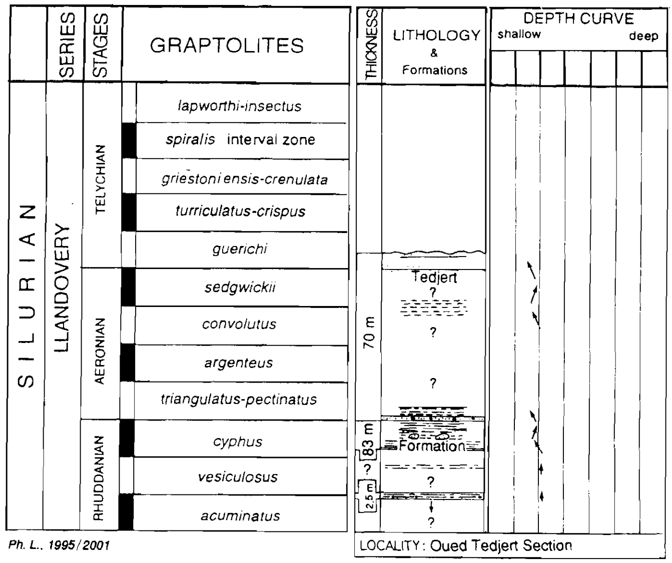


FIGURE 8 — Oued Tedjert section (Algeria); lithology, formations, and depth curve. Correlations between local and British graptolite zonation approximate; correlations tentative in upper part of section. After Collin et al. (1960 unpublished) and Legrand (1977 unpublished, 1999).

and the Tedjert Formation is visible (Fig. 1, locality 19; Fig. 8). The Tedjert Formation is about 180 m thick. There are practically no mud shales, but rather silty shales and siltstones (Lessard, 1961). Some beds are very rich in bivalves. The age of the lower beds may be Late Ordovician or Early Silurian. Above these, the interval equivalent to the *Cystograptus vesiculosus* Zone that has *Neodiplograptus africanus africanus* is thin in comparison to the beds of the *Coronograptus cyphus* Zone with *Glyptograptus (G.) tamariscus* and *Neodiplograptus fezzanensis.* The upper part of the section has *Normalograptus? libycus,* then *Coronograptus* sp aff. *C. gregarius,* and is middle Llandovery (Legrand, 1999). To the southwest, the Tedjert Formation may be 260 m thick, and the *Cystograptus vesiculosus* Zone is much thicker than elsewhere; however, accurate data are lacking.

Ti-n-Seririne Basin — The Silurian is found in bore holes in the Ti-n-Seririne Basin (Fig. 1, loc. 18). Its thickness is 165 m. Because of the lack of fossils, its precise age is unknown (Claret and Tempère, 1968).

Western Tassili Ouan Ahaggar — In the In Guezzam area (Fig. 1, locality 18), the Tedjert Formation decreases in thickness (80? m). Further west, outcrops are limited and discontinuous, and the Tedjert Formation is ca. 100 m. Useful fossils are lacking, but there is probably a thinning or disappearance of the lower biostratigraphic zones (Lessard, 1961; Legrand, 1999).

Synthesis — Outcrops in the Tassili Ouan Ahaggar and bore holes in the Ti-n-Seririne Basin are very significant paleogeographically. They confirm marine deposition from the latest Ordovician in a southern part of the African margin of Gondwana. The local graptolite species are endemic. However, their similarity with later faunas of the eastern Tassili N'Ajjer and the western Tassilis (Legrand, 1970; see below) constitutes a convincing argument for an intracratonic basin in the present Ahaggar Highlands area that persisted until the middle Llandovery (Legrand, 1995a, 1999, 2000).

Jebel Gargaf, Tihemboka, and Illizi Basin; Central Tassili N'Ajjer

Jebel Gargaf is close to the Murzuk Basin to the north, and the Tihemboka lies both on the northwest border of the Murzuk Basin and on the eastern border of the Illizi Basin. The central Tassili N'Ajjer is on the south border of the Illizi Basin.

Jebel Gargaf — North of Al Awaynat (Fig. 1, locality 11), Silurian shales are known from three small outcrops at Jebel Gargaf. The Aouinet Ouenine outcrop (Fig. 1, locality 21) is the largest and was discovered in 1956 (Massa and Collomb, 1960; Collomb, 1962). Unfortunately, only a schematic of the section has been published (Massa and Jaeger, 1971; Massa, 1988). The overlying unit is the Mamuniyat Formation (or Memouniat Formation). However, Klitzsch (1981) terms it the Haouaz Formation?, and reports a bed of quartzitic sandstone (3–5 m thick) with numerous impressions of Diplograptidae, which indicates a possible Silurian age. The majority of the formation consists of 30–35 m of graptolitic shales, probably of late Aeronian age (*Monograptus segwickii* Zone). Further south in the Bir el Gasr outcrop (which was the first Silurian outcrop discovered in this area) (Fig. 1, locality 20; see Lelubre, 1946; Freulon, 1951), the fauna should be more or less similar (Klitzsch, 1981). At a more eastern outcrop north of Wadi Cneir (Fig. 1, locality 21), the Silurian consists of 40–50 meters of shale, the base of which is covered, and which yields Aeronian graptolites (*Coronograptus gregarius* Zone) that are a little older than faunas at the more northern localities (Collomb, 1962 ; Klitzsch, 1981).

TIHEMBOKA, CENTRAL TASSILI N'AJJER — The outcrops at Tihemboka (Fig. 1) are rather poor and have not received detailed study (Freulon, 1964; Legrand, 1985b). The outcrops at Adrar Ikohahoène (Barlier et al., 1960; Fig. 1, locality 23) and those in the central Tassili N'Ajjer (Fig. 1) have been the object of many studies. These studies have often remained unpublished or have been published in a summary way, and give a very superficial view of these sections (Remack-Petitot, 1960; Legrand, 1962; Dubois et Mazelet, 1965). The age, determinations, and interpretation of these sections are often inaccurate. New correlations have been proposed (Legrand, 1985b) that have been clarified after revision of the diplograptids (Legrand, 1999). However, studies of the monograptids are in progress. A reference for this area is the Fadnoun Akba section (Fig. 1, locality 24; Fig. 9). About 5 m below the top of the Felar-Felar Formation, *Hirnantia sagittifera* occurs (Beuf et al., 1985, unpublished data; Legrand, 1995b). On the surface of the highest sandy bed of the Felar-Felar Formation, unindentified diplograptids are found. They are probably Silurian (Legrand, 1999). The overlying shaly Oued Imirhou Formation is about 330 m thick. At the base are some graptolite species known from Bohemia. Higher up, the faunas become scarcer and more endemic, and it is difficult to recognize the base of the *"Monograptus" sedgwickii* Zone. The facies become more silty upward (argillaceous–arenaceous Gara Maret Member), and then sandy (Tsitarene Submember) (Legrand, 1985b). It is believed that this coarser sedimentary unit corresponds to the *Monograptus turriculatus* Zone, but there is no paleontological evidence. The overlying Oued Ouret Formation (ca. 130 m) includes a lower interval of alternations of siltstones and silty shales. *Normalograptus? flamandi* in the nearby section at Oued Imirhou indicates the *Monoclimacis griestoniensis* Zone (upper Llandovery). In the somewhat higher Akba Fadnoun section, *Retiolites geinitzianus angustidens* (Dubois and Mazelet, 1965) indicates the transition to the *Monograptus tullbergi* Zone or *M. spiralis* Zone (upper Llandovery). Only fragments of graptolites have been found at one level in the overlying beds; these beds could still be upper Llandovery (Taran-

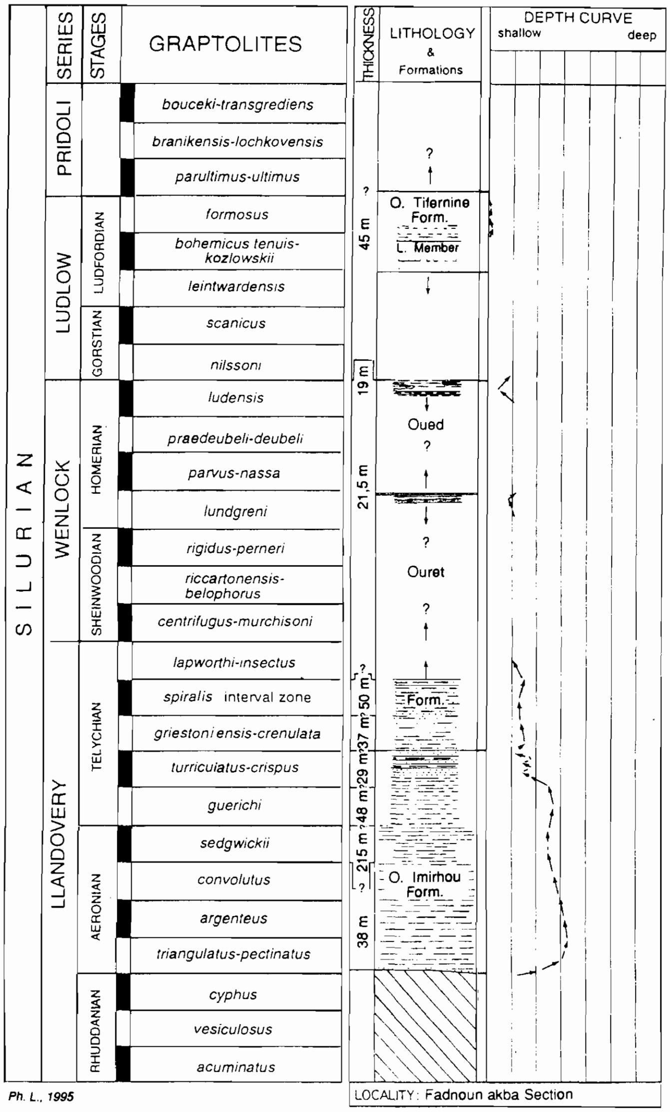


FIGURE 9 — Fadnoun Akba section (Algeria); lithology, formations, and depth curve. Correlations between the local and British graptolite zonation approximate; correlations tentative in upper part of section. After Legrand (1967 unpublished; 1985b, 1999).

non auctorum; Legrand, 1962, 1985b). In the highest 50 m of the Oued Ouret Formation, the facies changes indicate shoaling and emergence. Bivalves suggest the upper Wenlock or Ludlow, but a gap of part of the Wenlock or Ludlow is possible (Legrand, 1962, 1985b). Higher sandstones of the Oued Tifernine Formation (Dubois et al., 1967) are assumed to be Upper Silurian (Ludlow or Pridoli) (Legrand, 1985b), and not Devonian as indicated earlier by an error in a figure (Legrand, 1999). An unconformity is noted at the base of the Oued Tifernine Formation, and continental conditions extended over a broad area (Beuf et al., 1971). The abundance of sandy beds decreases westward, but the Lower Silurian transgression occured later in the west. At Oued Samene (Aïn el Kahla section, Fig. 1, locality 24), the transgression is probably dated in the *"Monograptus" segdwickii* Chron (Legrand, 1999). In the upper part of the sequence, the fluviatile sandstones of the lower member of the Oued Tifernine Formation are thinner, and the more open-marine upper member is thicker (Dubois et al., 1967).

ILLIZI BASIN — In the southern part of this basin (Fig. 1), the succession includes a lower shaly formation (up to 400–500 m), an overlying silty-shale formation (up to 250 m), and an upper sandy formation. At the base, it seems that the *Coronograptus gregarius* Zone lenses out quickly to the north, as does the *"Monograptus" convolutus* Zone. Farther north, the base of the Silurian is near the *"Monograptus" convolutus–"M."* sedgwickii zonal boundary, or in the *"M." sedgwickii* Zone. There is more sand in the upper part of the sequence; this increase in siliciclastic material probably appears later in this region than in the south, although this interpretation remains to be proven. Palynological studies in the Eal1 and Trn3 bore holes show that the top of the argillaceous–arenaceous sequence just below the unconformity with the Lower Devonian is Pridoli (Boumendjel, 1987; Boumendjel et al., 1988), but this result has to be confirmed elsewhere to be integrated in a paleogeographic reconstruction. Sequence stratigraphic analysis has been tried, but lacks biostratigraphic control (Assès, 1987; Assès and Delfaud, 1987; Talah et al., 1995).

REGIONAL SYNTHESIS — Marine deposits with a *Hirnantia* fauna follow the Late Ordovician glacial sequence. In the central Tassili N'Ajjer, the transgression appears to be middle Llandovery at the section west of Serdeles. There is probably a gap corresponding to the lower Llandovery. This transgression is contemporaneous with the beginning of regression on the Algerian–Libyan border, and it seems difficult to explain this by eustasy. The blanket of shaly rocks becomes younger from the southeast to the northwest in the Illizi Basin. In the central Tassili N'Ajjer, muddy deposition ended in the early Telychian (late Llandovery); then, following a possible hiatus, began again in the middle Telychian before being succeeded by Wenlock and/or Ludlow coastal deposits.

EASTERN TINRHERT, GHADAMÈS BASIN, BERKINE BASIN, AND TRIPOLITANIA, SOUTHERN TUNISIA

This large area is bounded in the south by the Gargaf uplift. The Ahara Arch (eastern Tinrhert), which was in existence from the Ordovician (Legrand, 1974; Latrèche, 1982), separates the Ghadamès Basin from the Illizi Basin. The Berkine Basin is the westward continuation of the Ghadamès Basin and the northern Tripolitania area. The greater part of this area is covered by the Great Eastern Erg (e.g., sand sea).

EASTERN TINRHERT — East of the Tesselit–Ohanet high (Fig. 1, locality 22), there is no data on the age of the base of the argillaceous sequence that can reach 600 m in thickness. Somewhat higher strata have lower Wenlock graptolites (*"Monograptus" riccartonensis* Zone). Alternating argillaceous–arenaceous beds in the upper part of the sequence may be Ludlow. Whether the Pridoli appears below the unconformably overlying Lower Devonian (Pragian) sandstone is uncertain (Jardin and Yapaudjian, 1968).

GHADAMÈS BASIN, SOUTHERN TRIPOLITANIA — This basin (Fig. 1), which is the eastern continuation of the eastern Tinrhert, has a somewhat different stratigraphy. According to Massa and Jaeger (1971) and Massa (1988), who described a composite section from several wells, the base of the "Formation des argiles principales" is lowest Llandovery, at least in bore hole F1-66. The shales just above this "formation" seem to be upper Llandovery (*"Monograptus" sedgwickii* and *"M." turriculatus* Zones), and this correlation seems appropriate for a certain part of the shale sequence. *Petalograptus* sp. cf. *P. altissimus* and *Climacograptus* sp. aff. *C. innotatus* (not *C. innotatus brasiliensis*) cited at this level and higher could be the endemic species *Petalolithus meridionalis* and *Normalograptus flamandi* of the *Monoclimacis griestoniensis* Zone, but this must be confirmed. The upper Rhuddanian and lower Aeronian are not recognized, whereas the upper Aeronian and Telychian could reach 150 m in thickness. The presence of the Wenlock is only postulated. The lower part of the overlying "Alternances argilo-gréseuses" is lower Ludlow. The rest of the formation is also assumed to be lower Ludlow, but without any paleontological evidence.

BERKINE BASIN — In the central part of this basin (Fig. 1), the Silurian lies at a great depth. As far as I know, no bore hell has reached it. To the north, Silurian shales are about 250 m thick and Ludlow graptolites are found near the base (well Zar1, P. Legrand, unpublished data). It is likely that the base of the shales is upper Wenlock, as in most north Saharan basins (Legrand, 1985b, 1994). An overlying argillaceous, silty, and sandy interval could be Pridoli in its lower part.

NORTHERN TRIPOLITANIA — The lithostratigraphic terminology used in northern Libya (Fig. 1) is that proposed for the Ghadamès Basin. The subsurface sequence described by Massa and Jaeger (1971) and Massa (1988) is a composite succession. Although the lower and middle Llandovery have not been recognized in bore holes, not enough data are available to allow me to assume that this interval comprises a hiatus. However, this gap may exist in eastern Tripolitania (discussed below). Graptolites in the "Formation des argiles principales" allow recognition of the top of the Llandovery and the Wenlock. The base of the Ludlow (according to the international definition) is near the base of the "Alternances argilo-gréseuses,"and the lower part of this formation is lower Ludlow. However, graptolites are unknown above this interval, and the Pridoli can only be demonstrated by chitinozoans (Jaglin and Massa, 1985; Jaglin, 1986). Fragments of the land plant *Cooksonia* occur (Daber, 1971), as well as numerous spores (Buret and Moreau-Benoit, 1986).

EASTERN TRIPOLITANIA — This region (Fig. 1) is the eastern part of the Ghadamès Basin and features the easterly continuation of strata found in northern Tripolitania. A synthesis of the Paleozoic of this area is in Belhaj (1996). No graptolites are known, and correlations are based on palynological analysis. The succession again includes a lower shaly unit (Tanezzuft Formation) and an overlying argillaceous, silty, or sandy unit (Acacus Formation). There is no evidence of the lower and middle Llandovery, although the upper Llandovery is thick (up to 300 m). The Wenlock, with a maximum thickness of 300 m, can be divided into two units: the lower unit consists of silty shale, and the upper unit consists of middle Wenlock shale and sandstone. It seems that the Wenlock is more aerially extensive than the Llandovery. The lower Ludlow is only present in the northwest part this area, and consists of interbedded shale and sandstone with a maximum thickness of 225 m. The upper Ludlow–Pridoli seems less areally extensive, and is an essentially arenaceous interval not exceeding 200 m in thickness.

SOUTHERN TUNISIA — In southern Tunisia (Fig. 1), the Silurian has been recognized only in drill cores (Bonnefous, 1963; Jaeger et al., 1975; Massa, 1988). Regretfully, the data published are often inaccurate. The lithostratigraphy is the same as in Tripolitania. The lowest Silurian unit is the "Formation des argiles principales." This fine-grained sandstone, ca. 10–20 m thick, has been assigned to the lowermost Llandovery, but the fossils seem to indicate the higher *Cystograptus vesiculosus* Zone. Above this sandstone, the existence of the middle Llandovery is uncertain, and the upper Llandovery (Telychian) does not exceed 20 m, although it may be absent in some drill holes. The Wenlock is characterized by many specimens of *Cyrtograptus* and is quite thin (15–30 m). The Wenlock–Ludlow boundary is somewhere in a 30–100? m transition interval, with the boundary placed at the base of this interval, if Jaeger's (1991) definition is followed, and at the top according to the international definition. This boundary seems to lie within the "Argiles principales" to the northeast, and elsewhere near the base of the "Alternances argilo-gréseuses." The overlying "Formation des alternances argilo-gréseuses" is lower Ludlow (Gorstian), and apparently extends into the upper Ludlow (Ludfordian). The highest "Alternances argilo-gréseuses" (more than 300 m) has not yielded graptolites, and its correlation as Pridoli is based on palynological data. Fragments of psilophytes have been observed from this highest unit.

SOUTHERN TUNISIA SYNTHESIS — At present, the Silurian is known only from bore holes, and a useful stratigraphic synthesis of this region is impossible. Most biostratigraphic data (particularly graptolites, chitinozoans, miospores, acritarchs) and sedimentologic descriptions are unpublished or imprecise. Misidentifications of endemic graptolite species are common. Confusion between bio- and lithostratigraphic units variously defined from outcrops or bore holes (Bekkouche and Perriaux, 1991); a likely diachronism of facies from south to north (Boumendjel, 1987); and a scarcity of cores help explain correlation problems (Legrand, 1985b). The entire Silurian is probably present in the cental part of this region, whereas the Lower Silurian is absent in the north, except in south Tunisia. The eastern and northern Tripolitania regions will be useful for future comparison with the Berkine Basin.

WESTERN TASSILI N'AJJER, ISSAOUANE-TIFERNINE BASIN, WESTERN TINRHERT, AND EASTERN EL BIOD HIGH

These areas lie in a hinge zone between the Illizi Basin, the eastern Tinrhert, and the Berkine Basin to the east and the basins of the Algerian Sahara to the west.

Western Tassili N'Ajjer — The hiatus prior to the Silurian transgression is longer here than in the central Tassili N'Ajjer. At Foum Ennemil (Fig. 1, locality 25; Fig. 10), the base of the Foum Ennemil Shale is upper Llandovery (Telychian, *"Monograptus" guerichi* Zone). However, in some places, thin beds of sandstone are present below the lowest shales. They belong to the upper *Coronograptus cyphus* or lower *Coronograptus gregarius* Zones and suggest shoreline deposits of a previous transgression in the lower–middle Llandovery boundary interval (Legrand, 1985b, 1999). To the east, the upper Llandovery Tsila Sandstones appear within the Foum Ennemil Formation (Legrand, 1985b, 1999). In the upper part of this latter formation, siltstone are more and more prominent, and their age is late Wenlock (Legrand, 1968, 1969a, 1969b, 1985b, 1994). In overlying strata, coarser-grained silicilastics become more and more important (Esker Nehed Forma-

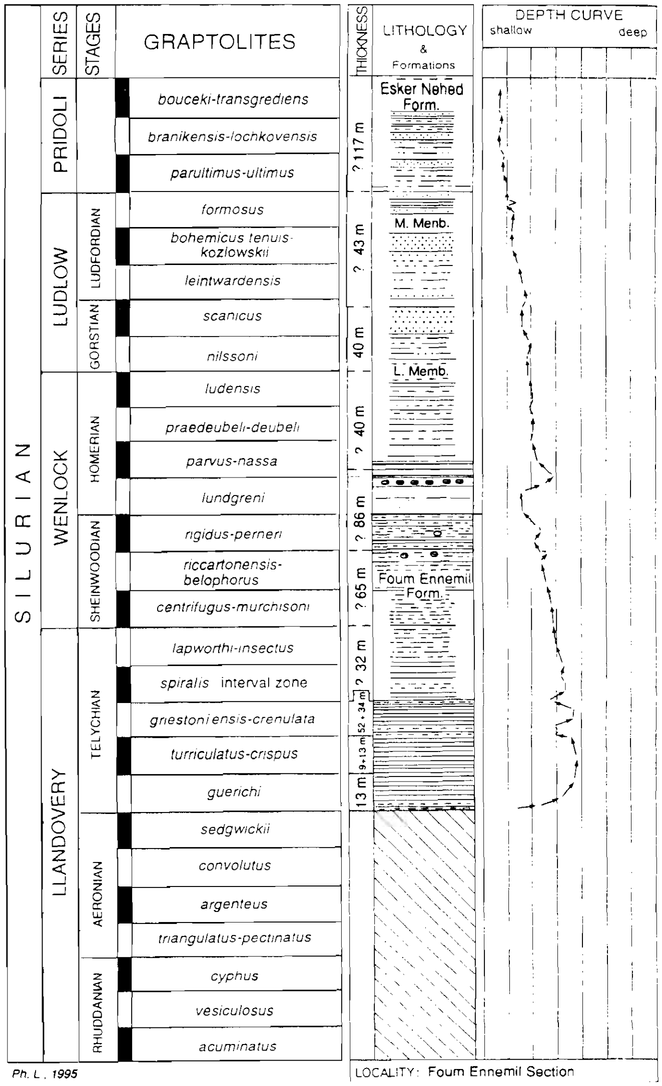


FIGURE 10 — Foum Ennemil section (Algeria); lithology, formations, and depth curve. Correlations between the local and British graptolite zonation approximate; correlations tentative in upper part of section. After Dubois and Mazelet (1965) and Legrand (1967, corrected 1985; 1999).

tion, with a latest Wenlock base). After a return to shale deposition, siltstones and sandstones dominate, but biostratigraphic data are rare on the probable Ludlow and Pridoli parts of the Silurian (Legrand, 1985b, 1994).

ISSAOUANE–TIFERNINE BASIN — In this basin (Fig. 1, locality 26), the Silurian has been recognized only in bore holes, except for the small Ta-n-Elak outcrop (Claracq et al., 1958; Legrand, 1962; Fig. 1, loc. 26). The Silurian is very similar to that observed in the western Tassili N'Ajjer, especially in the development of upper–uppermost Wenlock sandstone.

WESTERN TINRHERT — In the western Tinrhert (Fig. 1), as in the Issaouane Tifernine, the Silurian is represented by a lower argillaceous sequence, the Tin Fouy Formation (claystones), overlain by an argillaceous–arenaceous succession subdivided into the Mederba Formation and the Oued Tifist Formation (Jardine and Yapaudjian, 1968). Little has been published, but the Silurian transgression is known to be Telychian, at least in some bore holes (Legrand, 1999). Upper–uppermost Wenlock sandstone is also found.

EASTERN FLANK OF EL BIOD HIGH — The Silurian in this area (Fig. 1) is known only in bore holes. The lower part of the sequence is argillaceous, but the age of the base is variable locally and can be late Llandovery or late Wenlock, as elsewhere in a large part of the Sahara in this area. The resumption of the sandy sedimentation was latest Wenlock or younger, and an understanding of Silurian stratigraphy has been complicated by Hercynian erosion (Legrand, 1985b). Only palynological analysis allows a basis for correlation.

REGIONAL SYNTHESIS — The evolution of Silurian deposition in the western Tassili N'Ajjer, Issaouane–Tifernine Basin, and western Tinrhert was related to developments that can be observed further east. There is evidence of an east–west transgression. The sea was a little deeper and the fauna more abundant and less endemic in the east. In the north, this transgression was complicated by another event, the "Saharan basculation" (Legrand, 1969a, 1994) in the late–latest Wenlock, the effects of which were beginning to be felt in the northern part of the Berkine Basin.

AMGUID ANTICLINORIUM, WESTERN TASSILIS, AND ADJACENT BASINS

The Amguid, western Tassilis and the Habadra and Oued Djaret Basins are a very poorly described region.

AMGUID ANTICLINORIUM AND MOUYDIR BORDER — In the Amguid area (Fig. 1, locality 27), the Silurian is similar to that in the western Tassili N'Ajjer. However, the Silurian transgression occurred earlier on the western flank of this anticlinorium (*"Monograptus" sedgwickii* Zone). This is well illustrated near Felar-Felar on the western flank of the Amguid spur. Most of the lowest argillaceous sequence, the Tiounkeline Formation (claystones), is late Llandovery–Wenlock, and sandstone sedimentation reappeared in the late Wenlock as the Adrar Tassedit Formation (Legrand, 1985b). Higher biostratigraphic data are lacking until the top of massive Praguian sandstone. The Silurian on the border of Mouydir (Fig. 1, locality 28) has not been studied.

JEBEL IDJERANE ANTICLINE–JEBEL AZAZ–ERS OUM EL LIL — These outcrops (Fig. 1, localities 30–32) are very interesting (Fig. 11). The terminal thin sandstone bed of the

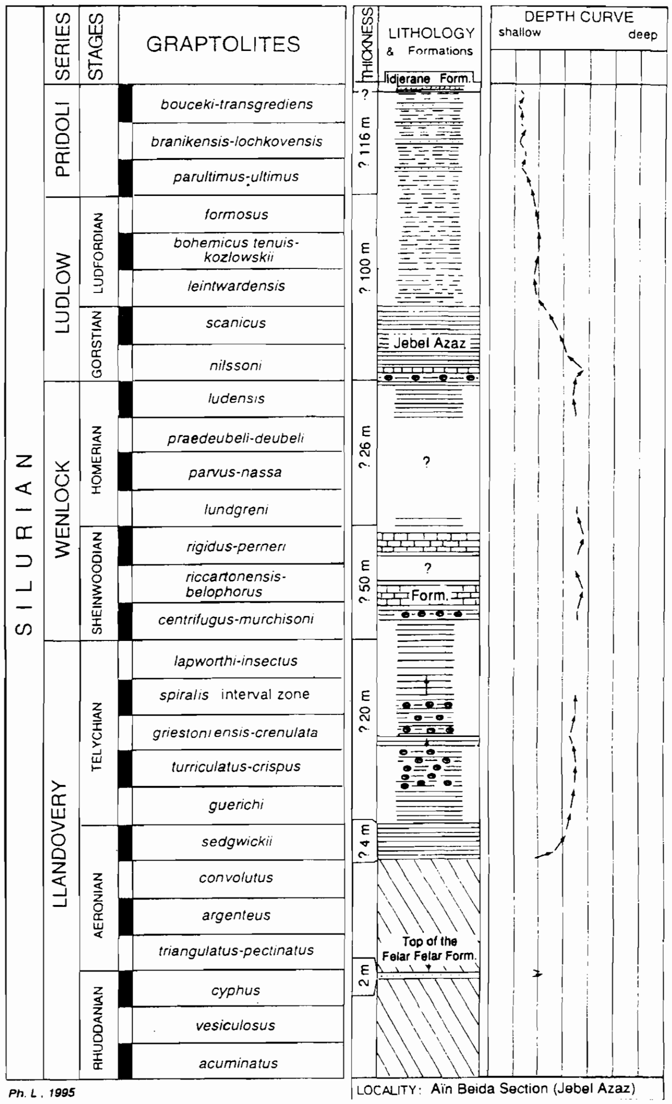


FIGURE 11 — Aïn Beïda (or Jebel Azaz) section, Algeria; the lowest part of the section is at Ers Oum el Lil; lithology, formation, and depth curve. Correlations tentative in upper part of section. After S. Rouaix et al. (*in* Legrand, 1962) and Legrand (1985b, 1994, 1999).

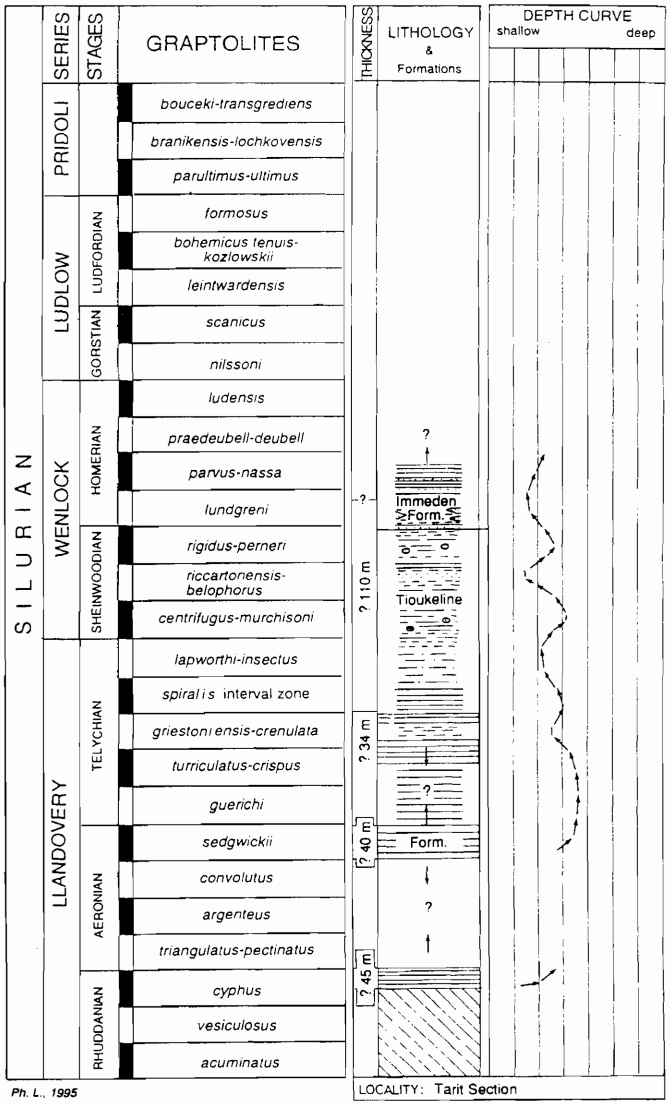


FIGURE 12 — Tarit section, Algeria; lithology, formations, and depth curve. Correlations tentative in uppermost part of section. After Legrand (1966, unpublished; 1970, 1985b, 1999).

Felar-Felar Formation, referred to the upper *Coronograptus cyphus* Zone (Legrand et al., 1959; Legrand, 1994, 1999), is significant because it is the benchmark used in reconstructing and dating this first transgression. The age of the overlying shales of the Jebel Azaz Formation is probably late Llandovery (*"Monograptus" sedgwickii* Zone or *"M." guerichi* Zone). This is the eastermost section, where limestones are somewhat developed. There is no uppermost Wenlock sandstone. The lower Ludlow has been recognized on the basis of graptolites. The total thickness is relatively small (200–250 m). The overlying sandstone of the Jebel Idjerane Formation is present. Its lower member with brachiopods and trilobites is Ludlow and/or Pridoli (Legrand, 1981, 1985b).

JEBEL SETTAF, FOUM BELREM, TASSILI OF TARIT — To the east of Foum Belrem (Fig. 1, locality 33), the Silurian does not show the important changes observed along the Tassilis further east. However, the age of the base of the Tioukeline Shale varies from place to place (*"Monograptus" turriculatus* Zone to *Monoclimacis griestoniensis* Zone) (Legrand, 1999). The coarser siliciclastic sedimentation that elsewhere announced the end of the Wenlock can be observed. To the west of the main fault at Foum Belrem in the Tassili of Tarit section (Fig. 1, locality 34; Fig. 12), the older *Coronograptus cyphus* Zone reappears at the base of the argillaceous sequence (Legrand, 1970, 1999). The presence of the middle Llandovery has not been demonstrated. Higher up, the upper Llandovery and the Wenlock are poorly represented, and their thickness decreases. The coarser sandstones of the upper Wenlock, however, are always present. The last outlier of the *C. cyphus* Zone, a sandstone bed, is observed further west at the Adrar Tikkadouine, east of Ouallène (Fig. 1, locality 34). The middle Llandovery is absent or very thin. The upper Llandovery shale and the upper Wenlock sandstone are both thick. Finally, to the west of Ouallène, Silurian that may begin in the *"Monograptus" turriculatus* Zone is much thinner, and the shales do not exceed 125 m (Legrand, 1995a). Above these shales, coarser siliciclastics become more and more important (Foum Immeden Formation). Lithostratigraphic subdivisions have been defined (Biju-Duval et al., 1968), but the proposed Ludlow, Pridoli, and Lochkow ages of these units are hypothetical.

Habadra and Oued Djaret Basins — The Silurian is known as the Meredoua Formation (claystone) in the Habadra (Fig. 1, locality 29) and Oued Djaret Basins (Figure 1, locality 35) from bore holes (Legrand 1985b). However, cores are few in number, and there is practically no biostratigraphic data available. Sequence stratigraphic analysis has been tried, but without biostratigraphic control (Talah, 1987; Talah and Delfaud, 1989; Talah et al. 1995).

Regional synthesis — The Amguid anticlinorium is a major structural feature of the Algerian Sahara. In the Silurian, it seems to have played a minor role in deposition, in comparison to the succession on its two flanks. To the north on the El Biod High, the extent of the "Saharan basculation" appears to have been considerable. Along the western Tassilis, there is limited evidence for an initial early Llandovery transgression that probably came from the southeast. A second transgression in the late Llandovery came from the east. A middle Llandovery hiatus is possible. A return to a sand deposition before the end of the Wenlock is the rule.

Azzel Matti and Tanezrouft

The Azzel Matti and Tanezrouft regions (Fig. 1, locality 36) occupy a significant place between the Eglab Massif and the northern extension of the Ahaggar uplift, and have distinctive characters in the Lower Paleozoic. The presence of lower Llandovery (*Coronograptus cyphus* Zone) argillaceous–arenaceous facies in the upper Ain ech Cheikr Sandstone should be noted, as well as the occurrence of the middle Llandovery and the great thickness of the Wenlock (Fig. 13; Legrand, 1985b, 1999). Faunal associations show still some species known only in the Tassilis (Legrand, 1999). There was no sandstone deposition during the late Wenlock, and shale deposition (Ain ech Cheikr Formation) predominated (Legrand, 1985b). Above the Ludlow limestones, there are no diagnostic fossils of the Pridoli, and none from the Silurian–Devonian boundary interval. To the east, drill cores show the Silurian in the Azzel Matti Basin and also in the northern Tanezrouft to the west (Remack-Petitot, 1960). However, the Lower Silurian pinches out in the Eglab Massif, and there are no outcrops of Silurian at Bir Ould Brini (Fig. 1, locality 36), where Lower Devonian sandstones overlie the Ordovician.

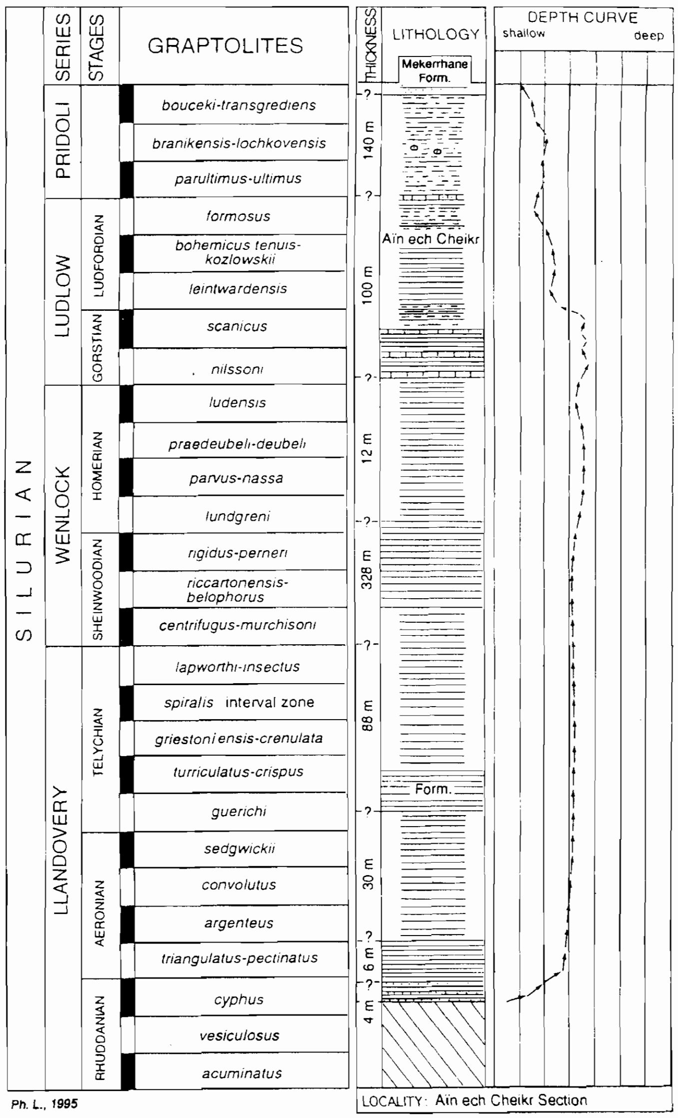


FIGURE 13 — Aïn ech Cheikr section, Algeria; lithology, formations, and depth curve. Correlations tentative in upper part of section. After Legrand (1966, unpublished; 1985b, 1999).

Touat, Gourara, and the Ougarta Range

These are three areas with small outcrops of Silurian between the Azzel Matti and the Moroccan Anti-Atlas.

Touat — Only the Upper Silurian is present in outcrops of the Touat (Legrand, 1962, 1981, 1985b; Fig. 1, locality 37). Above a limestone bed with *Pristiograptus* sp. and *Saetograptus chimaera*, an argillaceous–arenaceous sequence with two limestone beds has been observed. It is overlain by the Touat Formation, a sandstone unit which is very argillaceous in its lower and middle parts. The lower part is Pridoli, but the middle part is likely Lower Devonian. The thickness of the Ludlow and Pridoli does not exceed 200 m.

Gourara — In the Gourara (Fig. 1, locality 38), there is some reduction of thickness in the Silurian, and a greater development of limestone beds, particularly at the level of the Pridoli *"Scyphocrinites"* beds. The lowest known graptolites were collected 10 m above the top of the *Hirnantia*-bearing Ordovician sandstone. These graptolites belong to the *"Monograptus" turriculatus* Zone. Thus, argillaceous sedimentation of the Fegaguira Formation (claystones) began in this or the preceding zones (*"Monograptus" guerichi* or *"M." sedgwickii* Zones) (Legrand, 1985b, 1999). A major subsidence took place in the late Wenlock. Higher limestone beds yield a lower Ludlow graptolite fauna, and the passage to the Devonian occurs within a 200 m-thick argillaceous sequence. The lowest Lochkowian trilobites and brachiopods are known in the base of the overlying Benyassin Formation sandstones (Legrand, 1985b).

Ougarta Range — The Silurian of the Ougarta Range was discovered by Rey (1914) and described by Menchikoff (1930) and Poueyto (1952). Shaly deposition of the Oued Ali Formation began in the Ougarta Range (Fig. 1, locality 39) during the *"Monograptus" convolutus* Chron. However, these shales are underlain by a few meters of sandy–shaly deposits that may be contemporaneous with the *Coronograptus gregarius* or even the *C. cyphus* Zones (Fig. 14). Unfortunately, the rare graptolites at this level are indeterminable (Legrand, 1999). A considerable thickness of upper Wenlock is present, with important lateral changes possibly caused by tectonism. The upper–uppermost Wenlock is silty or sandy with numerous trilobites (Legrand, 1969a). The Ludlow includes several beds of limestone. The Pridoli is also present, and includes *"Scyphocrinites"* horizons (this is the type locality of *Marhoumacrinus*). This area illustrates the Silurian–Devonian boundary (Legrand, 1977).

Regional synthesis — The Silurian of this region is different from that of the Tassilis, as it has limestones in the Wenlock, Ludlow, and Pridoli. Nevertheless, evidence for most of the sedimentologic events recorded in the Tassilis can be found here.

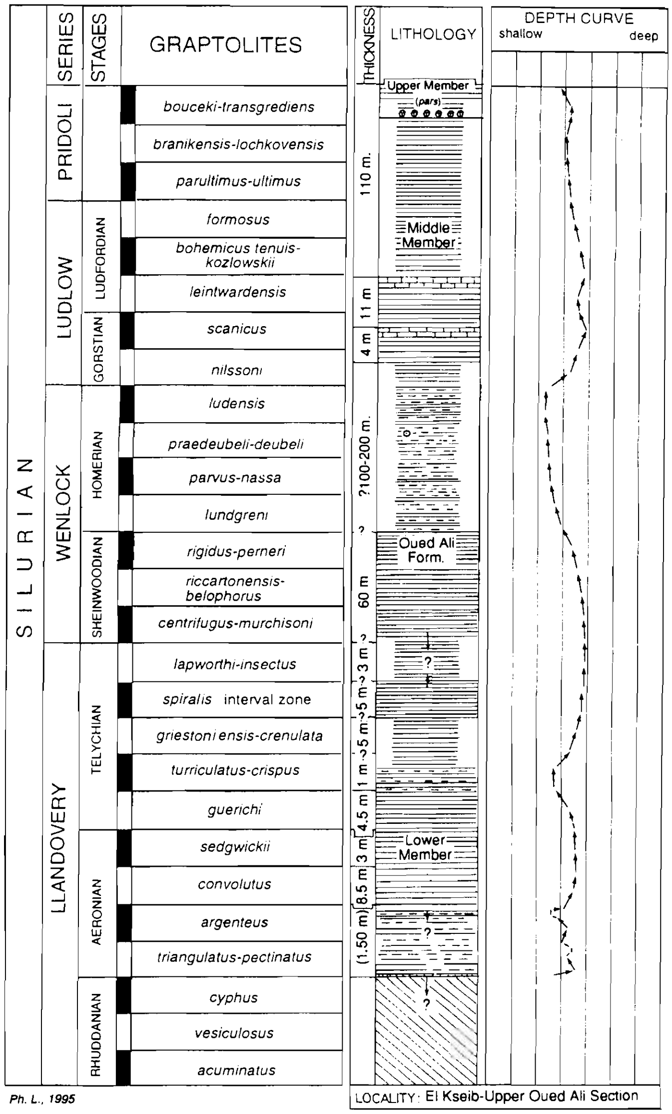


FIGURE 14 — El Kseib–upper Oued Ali section, Algeria; lithology, formations, and depth curve. After Menchikoff (1930), Poueyto (1952), and Legrand (1969a, 1985b, 1988, 1999).

Great Western Erg and North Saharan Basins

The Silurian is recognized in drill cores in the very large region of the north Saharan basins (Fig. 1). In these basins (apart from the Oued Rharbi region, where the base of the lower member of the Oued Mehaiguene Formation is at least upper Llandovery; Fig. 1, locality 39), numerous drill cores show a late Wenlock (*Cyrtograptus rigidus* Chron) transgression (Legrand and Nabos, 1962; Willefert, 1962; Legrand, 1981, 1985b, 1994). The upper member of the Oued Mehaiguene Formation has upper Wenlock graptolites and overlies Upper Ordovician rock with Ashgillian chitinozoans on the M'Kratta sheet (Oulesbir and Paris, 1995). The stratigraphy of the Oued Mehaiguene is poorly known, and the formation extends into Devonian (Lochkovian) rocks with graptolites (Legrand, 1964, unpublished) and palynomorphs (Magloire, 1967; Boumendjel, 1987). To the east near the El Biod High, the Silurian is very condensed and has the southeasternmost *Scyphocrinites* beds. There is no abrupt break with resumption of the sandstone deposition, and there is no evidence of the emergence of the Hassi Messaoud region (Fig. 1, locality 98; Legrand, 1985b).

Bechar Region, Algeria, and Eastern Moroccan Anti-Atlas

This region is at the junction of the Anti-Atlas, the Ougarta Range, and the South Atlas fault.

Bechar region — The Ben Zireg area (Fig. 1, locality 41; Fig.15) shows the transgression of the uppermost Wenlock (*Cyrtograptus lundgreni* Zone) across the Ordovician. Up to the Lochkov, the Silurian is 240 m thick. More to the north, a similar sequence is found at Teniet el Ghenia (Fig. 1, locality 90), but the thickness is only 110 m (Jaeger and Massa, 1965). The Silurian transgression occurred later than elsewhere in the north Saharan basins. Southeast of Bechar (Fig. 1, locality 40), some bore holes show a very thin Upper Silurian sequence.

Eastern Anti-Atlas — The Silurian is present chiefly in the Tazzarine Depression (Fig. 1, locality 48), the Maïder Basin (Fig. 1, locality 47), and the Tafilalt Basin (Fig. 1, locality 43). This is a very interesting area with significant facies changes. Consequently, it is difficult, at least in the Lower Silurian, to construct a composite section. The stratigraphy (see Destombes et al., 1959; H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 256–264) requires a few comments. The lower Llandovery is known to the east, where it is confined to a "subsidence channel" (Tizi Ambed Formation; Fig. 1, locality 45) along the western edge of the Tafilalt (Fig. 16). Higher in the sequence, locally coarse-to-argillaceous sandstone of middle Llandovery age (Aeronian) is found. Elsewhere, the transgression was later, and, at least, Telychian. The sequence becomes very homogeneous through the argillaceous or argillaceous–calcareous Tamaghrout Formation (Fig. 1, locality 46), the upper part of which is Wenlock. Higher strata have limestones beds in the shales, and the sequence seems to be condensed, as at the

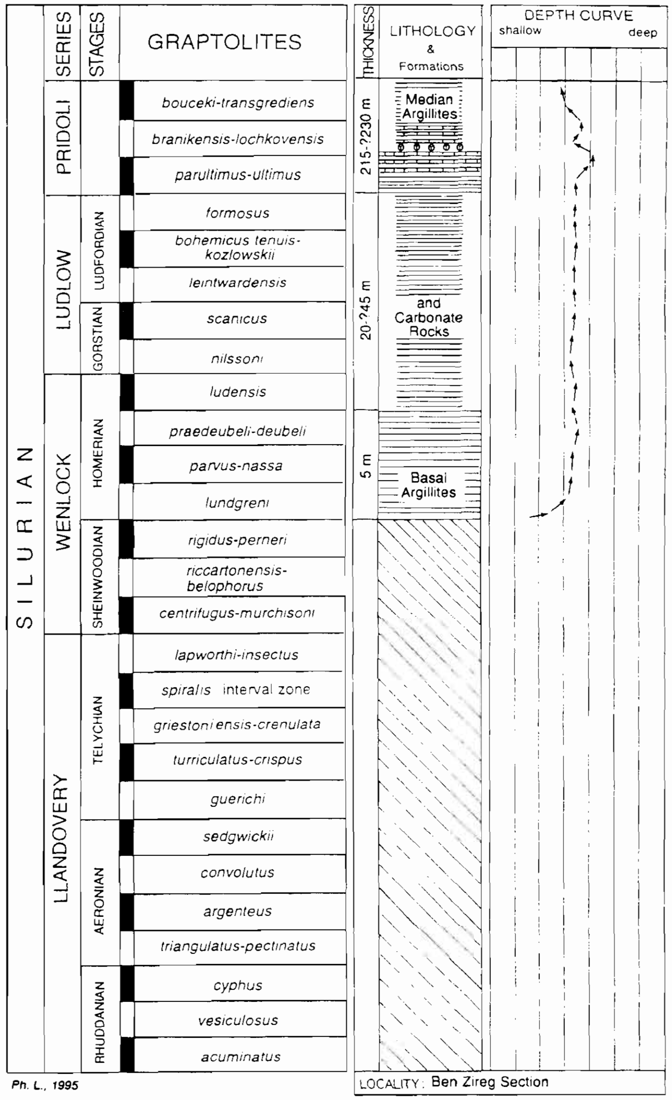


FIGURE 15 — Ben Zireg section, Algeria; biostratigraphy, lithology, formations and depth curve. After Massa et al. (1965) and Jaeger and Massa (1965).

Hamar Laghdad section near Erfoud (Fig. 1, locality 42) with its "*Scyphocrinites*" beds (Fig. 17) (Massa et al., 1965; Jaeger and Massa, 1985; H. Hollard and S.Willefert in Destombes et al., 1985, p. 259–262).

Regional synthesis — Although discontinuous exposures make correlations difficult, this region illustrates the role of local tectonics in the Silurian transgression.

Central and Western Moroccan Anti-Atlas

The southern slopes of the Anti-Atlas border the Dra Plain (Fig. 1). Small Silurian outcrops occur along this

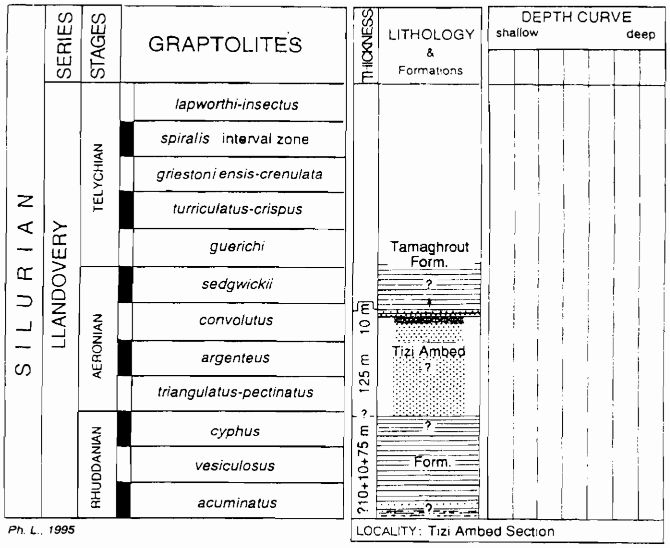


FIGURE 16 — Tizi Ambed section, Morocco; lithology and formations. Correlation of lower 20 m uncertain as sparse graptolites in the sandstones; the age of the uppermost ferruginous sandstone problematical. After Destombes et al. (1959) and H. Hollard and S. Willefert (*in* Destombes et al., 1985).

1000 km-long belt.

Central Anti-Atlas — The regional reference for the Silurian is the Iriqui section (1,150 m thick), which is located east-southeast of Foum Zguid (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 245–252; Fig. 1, localities 47, 49; Fig. 18). The Silurian begins with sandstones that fill erosional irregularities. These sandstones (lower Aïn Chebi Formation) are middle Llandovery (locally *"Monograptus" convolutus* or *"M." sedgwickii* Zone). However, a little to the east near Mel'Ag (Fig. 1, locality 48), a shaly intercalation discovered by Destombes (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 253) is lower Llandovery (*Cystograptus vesiculosus?* and *Coronograptus cyphus* Zones). Muddy sedimentation began in the Telychian (*"Monograptus" turriculatus* Zone, upper Aïn Chebi Formation). The lower Telychian is especially rich in graptolites (Foum Zguid and Jebel Amsaïlikh sections; Figure 1, localities 47, 50) (Waterlot, 1941; Jacquemont and Hollard, 1956). Clays with calcareous concretions comprise the lower Amsaïlikh Formation, which has Wenlock graptolites. The Ludlow (upper Amsaïlikh Formation) is a relatively thick, silty shale with rare limestones only at the top. The overlying, thick Iriqui Formation (Pridoli) has ferruginous limestones intercalated with laminated clays and has *"Scyphocrinites"* near its top.

To the west, the thickest Silurian section lies in the Oued Aguemamou near Jebel Addana (Fig.1, localities 51, 52). Another section is along the Oued Akka (Fig. 1,

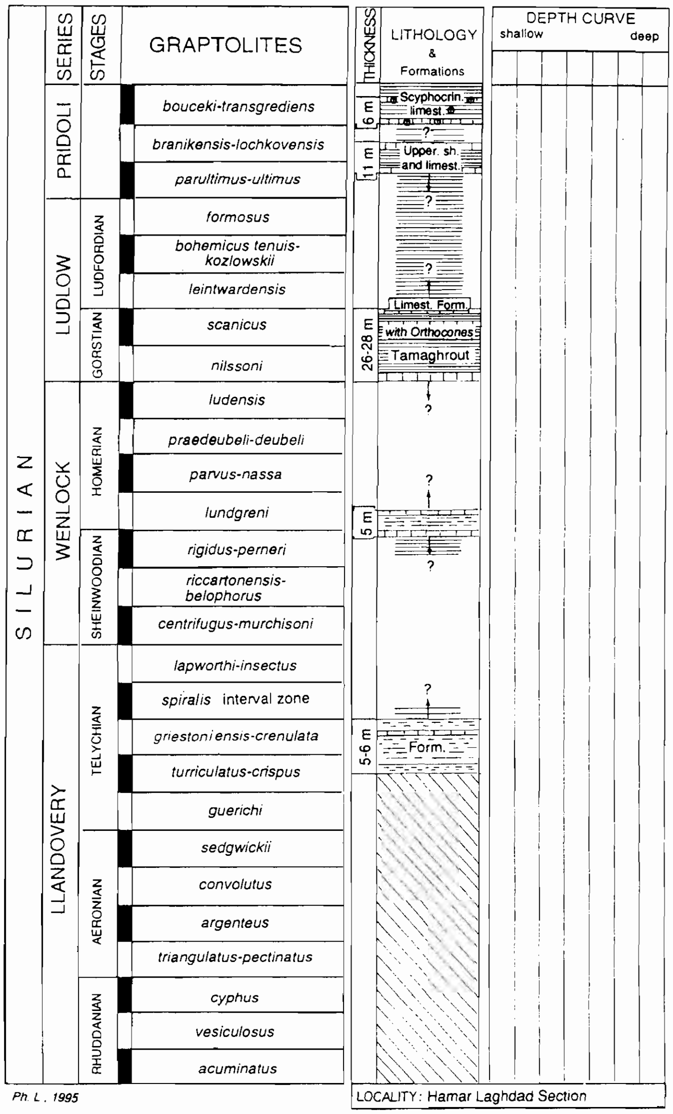


FIGURE 17 — Hamar Lagdad section, Morocco; lithology and formations. After Massa et al. (1965), Jaeger and Massa (1965), and H. Hollard and S. Willefert (*in* Destombes et al., 1985).

locality 53). Generally speaking, there are more calcareous deposits near the Wenlock–Ludlow boundary, and some lithofacies change in the Iriqui Formation. Thus, the limestones are developed in the middle part of the formation amidst sandy intercalations (for example, between Akka and Assa; Fig. 1, localities 53, 55).

Western Anti-Atlas — The representative Silurian section is at Aïn Ouïn-n-Delouine (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 240 et sq.; Fig. 1, locality 56; Fig. 19). This section is noteworthy because of the presence, at the base of the Silurian shales, of *Parakidograptus acuminatus.* On the other hand, the upper lower and middle Llandovery zones have not been recognized.

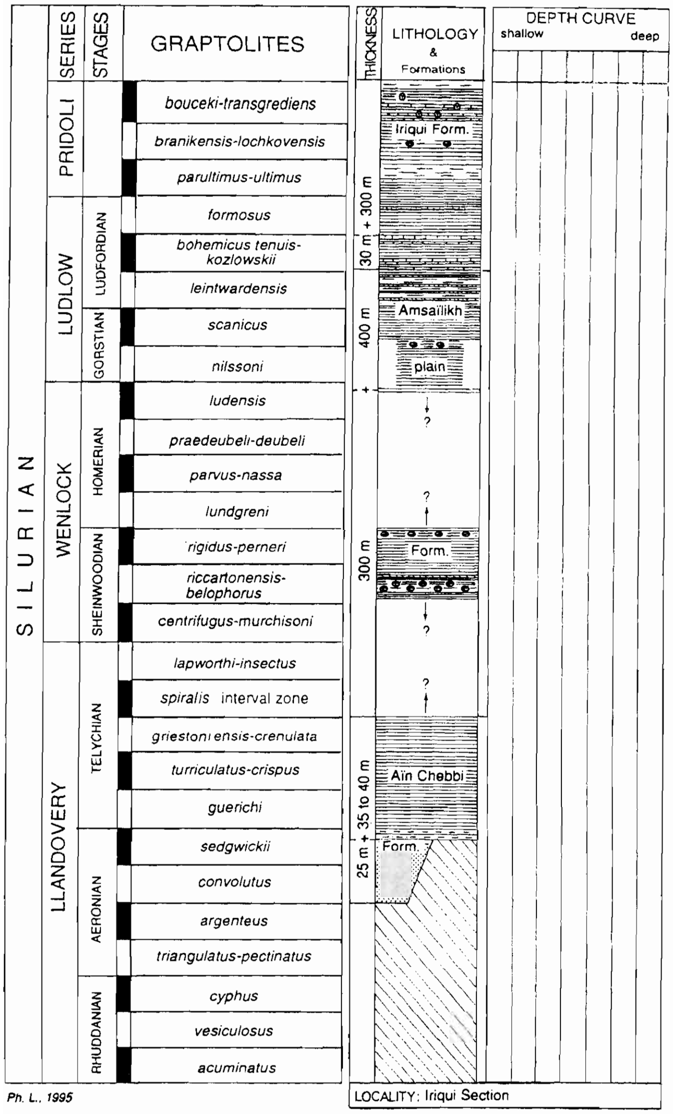


FIGURE 18 — Iriqui section, Morocco; lithology and formations. After H. Hollard and S. Willefert (*in* Destombes et al. (1985).

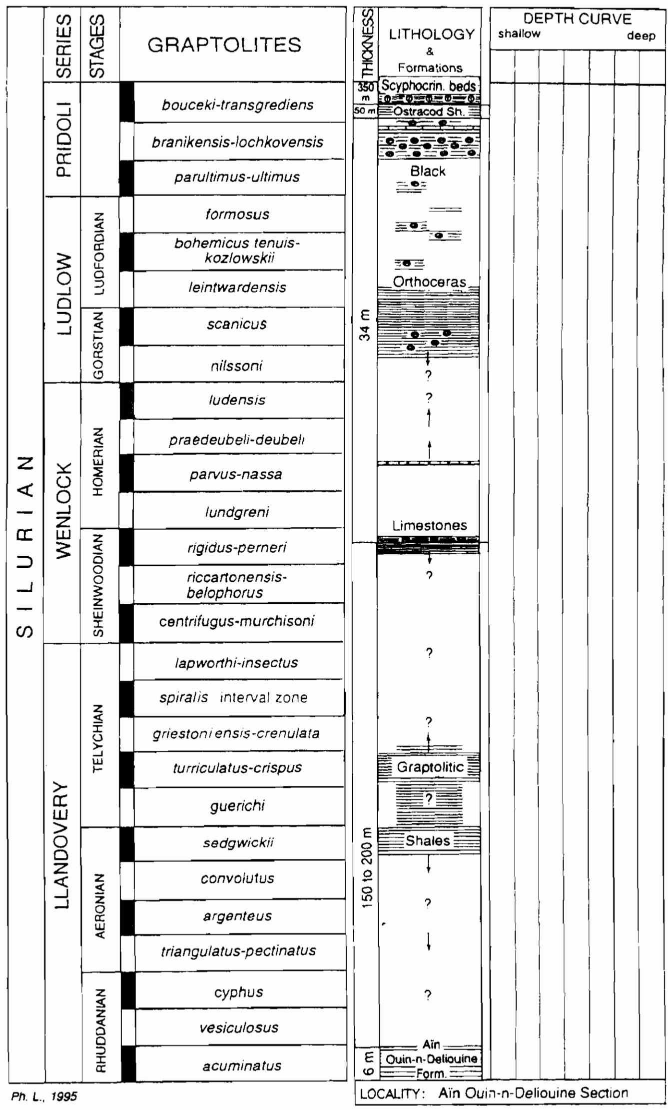


FIGURE 19 — Aïn Ouin-n-Deliouine section, Morocco; lithology and formations. After H. Hollard and S. Willefert (*in* Destombes et al., 1985).

If these zones are present, they have a reduced thickness. Only the top of the Aeronian and lower Telychian are present. The lower Wenlock has not been identified, but a very condensed uppermost Wenlock has been well dated (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 242). The Ludlow is particularly argillaceous, and has levels of calcareous nodules. Carbonate deposits begin at the top of the Ludlow and continue into the lower Pridoli. *"Scyphocrinites,"* the development of sandstones in the upper part, and its great thickness are noteworthy features in the Pridoli.

Regional synthesis — The description of these two sections may give the impression that the local sequence is always the same. In fact, there are many subtle variations. According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 255, 256), local depositional history included Llandovery transgression and the deposition of black shales over a large area, followed by the appearance of the first limestones. Lateral facies are related to appearance of a central zone in which fine sand accumulated, and to two small basins. In the Pridoli, the small southwest basin lay at the foot of a slope and received *"Scyphocrinites"* debris and some turbidites. The small northeast basin was open to the northern seas and received fine sands.

Tindouf Basin

In the central Tindouf Basin, only a few bore holes have given useful information on the Silurian (i.e., depth, thickness, and facies), but little other stratigraphic data are available. On the other hand, the Silurian crops out along the southern border of the basin (Gevin, 1960; Sougy, 1964; Legrand, 1969b, 1985b).

EASTERN PART — The Silurian in the eastern Tindouf Basin is poorly known. At Bou Bernous (Fig. 1, locality 57), it consists of Llandovery shale and a calcareous Wenlock interval, and is overlain by shales of unknown age. The whole 120 m-thick sequence is capped by a conglomeratic bed assigned to the Lower Devonian (Gevin, 1960). In the Aouinet bel Legrâ section (Fig. 1, locality 58), the Silurian consists of about 10 m of shale with thin siltstone beds, most of which must be assigned to the Wenlock. These shales are overlain by a conglomeratic bed of late or latest Wenlock age (Legrand, 1969b). Above this conglomerate, a shaly–sandy sequence occurs; part of which is probably Ludlow and Pridoli. The boundary with the Devonian is uncertain. Recently, evidence from a shallow bore hole demonstrated that the basal shales (not visible outcrops) are lower Llandovery. A ferrugineous bed marks the base of the Wenlock shales and is the upper bracket of the hiatus (Paris et al., 1995). Finally, the Oued El Hamra section (Gevin, 1960) further west (Fig. 1, locality 59) needs to be re-examined.

WESTERN PART — Silurian rocks are present in the "Zemmour noir" section (west of Gara Bouya Ali; Fig. 1, locality 60). They begin with black shales that apparently thin from north (30 m) to south (6 m). These shales, referred to the *"Monograptus" triangulatus* Zone and to the uppermost *"Monograptus" sedgwickii* Zone (Willefert, 1988), are overlain by Wenlock limestones, and then by shales and alternating shales and limestones. The next interval consists of calcareous beds with *"Scyphocrinites;"* a thick, argillaceous unit with another bed of *"Scyphocrinites;"* and then more shales. The Silurian is about 90 m thick (Sougy, 1964). Precise biostratigraphic data is needed to correlate these sections more accurately.

REGIONAL SYNTHESIS —The thickness of the Silurian varies along the southern border of the Tindouf Basin and is very thin in comparison to the Silurian of the southern Anti-Atlas. The Reguibat shield does not seem to have subsided much in the Silurian. Because of the lack of biostratigraphic data in the central basin, paleogeographic maps in this area are purely hypothetical.

Taodenni Basin and Its Borders

The Silurian of this immense basin is poorly known. On the northern border, the Silurian seems absent in the eastern part from the 7° meridian. Two bore holes in the central part remain unpublished. In the south, some information on a bore hole in Mali is known (Legrand, 1999). In summary, it is only in the western part of the basin that useful sections have been described.

MAURITANIAN ADRAR — The stratigraphy suggested by Monod (1952) is different from Trompette's (1973), which I follow here, though some problems remain unresolved. The Silurian of the Mauritanian Adrar (Figure 1, locality 61) consist of the 70 m-thick Oued Chig Group (Trompette, 1973), which may be subdivided into three sequences. The composite section (Figure 20) is rather schematic, and includes biostratigraphic data from about twenty sections. Consequently, the proposed sedimentary breaks must be accurate. The Silurian begins in the lower? or middle Llandovery, and the lower Telychian may be absent or extremely thin. A break is possible within the Wenlock. The position of the Wenlock–Ludlow boundary is not known precisely, but the lower Ludlow is present. The Pridoli is absent, probably because of pre-Devonian erosion and unconformity. It is noteworthy that the shales are always silty or sandy, and that the sandy intercalations of Middle and Late Silurian age are often rich in iron oxides and phosphates.

NORTHWEST MARGIN OF TAOUDENNI BASIN — The first Silurian outcrop with numerous monograptids was found by Monod in 1935 (Menchikoff and Monod, 1936) near Tinioulig dans la région de Mejahouda (Fig. 1, locality 62). Other fossiliferous outcrops are known in this-area, and may represent this same interval. The Silurian here consists of 20 m of shales referred to the Mlahès Group. Lower and middle Llandovery and upper Wenlock graptolites are found in these shales (Deynoux et al., 1985; Willefert, 1988). Further northeast, the Silurian is absent, and the Middle Devonian rests on the Mejahouda Group, of supposed Ordovician age (Villemur, 1967).

TAGANT, ASSABA, AFFOL, AND HODH — In the Tagant, northern Assaba, and Affol (Fig. 1, locality 64), the Silurian successions are similar. According to J. Delpy (1967, unpublished data, *in* Deynoux et al. 1985, p. 387), there are two formations at the Akjat section. The Itilen Formation, about 25 m thick, is made up of coarse, ferrugineous sandstone. If this formation overlies the Dikel Group, which is partly of glacial origin, it could be Silurian (or uppermost Ordovician). The Charania Formation, with an average thickness of 5–10 m, has two graptolite faunas. The lower could be late Aeronian–early Telychian,

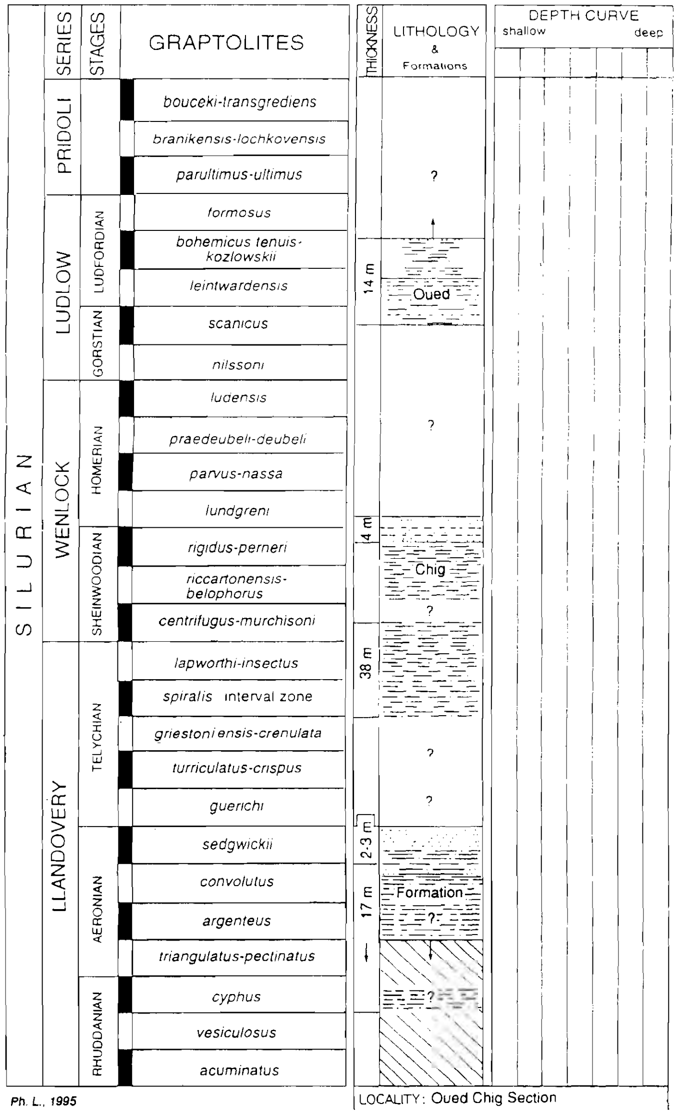


FIGURE 20 — Oued Chig section, Mauritania; composite section; lithology and formations. After Trompette (1973).

the higher, because of faulty identifications, cannot be evaluated.

The Khatt region of Mauritania lies in the transition between the Adrar and the Tagant region (Fig. 1). The Silurian crops out extensively, but paleontological data are fragmentary. Nicklès (1947) noted graptolites collected by Jacquet in 1935 (Jacquet and Monod, 1935) of early Telychian age. J. Delpy (1967, unpublished data; *in* Deynoux et al., 1985, p. 388) reported shales with Ludlow graptolites. Trompette (1973) described a section of Aeronian–lower Telychian shales, a sandy sequence of undetermined age, and a third interval of lower Wenlock shales.

Further east in the Hodh (Fig. 1), graptolitic shales are found east of Tichit and Oulata (Deynoux, 1980; Deynoux et al., 1985). Glacio-marine rocks transitional into overlying graptolitic shales are found near Aratane (Fig. 1, locality 63). The maximum thickness of the Aratane Group is about 80 m. Preliminary study by Willefert (1988) noted Ashgillian and Lower Silurian graptolites, and recognized some similarities between these faunas and the graptolites of the eastern Tassili N'Ajjer. Underwood et al. (1998) reported the *"Glyptograptus" persculptus, Parakidograptus acuminatus*, and *Atavograptus atavus* Zones, although the eponymous species were not found. More detailed studies are required. Chitinozoans are associated with these graptolites (Paris et al., 1998). Graptolites collected from higher shales range from the lower Llandovery (upper Rhuddanian) to middle Llandovery (Aeronian) (Nicklès, 1947; S. Willefert *in* Deynoux et al., 1985, p. 391). Although the Silurian thins eastwards, it cannot be determined whether or not it persists east along the axis of the Taoudenni Basin. Indeed, the Silurian east of the Hodh is suggested by a drill hole in northern Mali at the southern edge of the basin.

Regional synthesis — A better knowledge of the Silurian of the Taoudenni Basin is essential to understanding the paleogeography of the Silurian on the north African border of Gondwana and its relations with South and North America. More biostratigraphic study would be useful, but requires the release of information on new bore holes.

Bove Basin, Guinea, and Guinea Bissau

The Bove Basin (Fig. 1) is difficult to study because the outcrops are discontinuous and difficult to correlate. The Silurian has been known here for a long time (Sinclair, 1918; de Chetelat, 1938), but it is only relatively recently that Silurian stratigraphy has been studied (Villeneuve, 1984; Villeneuve and Da Rocha Araujo, 1984; Deynoux et al., 1985; Villeneuve et al., 1989; Racheboeuf and Villeneuve, 1992). Unfortunately, the data, collected at different times, are not always easy to compare, and the results sometimes conflict. Nevertheless, we consider a generalized column to be relatively useful (Fig. 21). The Silurian consists of the Télimélé Group, with three formations and a 150–330 m thickness. Shales are predominant, but three sandy levels occur: the Bambaya, Sangui, and Dantara Sandstones. To the north of the basin, the Kolda 1 (Ko 1) bore hole has 150 m of graptolitic shales first attributed to the Ludlow (but now to the uppermost Wenlock) (Marquis and Couppey, 1961, unpublished data). These faunas require restudy.

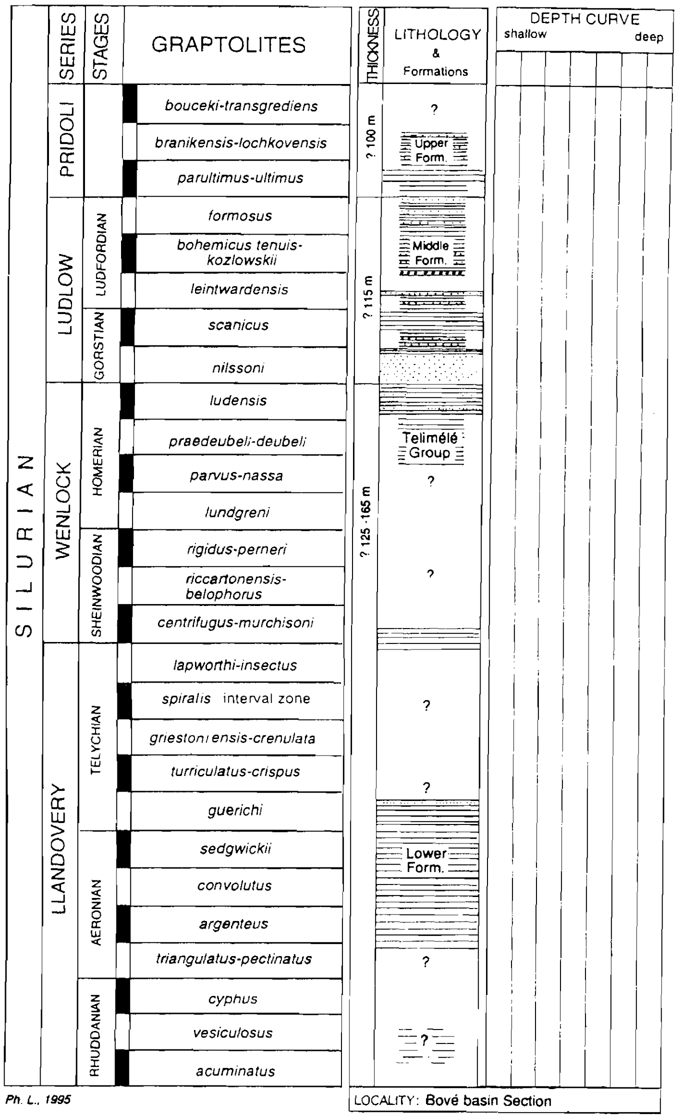


FIGURE 21 — Bov Basin section, Guinea–Guinea-Bissau; composite section; lithology, and formations. After Villeneuve and Da Rocha Araujo (1984) and Villeneuve et al. (1989).

REGIONAL SYNTHESIS — The Bove Basin is in a very important area because of its location between the Mauritanides and Rokelides chains. It constitutes the last outcrop belt in Africa that earlier faced the Suwannee Basin in subsurface Florida (Villeneuve, 1988). Consequently, its fauna is particularly significant.

HIGH ATLAS AND DJEBILET, MOROCCO

Silurian outcrops of the High Atlas and Djebilet may be grouped into four geographic areas from southwest to northeast. These are the western High Atlas, central High Atlas (Adrar N'Dgout and Skoura country), eastern Djebilet (and northern slopes of the High Atlas), and eastern High Atlas (Sebbah-Kebir and Tamlelt).

WESTERN HIGH ATLAS — The Paleozoic of the High Atlas (Fig. 1, locality 65) has been strongly affected by tectonic disturbance, and the normal sequence of the Silurian has been destroyed. Thus, no satisfactory section is known. According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 264–267), the following parts of the Silurian can be identified: 1) middle to basal upper Llandovery shale (Bäcker collection, but the precise zonation is not clear); 2) possible Wenlock shale and nodular limestone (de Koningk collection); and 3) argillaceous limestone, which ranges from Ludlow (*Neodiversograptus nilssoni-Lobograptus scanicus* Zone) to Pridoli (*Monoclimacis ultimus* Zone) (Schaer collection). The other parts of the Silurian are not necessarily absent, as a continuous section is lacking.

ADRAR N'DGOUT AND SKOURA COUNTRY — The Silurian is practically the same in these two areas (Fig. 1, localities 81, 82). In the Skoura inlier (Taliouine and Affela n'Irhil sections, Fig. 22) (Roch, 1939; F. Duffaud, 1966; J. Destombes, 1966, *in* J. Destombes et al., 1985, p. 297, 307, 308), the vertical succession includes: 1) sandstone and shale with upper Llandovery graptolites; 2) shale, then shale with calcareous nodules and a bed of arenaceous limestone with upper Wenlock graptolites; 3) Ludlow shale; and 4) Pridoli shale with calcareous nodules. Further west on the northwest slope of the Adrar N'Dgout at Tizi n' Tichka, the sequence is similar (Gigoult, 1937; Roch, 1939) but includes lower Llandovery (upper Rhuddanian) (Ouanaimi, 1998). In this same area near Tiwghaza, lower Llandovery (lower Rhuddanian) biotas have been found at the base of the sandy shales (Destombes and Willefert, 1988).

EASTERN DJEBILET AND NORTHERN ATLAS — The Silurian of the eastern Djebilet has long been known (Gentil, 1905). However, the succession has only begun to be understood as a result of structural studies by Huvelin (1967; Fig. 1, locality 80). That is also true for the Silurian of the allochtonous inliers of Aït Mallah and Aït Mdioua (Fig. 1, locality 82). Considerable work on graptolites by S. Willefert (*in* Destombes et al., 1985, p. 296–306) allows recognition of practically all of the Lower Silurian zones, including the oldest, the *Parakidograptus acuminatus* Zone, at Moulay bou Anane in the eastern Djebilet and at Ghogoult in Aït Mallah (Destombes and Willefert, 1988). The Lower Silurian consists of black graptolitic shale with siliceous beds (or "phtanites") near the base. The thickness of the Lower Silurian is probably minimal. The Wenlock is represented only by the lowest and highest zones. Lower Ludlow graptolites have been identified. The sec-

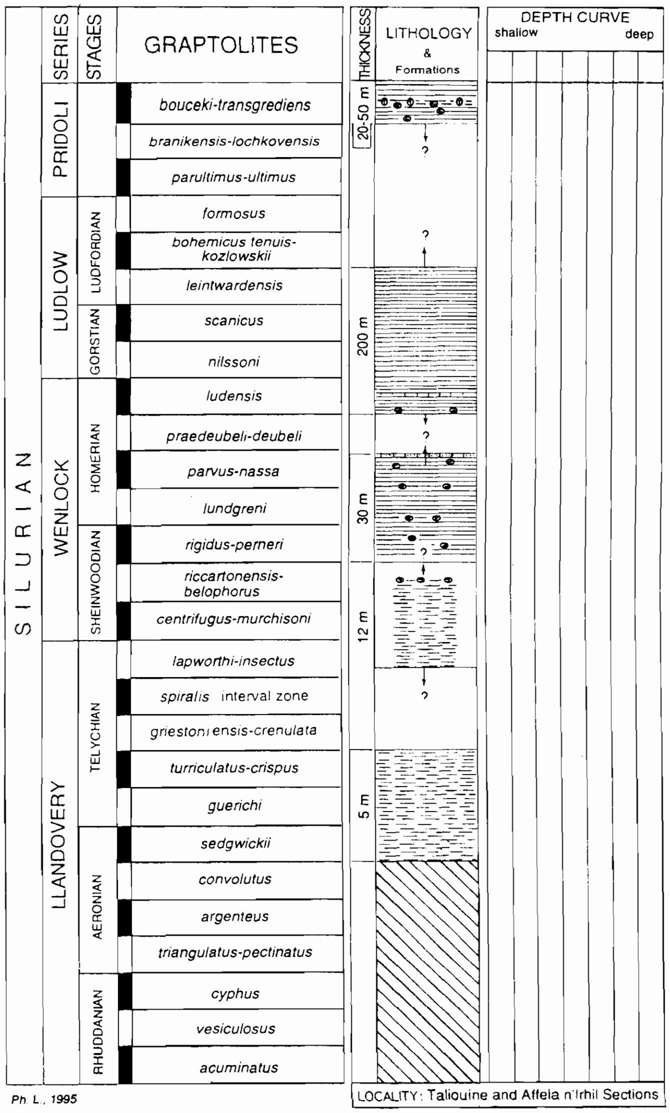


FIGURE 22 — Taliouine and Affela n'Irhil sections, Morocco; lithology. After H. Hollard and S. Willefert (*in* Destombes et al., 1985).

tions show affinities with the Silurian to the south and east in the Anti-Atlas. However, the absence of significant limestone intercalations in the upper Wenlock and Ludlow is notable.

SEBBAB-KEBIR — In this area (Fig. 1, locality 83), the Silurian is represented only by shales. Graptolites allow recognition of the upper Llandovery *("Monograptus" spiralis* Zone), upper Wenlock, and lower Ludlow. The lower and middle Llandovery and Pridoli have not been recognized, but it is not possible to conclude that these intervals correspond to stratigraphic gaps. The thickness of the sequence is unknown (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 308).

TAMLELT — In several small outcrops (Aïn el Orak, Jebel Korima, Jebel Aziza, El Hazma, Ez Zroug, El Atchana, Guelb Renazi; Fig. 1, localities 84–89), the Silurian consists of lower shale, siliceous shale, and graphitic microquartzite (or "phtanite") of middle Llandovery (*Coronograptus gregarius* Chron) to Wenlock age. Higher strata include shale with limestone attributed to the Wenlock and lower Ludlow (Rey, 1911, 1914; Doll, 1913; unillustrated Du Dresnay graptolite collections identified by Godfriaux-Delcroix *in* Destombes et al., 1985, p. 309, 310). The lower Llandovery has not been recognized, but this may not indicate a stratigraphic gap because of the discontinuous outcrop. There is no indication of the thickness of the Silurian.

REGIONAL SYNTHESIS — The discontinuity of the outcrops as a result of Hercynian structural complexity does not allow a complete picture of the Silurian in the High Atlas. The lack of some graptolite zones may only reflect structural complexity, and a complete succession may exist. On the other hand, the lack of calcareous rocks in the eastern High Atlas must be emphasized. However, the High Atlas Silurian differs from the shaly Silurian at Tamlelt, which ranges from the Llandovery to Wenlock or lower Ludlow, and the mainly calcareous Silurian of the Ben Zireg area (see above), which ranges from upper Wenlock to Lochkov with a Llandovery–lower Wenlock gap. These differences lead to paleogeographic questions about the location of these areas in the Silurian.

ESSAOUIRA BASIN

The Essaouira Basin is bounded on the south by the western High Atlas and on the north by the western Djebilet (Fig. 1, locality 66). The Silurian in the Essaouira Basin is known only in bore holes MKL102 and MKL 104. The thickness is about 130–160 m (Lüning et al., 2000a). No biostratigraphic data have been published, and the sequence may be incomplete.

WESTERN MOROCCO

COASTAL MESETA — In the Paleozoic region of Ben Slimane (Fig. 1, locality 67) east of Casablanca, J. Destombes (in Destombes and Jeannette, 1966) described the Silurian in the Aïn Sidi Larbi section (Fig. 23). It is 100 m thick and fossiliferous from the upper Wenlock to the Lochkovian (Willefert, 1966). The age of the lower 25 m of shales remains doubtful, and could be Ordovician. The facies here are similar to those of the southern Anti-Atlas, although the thicknesses are much reduced (H. Hollard and S. Willefert in Destombes et al., 1985, p. 267–269).

DOUKKALA AND THE REHAMNA — Between the west-

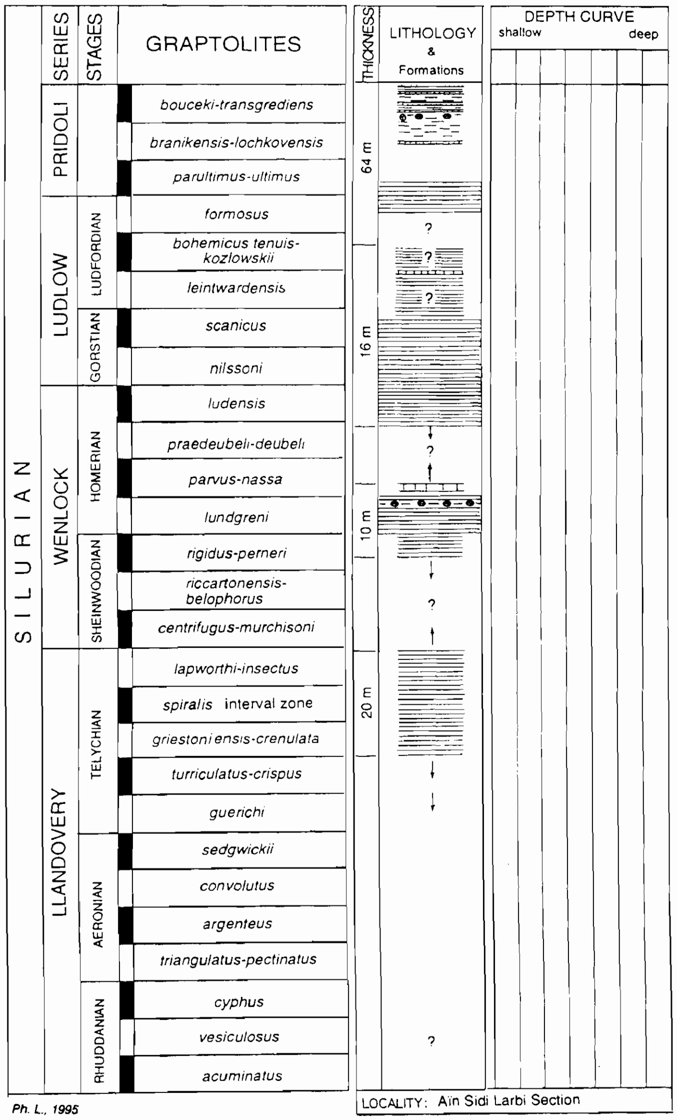


FIGURE 23 — Aïn Sidi Larbi section, Morocco; lithology and formations. Lower 20 m lack fossils and have problematical age. After Destombes (1966, unpublished) and H. Hollard and S. Willefert (*in* Destombes et al., 1985).

ern Djebilet and the Coastal Meseta, the Silurian crops out in the Doukkala and Rehamna (Fig. 1, localities 68, 68). The outcrops are scattered but have yielded interesting data. At Djebel Krarou (Fig. 1, locality 69), 25 m of black shale assigned to the upper Llandovery (Telychian) overlie Upper Ordovician sandstones (Gigout, 1951; Bäcker, 1965, unpublished data *in* Destombes et al., 1985, p. 269, 270). Near Mechra-ben-Abbou (Fig. 1, locality 69), another shaly sequence intercalated between sandstones and limestones with graptolites is Ludlow (lower Ludfordian) (Bäcker, 1965, unpublished data *in* Destombes et al., 1985, p. 270). In addition, a shale–sandstone succession with *Scyphocrinites* limestone lenses is probably uppermost Silurian. Together, these latter two successions may be more than 200 m thick. About 60 km to the northwest are the Silurian outcrops of the Oulad-Abbou syncline (Gigout, 1951, 1954; Fig. 1, locality 70). The Silurian here is about 100 m thick, but only the Ludlow has graptolites. According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 270), the Pridoli seems to be absent, and the Lochkov might be transgressive on the Ludlow. In the Doukkala Basin, several bore holes, such as DOT1, have cut the Silurian, but nothing is known about the stratigraphic sequence. Between Oulad-Abbou and Mechra-ben-Abbou, the interesting Oulad-Saïd boring 1 (OS 1) shows an almost complete Silurian succession (Fig. 24). The Silurian is thin because the upper Wenlock is transgressive on the lower Llandovery and because of the slow rate of sediment accumulation during the Ludlow (H. Hollard and S. Willefert *in* Destombes et al. 1985, p. 271, 272).

RABAT AREA — The Silurian of the Oued Bou-Regreg Valley (Fig. 1, locality 71; Fig. 25) has been studied by many workers (Savornin, 1921; Lecointre, 1926; Cogney, 1957; Rousselle, 1961; Alberti, 1969, 1970). According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 272–276), the base of the section is not clear because the underlying "Ordovician" quartzitic sandstones are not dated, because the contact might be tectonic, and because the overlying 17 m of shales lack fossils. However, subsequent study showed that the contact between the Ordovician(?) and Silurian is an erosional surface overlain by a microconglomerate, and that the first graptolites appear 8 m above the microconglomerate (El Hassani, 1991). Only the Ludlow and the Pridoli are dated by graptolites. The trilobites of these beds confirm the strong affinities of the Ludlow of Morocco with the Budnanian of Bohemia (Alberti, 1969, 1970).

Some 30 km to the east of the Bou Regreg sections, the Silurian crops out at El Khaloua and Oued Satour. S. Willefert (*in* Destombes et al., 1985, p. 276, 277) demonstrated that the graptolites are Pridoli and that the proposed "Tarannon" and Ludlow of this region (Lecointre, 1931, 1933) were the result of graptolite misidentifications.

REGIONAL SYNTHESIS — In western Morocco, two types of adjacent Silurian successions are observed. The first is nearly complete, with possible minor gaps or condensation, and features sandy intervals in a mainly shaly or calcareous shaly sequence. The second type of succession begins in the upper Wenlock, is comprised of argillaceous limestone and shales, and resembles the Silurian in the Ben Zireg area or in the north Saharan basins (see above).

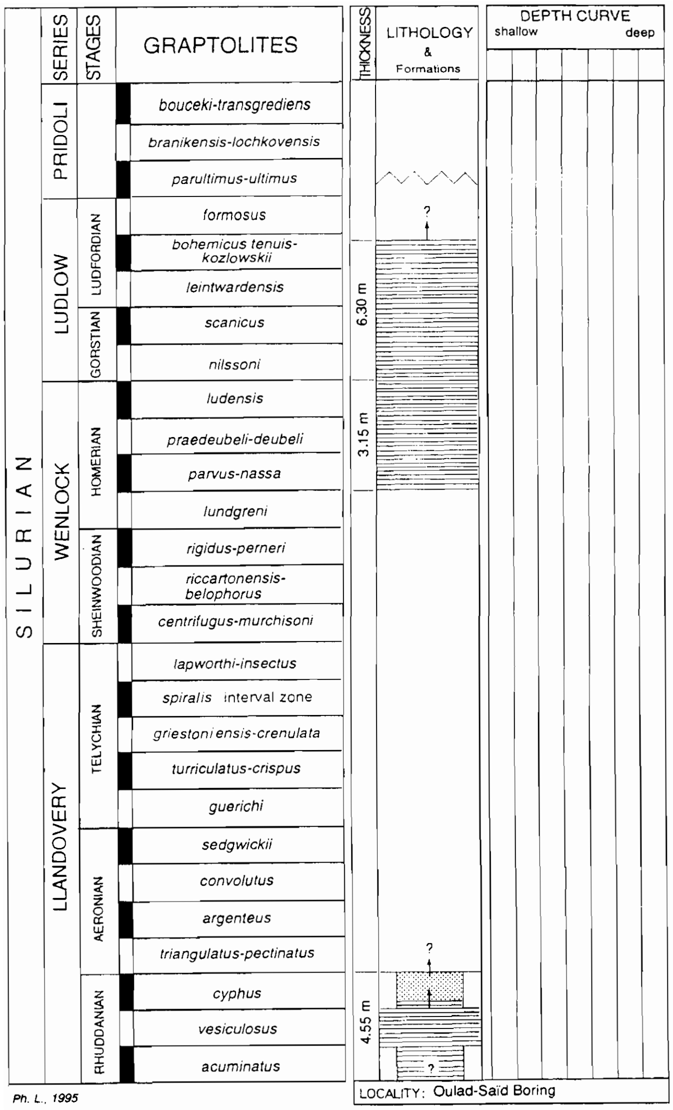


FIGURE 24 — Oulad-Said core, Morocco; lithology. After H. Hollard and S. Willefert (*in* Destombes et al., 1985).

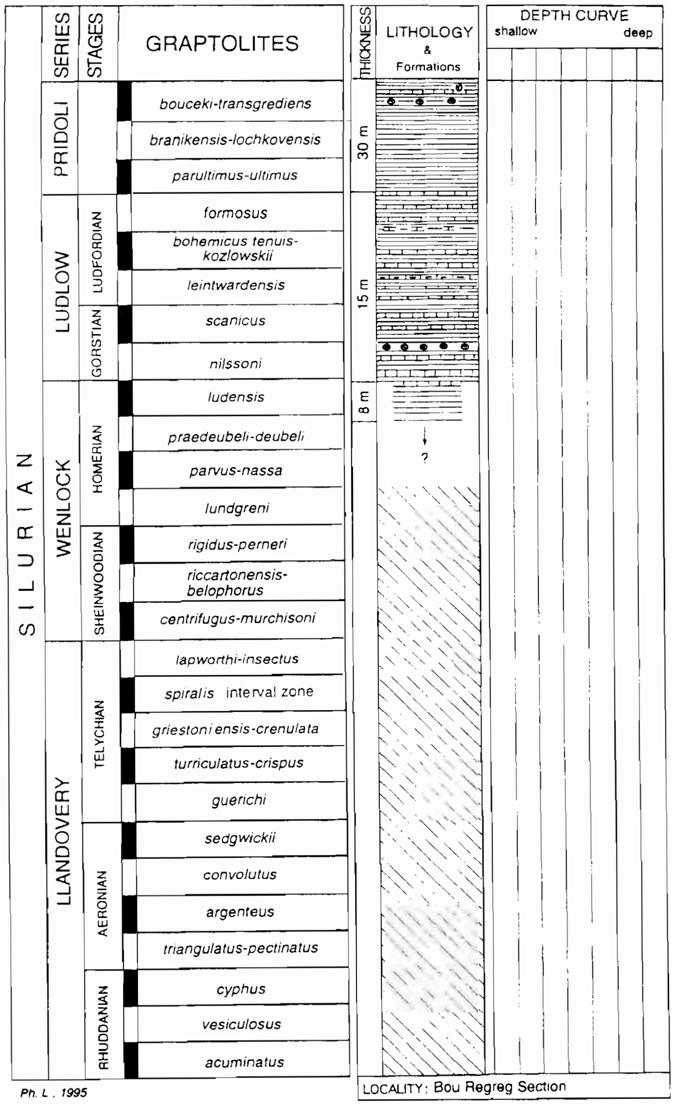


FIGURE 25 — Bou Regreg section, Morocco; lithology. Lower 8.0 m above Ordovician(?) lack fossils for correlation. After H. Hollard and S. Willefert in Destombes et al., 1985) and El Hassani (1991).

CENTRAL HERCYNIAN MASSIF, MOROCCO

This region includes two very significant Silurian areas in Morocco. These include the Kouribga-Oulmes anticlinorium and the Kasbah Tadla–Azrou "anticlinorium." The Tazzeka area, just to the south of the pre-Rifan frontal nappes, will also be considered herein. Reconstructing the Silurian stratigraphy in the central Hercynian massif is difficult. This is particularly true for the eastern district, where thrust slices have been demonstrated (Huvelin, 1970, 1973).

KOURIBGA–OULMES ANTICLINORIUM — According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 277–285), the succession of the Kouribga–Oulmes anticlinorium (Fig. 1, locality 72) incudes lower, massive quartzite (average 15 m thick). There is a gradual upward passage from these quartzites to Telychian shale (G. Suter, 1957, unpublished data, *in* Destombes et al., 1985, p. 279). These quartzites might be Silurian transgressive deposits (Termier, 1936). The immediately overlying shales (80–100 m thick fide Termier, 1936; 20–30 m fide G. Suter, 1957, unpublished data, *in* Destombes et al., 1985, p. 279) have graptolites. These graptolites (G. L. Elles *in* Termier, 1936; G. Waterlot *in* Van Leckwijck et al., 1955; J.

Godfriaux-Delcroix, 1957–1958, unpublished data, *in* Destombes et al., 1985, p. 279) are middle or upper Llandovery (lowest Telychian). These shales are locally upper Telychian (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 283). These latter shales are succeeded by a shale–limestone sequence (50 m), in which the shales become red or yellow upward and have calcareous and local sandstone beds. Graptolites (determinations by G. L. Elles, G. Waterlot, and J. Godfriaux-Delcroix) from the base of this latter unit are upper Wenlock–lower Ludlow. The overlying higher Silurian unit is the Sidi-M'Bellej Shale. The Upper Silurian and the transition into the Devonian are poorly known. Shales with limestones containing *Scyphocrinites* sp. are assigned to the Pridoli, but the upper Ludlow and the Ludlow–Pridoli passage have not been observed.

KASBAH TADLA–AZROU "ANTICLINORIUM" — The Silurian is poorly known in this eastern part of the central Hercynian massif (Fig. 1, locality 73). The basal quartzites and the "El Krad Flysch," once considered Lower Silurian, are now placed in the Ordovician (Allary et al., 1972a, 1972b; Huvelin, unpublished data *in* Destombes et al., 1985, p. 285, 287). There is no continuous section, but a composite section has been constructed. The Silurian at Ito (Fig. 1, locality 74) has a lower siliceous facies ("phtanites") known as the "flaggy Mokattam facies" in the Kasbah Tadla–Azrou "anticlinorium" (Agard et al., 1958). The thickness of the Mokattam Formation varies from 15–25 m. Willefert (1963b) and H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 285, 286, 288–290) described the Mokattam Formation in the Eguer–Iguiguena section near Ito. It includes lower siliceous shales (8–10 m) of early Llandovery (late Rhuddanian) to middle Llandovery age (early Aeronian). Overlying argillaceous shales include 3–4 m of middle Llandovery (middle Aeronian), 2.5 m of upper Llandovery (upper Aeronian), and 4+ m of upper Llandovery (lower "Tarannon")–lower Telychian (upper "Tarannon") at the top. These latter shales mark the beginning of an argillaceous sequence with evidence of a late Telychian age.

In the Silurian at Touchchent (Fig. 1, locality 75), siliceous shales that alternate with mud shales (lower Telychian) are known. Higher black shales have Telychian and middle Wenlock graptolites (Willefert, 1963). Black shales with later Wenlock faunas were reported by Morin (1957). Elsewhere, the upper horizons consist of limestones with *Scyphocrinites* and are probably Pridoli (H. Hollard and S. Willefert *in* Destombes, 1985, p. 286, 287, 291).

The Silurian of the Azrou region (Fig. 1, locality 76) is similar to that at Ito. According to M. Diouri (*in* Destombes et al., 1985, p. 287, 290, 292), the Mokattam Formation begins in the *Parakidograptus acuminatus* Zone and spans the Llandovery (Rhuddanian and Aeronian) in a 30 m interval. In other sections, black graptolitic shales are upper Telychian, lower and upper Wenlock (10–20 m), and lower Ludlow (20–30 m). Shales with *Scyphocrinites* limestone cap the Silurian.

SOUTHWEST NOSE OF KASBAH TADLA-AZROU "ANTICLINORIUM" — According to H. Hollard and S. Willefert (*in* Destombes et al., 1985, p. 292), the Silurian in this region (Fig. 1, locality 77) has been encountered in drill holes. The facies are similar to those in surface outcrops. The *Parakidograptus acuminatus* Zone forms the base of the Silurian. The thickness of the Llandovery–uppermost Wenlock is less than 120 m, and the total thickness of the Silurian seems to be less than 200 m.

TAZZEKA — The Silurian crops out near the northern extremity of the Middle Atlas (Termier, 1936; Van Leckwijck and Termier, 1938; J. Destombes and M. Abbès, unpublished data *in* Destombes et al., 1985, p. 292–294). In the Souk-et-Tleta and Souk-el-Khemis section of the Tazzeka area (Fig. 1, locality 78), the Silurian is complete but thin (less than 100 m). The probable absence of limestones and the complete absence of sandstones and quartzites should be noted (Fig. 26).

REGIONAL SYNTHESIS — Available data suggest that there are no fundamental differences between the Silurian successions south and northwest of the South Atlas Border fault in western Morocco. However, the north Atlas facies change to the east and north with the shales often replaced by siliceous beds ("phtanites") rich in radiolarians (as at the Tifrit horst in northwest Algeria) and with the disappreance of all carbonate units.

GHAR ROUBAN MOUTAINS AND TIFRIT RISE

The Oujda area, the Ghar Rouban Mountains, and the Tifrit Rise are highly tectonized regions (Fig. 1, localities 91, 91, 94). Rare outcrops of shale and siliceous beds ("phtanites") contain middle Llandovery, upper Llandovery, and upper Wenlock graptolites (Lucas, 1938, 1942, 1948; Owodenko et al., 1938; Waterlot, 1941b).

RIF AREA, MOROCCO

The Paleozoic is known only from the inner structural belts of the Rif in the Tetouan area (Fig. 1, localities 92, 93). The Silurian in this area is composed of upper Llandovery (upper Aeronian and Telychian) muddy and siliceous shales ("phtanites") (Kornprobst, 1974; H. Hollard and S. Willefert in Destombes et al., 1985, p. 311, 312).

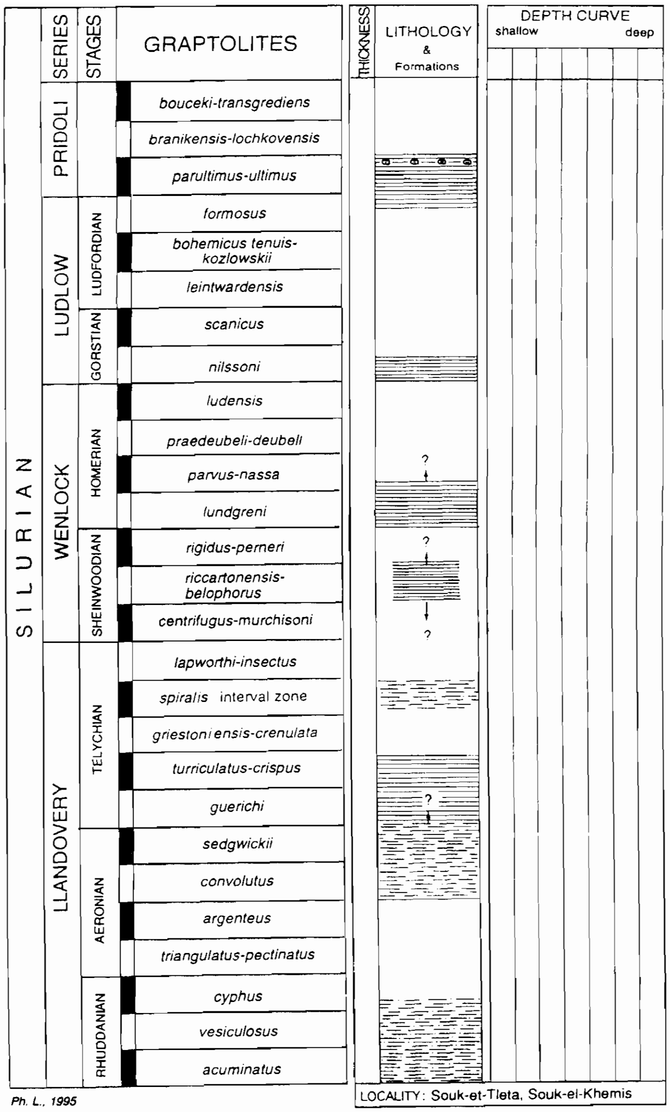


FIGURE 26 — Souk-et-Tleta and Souk-el-Khemis sections, Morocco; lithology. Thicknesses are difficult to measure. After H. Hollard and S. Willefert (in Destombes et al., 1985).

Northern Algeria

The Silurian is known in Beni Affeur (Petite Kabylie in Fig. 1, locality 97; Ehrmann, 1922; Durand-Delga, 1952, 1955). In 1974, I suggested that the stratigrapic sequence requires revision (e.g., Legrand, 1985b). Since then, Alberti (1980) showed that the sequence was tectonized. Moreover, the presence of Llandovery graptolites now seems doubtful. On the other hand, the Ludlow and Pridoli ages of the limestones with trilobites is well established (Alberti, 1980). In the "Grande Kabylie" (Fig. 1, locality 96) at Ihamziène (Barbier et al., 1948), a Bohemian-like facies with platy, silty limestone, perhaps with *"Scyphocrinites"* and with doubtful graptolite fragments is Pridoli or Lochkovian (Gelard et al., 1978; Legrand, 1985b). Silurian siliceous shale ("phtanite") also occurs in the Traras massif (Guardia, 1967; Fig. 1, locality 95). One level is Telychian, and the other is lower Ludlow.

Depositional Environments and Lithostratigraphy

Regional siliciclastic mudstone deposition across North Africa poses problems related to the thickness of these deposits, the large amount of organic matter, and the presumed anoxic to strongly dysaerobic conditions. There is no convenient recent model for it (Legrand, 1994, 1999). Mineralogical and chemical analyses are not particularly useful in explaining this facies. However, the geochemical aspect of Silurian shales, trace elements, and radioactive elements have been the subject of a few studies (Stevaux and Kulbicki, 1967; Pelet et Tissot, 1970; Hassan and Massa, 1975; Hassan, 1976; Combaz, 1986). Organic matter content and maturation have been extensively studied, but the conclusions have rarely been published (Lagoun-Defarge, 1989; Drid, 1989; Meister et al., 1991; Chaouche 1992). New approaches (e.g., Lüning et al., 2000a) seem debatable.

On the other hand, the analysis of the sandy sedimentation and reconstruction of the paleoenvironments pose fewer problems if one takes into consideration only the broad syntheses allowed by this approach (Beuf et al., 1971). Detailed studies are more time-consuming because of the large areas and the considerable thicknesses involved, and are consequently few and limited. The common lack of biostratigraphic control also limits the significance of regional syntheses. Alternative interpretations of whether sand deposition intervals were related to erosional events associated with sea-level, isostasy, or tectonic controls and their extent and timing are possible. Facies changes may also be diachronous. Considerable volumes of sediment (mainly quartz sand) were involved, and their provenance from the east or south needs further consideration.

Other questions involve the iron oolites, which are plentiful in the Ludlow and Upper Silurian (Chauvel and Massa, 1981), and the siliceous shales and the graphitic microquartzites ("phtanites") that are common north of the South Atlas Border Fault. Waterlot (1941b) has drawn attention to these "phtanites," but their significance is still a subject of discussion (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 313).

Finally, the onset of carbonate deposition (mostly fossil hash-rich), chiefly during the Wenlock, Ludlow, and Pridoli, suggests warming episodes. However, this seems true only in part. Calcareous nodules are known at the Ordovician–Silurian boundary in the eastern Tassili N'Ajjer, which was near the South Pole at this time. Carbonate sedimentation also depends on local geochemical disequilibrium events, but no detailed studies are available on this matter (Legrand, 1994). The *Scyphocrinites* limestones constitute a special case, and problems remain about how this widespread and uniform facies formed.

Patterns in Lateral and Vertical Community Biofacies Change

There is no evidence of biological community successions, mainly because of the scarcity of benthic faunas. Moreover, studies are not enough advanced in most regions to tackle this type of problem. It is possible, however, to characterize the typical associations between lithologic facies and fossil groups. For example, sandstones often have land plants; microconglomerates with phosphatic pebbles have inarticulate brachiopod fragments; siltstones often have eurypterids, and green shales often have molluscs. Other types of sandstones have homalonotinid trilobites and rare brachiopods. Some limestones have orthoceratids, bivalves, and *Scyphocrinites*. The siliceous shales ("phtanites") feature radiolarians. These associations do not allow calculation of precise depths, but trends can be outlined.

Climatic Indicators and Variations

In the southern part of the study area, there are early Llandovery indications of possibly persistent peri-glacial conditions and local glaciers. In the Illizi Basin, a middle–upper Llandovery graptolite fauna has been found in the upper member of the Gara Louki Formation, a typical glacial deposit (B.R.P. et al., 1964, bore hole Daïa). However, similar occurrences must be found to corroborate this persistence of glaciation. Semtner and Klitzsch (1994) thought that there is "evidence for a second phase of glacial advance in Early Silurian times" after observations in the Ennedi (Chad) and the Djebel Taguru in Sudan (Fig. 1), but their evidence is debatable (Legrand, 1995).

Carbonates, locally observed near the Ordovician–Silurian boundary, are more abundant in the Wenlock, Ludlow, and upper Pridoli. This feature has been interpreted to reflect either an episode of global warming or a northward paleogeographic migration of the region. The iron oolites could test these hypotheses, which remain tentative.

Paleogeography

The paleogeographic reconstruction for such a large area is difficult. In the case of widely separated outcrop and bore hole observations, the proposed correlations may not be supported by actual data. Of course, in some basins, the subsurface information is rare or absent; this is true for the Taoudeni Basin, for example, while in other basins precise informations remains unpublished (e.g., Ghadamès and Berkhine Basins). The Silurian has been locally deeply buried (more than 10 km in the Saharan Atlas just north of the South Atlas Border Fault), or uplifted and eroded recently. Such a situation is believed to exist in the Ahaggar, which sometimes is wrongly considered to have been a barrier during the Paleozoic. This area was more probably a basin, at least during the Llandovery and Wenlock (Legrand, 1970, 1989, 1995a, 1999). Further considerations include such problems as the fact that precise dating is not possible for the continental series, while very thin deposits that correspond to a biostratigraphic zone may remain unnoticed, especially in shoreline areas where they may have been destroyed during subsequent transgression. Finally, Silurian paleogeographic maps involving the whole of North African Gondwana have never been compiled. Regional maps are rare. In Morocco, apart from early reconstructions (Termier, 1936; Roch, 1950; Choubert, 1952), only two small but valuable maps have been published (H. Hollard and S. Willefert *in* Destombes et al., 1985, p. 318, 319). In Libya, the essays by Klitzsch (1970, 1981) and Klitzsch and Semtner (1993) are limited to shoreline reconstruction. However, recent maps have been published for the Ghadamès and western Sirte Basins (Belhaj, 1996). In Algeria, I published a series of Silurian facies maps on a stage-by-stage scale (Legrand, 1981, 1985b).

With subsequent improvements in understanding regional geologic evolution, for instance during the Llandovery in the Algerian Sahara, it appears that maps that involve data from one or two graptolite zones must be prepared (Legrand, 1989) in order to avoid a mix of conflicting data. This requires a large amount of consistent biostratigraphic information to set up a framework by which sedimentary and paleogeographical data can be integrated. With the present state of knowledge, it is not possible to produce a map for every graptolite zone. I shall, however, present nine maps, of which only five meet the requirements set forth above.

Paleogeography of the *Akidograptus ascensus-Parakidograptus acuminatus* Zone and equivalents

(Regional substage g1a1, g1a2–a3) — The early Llandovery (Fig. 27) poses a particular problem. What was the paleogeography at the onset of the Silurian, or more precisely, at the end of the Hirnantian? The continental ice cap began to melt during the Rawtheyan, at least in the Djado, and possibly a little later in such regions as the eastern Tassili N'Ajjer. In the Oued In Djerane region (Fig. 27, locality 13), if the dating is correct, sedimentation was continuous across the Ordovician–Silurian boundary, as in the Hodh (Fig. 27, locality 63). In other areas such as the central Tassili-n-Ajjer or the Ougarta, it is likely that a final regression occurred just before the Silurian. On the map, some points are worthy of note. The graptolite *Parakidograptus acuminatus* is only found in Morocco and north and south of the South Atlas Border Fault. In other areas where deposits of this age are presumed to exist, the species' occurrence is less certain but possible. In a large part of the study area, there are no deposits of this earliest Silurian age, even in the Algerian Sahara where the basal Silurian is well dated. This leads to a proposal that three regions were invaded by the sea at this time: 1) a part of Morocco that includes three areas in the Anti-Atlas and the northern border of the Tindouf Basin; 2) a part of the Ghadamès Basin of Libya and perhaps northern Tripolitania; and 3) a large area of the Murzuk Basin (perhaps extending from the Tibesti or from the Djado) to the Hodh, although it should be remembered that nothing is known of the Silurian in the central Taoudeni Basin.

Paleogeography of the *Cystograptus vesiculosus* and *Coronograptus cyphus* Zones and equivalents (Regional substage g1a4) — *Cystograptus vesiculosus* rarely occurs in North African Gondwana, and *Coronograptus cyphus* has never been positively identified. These upper Rhuddanin rocks cover a larger area than the lowest Silurian, but have different limits (Fig. 28). I think that the shales in the Kufra Basin and on the Libyan–Sudanese border are coeval, but the evidence is weak. Regression in the Oued In Djerane region (Fig. 28, locality 13) took place

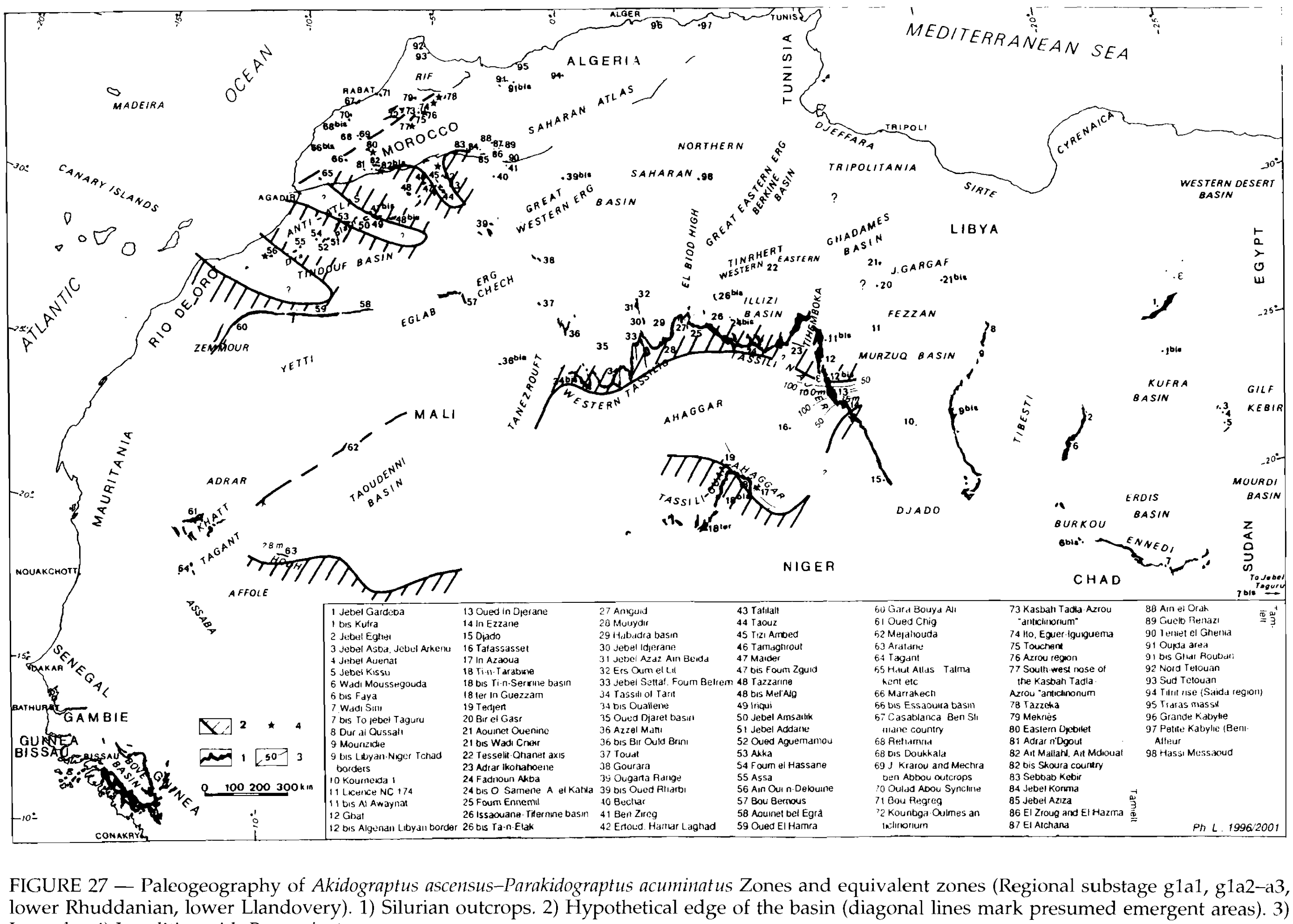


FIGURE 27 — Paleogeography of *Akidograptus ascensus–Parakidograptus acuminatus* Zones and equivalent zones (Regional substage g1a1, g1a2–a3, lower Rhuddanian, lower Llandovery). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark presumed emergent areas). 3) Isopachs. 4) Localities with *P. acuminatus*.

during transgression in the Algeria–Libya border region to the north (Fig. 28, locality 12). There is evidence for this invasion in the northern and southern Tassilis, which demonstrates the existence of the Ahaggar Basin in the Silurian. There is an indication of the *C. vesiculosus–C. cyphus* Zones in the southern Tindouf Basin, and they are presumed to have been laid down in the Ougarta range (Fig. 28, p. 39). The extent of the Silurian in the Moroccan Anti-Atlas differs from the preceding map (Fig. 27). This reflects a new transgression, which was accompanied by the influx of such foreign species as *Atavograptus atavus*. In my opinion, it is reasonable to suppose that a communication existed between the Ahaggar Basin and the Hodh.

PALEOGEOGRAPHY OF THE *CORONOGRAPTUS GREGARIUS* AND *"MONOGRAPTUS" CONVOLUTUS* ZONES AND EQUIVALENTS (REGIONAL SUBSTAGE G1B1–G1B2) — In the middle Llandovery (early Aeronian), the sea continued to transgress the central Tassili N'Ajjer (Fig. 29, locality 24), while it withdrew from the eastern Tassili N'Ajjer. Shoreline to deltaic facies were laid down in the east. The transgression reached the southern Illizi Basin and was less extensive in the western Tassili N'Ajjer. Marine deposits of this age have been recorded in the Ougarta range (Fig. 29, locality 39). Because of the phyletic relationship of *Normalograptus (N.?) libycus* and *N. (N.?) brasiliensis*, the central Tassilis and Brazil have been presumed to be contiguous (Jaeger, 1976; Legrand, 1999), even though these species are unknown in Mauritania and in the Bov Basin. An east-to-west transgression in the Tassili N'Ajjer and western Tassilis was accompanied by an uplift of the eastern border that continued during the *"Monograptus" sedgwickii*, *"M." guerichi*, and *"M." turriculatus* Chrons (Legrand, 1989). In Libya, the sea invaded the Jebel Gargaf area.

PALEOGEOGRAPHYOF THE *MONOGRAPTUS SEDGWICKIII* ZONE AND EQUIVALENTS (REGIONAL STAGE G1C) — Late Llandovery (late Aeronian) sandy sedimentation per-

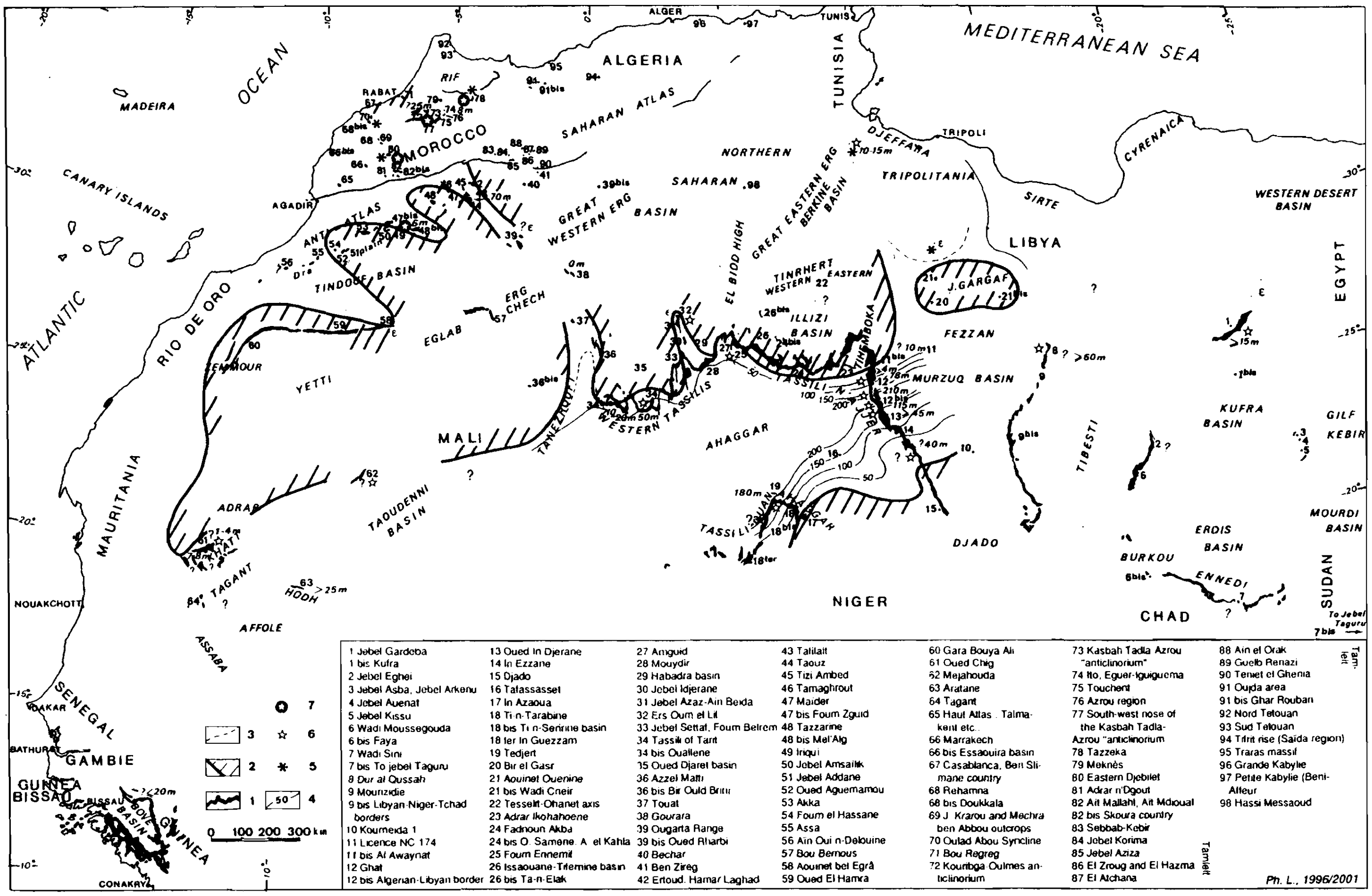


FIGURE 28 — Paleogeography of *Cystograptus vesiculosus* and *Coronograptus cyphus* Zones and equivalents (Regional substage g1a4, middle–upper Rhuddanian, lower Llandovery). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark areas emergent areas) 3) Alternative or very doubtful isopachs. 4) Isopachs. 5) Localities with C. vesiculosus. 6) Localities with *Neodiplograptus fezzanensis*. 7) Localities with C. *cyphus*.

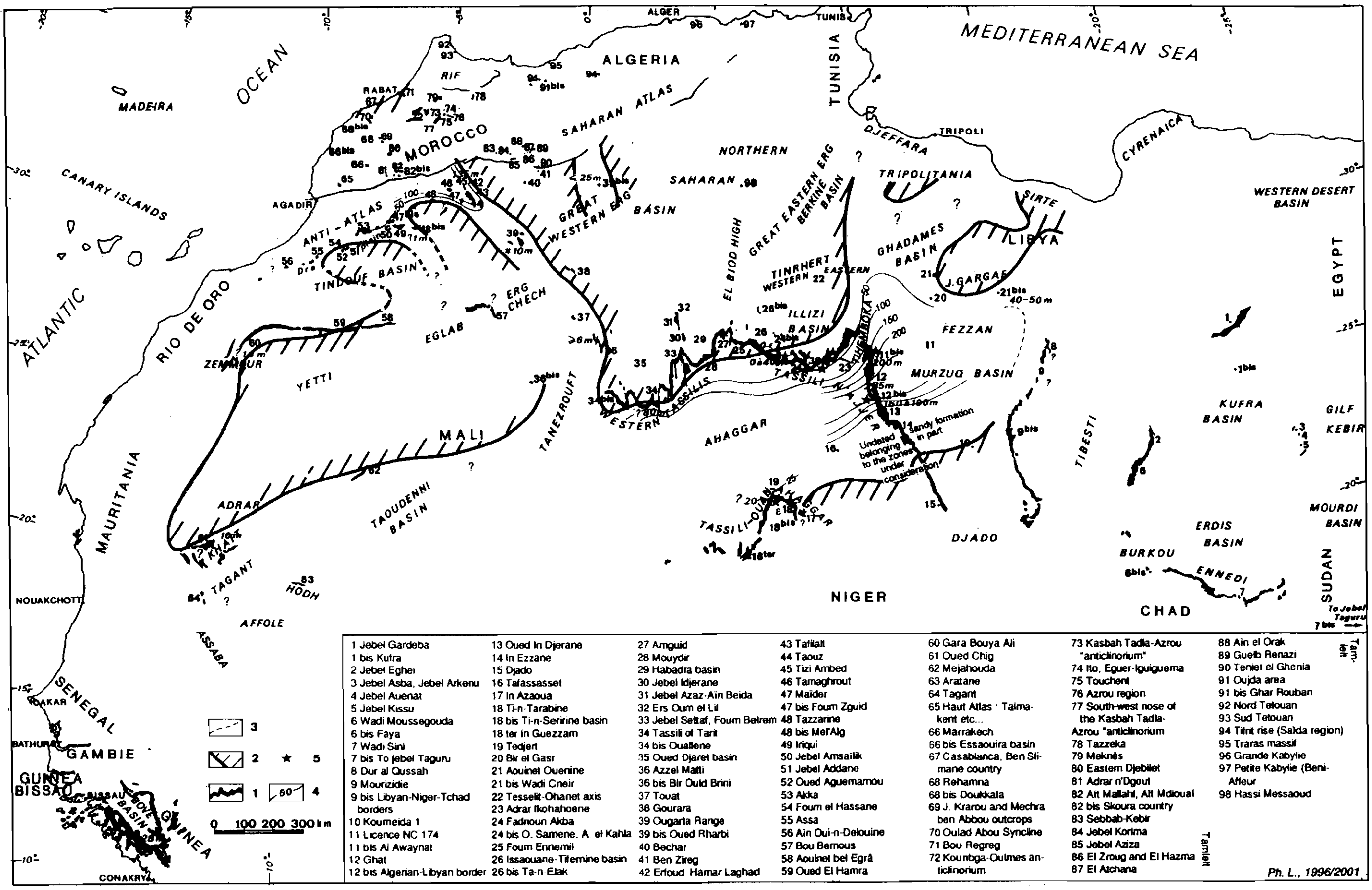


FIGURE 29 — Paleogeography of *Coronograptus gregarius* and *"Monograptus" convolutus* Zones and equivalents (Regional substage g1b1–g1b2, lower Aeronian, middle Llandovery). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark areas emergent areas). 3) Very doubtful isopachs. 4) Isopachs. 5) Localities with *Normalograptus? libycus* and *N.?* sp. aff. *N.? libycus*.

sisted in the eastern Tassili N'Ajjer and, possibly, in a large part of the Murzuk Basin (lower Acacus Formation) (Fig. 30). Similar developments were likely in the Tassili-Ouan–Ahaggar, but no interval higher than the *Coronograptus gregarius* Zone is known in this country below the sub-Devonian unconformity. In the central Tassili N'Ajjer (Fig. 30, locality 24), the muddy sediments were more silty, and the thicknesses are much greater. At the same time in the Illizi Basin, the sea transgressed to the northwest. On the other hand, this interval is only a few meters thick in southern Tripolitania, is not identified (either due to a gap or condensation) in northern Tripolitania, and is doubtfully present in south Tunisia. Further west in the western Tassili N'Ajjer and Amguid area, the sandstones and shales of the *M. sedgwickii* Zone pinch out, and are absent further west (except, perhaps, in Tassili of Tarit). It is likely that the sea covered the northeast Tanezrouft (Fig. 30, localities 36, 37). In the Ougarta range (Fig. 30, locality 39), this zone consists of a few meters of silty shales. In Morocco, this zone almost always seems very thin. It is found in the Oued Chig area, (Mauritanian Adrar; Fig. 30, locality 61) in sandstone, and in Guinea.

Paleogeography of the *"Monograptus" guerichi–"M." griestoniensis* Zones (Regional stage g2a) — It is likely that siliciclastics of this upper Llandovery (Telychian) interval are present in the eastern Tassili N'Ajjer and in the Murzuk Basin, but this is speculative because of the absence of biostratigraphic data (Fig. 31). In the central Tassili N'Ajjer, developments include a short regression, followed by a more open marine episode, and then increasingly near-shore deposition (Fig. 31, locality 24). At the same time, marine shales of the Tin Fouye Formation onlap the uppermost Ordovician further north in the western Tinrhert. In southern Tripolitania, these zones comprise a relatively thin shale succession. In northern Tripolitania and south Tunisia, these zones are not identified (due to either a gap or condensation?). To the west, the sea onlapped the western Tassili N'Ajjer and

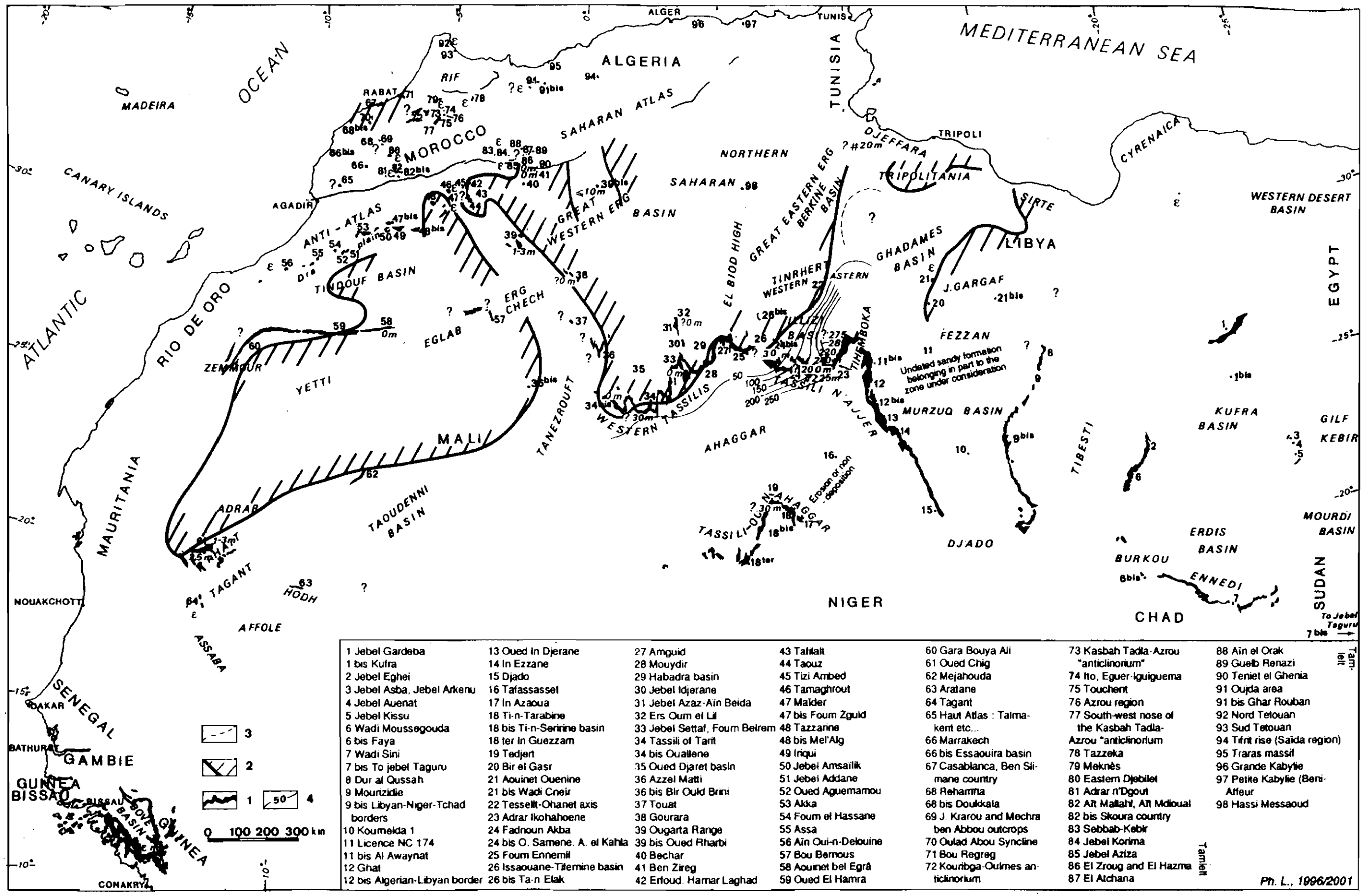


FIGURE 30 — Paleogeography of *"Monograptus" sedgwickii* Zone and equivalents (Regional stage g1c, upper Aeronian, upper Llandovery). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark emergent areas). 3) Very doubtful isopachs. 4) Isopachs.

deposited the thick Foum Ennemil Shale with its Telychian graptolites (Fig. 31, locality 25). These developments are also found in the western Tassilis. This was likely the first time since the terminal early Llandovery that the Jebel Azaz (Fig. 31, locality 31) region was submerged. Similar developments occurred west of and near Ouallene (Fig. 31, locality 34). However, the shoreline did not migrate significantly at the Tassili of Tarit (Fig. 31, locality 34), probably because of recurrent movement on the Foum Belrem fault (Fig. 31, locality 33) (LeGrand, 1970; Beuf et al., 1971). The whole region from the Azzel Matti (Fig. 31, locality 36) to the Ougarta range (Fig. 31, locality 39), as well as the Gourara uplift (Fig. 31, locality 38), was submerged. The shoreline progressed across the northern Tanezrouft and may have reached the eastern Eglab massif. Telychian outcrops are known in Mauritania and Guinea, but are thin. In the Anti-Atlas, the thicknesses are much greater. In the eastern Anti-Atlas (Fig. 31, localities 43, 47), the Telychian transgression was important, and a uniform Telychian shale facies caps the numerous facies changes of the lower and middle Llandovery (H. Hollard and S. Willefert *in* Destombes et al. 1985, p. 259). These Telychian shales thicken west and can reach 200 m. At the north of the Atlasian flexure, the Telychian is known in numerous outcrops in Morocco and Algeria (Traras massif) (Fig. 31, locality 95).

FACIES MAPS OF LOWER, UPPER, AND UPPERMOST WENLOCK — In many areas in the North African margin of Gondwana, Wenlock biostratigraphic data are too rare or incomplete to allow the drafting of paleogeographic maps with isopachs. The absence of guide fossils in the Wenlock is comparable to the lack of guide fossils in the lower Llandovery. For example, *Cyrtograptus* species that are used to define the Wenlock zones are very rare south of the South Atlas fault, except in the Ougarta range, the northern Sahara basins, or southern Tunisia. In addition, uppermost Wenlock graptolites have long been taxonomically confusing, and different definitions of the Wen-

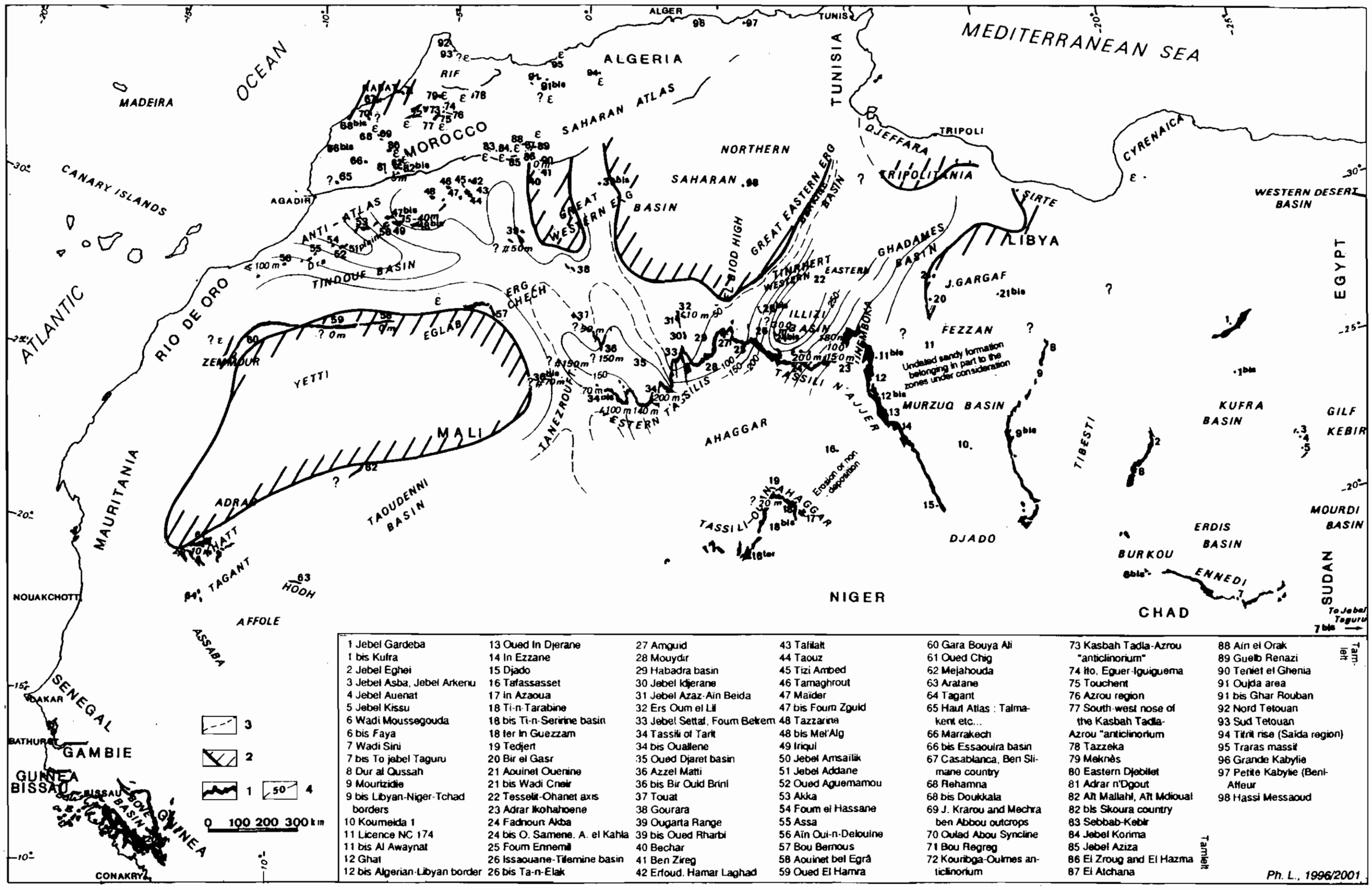


FIGURE 31 — Paleogeography of *"Monograptus" guerichi–"M." griestoniensis* Zones (Regional stage g2a, Telychian, upper Llandovery). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark emergent areas). 3) Very doubtful isopachs. 4) Isopachs.

lock–Ludlow boundary have complicated the problem. On the other hand, as the late Wenlock was a very important time in North African geology, it is important to review the depositional history.

Early Wenlock sedimentation (Fig. 32) often involved continuation of the preceding Telychian sedimentation. Coastal or deltaic sandy sedimentation persisted on the southeastern platform. Silty shales, typical of a low-energy, outer neritic environment, dominated further offshore. Mud shales, sometimes with limestone beds, are observed only in the northwest. Marine conditions seem more evident in the Tinrhert and probably in the Berkhine and Ghadamès Basins. An important event (Saharan tilting) affected the Algerian Sahara between the late and latest Wenlock (Figs. 33, 34). This event led to a transgression in the Ben Zireg section (Jaeger and Massa, 1965) and northern Saharan basins (Legrand 1969, 1981, 1985b, 1994) and to sandstone deposition in the Tassilis. The effects of Saharan tilting are found again in the Ougarta, and even on the south border of the Tindouf Basin. This is a major event that was expressed differently in different regions. Thus in south Tunisia, the uppermost Wenlock is very thick, whereas it is thin in the northern Saharan basins. In the adjacent regions, it is less evident, perhaps because of a lack of data and because these regions were far from the shorelines. However, this event is observed in the Casablanca region of Morocco and locally in Mauritanian Adrar. Any explanation (epeirogenic movement, isostasy, reorganization of intraplate stress fields), requires a precise chronology and, consequently, an improved biostratigraphy.

MAP OF THE PRIDOLI — More work is required for an adequate Ludlow paleogeographic map. A definition of the Ludlow–Pridoli boundary requires abundant graptolites, a condition met only in the northern Saharan basins. This boundary cannot be defined within the sandy littoral facies of the Tassilis and further east. In some bore holes, thin clay beds rich in chitinozoans and acritarchs

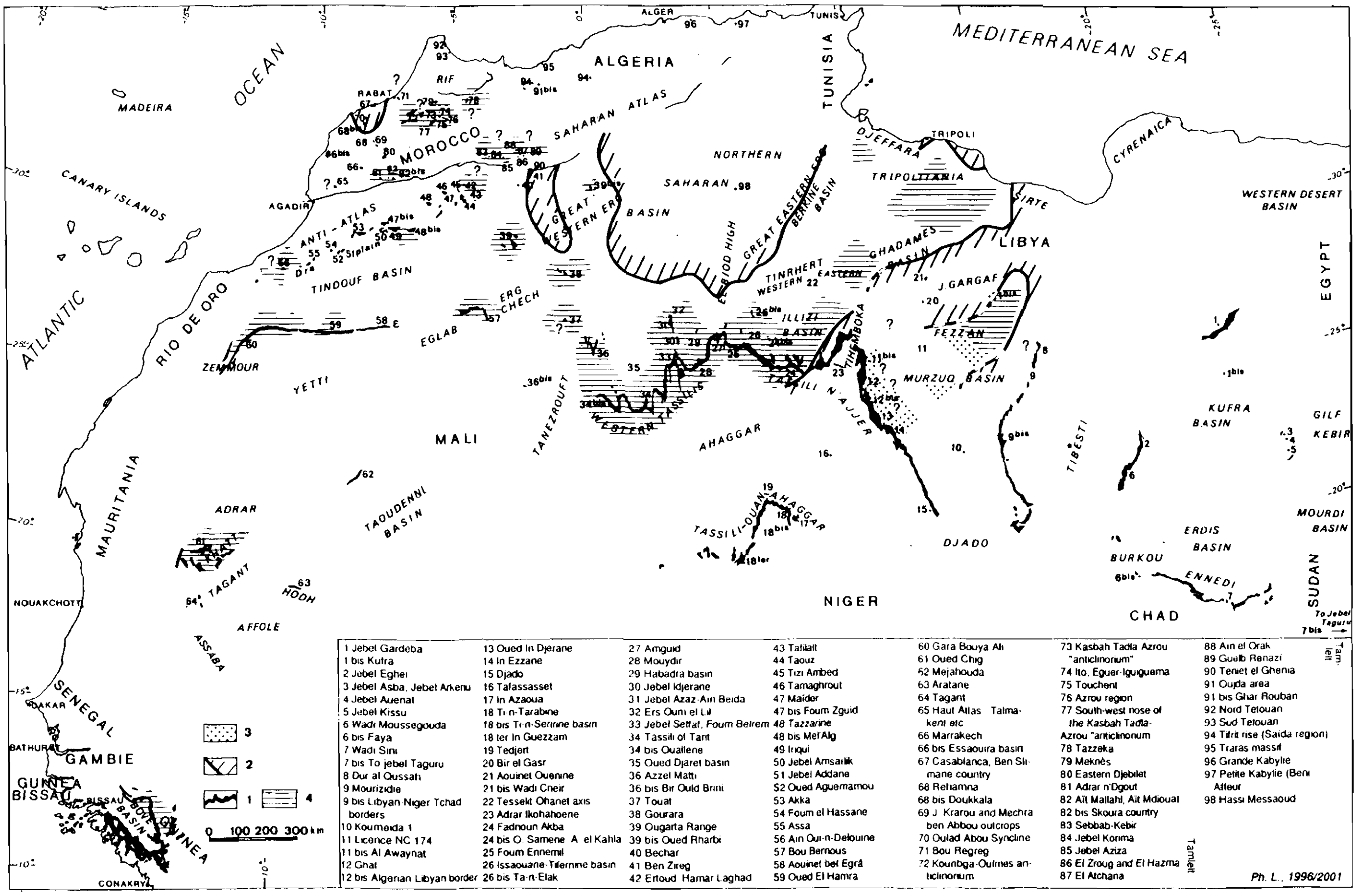


FIGURE 32 — Paleogeography of lower Wenlock. 1) Silurian outcrops. 2) Edge of the upper Wenlock. 3) Arenaceous and argillaceous–arenaceous facies. 4) Argillaceous facies.

can be used to approximate the level of the boundary. A gap at the base of the Pridoli could exist in the Illizi Basin and Tinrhert (Jardine and Yapaudjian, 1968), and erosion of the Lochkov and upper Pridoli below the Devonian unconformity is known. In other areas, the Silurian–Devonian boundary is difficult to determine. As a result, Pridoli thicknesses and isopachs are only tentative. On the Pridoli map (Figure 35), the *"Scyphocrinites"* beds are restricted to northwestern areas (Regnault, 1985), and there was strong subsidence in northern Tripolitania, possibly the Berkhine Basin, and in the western Anti-Atlas. An emergence in the south remains hypothetical.

Conclusions

Because of its extent, the Silurian of North Africa requires more precise and extensive research. Available data suggest a number of conclusions.

Structural activity seems to have played an important role in the Silurian history of the North African margin of Gondwana. In fact, some of the activity that began in the Late Ordovician (middle Caradocian?) seems to be related to the Taconic orogeny. For example, an important North African Gondwanan structural unit is the presumed east–west trough in the Ahaggar and Taoudeni regions. It was probably inherited from older east–west structures (a similar direction to the structural trend of the Reguibat shield), and was likely active before and during the onset of the glaciation. However, the focus on Ordovician glaciation has led workers to neglect the effects of potentially significant epeirogenic movements. During the Silurian, several structural units, such as the Eglab Massif, had an evident role on the Silurian transgression. However, this role seems less evident in the northern Saharan basins.

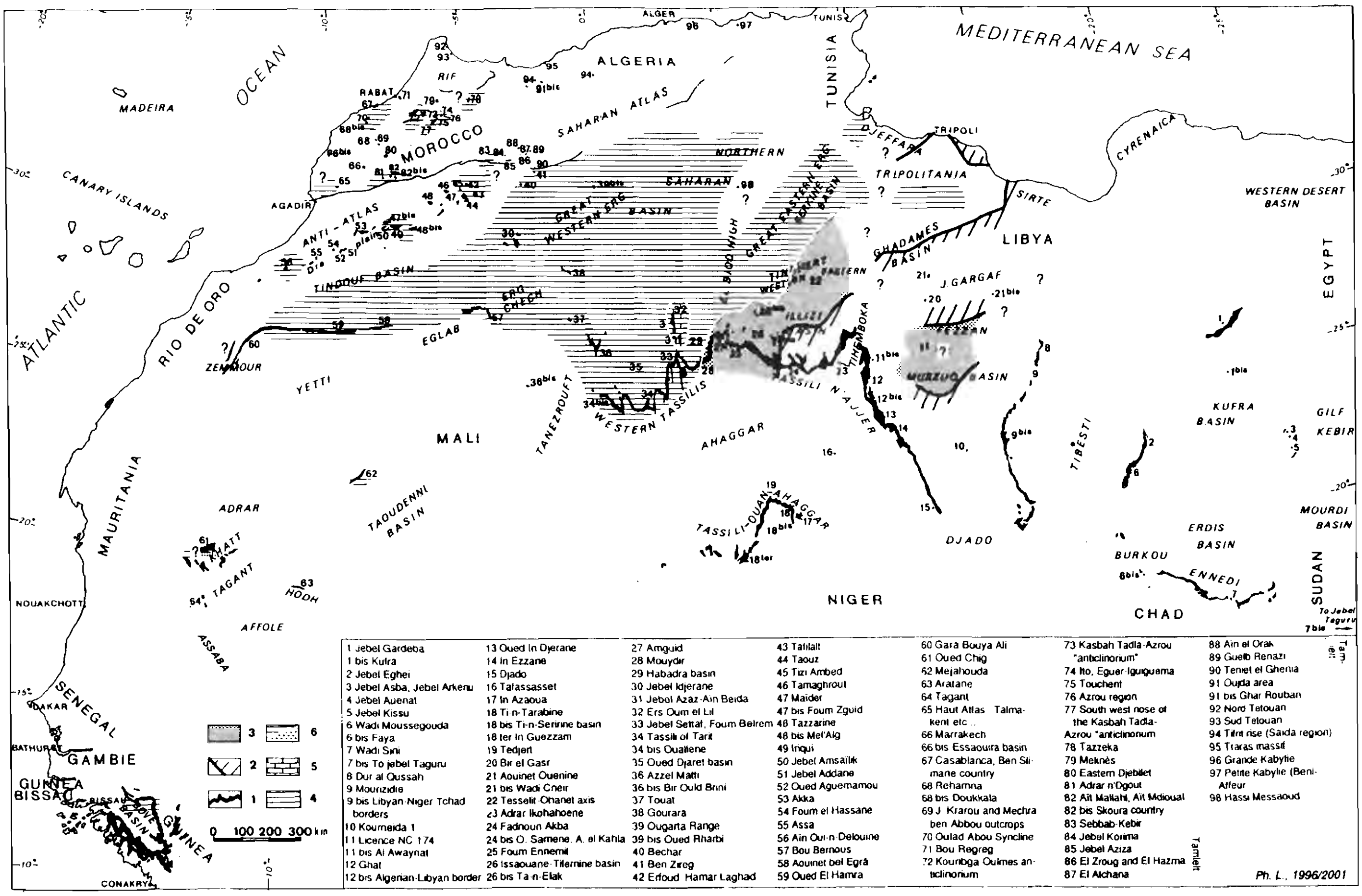


FIGURE 33 — Paleogeography of upper Wenlock. 1) Silurian outcrops. 2) Edge of the lower Wenlock or sub-upper Wenlock unconformity. 3) Arenaceous and argillaceous facies. 4) Argillaceous facies. 5) Argillaceous–calcareous facies. 6) Limit of facies.

Several Silurian paleogeographic belts can be distinguished in the Silurian on the North African margin of Gondwana. Some belts coincide with the structural zones, but others do not. These belts often trend roughly east–west, but their boundaries are blurred. In fact, only the most southern belt is clearly defined. In this belt, Llandovery deposition was controlled by uplift of the Egyptian–Sudanese High, and consisted of increasingly sandy deposits. They are mostly sand-rich and dominant from the late Wenlock on. The Silurian of the southern and eastern Libyan basins, the Tassilis, and probably the Taoudeni Basin and its western end all lie in this belt. The Bov Basin seems to be different, as it has a thick Ludlow and a fauna with Malvinokaffric affinities (Racheboeuf and Villeneuve, 1992). To the north, comparable paleogeographic belts are less clearly defined. This is probably a result of the presence of opposing structural features. Other regions where Llandovery and early Wenlock deposition was minor or absent can be distinguished. In these areas, sedimentation was very low just before or beginning in the late Wenlock, and became important during the Ludlow or Pridoli. This is the case in the Berkhine Basin, Ghadamès Basin, and north Tripolitania. On the other hand, the Ludlow and the Pridoli are thin in most of the northern Saharan basins, including the Bechar area. However, adjacent regions where the lowest Silurian is more or less complete do exist. The Silurian from the Ougarta to the Azzel Matti is one such area, as is the Anti-Atlas, at least in part.

"Late Ordovician glaciation" is often given as an explanation of Ordovician–Silurian boundary events, even though sufficient data may not be available to support this concept. The glaciation was a very complicated event (Legrand, 1995b). The "microconglomeratic clays" probably began to be deposited with glacial still-stand, and the melting of the Late Ordovician inland ice likely

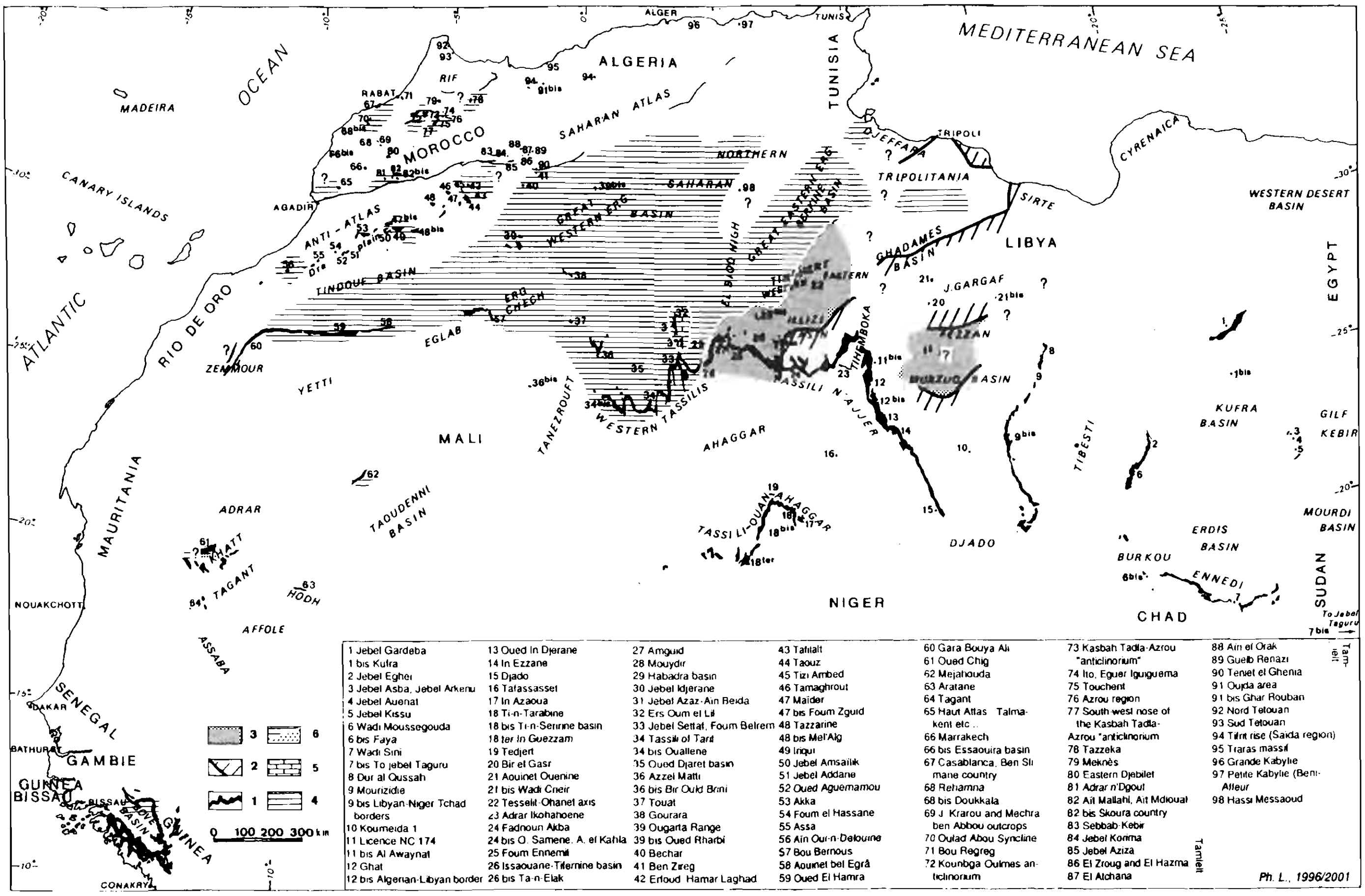


FIGURE 34 — Paleogeography of uppermost Wenlock. 1) Silurian outcrops. 2) Edge of uppermost Wenlock. 3) Arenaceous and argillaceous facies. 4) Argillaceous facies. 5 Argillaceous–silty facies. 6) Argillaceous–calcareous facies. 7) Limit of facies.

increased the rate of deposition of these "microconglomeratic clays." This rock type is found on the North African border of Gondwana, in Arabia, and in some European regions (e.g., Thuringia; Katzung, 1961, and Blumentsgel, 1965). These clays were probably deposited from north to south with transgression, and sea level was linked to deglaciation. The return of Hirnantian faunas after this glacio-marine event, and the scarcity of glacial quartz grains in uppermost Ordovician and Lower Silurian shales, demonstrate that sediment input directly from the melting of continental ice was minor. Indeed, the remnant continental ice cap was distant (perhaps in South America), and played only a secondary role in deposition in North Africa.

Another paradox is related to the isopachs and facies distribution during the Llandovery. The position of the South Pole at the end of the Ordovician is hypothetical, and has been suggested to lie in areas ranging from the Gulf of Guinea to the Touggourt (Torsvik et al., 1996). This imprecision is aggravated by the uncertainty of the ages of assumed polar positions and the frequent confusion between the different types of poles (magnetic vs. geographic), which are now separated by 1200 km. Consequently, it is foolish to pretend that facies and isopachs recorded zone by zone through the Llandovery somehow directly reflect a single South Pole. Moreover, our reconstructions may be distorted by the absence of the Silurian to the south of the Sahara (i.e, southern parts of Mali, Niger, and Chad; Burkina Faso; Ivory Coast; Benin; Nigeria). Some facts, however, remain unexplained, in particular the differences in facies and rates of subsidence between Mauritania, Libya, and Niger, which are all regions with traces of glaciation. I postulate that regional structural activity was important enough to obscure the rise of sea levels following deglaciation.

Another problem is that the melting of the ice caps was finished (or almost finished) at the end of the Ordovician, and cannot have been responsible for Si-

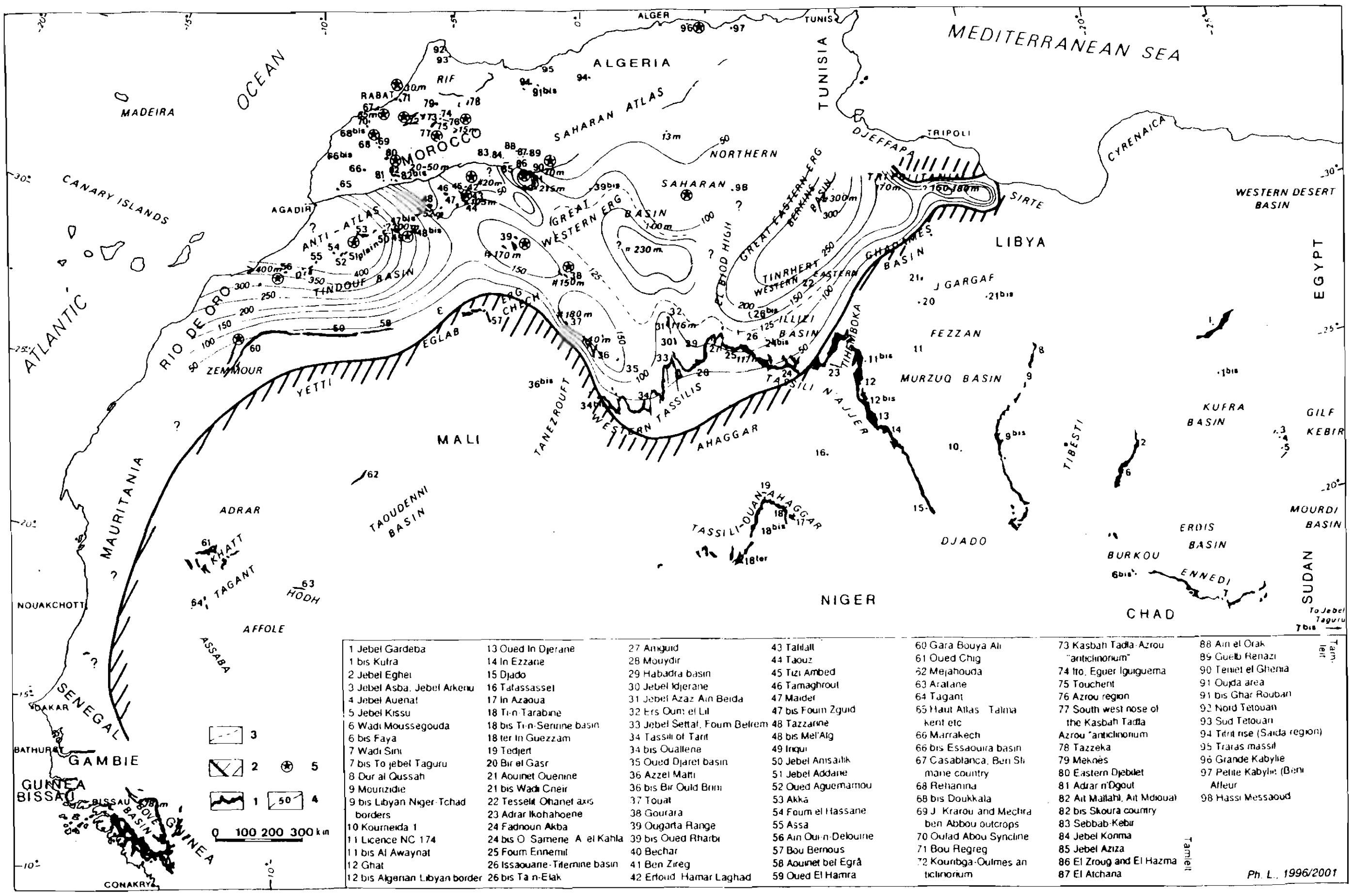


FIGURE 35 — Paleogeography of the Pridoli (Regional stage g3c). 1) Silurian outcrops. 2) Hypothetical edge of the basin (diagonal lines mark emergent areas). 3) Intermediate isopachs. 4) Isopachs. 5) Localities with *"Scyphocrinites."*

lurian transgression on the African border of Gondwana. Some areas probably underwent epeirogenic movements (an uplift of the Egypto–Sudanese High), which were responsible for the gradual transgression of the Silurian sea from east to west in the Tassili N'Ajjer and adjacent basins. The apparent transgression direction is masked by the lack of data on the Ahaggar itself, but a south-to-north movement is likely. It also appears as an uplift of the continent on the east and south, which caused the sea to move northwards and westwards onto other emerging lands. Such a depositional scheme is totally different from the classic models of transgression, because transgressive clays correspond to prograding sandstones. Llandovery history, at least in the southern Sahara, can be summarized as the propagation of a transgressive wave from east to west and from south to north, in association with an uplift of the area invaded by the sea during the preceding stage. This is more difficult to see to the north. It is difficult to have a comprehensive view of what happened in Morocco because of the isolated sections and outcrops and the effects of later structural activity.

Glacial activity in the Llandovery is doubtful in Africa (e.g., Legrand, 1995b), but not in Brazil and Argentina, where several glaciations have been described (Caputo and Crowell, 1985; Grahn and Caputo, 1992; Caputo, 1998). This is worth further consideration, since in some cases, two different events may have become conflated. On one hand is the glacial activity as observed in today's interglacial interval near the poles or in some mountain ranges — something which today would be called "normal." On the other hand, there are periods of global glaciation that included the development of continental ice caps and a cooling of the earth. Can one make some geological sense from such observations? The traces of glaciation observed in the Llandovery are much less important than those observed in the North African uppermost Ordovician. Do they indicate continent-wide or regional glaciation?

The quasi-extinction and the re-radiation of a number of biotic groups, including graptolites, have been related to Late Ordovician glaciation and deglaciation. This contemporaneity of events, at least with deglaciation, appears obvious, but the mechanisms linking them are purely hypothetical. Thus, is it reasonable to correlate glaciation history and the consequent eustatic events to faunal impoverishment and a subsequent diversification? On the North African border, particularly in the Sahara, it seems that fossil distributions suggest that many others factors have intervened, and it is difficult to recognize the dominant one.

A major event occurred in the late Wenlock; the Saharan platform tilted around an east–west trending axis that was limited by north–south- or northwest–southeast-oriented structural features. The resultant facies distribution changed totally, at least in the Algerian Sahara. There was another transgression to the north and a regression in the south, along with the return of sandy sedimentation and some local subsidence. The southern regression probably took place a little later than the northern flooding. The sea was probably more extensive than previously believed (for example, in Morocco near Casablanca), with possible minor variations in the timing of transgression. The cause of this diachronous submergence is not merely regional epeirogeny, but something more important that can be compared with what is observed — for instance, in the upper and uppermost Wenlock of the Bohemian massif. At the same time, a slight global warming, as noted above, seems to have occurred.

Finally, the Pridoli has often been neglected in North African paleogeographic syntheses because of the difficulty recognizing its boundaries. However, its areal extent is surprising, as is the amount of subsidence involved in this epoch. The subsidence is not associated with any important tectonic activity, and the Pridoli Series is locally depositionally continuous with the Lochkovian. However, there is a possible gap at the base of the Pridoli in the eastern Sahara. At the end of the Pridoli and Silurian, the huge sand apron that covers nearly the entire North African margin of the Gondwanan continent must be related to a new Early Devonian epeirogenic phase of regional extent.

One mechanism cannot explain all of the geologic and biotic features. In addition, the current state of knowledge may explain such features as the apparent absence of any rhythmic sedimentation in the Silurian. This feature may reflect the great thicknesses of the Silurian and the earlier descriptive and explanatory approaches to this interval. However, in the entire North African part of Gondwana which was affected by the Late Ordovician glaciation, the structural features and epeirogenic movements appear to have been dominant over the eustatic variations during the Silurian.

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SILURIAN STRATIGRAPHY AND PALEOGEOGRAPHY OF GONDWANAN AND PERUNICAN EUROPE

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ABSTRACT — The modern geographical distribution of north Gondwanan regions with Silurian sediments and faunas shows a rather complex and disorganized picture. This apparent disorganization mainly resulted from tectonism during the Variscan orogeny and, to a minor degree, later modifications by the Alpine orogeny. We infer that the pre-orogenic paleogeography of the north Gondwanan Silurian featured a continuous shelf that can be divided into proximal and distal parts. The Perunica microcontinent (i.e., the present-day Bohemian Massif) was incorporated into the Variscan orogen as terrane between the Gondwana and Baltica continents. The existence of Perunica is supported by paleomagnetic data and by the distribution of Ordovician and Silurian fossil communities.

INTRODUCTION

North African Gondwana was influenced in the Early Silurian by eustatic rise that resulted from the melting of the Gondwana ice cap, and probably also by related deep-water anoxia which was ameliorated slightly during the late Wenlock (Paris and Robardet, 1990). Later development of carbonates, starting in the Wenlock and particularly in the Ludlow and Pridoli, indicates a better ventilation of bottom waters and a climate warming. This area had a Silurian paleoposition of ca. 40°–50° S latitude (J. Kříž, unpublished data, 1996).

Close faunal relationships exist between the Silurian of Morocco, Spain and the Pyrenees, the Mouthoumet Massif, Montagne Noire, the Aquitaine Basin, the Armorican Massif, Sardinia, and the Carnic Alps (Kříž, 1996, 1999; Kříž and Serpagli, 1993). These relationships suggest the location of these areas on the northern margin of Gondwana during the Silurian (Fig. 1). The northern position of the Perunica microcontinent (Havlíček et al., 1994) in the Silurian is reflected by carbonate facies that have very rich brachiopod-dominated communities with other benthic elements (e.g., corals, crinoids, trilobites, and mollusks). Bivalve-dominated communities of Sardinia and the south Armorican Domain, Prague Basin, Carnic Alps, Mouthoumet Massif, and Montagne Noire (Kříž and Paris, 1982; Kříž, 1991, 1996, 1997a, 1997b, in press; Kříž and Serpagli, 1993)) show closer relationships along the northern margin of Gondwana than to communities from the Prague Basin on the Perunica microcontinent. Moreover, bivalve-dominated communities show a generally higher diversity in Perunica than in Gondwana.

SEDIMENTARY ROCKS

The Silurian of North African Gondwana and Perunica is characterized by two types of successions. The first succession, which occurs in most of the Armorican regions (except the south Armorican Domain) and in most of the Hesperian Massif (except the Ossa-Morena Zone), is

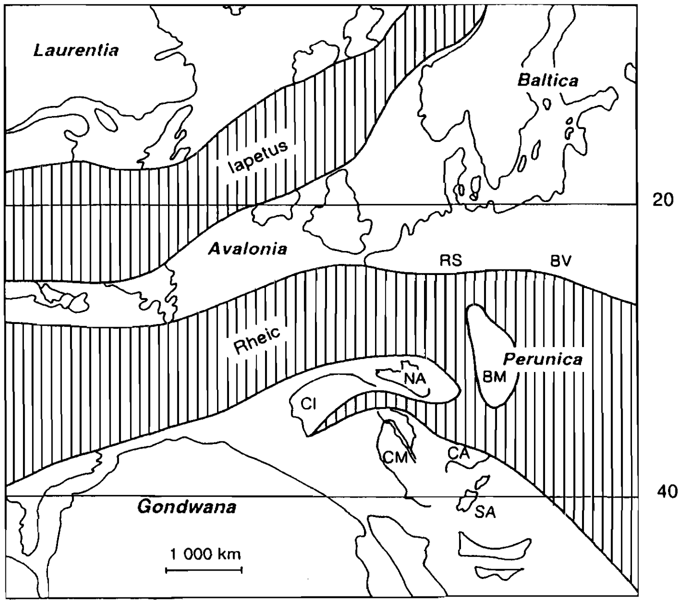

FIGURE 1 — Middle Silurian paleogeography of North African Gondwana, Perunica, Baltica, and Avalonia [based on Cocks and Fortey (1982), Paris and Robardet (1990), and Havlíček et al. (1994)]. Abbreviations: BM, Bohemian Massif; BV, Brunovistulicum; CA, Carnic Alps; CI, Central Iberian Domain; CM, Cantabrian Mountains; NA, North Armoricain Domain; RS, Rheinisches Schiefergebirge (Rhenish Slate Mountains of text); SA, Sardinia.

largely characterized by siliciclastics (see Gutiérrez-Marco et al., 1998). These successions typically begin with sandstones that probably include the Ordovician–Silurian boundary and "lower" Llandovery; continue as graptolitic black shales through the Llandovery, Wenlock, and Ludlow; and end with sandstone–siltstone alternations (Ludlow and Pridoli) that pass into Lower Devonian (Lochkovian) sandstones. The important characteristic of this succession is the predominance of rather coarse-grained siliciclastics that were derived from emergent land areas. This terrigenous influx probably indicates that these depositional areas were on the proximal parts of the North Gondwana shelf (J. Kříž, unpublished data, 1996).

The second type of Silurian succession characterizes all other Gondwanan European regions and reflects low terrigenous influx. It is typically composed of black shales or calcareous shales and limestones (Berry and Boucot, 1967). Such successions apparently developed in more distal, outer-shelf environments. Within this type of succession, it is possible to distinguish three main types of facies development: the "Shelly Fauna," Prague Basin, and Thuringian facies.

The shallow-water, "Shelly Fauna Facies" is often characterized by great thicknesses of calcareous shales with limestone nodules and lenses with a shelly fauna. Especially important features include cephalopod limestones that occur through the Homerian–Gorstian boundary and in the upper Gorstian, lower and upper Ludfordian, and lowest and uppermost Pridoli (Kříž, 1998). This lithofacies was deposited below wave base, but within the reach of surface currents which ventilated the sea bottom. It is characterized by the presence of recurring communities dominated by bivalves (*Cardiola* Community Group, Kříž, in press), and by very abundant cephalopods and other mainly molluscan faunal elements whose larvae were transported by surface currents across the north Gondwana and Perunica basins (Kříž, 1998). This facies is developed, for example, in the Prague Basin (Bohemia) (Kříž, 1991); the Carnic Alps (Austria and Italy) where it is known as the Plöcken facies and Wolayer facies (Schönlaub, 1980); in southwest Sardinia (Italy); in the Montagne Noire (France); in the Central Iberian Zone and the Catalonian Coastal Ranges (Spain), and in Germany (Bavarian facies; Kurze and Tröger, 1990). Correlations of the Shelly Fauna Facies are based on graptolites, conodonts, chitinozoans, bivalves, and cephalopods.

The very shallow-water, shell-rich carbonate "Prague Basin Facies" apparently represented favorable conditions for distal-shelf benthic communities dominated by brachiopods, crinoids, corals, and trilobites (J. Kříž, unpublished data, 1996). This facies developed on the gentle slopes of the Wenlock–Ludlow volcanic archipelago in the Prague Basin (Kříž, 1991). Similarly, this facies appeared on the slopes of a Ludlow volcanic center near Graz, Austria (Fritz and Neubauer, 1988; Neubauer, 1989), and in the middle Llandovery of the Cantabrian Zone (Spain), where the facies developed on basaltic volcanoes (see Truzols and Julivert, 1983). Correlations of the Prague Basin Facies with the Shelly Fauna Facies and Thuringian Facies (described below) can only be made by a combination of conodont and chitinozoan successions. Direct correlation is also possible between the Shelly Fauna and the Prague Basin facies with graptolites, chitinozoans, ostracodes, cephalopods, and bivalves which occur in both facies.

The deep-water Thuringian Facies is represented by relatively thin sequences of Lower Silurian shales with chert (Jeger, 1976, 1977b). The absence of current orientation and the character of the rocks indicate a calm, relatively deep-sea environment (Jaeger, 1976, 1977). This anoxic or dysaerobic environment changed during the later Silurian with deposition of the Ockerkalk (a blue-gray, argillaceous limestone that weathers to ochre and is characterized by thick beds with thin quartzite intercalations). In Germany, these quartzites may be interpreted as distal turbidites. This facies is developed in the Saxothuringian Basin (Germany), Bohemia, Carnic Alps (Bischofalm facies), Ossa-Morena Zone (Spain), and southeast Sardinia (Italy). Correlations of the Thuringian facies are based predominantly on graptolites, and in the Upper Silurian on conodonts (Jaeger, 1976, 1977a). The development of the Thuringian and Shelly Fauna facies suggests that most of the Variscan regions (e.g., Germany, Carnic Alps, Sardinia) were close to the north Gondwana margin and represented an unstable shelf region (Hammann, 1992).

The Thuringian, Shelly Fauna, and Prague Basin facies all developed on the Perunica microcontinent, but the Thuringian Facies developed across the majority of the area. The Shelly Fauna facies is known only from the Prague Basin, and a unique Prague Basin facies is developed in the Prague Basin on volcanic edifices (Kříž, 1991).

Correlations

Silurian graptolitic shale is readily correlated by graptolites and chitinozoans. In the carbonate facies where graptolites are rare or missing, conodonts and chitinozoans may be used for correlations. The wide distribution of the cephalopod limestone lithofacies allows correlation of the Shelly Fauna facies in north Gondwana. Cephalopod limestones form horizons in the upper Wenlock (*Testograptus testis* Zone), lower Ludlow (*Colonograptus colonus* Zone), Gorstian (upper *Saetograptus chimaera* Zone), lower Ludfordian (lower *Saetograptus linearis* Zone), Ludlow–Pridoli boundary interval (*Pristiograptus fragmentalis* and *Pristiograptus ultimus* Zones), and in the Pridoli (upper *Monograptus transgrediens* Zone) (Kříž, 1998).

Deposition of the cephalopod limestones (Gnoli *et al.*, 1980; Kříž, 1991, 1992, 1998, in press; Ferretti and Kříž, 1995) took place below normal wave base, where surface currents reached the bottom. The biofacies originated during early phases of Silurian sea-level rises (Kříž, 1998), when the bottom was affected periodically by laterally shifting surface currents that transported cephalopods, bivalves, and larval stages of other organisms into the basins. Current-oriented cephalopod shells in the Prague Basin reflect southwest–northeast surface currents from late Wenlock to the latest Pridoli (Ferretti and Kříž, 1995). These currents ventilated the originally anoxic or dysaerobic bottom and periodically slowed the rate of sediment accumulation by winnowing out the lime mud. Thick beds of fossiliferous limestones alternate with thin micritic limestones, the latter of which probably represent long periods of quiet sedimentation in contrast to the rapid episodes of higher energy, fossil hash sediment accumulation.

Surface currents were apparently responsible for the distribution of the cephalopod limestones in northern Gondwana (Morocco, Algeria, Montagne Noire, Carnic Alps, Sardinia) and Perunica (Prague Basin) (Kříž, 1984, 1991, 1992, 1996, 1998; Kříž and Serpagli, 1993; Ferretti and Kříž, 1995). The distribution of the Bohemian-type bivalve fauna in Gondwana (Bolivia, Florida, Morocco, Algeria, Guinea, Spain, Montagne Noire, Armorican massif, Carnic Alps, Sardinia, eastern Serbia, Turkey), Perunica (Prague Basin), Baltica (southern Sweden, Poland), the Taimir block, the Tien Shan region, and the Caucasus are also linked to surface currents (Kříž, 1996, 1998; Kříž and Bogolepova, 1995). These surface currents provided a connection with the eastern Australian Yass Basin, where the Bohemian bivalve fauna has its easternmost occurrence (Rainbow Hill section, Black Bog Formation, Booroo Ponds Group; D. L. Strusz, personal commun., 1996). The currents probably derived from the South Tropical Current (Wilde *et al.*, 1991), which reached the northwestern Taimir Basin in eastern Siberia where a Bohemian-type fauna also occurs (Kříž and Bogolepova, 1995). This circulation agrees with the reconstruction of the summer South Subpolar and South Tropical Currents during the Ludlow (Wilde et al., 1991).

Silurian of the Iberian Peninsula

On the Iberian Peninsula (Fig. 2), Silurian rocks occur in the Hesperian Massif (or Iberian Massif) in the southwest Variscan fold belt. These rocks are known from all of the distinct tectonostratigraphic "Zones" defined by Lotze (1945) and subsequently modified (Julivert et al., 1974; Robardet and Gutiérrez-Marco, 1990b) in Iberia, with the exception of the South Portuguese Zone (Fig. 2G). The South Portuguese Zone shows outcrops only of Upper Devonian and Carboniferous. Silurian rocks occur in other Variscan areas that were also affected by the Alpine orogeny, such as in the Pyrenees, Catalonian coastal ranges, the Iberian Cordillera in the northeast, and the Betic Cordillera in the south. During the Silurian, the Iberian Peninsula was part of the northern African margin of Gondwana. Detailed reviews on Silurian stratigraphy and paleogeography of the Iberian Peninsula are available (Gutiérrez-Marco et al., 1998; Robardet et al., 1998; Štorch et al., 1998).

The latitude and climate of Iberia cannot be defined precisely from Silurian lithofacies (almost entirely terrigenous) or faunas (mainly graptolites). However, latitudinal position can be estimated as roughly intermediate between the rather high latitudes of this area during the latest Ordovician (i.e., Hirnantian glacio-marine sediments with dropstones derived from the African ice cap) and the warm temperate to subtropical latitudes of Early

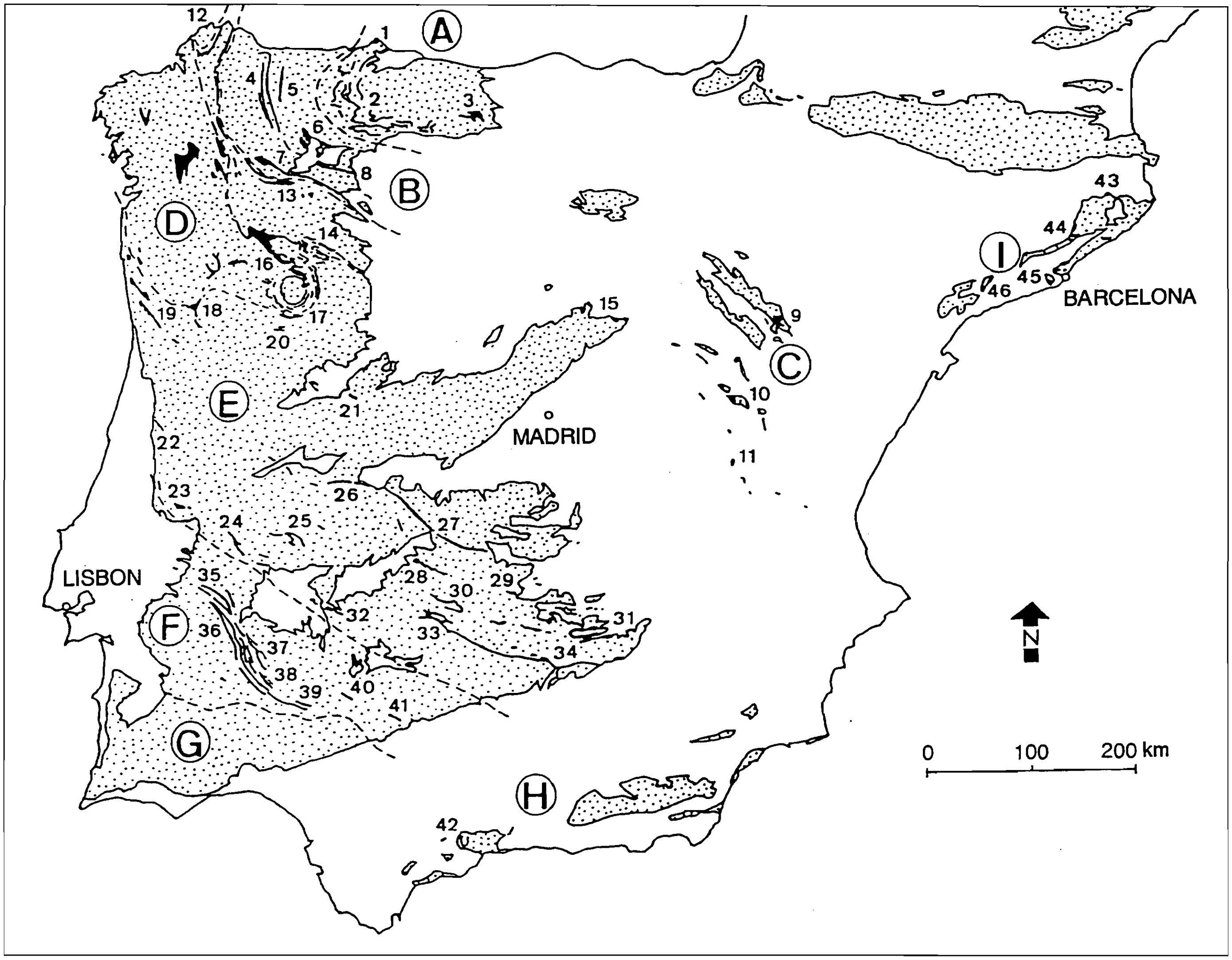

FIGURE 2 — Geological map of the Iberian Peninsula showing Silurian (in black) and main Precambrian and Paleozoic exposures (stippled). Abbreviations: AG, Hesperian Massif (A, Cantabrian Zone [Z.]; B, West Asturian–Leonese Z.; C, Iberian Cordillera Z.; D, GaliciaTrás os Montes Z.; E, Central Iberian Z.; F, Ossa–Morena Z.; G, South Portuguese Z.); H, Betic Cordilleras; I, Catalonian Coastal Ranges. Silurian of the Pyrenees omitted. 1–46, Main Silurian fossil localities and reference sections in Spain and Portugal: 1, Cabo Penas; 2, Silurian outcrops in "fold and nappe" region; 3, Ibid. in Palentian region; 4, Rececende and Villaodrid synclines (Mondonedo nappe); 5, Los Oscos thrust sheet; 6, Vega de Espinareda synclinorium; 7, Caurel-Penalba syncline; 8, Castrillo syncline; 9, Eastern Iberian Chains; 10, Albarracín anticlinorium; 11, Serranía de Cuenca anticlinorium; 12, Cabo Ortegal area; 13, Sil and Truchas synclines; 14, Alcanices synclinorium; 15, Riaza and Atienza areas ; 16, Verín–Bragança region; 17, Mogadouro and Morais area; 18, Vila Real (Marão); 19, Porto-Valongo; 20, Moncorvo syncline; 21, Tamames syncline; 22, Serra do Buçaco; 23, Dornes–Maçâo areas; 24, Portalegre; 25, Sierra de San Pedro and Cáceres syncline; 26, Canaveral syncline; 27, Guadarranque syncline; 28, Herrera del Duque syncline; 29, Corral de Calatrava; 30, Almadén syncline; 31, Torre de Juan Abad; 32, Alange; 33, Cabeza del Buey–San Benito; 34, El Centenillo-Guadalmena; 35, Estremoz anticline; 36, Terena syncline; 37, Villanueva del Fresno; 38, Barrancos area; 39, Hinojales area; 40, Valle syncline; 41, Cerrón del Hornillo syncline; 42, Malágulde region; 43, Guilleries; 44, Montseny; 45, Barcelona; 46, Serra de Miramar.

Devonian limestones and reefs. On the other hand, no diagnostic or unequivocal Silurian paleomagnetic data are available for Iberia. Indeed, Silurian volcanics from the Almadén area (southern Central Iberian Zone) were remagnetized just prior to the Variscan orogeny (i.e., probably in the Late Devonian or Early Carboniferous; see Perroud et al., 1991; Pares and Van der Voo, 1982). The paleogeography proposed in the world maps of Scotese and McKerrow (1990), which is mainly based on Gondwanan lithofacies (Scotese and Barrett, 1990), is acceptable for the Iberian Peninsula, and suggests movement from ca. 45–50° S in the Early Silurian to ca. 35–40° S in the Late Silurian.

Cantabrian Zone.—The Cantabrian Zone in the Hesperian Massif (Fig. 2A) shows thin-skinned tectonic structures with complex thrust systems (Pérez Estaún and Bastida, 1990). Silurian rocks, which have been recognized since the end of the last century (see Truzols and Julivert, 1983), occur both in the "fold and nappe" region and in the allochthonous Palentian region (=Pisuerga-Carrión Unit) in the southeast (Fig. 2, localities 2 and 3).

In the "fold and nappe" region, the oldest Silurian rocks occur only in the Cabo Peñas area (Fig. 2, locality 1). In this area, the upper El Castro Formation (i.e., Viodo Member limestones, 30 m) overlies basalt volcanics and has a rich, diverse fauna with brachiopods, bryozoans, tabulate and rugose corals, trilobites, sponges, ostracodes, and conodonts (Fig. 3). Conodonts indicate that the lower and middle Viodo Member is Rhuddanian (*Distomodus kentuckyensis* Zone) and the upper part is Aeronian (*Distomodus staurognathoides* Zone) (Sarmiento et al., 1994). Brachiopods, which are mostly new species, are considered an early Telychian *Stricklandia* fauna (Villas and Cocks, 1996).

The Formigoso Formation (Fig. 3) lies above the Viodo Member in the Cabo Peñas area and directly overlies Lower or Middle Ordovician rocks in most of the Cantabrian regions. The formation consists of 100–300 m of black and gray siltstone and shale with a few thin, sandy intercalations. The Formigoso Formation, with abundant graptolites and palynomorphs and rare brachiopods, bivalves and trilobites, extends from the upper Aeronian *Demirastrites convolutus* Zone to the lower Sheinwoodian *Cyrtograptus centrifugus-Cyrtograptus murchisoni* Zone (Truyols et al., 1974; Aramburu et al., 1992). These age assignments are based on graptolites and are confirmed by such organic-walled microfossils as chitinozoans (Truyols et al., 1974).

The Formigoso Formation is overlain by the Furada (=San Pedro) Formation (Fig. 3), which consists of 80–200 m of gray or reddish ferruginous sandstones with shale intercalations, thin oolitic ironstones, and sandy limestone lenses in the upper part. The Furada Formation yields late Wenlock palynomorphs in its lower part and Ludlow brachiopods, graptolites, and palynomorphs in its middle part. Ichnofossils are very common and diverse at many levels. The uppermost 30 m contains Pridoli chitinozoans (Priewalder, 1997). Higher levels with Lochovian brachiopods, trilobites, and conodonts indicate that the Silurian–Devonian boundary lies in the upper Furada Formation (Truyols et al., 1974; Aramburu et al., 1992).

In the Palentian region (Fig. 2, locality 3), the lithologic succession is different. The white sandstones of the Robledo Formation (ca. 160 m) are certainly Silurian and yield Wenlock chitinozoans near their top. The overlying siltstones and black shales of the Las Arroyacas Formation (ca. 300–450 m) have benthic faunas, graptolites, and chitinozoans that are Wenlock, Ludlow, and questionably Pridoli. The sandstones and carbonates of the Carazo Formation (sensu stricto, 250–380 m) have Pridoli chitinozoans and higher Lochkovian brachiopods and scyphocrinoids (Schweineberg, 1987; García Alcalde et al., 1990; Aramburu et al., 1992).

West Asturian–Leonese Zone — The Silurian in the West Asturian–Leonese Zone (Fig. 2B) conformably overlies a presumed uppermost Ordovician quartzite unit (Vega Quartzite and equivalents), except in the Los Oscos region (Fig. 2, locality 5) where the Silurian disconformably overlies Arenigian quartzites. The succession comprises a thick unit (La Garganta Beds, up to 500–600 m) of black shales with sparse intercalations of sandstone, argillaceous nodules, and chloritoid slates. The La Garganta Beds are overlain by at least 40 m of ferruginous sandstone (Queixoiro Beds), a possible equivalent of the Furada Formation of the Cantabrian Zone. Known fossils are restricted to the lower half of the La Garganta Beds and include abundant graptolites of the *Cystograptus vesiculosus–C. lundgreni* Zones (Rhuddanian–basal Homerian) in the Mondoñedo nappe (Walter, 1968; Romariz, 1968; Fig. 2, locality 4), Los Oscos thrust sheet (Marcos and Philippot, 1972), and Vega de Espinareda synclinorium and Castrillo syncline (Pérez Estaún, 1978; Gutiérrez-Marco and Robardet, 1991; Fig. 2, localities 6, 8).

Silurian is known in the core of the Caurel–Peñalba syncline (Fig. 2, locality 7) at the southern limit of the West Asturian–Leonese Zone. The lithologies and faunas of this region are similar to those of the Silurian in the northern Central Iberian Zone and include ca. 200 m of black chloritoid shales that overlap different Ordovician formations The black shales yield late Ludlow trilobites with Bohemian affinities in their upper part (Rábano et al., 1993).

Iberian Cordillera — Silurian rocks crop out in the western (Castilian Branch) and eastern (Aragonian Branch) parts of the Iberian Cordillera (Fig. 2C) as part of the Hesperian Massif under the Mesozoic and Cenozoic. The eastern and western Iberian Cordillera (Fig. 2, localities 9–11) can be simply considered as eastern extensions of the West Asturian–Leonese and Cantabrian Zones.

The Silurian of the Iberian Cordillera (see Gutiérez-Marco et al., 1998) begins with the massive, white Los Puertos Quartzite, with an average thickness of 20–35 m, that unconformably overlies Late Ordovician (Hirnantian) diamictites of the Orea Formation. The overlying Bádenas Formation, predominantly black shale with nodules and frequent arenaceous horizons toward the top, reaches a thickness of 850–1,400 m in the Aragonian Branch. Two main sandstone intervals divide the Báde-

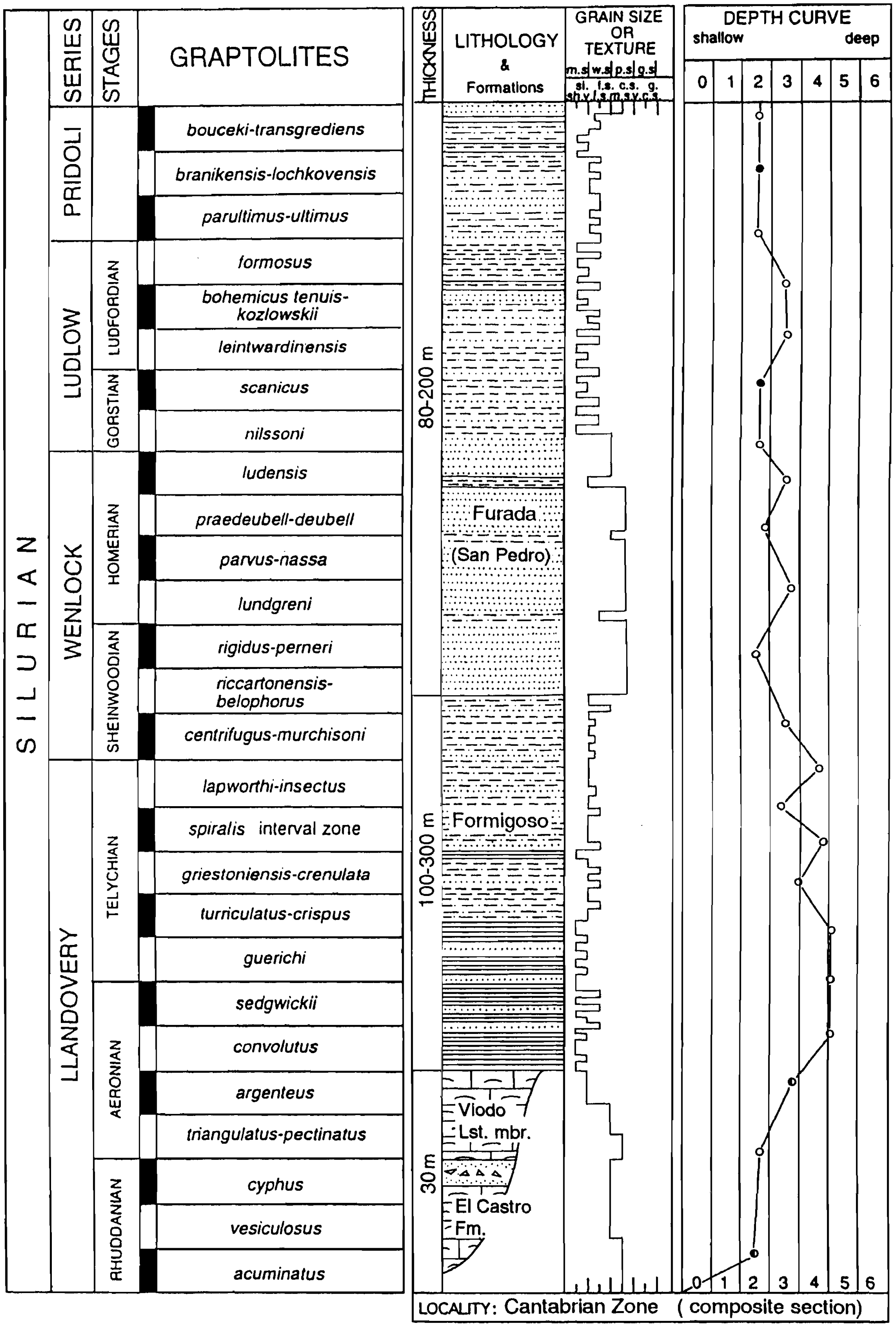

FIGURE 3 — Silurian of the Cantabrian Zone (composite section of the "fold and nappe region") in the Iberian Peninsula.

nas Formation into five members. In the western Iberian Cordillera, the Bádenas Formation is thinner (maximum 300–400 m) and crops out discontinuously in the core of the Alpine anticlinoria of Albarracín and Serranía de Cuenca (Fig. 2, localities 10, 11). Overlying the Bádenas Formation, the Silurian continues in the Eastern Iberian Chains as a dominantly sandstone unit (Luesma Formation, ca. 200 m) with quartzites, sandy shales, and ferruginous sandstones and with sparse limestone lenses in the upper part. The unit correlates with the Alcolea Formation of the "Central System" of the Central Iberian Zone, and with the Furada Formation of the Cantabrian Zone.

The oldest Silurian fossils in the above succession were found in shale intercalations within the upper Los Puertos Quartzite, and include Rhuddanian and Aeronian graptolites (*Parakidograptus acuminatus–Coronograptus cyphus*, *Demirastrites triangulatus*, and *Demirastrites convolutus* Zones). The black shales of the Bádenas Formation have abundant graptolites that range from the basal Telychian *Spirograptus guerichi* Zone to the basal Ludfordian *Saetograptus leintwardinensis* Zone. The non-graptolite fauna of the black shales and nodules of this formation includes brachiopods, bivalves, cephalopods, eurypterids, phyllocarids, peltocarids, cornulitids, trilobites, and conodonts. In the upper sandstones of the Bádenas Formation in the Aragonian Branch, shallow-water brachiopods, echinoderms, conodonts, trilobites, and ichnofossils occur. Finally, the Luesma Formation of the Eastern Iberian Chains has Pridoli brachiopods and Lochkovian conodonts and brachiopods in its upper part (Carls and Gandl, 1967; Truyols and Julivert, 1983; Gutiérrez-Marco and Štorch, 1998; Štorch, 1998).

CENTRAL IBERIAN ZONE — This zone (Fig. 2, region E) comprises two structural and paleogeographic parts (or domains) with different types of Silurian successions. These include the "Domain of Recumbent Folds" in the north and the "Domain of Vertical Folds" in the south (Díez Balda et al., 1990). The first domain includes at least three types of Silurian successions, and the latter is lithologically more uniform.

The first type of sequence crops out east and north of the Ollo de Sapo anticlinorium and comprises a lower unit of graptolitic black shales (Llagarino Formation, 80–150 m) that is overlain by a very thick unit of chloritoid shales with calcareous nodules and some sandstone beds (Salas Beds, ca. 1,000 m; Piçarra et al., 1998). Between these units, a sandstone (Yeres Quartzite, 5–25 m) occurs, but only in the eastern part of the Sil syncline (Fig. 2, locality 13). Graptolites show that the base of the Silurian succession is markedly diachronous and overlies different Ordovician formations (see Rábano et al., 1993, for references). However, in the Truchas synclinorium (Fig. 2, locality 13), the Llagarino Formation lies conformably on Hirnantian sandstones (Losadilla Formation) and has *Parakidograptus acuminatus* Zone graptolites at its base (Gutiérrez-Marco and Robardet, 1991). The upper Llagarino Formation is Gorstian (*Neodiversograptus nilssoni* Zone; Rábano et al., 1993). The Yeres Quartzite has trilobites, brachiopods, echinoderms, solitary rugose corals, bivalves, and cephalopods comparable with uppermost Ludlow assemblages of the Kopanina Formation in Bohemia (Rábano et al., 1993). Pridoli graptolites and scyphocrinoids occur in the Salas Beds, which must contain the Silurian–Devonian boundary in their upper part, and are conformably overlain by Pragian limestones.

West and south of the Ollo de Sapo anticlinorium, the second type of Silurian succession includes a thick unit of black shales with intercalations of chert (lydite), quartzites, conglomerates, and limestones (Manzanal del Barco Formation, ca. 200 m). The graptolite of this formation (Jiménez Fuentes and Quiroga, 1981) are Aeronian to lower Homerian (*Monograptus argenteus* to *Cyrtograptus lundgreni* Zones). The Manzanal del Barco Formation is overlain in the eastern Alca_ices synclinorium (Fig. 2, locality 14) by 300 m of gray shales and limestones of the Almendra Formation, with some flysch intervals that have yielded Ludlow, Pridoli, and Lower Devonian conodonts (García-López et al., 1996). This formation is replaced westwardly in this synclinorium by a thick flysch succession (San Vitero Formation, over 1,500 m of greywacke, shale, and conglomerate with olistoliths and transported plant remains).

A much shallower type of Silurian succession is located in the northeastern Central Iberian Zone near Atienza and Riaza (Fig. 2, locality 15), and is similar to the sequence known further east in the Iberian Cordillera. The Santibáñez Quartzite (20–30 m), supposedly of Rhuddanian age, lies conformably on Hirnantian shales and siltstones near Atienza and Riaza. The quartzite is succeeded by black shales and siltstones with quartzite interbeds in the upper part (Cañamares Formation, 190–250 m) with graptolites (Rhuddanian *Coronograptus cyphus* to lowest Gorstian *Neodiversograptus nilssoni* Zones). A thin limestone near the top of the Cañamares Formation has Pridoli conodonts. The Cañamares Formation is overlain by a 750–800 m sandy unit (Alcolea Formation) with upper Pridoli brachiopods and trilobites in its upper member. The Alcolea Formation is then overlain by a unit with Devonian assemblages (see Fernández Casals and Gutiérrez-Marco, 1985, for references).

The "Domain of Vertical Folds" includes the large Silurian outcrops in the southern Central Iberian Zone. Toward the northwest of this domain, the core of the Tamames syncline (Fig. 2, locality 21) has a succession with a variety of lithological types that resemble those on the south limb of the Ollo de Sapo anticlinorium. In the Moncorvo syncline (Portugal; Fig. 2, locality 20), a succession of Silurian black shales with a pelagic limestone rich in scyphocrinoids and orthocerids and with Pridoli conodonts has been described (Piçarra et al., 1995a).

The Silurian in the southern Central Iberian Zone begins with a lower unit of quartzites (Criadero Formation, 7–70 m) that rests conformably on Hirnantian shales and glacio-marine diamictites. The Criadero is overlain by 5–50 m of graptolitic black shales with sparse nodules and thin sandstone horizons near their top (Fig. 4). The next overlying thick sequence (150–400 m) includes nonfossiliferous sandstones and micaceous shales that continue into the Devonian. The entire succession shows intercalations of volcanic rocks, especially in the area around Almadén (Fig. 2, locality 30), with mercury mineralization related to Aeronian volcanism. Other areas with Silurian rocks in the Central Iberian Zone are the Portuguese outcrops near Porto, Serra do Bussaco, Maçáo-Dornes, and Portalegre (Fig. 2, localities 22–24), as well as the Spanish regions of Guadarranque, Cáceres, San Benito-El Centenillo, and the western Sierra Morena (Fig. 2, localities 25, 27; Sarmiento and Rodríguez Núñez, 1991; Štorch et al., 1998).

Faunas from the top of the Criadero Formation (and its partial equivalent, the Portuguese Vale da Ursa Formation) include rare assemblages of Rhuddanian and Aeronian graptolites (Coronograptus cyphus–Stimulograptus sedgwickii Zones; Brenchley et al., 1991; García Palacios et al., 1996a). The overlying black shales include abundant graptolites indicative of most of the zones between the basal Telychian and Homerian (*Spirograptus guerichi–Pristiograptus dubius parvus-Gothograptus nassa?* Zones; Romariz, 1962; García Palacios et al., 1996b) and sparse brachiopods, bivalves, cephalopods, trilobites, eurypterids, phyllocarids, peltocarids, cornulitids, and conodonts (Sarmiento and Rodríguez Núñez, 1991; García Palacios and Rábano, 1996). Graptolite biostratigraphy clearly shows a hiatus of variable duration at the base of the black shale unit and a clear diachroneity of the base of the overlying sandstone unit, which becomes older to the northwest (Rodríguez Núñez et al., 1989).

GALICIA–TRÁS-OS-MONTES ZONE — This zone (Fig. 2D) constitutes a great allochtonous sheet in the northwest Central Iberian Zone. Its uppermost ophiolitic sequences are overlain by exotic continental terranes (Cabo Ortegal, Órdenes, Bragança, and Morais complexes; Martínez Catalán et al., 1996). The autochthonous unit in this region is a very thick metasedimentary succession termed the "Schistose Domain of Galicia–Trás os Montes," which includes several Silurian intervals with rare, poorly preserved graptolites (see Farias, 1992, for references). Graptolites are more abundant around the Bragança Massif (Romariz, 1968; Fig. 2, locality 16). The local presence of these fossils allow Silurian rocks to be identified in the Nogueira Group (900–1,000 m, with Telychian graptolites in its lower part) and in the Paraño Group (2900–3200 m, with Llandovery–Wenlock graptolites in its lower part; Romariz, 1968).

OSSA–MORENA ZONE — Silurian rocks are known in the Spanish and Portuguese parts of the Ossa–Morena Zone in southern Iberia (Fig. 2F). The best-preserved and -documented reference successions are in Spain in the Cerrón del Hornillo and Valle synclines (Seville Province) in the eastern part of this zone (Fig. 2, localities 40, 41). In other areas of the Ossa–Morena Zone where outcrops are rarer and where tectonic structures are complicated, the Silurian is not as well known, even though the Barrancos area (Portugal; Fig. 2, locality 38) provides interesting information.

Lithology, faunas, and stratigraphy are almost identical in the Cerrón del Hornillo and Valle synclines (Jaeger and Robardet, 1979; Robardet and Gutiérrez-Marco, 1990a). The Silurian conformably overlies the late Ashgillian Valle Formation, which is composed of dark shales and siltstones that include clast-bearing horizons regarded as Hirnantian glacio-marine sediments (see Gutiérrez-Marco et al., 1998, p. 22). The Silurian–Lochkovian succession consists of 130–150 m of black, argillaceous shales with intercalations of siliceous slates and chert. The lowermost part of the succession shows sandy shale levels; a thin (0.5–0.8 m), black limestone with orthocerids and bivalves occurs in the Ludlow (Jaeger and Robardet, 1979). The most important lithologic change is shown by the Pridoli "*Scyphocrinites* Limestone" (S.Lst. in Fig. 5; 10–15 m). This unit of alternating limestones and shales divides the black shale succession into "Lower Graptolite Shales" (Fig. 5) and "Upper Graptolite Shales" (U.G.S. in Fig. 5). The Silurian–Lochkovian is fossiliferous throughout. The fauna is composed almost entirely of graptolites. Most of the Silurian–Lower Devonian graptolite zones have been recognized, and this indicates a continuous succession that extends from the base of the Rhuddanian (*Parakidograptus acuminatus* Zone), through the uppermost Pridoli (*Monograptus transgrediens* Zone), and into the Lochkovian (up to the *Monograptus hercynicus* Zone) (Jaeger and Robardet, 1979; Lenz et al., 1996; Piçarra et al., 1998). Orthocerids and bivalves are known in the Ludlow limestone; scyphocrinoids, trilobites, bivalves, conodonts, ostracodes, graptolites, and rare solitary corals and brachiopods also occur in the "*Scyphocrinites* Limestone." Anoxic or strongly dysaerobic conditions persisted through the earliest Rhuddanian–late Lochkovin, with the exception of the Pridoli, when bottom oxygenation probably increased.

The tripartite succession of the eastern Ossa–Morena

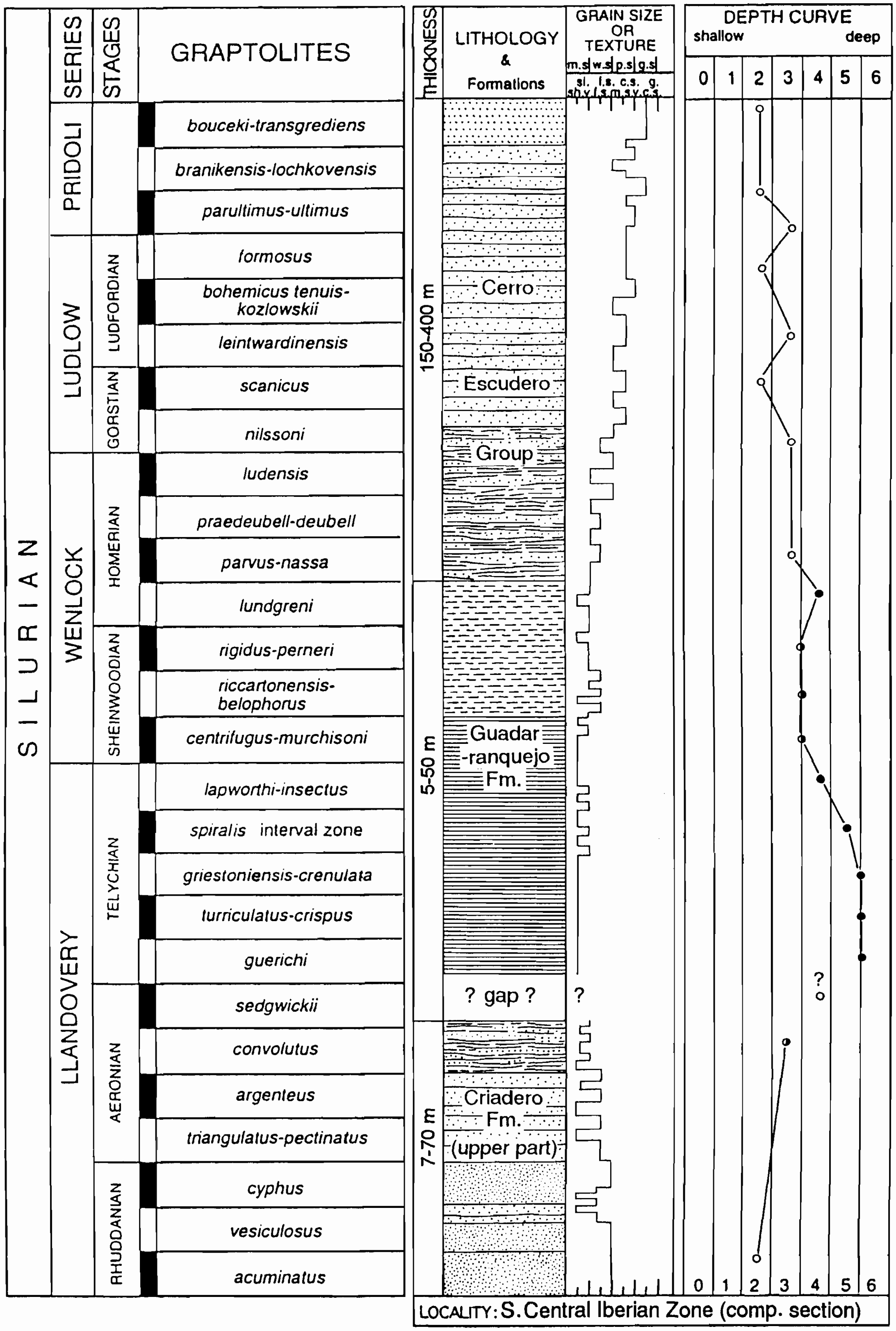

FIGURE 4 — Silurian of the Central Iberian Zone (composite section for the Almadén–Corral de Calatrava area, Iberian Peninsula.

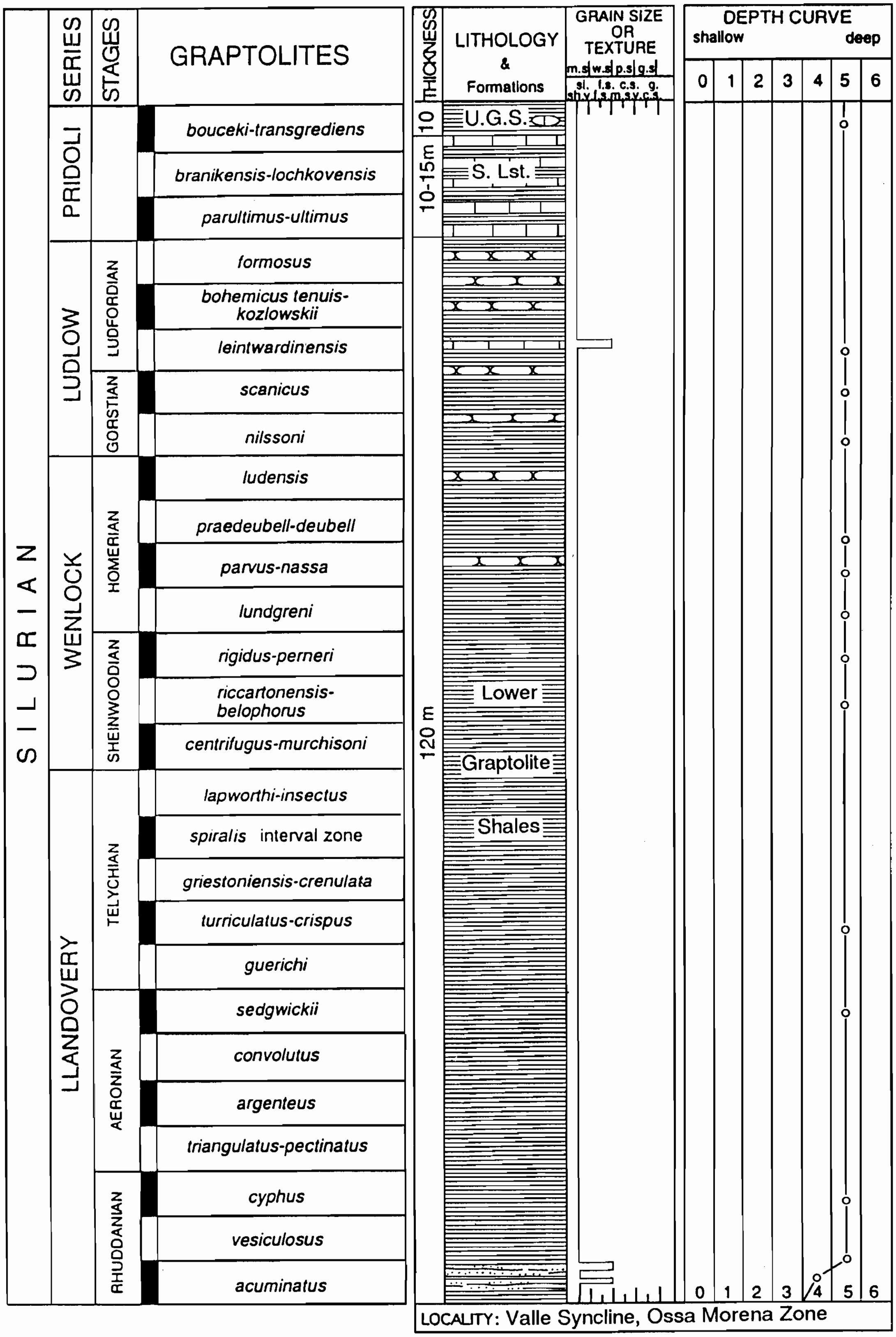

FIGURE 5 — Silurian of the Valle syncline, Ossa–Morena Zone (LGS, lower graptolitic shales; S. Lst., *Scyphocrinites* Limestone) in the Iberian Peninsula.

Zone is very similar lithologically and faunally to the typical "Thuringian triad" (Jaeger, 1976, 1977b). The "*Scyphocrinites* Limestone" appears to be the precise equivalent of the "Ockerkalk" in Germany (Robardet, 1982; Robardet and Gutiérrez-Marco, 1990a). Anoxic or strongly dysaerobic sedimentation ended in the eastern Ossa–Morena Zone and Germany after the late Lochkovian with deposition of green–brown shales and siltstones with trilobites, ostracodes, and brachiopods in the Cerrón del Hornillo and Valle synclines of the Ossa-Morena Zone (Racheboeuf and Robardet, 1986; Robardet et al., 1991).

To the west, Silurian rocks also occur in the Hinojales and Villanueva del Fresno areas (Spain; Fig. 2, localities 37 and 39), and in the Barrancos and Estremoz areas (Portugal; Fig. 2, localities 35 and 38). In these areas, the Silurian succession is not very well known. However, it must be noted that the "*Scyphocrinites* Limestone" has never been observed in these areas, and that unfossiliferous sandstones and quartzites occur in the lowermost part of the succession in the northwest (Villanueva del Fresno and Estremoz). In the Barrancos, Portugal, area, the base of the Silurian (*Parakidograptus acuminatus* Zone) has been recognized (Piçarra et al., 1995b) in the lowermost "Xistos com Nódulos" Formation (30–50 m of black shale and chert; Llandovery to Ludlow; *Parakidograptus acuminatus–Lobograptus scanicus* Zones). The succeeding "Xistos Raiados" Formation (100 m of banded chloritoid shales and siltstones; Oliviera et al., 1991) extends from the Ludfordian into the Lochkovian. Pridoli graptolites from the *Monograptus parultimus* to *M. bouceki* Zones were discovered in its lower part (Piçarra et al., 1998), and the uppermost 12 m yield *Monograptus uniformis* (see Robardet et al., 1998). The general succession of graptolite faunas from Barrancos is very similar to that of the Valle syncline, especially through the Homerian *Lundgreni* Event (see Jeppsson, 1998), when distinct lithologies and non-graptolite faunas developed in both areas (Gutiérrez-Marco et al., 1996; Rigby et al., 1997).

The Silurian succession in the Ossa–Morena Zone thus appears to be clearly different from the Silurian of other Iberian regions. It is particularly different from the successions in the now juxtaposed Central Iberian Zone (Robardet, 1976; Robardet and Gutiérrez-Marco, 1990b).

CATALONIAN COASTAL RANGES — In this area of northeast Spain (Fig. 2I), small Silurian outcrops are known in the Barcelona area. The Silurian overlies Upper Ordovician quartzites, siltstones, and black shales with late Caradocian to middle Ashgillian faunas (Villas et al., 1987). Hirnantian glaciomarine sediments have not been recognized, but the upper Ashgillian may be represented by several tens of meters of unfossiliferous dark shale (Julivert and Durán, 1992).

The Silurian features a lower black shale unit (150–300 m) with chert intercalations in its lower part, sandy levels in the middle, and limestone beds and lenses in the upper part. The base of the Silurian has not yet been determined, but may correspond to a hiatus or occur within a sequence of unfossiliferous quartzites. The graptolitic black-shale sequence extends from the *Cystograptus vesiculosus* Zone and possibly to the *Saetograptus leintwardinensis* Zone (i.e., from the Rhuddanian to the lower Ludfordian) (Julivert et al., 1985; Julivert and Durán, 1990).

The carbonate sequence that overlies the graptolitic black shales comprises two formations. The lower one (La Creu Formation) consists of 30–40 m of gray, massive, nodular limestones with thin shale interbeds. Crinoids, orthocerids, bivalves, and conodonts indicate that most of the La Creu Formation is Pridoli (*Ozarkodina eosteinhornensis* Zone). The lowermost beds of the formation are Ludlow, and the upper part is Lochkovian (García López et al., 1990).

The overlying 35–40 m of the Olorda Formation consist mainly of nodular limestones and marls with Lochkovian brachiopods, dacryoconarids, and conodonts. Graptolites from the *Monograptus uniformis* to *M. hercynicus* Zones occur in shales in the lowest part of the formation (Lenz et al., 1996), and higher faunas range from the Pragian into the Emsian.

BETIC CORDILLERAS — The Betic Cordilleras of southeast Spain (Fig. 2H; see references in Gómez Pugnaire, 1992) are part of the Mediterranean Alpine Belt and the so-called "Malaguide Complex" (= "Betic" of Málaga) and have Paleozoic rocks. The Morales Formation (minimum thickness 200 m) is Upper Ordovician to lowest Devonian. Its middle part (probably Llandovery–lower Ludlow) consists of shale and siltstone with chert and limestone intercalations; the latter lithologies have yielded tintinnids and conodonts. The upper Morales Formation (probably upper Ludlow–Lochkovian) consists of limestones with tentaculitids, cephalopods, and conodonts. Silurian conodonts have also been recorded from limestone olistoliths within the younger (Devonian–Carboniferous) Sancti Petri and Almogía Formations in the same area (García-López et al., 1996).

OVERVIEW OF IBERIAN SILURIAN — All the Silurian regions that now compose the Iberian Peninsula were part of the north Gondwanan marine shelf, which extended north of the African margin of Gondwana (Fig. 1). The distinct sedimentological and faunal features that characterize the different regions give an impression of their relative position on the north Gondwanan shelf and of pre-Variscan paleogeography.

The Ossa–Morena Zone (OMZ) is clearly distinct from the central Iberian regions now situated to the north. OMZ features include: 1) a continuous transition between the Ordovician and Silurian; 2) anoxic or strongly dysaerobic black shale sedimentation that started as early as the *Parakidograptus acuminatus* Chron and persisted through the Silurian into the Lochkovian; 3) chert interbeds; 4) an important lithofacies change in the Pridoli with deposition of the "*Scyphocrinites* Limestone," an exact equivalent of the "Ockerkalk" of other European regions; and 5) re-appearance of oxygenated conditions in the Pragian.

The Silurian of the OMZ is obviously similar to the Thuringian Facies, and the OMZ appears to have been located on the outer part of the north Gondwanan shelf. The present juxtaposition of the Ossa–Morena and Central Iberian zones along the Badajoz–Córdoba Shear Zone does not reflect Silurian paleogeography, and most probably resulted from sinistral strike-slip movements during the Variscan orogeny.

The Silurian of the southern part of the Central Iberian Zone is different from that of the OMZ and is characterized by: 1) sandy lithofacies near the Ordovician–Silurian transition, although the lack of fossils does not necessarily reflect either hiatuses or continuous sediment accumulation; 2) a first occurrence of Silurian graptolite faunas in the Aeronian or Telychian; 3) anoxic or strongly dysaerobic black shale sedimentation from the early Telychian to late Homerian that apparently took place in shallower environments than is usual for this type of facies; and 4) deposition of alternating sandstones, siltstones, and shales during the Ludlow and Pridoli in more normally oxygenated environments. This coarser terrigenous influx announced the initiation of Lower Devonian sand deposition. The sedimentary evolution of the Cantabrian Zone (with the exception of the Palentine Domain), parts of the West Asturian–Leonese Zone, and the Iberian Cordillera is relatively similar to that of the Central Iberian Zone.

Within the Hesperian Massif, there is a general trend from shallow deposits in the southern Central Iberian Zone to deeper, more distal sedimentation both in the northern Central Iberian Zone and in the southern West Asturian–Leonese Zone. Pridoli "*Scyphocrinites* Limestone" occurs in the Moncorvo area (Portugal) and in the northern Central Iberian Zone. In the boundary between the Central Iberian and West Asturian–Leonese Zones, Silurian successions are mostly shale, are silty, with a few limestones (Alcañices syncline), and have Ludlow trilobites with Bohemian affinities (Peñalba and Sil synclines). These features are more reminiscent of the Silurian of the Pyrenees and Catalonia. These Hercynian magnafacies are also recognized in the Palentian Domain of the Cantabrian Zone (which is supposedly of Asturian–Leonese provenance). If the San Vitero Formation is actually Silurian–Devonian (and not a younger flysch unit), its occurrence in a western position is an additional argument in favor of deeper environments to the west. Despite complications that resulted from Variscan tectonism (but with the exception of the Cantabrian Zone), there is a general trend to deeper and more distal areas from south to north and toward the northwest in Iberia.

A good biogeographic marker for the shallowest areas of the Silurian shelf is the graptolite *Metaclimacograptus flamandi* (LeGrand, 1993), which is abundant in Telychian black shale facies (from the *Monograptus crispus*–lower *Torquigraptus tullbergi* Zones) of the southern Central Iberian Zone, Iberian Cordillera, Central System, and Castrillo syncline of the West Asturian–Leonese Zone. This graptolite is unknown in coeval deposits of the Ossa–Morena Zone, the boundary region between the Asturian–Leonese and Central Iberian Zones, the Pyrenees, and Catalonian Coastal Ranges. This graptolite, which was not distinguished from *Paraclimacograptus innotatus brasiliensis* (Ruedemann) in older studies, has a wide distribution related to more inshore environments around the African part of Gondwana. In Iberia, Llandovery black shales with *M. flamandi* also show a number of features that probably reflect shallow conditions, and these black shales are probably not very different bathymetrically from the storm-influenced sandstones that usually underlie this facies (García Palacios et al., 1996b; Gutiérrez-Marco and Štorch, 1998). *Paraclimacograptus innotatus brasiliensis* sensu stricto is restricted to older levels (Rhuddanian–Aeronian) of the South American part of the Perigondwana shelf in Brazil, Argentina, and Paraguay, where it is indicative of shallow-shelf areas (Jaeger, 1976).

The Silurian of the Catalonian Coastal Ranges consists of black shales (Rhuddanian–Ludfordian) and limestones (upper Ludlow–lowest Lochkovian). These shales were probably deposited in outer-shelf environments, and this interpretation accords with the proposed paleogeographic reconstruction.

Silurian of the Pyrenees

Most of the Silurian occurs in the High Paleozoic Range, in the Paleozoic massifs of the north Pyrenean Zone, and in the Mouthoumet Massif (Figs. 6, 7). Silurian units, generally a dark-colored pelitic facies, provide stratigraphic markers in the Paleozoic of the Pyrenees. Unfortunately, the Silurian is highly tectonized, and its outcrops are generally poor. The lithosequence is problematical, and many outcrops are needed to reconstruct a composite lithostratigraphic column. Such a reconstruction is allowed by Llandovery and Wenlock graptolites and

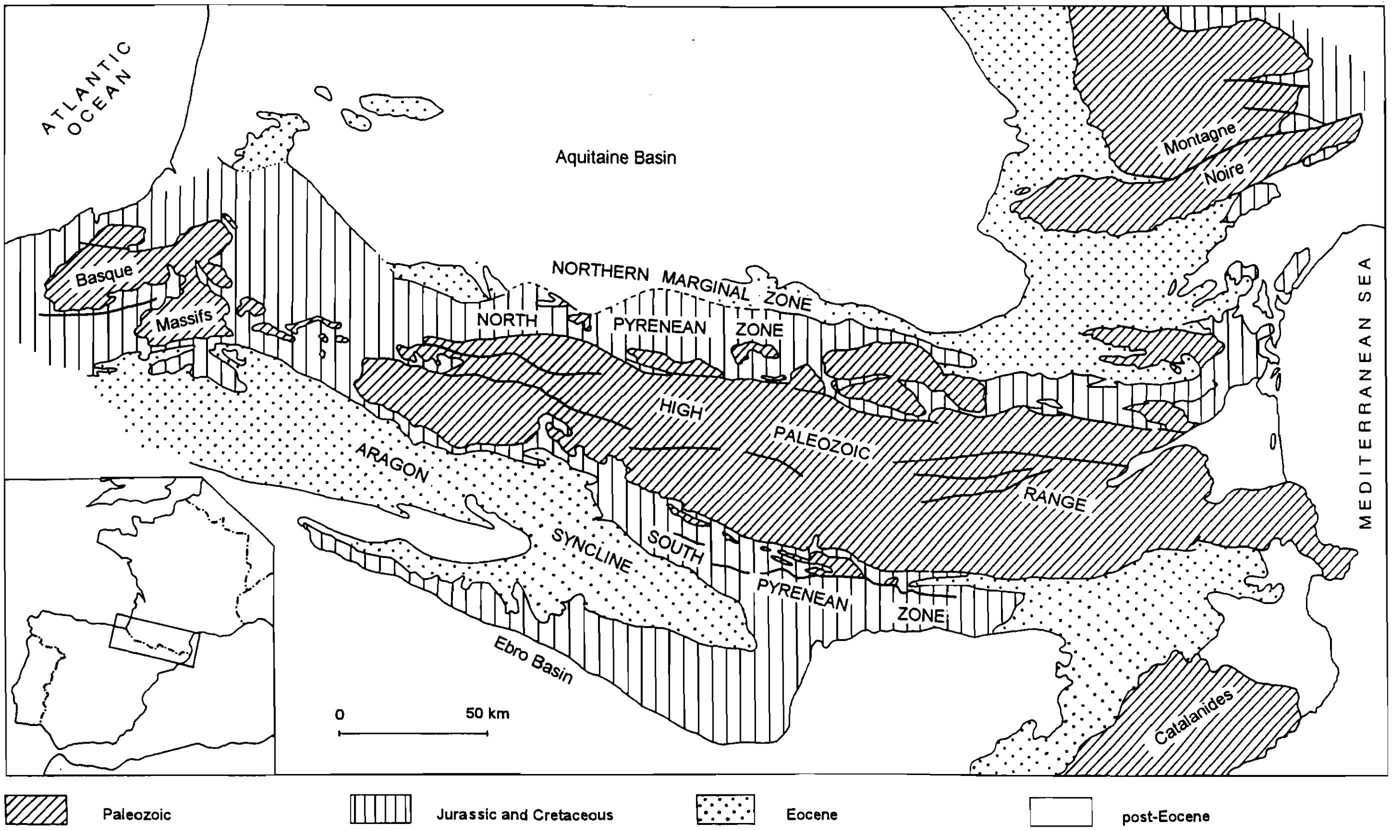

FIGURE 6 — Major structural divisions of the Pyrenees.

Upper Silurian conodonts (Dégardin, 1988).

LITHOSTRATIGRAPHY — Stratigraphic columns (Figs. 8–14) show significant changes in lithology and thickness. Black pelites are developed from the base of the Silurian up to the Wenlock. Towards the top of the Silurian, carbonate sediments appear and continue into the Lower Devonian.

Throughout the Pyrenees, the Lower Silurian (i.e., Rhuddanian and Aeronian) is characterized by organic-rich black pelites. However, some interbedded limestones are developed in the central Pyrenees, and a pelitic sequence with sandstone lenses is known from the the Camprodon area (Fig. 10). In general, pelitic sedimentation continued during the Telychian, except in Mouthoumet Massif, where carbonates were deposited.

During the Sheinwoodian and Homerian, pelitic sedimentation was ubiquitous across the Pyrenees, except in the eastern and central Pyrenees where carbonate sedimentation also took place. During the Gorstian and Ludfordian, important lateral facies changes took place. Toward the west in the Pays Basque, intercalations of sandstones occur within pelitic rocks. In the eastern Pyrenees, carbonate sedimentation took place locally with thin-bedded siltstone deposition. Except in the southern area (Sallent de Gallego region, Andorre, Bar-Toloriu, and Camprodon areas) where carbonate nodules are scattered in pelites, pelitic sedimentation prevailed in all other regions.

During the Pridoli, the diversity of the facies was an important feature. In the west, rhythmic pelite and sandstone sedimentation was continuous from the Pays Basque to the Argeles–Gazost area. Somewhat different facies in the central Pyrenees feature dominantly siliciclastic mudstones with carbonate intercalations in thin beds or nodule horizons. In the eastern Pyrenees, carbonate facies were abundant until the Early Devonian (Centéne and Sentou, 1975).

BIOSTRATIGRAPHY IN THE PYRENEES — Silurian biostratigraphy in the Pyrenees is based mainly on graptolites from the pelites and conodonts from the carbonates. Eighty-six graptolite species assigned to thirteen genera are recognized in the Silurian of the Pyrenees (Dégardin, 1988).

Nine graptolite zones are recognized: Aeronian *Demirastrites triangulatus*, *D. convolutus*, and *Stimulograptus sedgwickii* Zones; Telychian *Spirograptus turriculatus*,

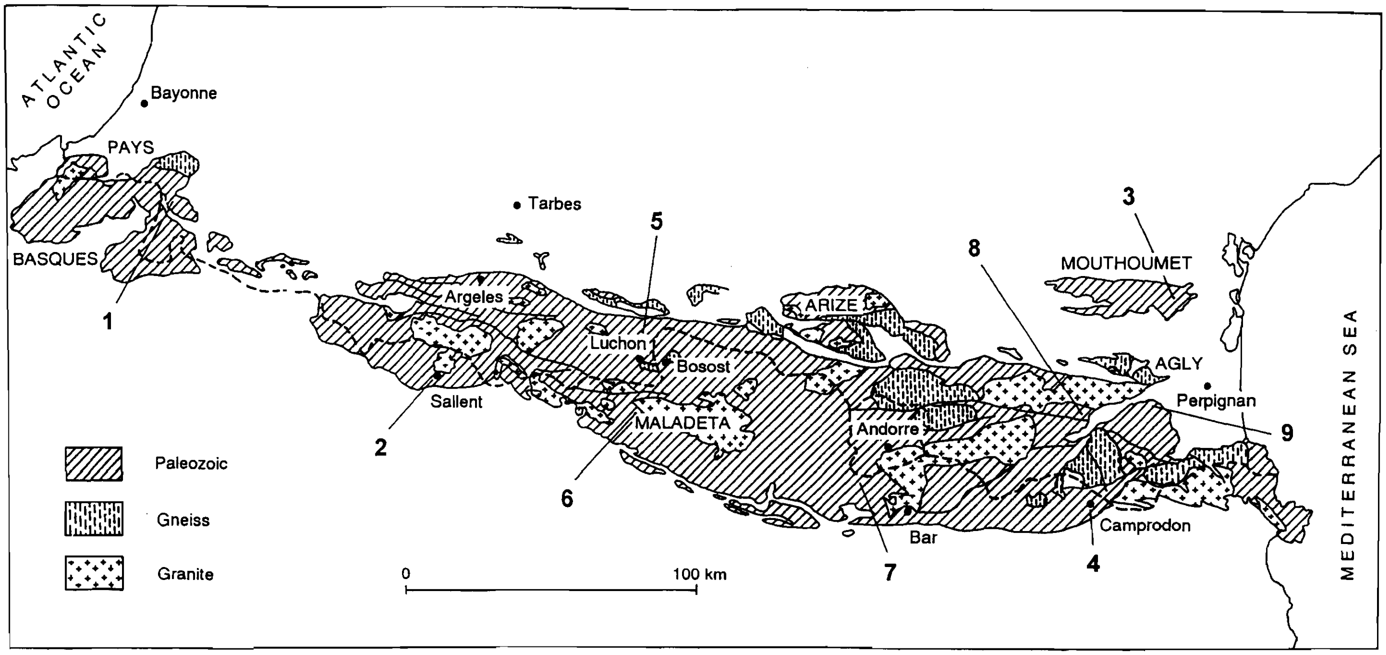

FIGURE 7 — Distribution of the Paleozoic and correlation charts in the Pyrenees (see Figs. 8–14) and the Mouthoumet Massif. Localities: *1*, St. Martin d´Arrosa; 2, Sallent de Gallego; *3*, Mouthoumet Massif; *4*, Camprodon; *5*, Pale de Burat; *6*, Sierra Negra; *7*, Querforadat-Toloriu; *8*, Villefranche; *9*, Castelnou.

Monoclimacis griestoniensis, and *Oktavites spiralis* Zones; Sheinwoodian *Monograptus riccartonensis* Zone; Gorstian *Neodiversograptus nilssoni* Zone; Ludfordian *Bohemograptus bohemicus* Zone. Correlations are based essentially on the association of species. The first association includes *Demirastrites triangulatus, D. delicatulus, Monograptus intermedius, M. lobiferus*, and *Campograptus communis communis*, which indicates the base of the Aeronian Stage of the Llandovery. This association is present in the Pays Basque, central Pyrenees, and Camprodon region. Another association with *Stimulograptus sedgwickii, M.? involutus*, and *Pristiograptus regularis* is upper Aeronian (Dégardin, 1988).

A later group of species with *Torquigraptus proteus, T. planus, Monograptus undulatus*, and *Spirograptus turriculatus* is lower Telychian, whereas *Oktavites spiralis* and *Stimulograptus halli* indicate the upper Telychian. These graptolites are frequent in the central Pyrenees where they are associated with *M. priodon* and *Retiolites geinitzianus geinitzianus* (Dégardin, 1988).

Monograptus lamarmorae, M. mutuliferus mutuliferus, and *M. uncinatus* var. *tariccoi* occur at the top of the Wenlock in the Pyrenees. This assemblage is present in the Pays Basque and central Pyrenees (Benasque area) with *M. flemingii*. The youngest group, with *Pristiograptus dubius, Monograptus uncinatus orbatus, Colonograptus colonus*, and *Neodiversograptus nilssoni*, characterizes the Gorstian (Dégardin, 1988).

Conodonts have been extracted only from the carbonates commonly found at the top of the Wenlock and in the Ludlow and Pridoli of the eastern Pyrenees, and on the Spanish side of the central Pyrenees. The faunas corroborate the biostratigraphy denoted by the graptolites in the pelitic levels.

Kříž (1996) described eight species of bivalves from the Mouthoumet Massif that are known or closely related to the bivalves of the Prague Basin, Bohemia. They range from late Wenlock to Pridoli and show the close faunal relationships between both regions. *Cardiobeleba thorali* and *C.* sp. aff. *C. ava* show very close relationships to upper Wenlock faunas of the Carnic Alps (Austria, Italy).

PALEOGEOGRAPHY — The Silurian of the Pyrenees includes inner-shelf deposits, as documented by the low thickness and local presence of such benthos as bivalves, cephalopods, brachiopods, and echinoderms. The environment was commonly anoxic, as shown by high pyrite and organic matter content.

On the Scotese et al. (1979, 1990a, 1990b) paleogeographic reconstruction, the Pyrenees are positioned on the northern margin of Gondwana. It is possible to place the Pyrenean area, as in the model of Pickering et al. (1988), somewhere between Brittany and Iberia, but its precise position remains uncertain. Furthermore, it seems that the north Gondwanan platform was tectonically

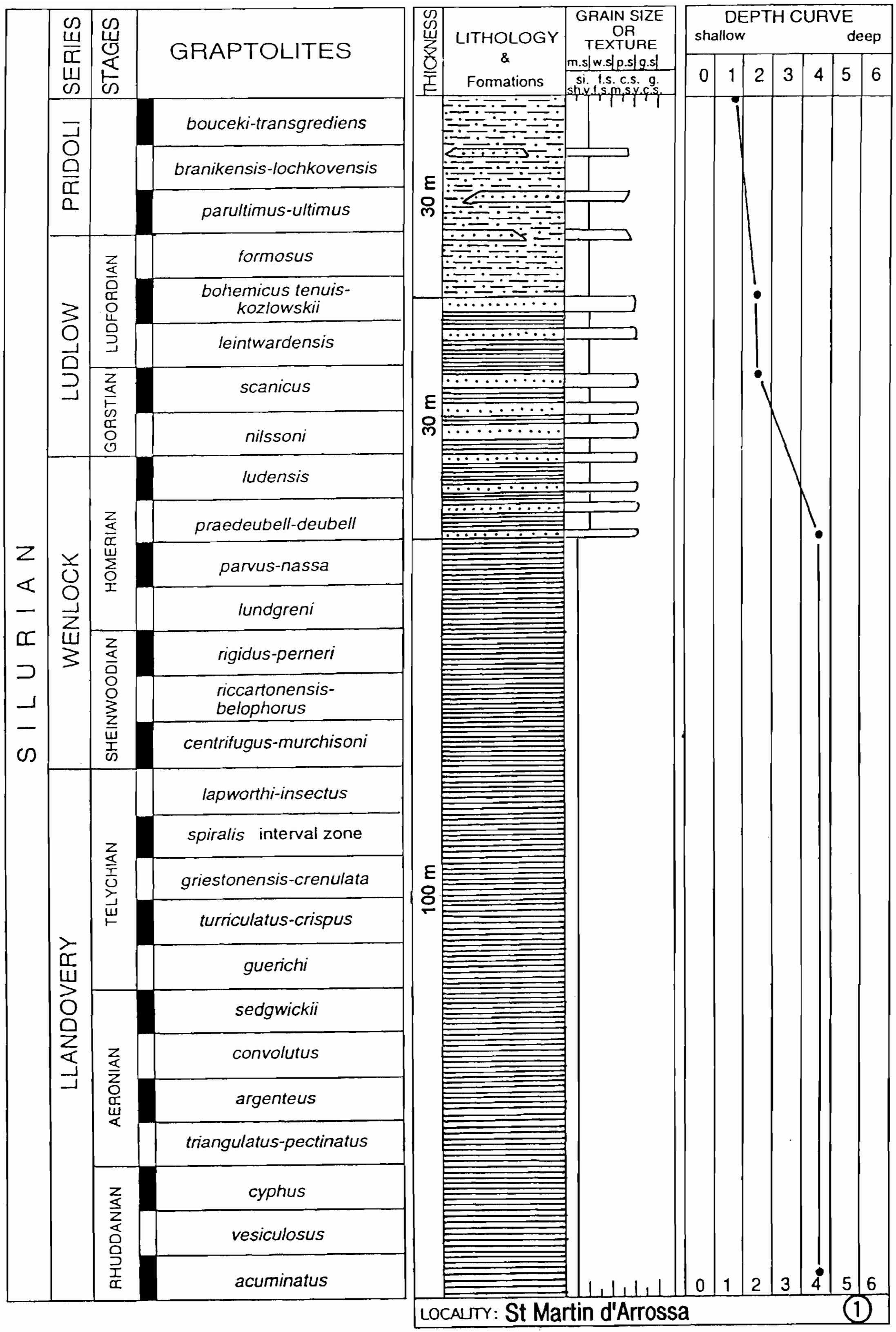

FIGURE 8 — Silurian correlation chart of St. Martin d´Arrosa. For location, see Fig. 7 (locality 1).

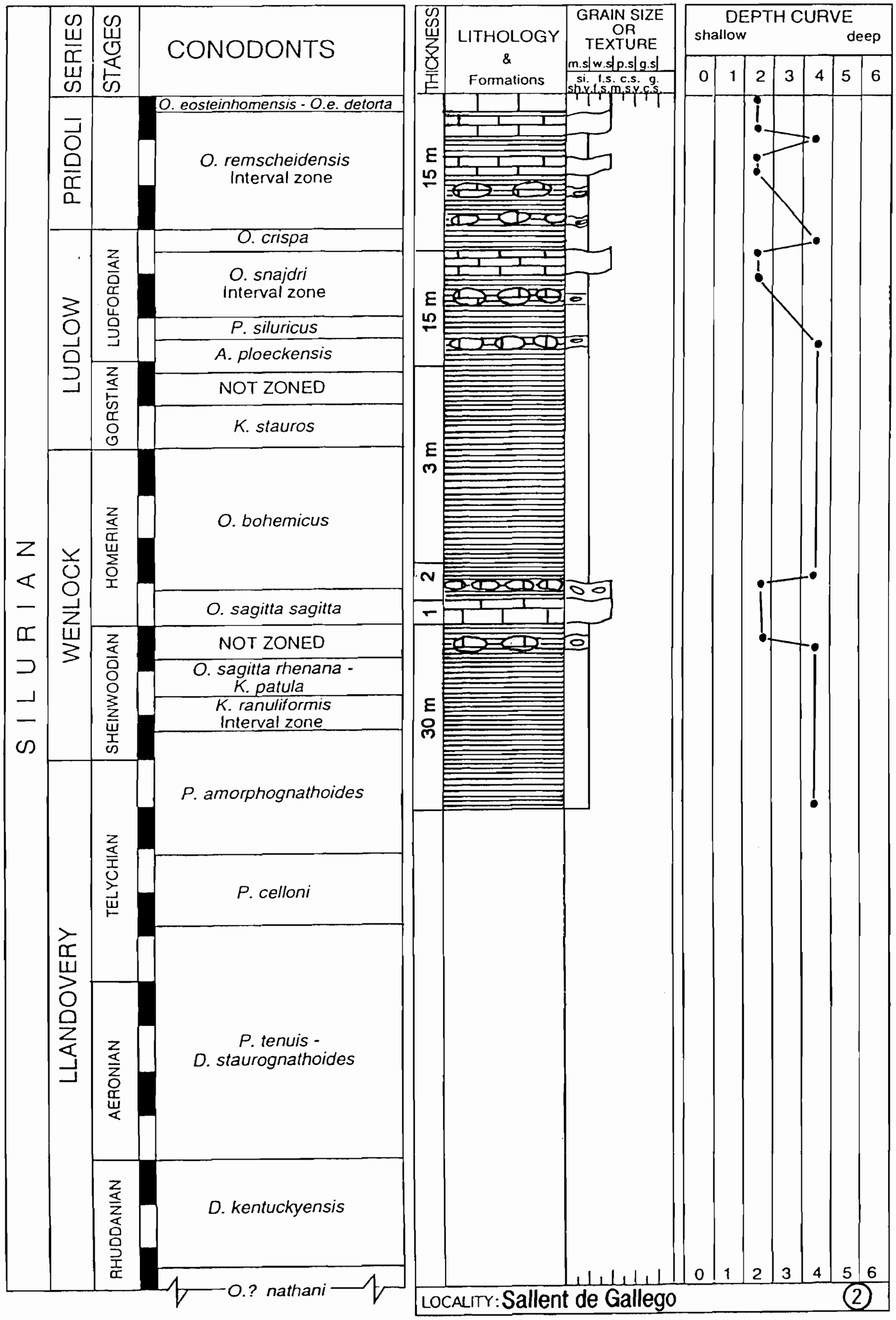

FIGURE 9 — Silurian of Sallent de Gallego. For location, see Fig. 7 (locality 2).

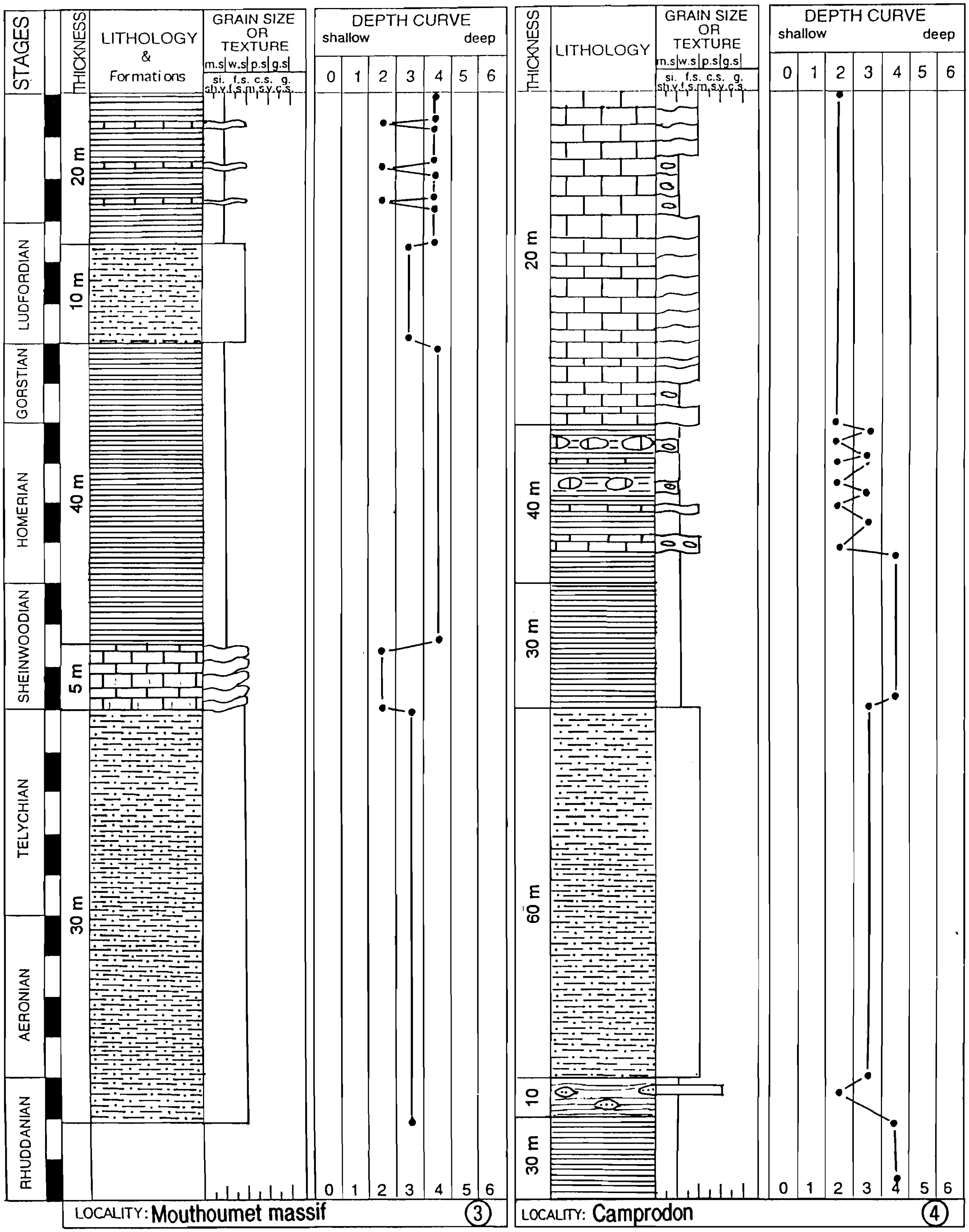

FIGURE 10 — Silurian of the Mouthoumet Massif and Camprodon. For locations, see Fig. 7 (localities 3, 4).

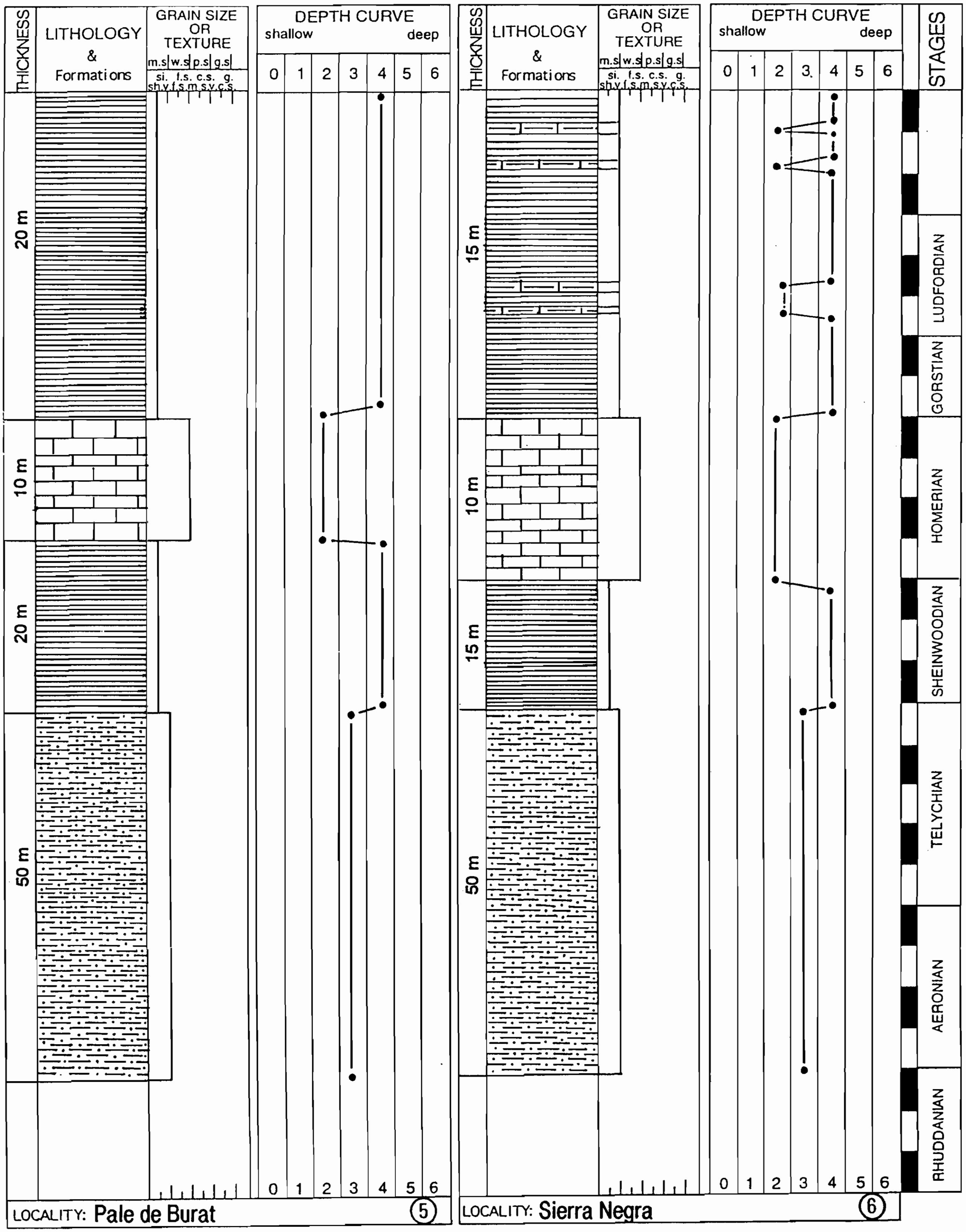

FIGURE 11 — Silurian Pale de Burat and Sierra Negra. For locations, see Fig. 7 (localities 5, 6).

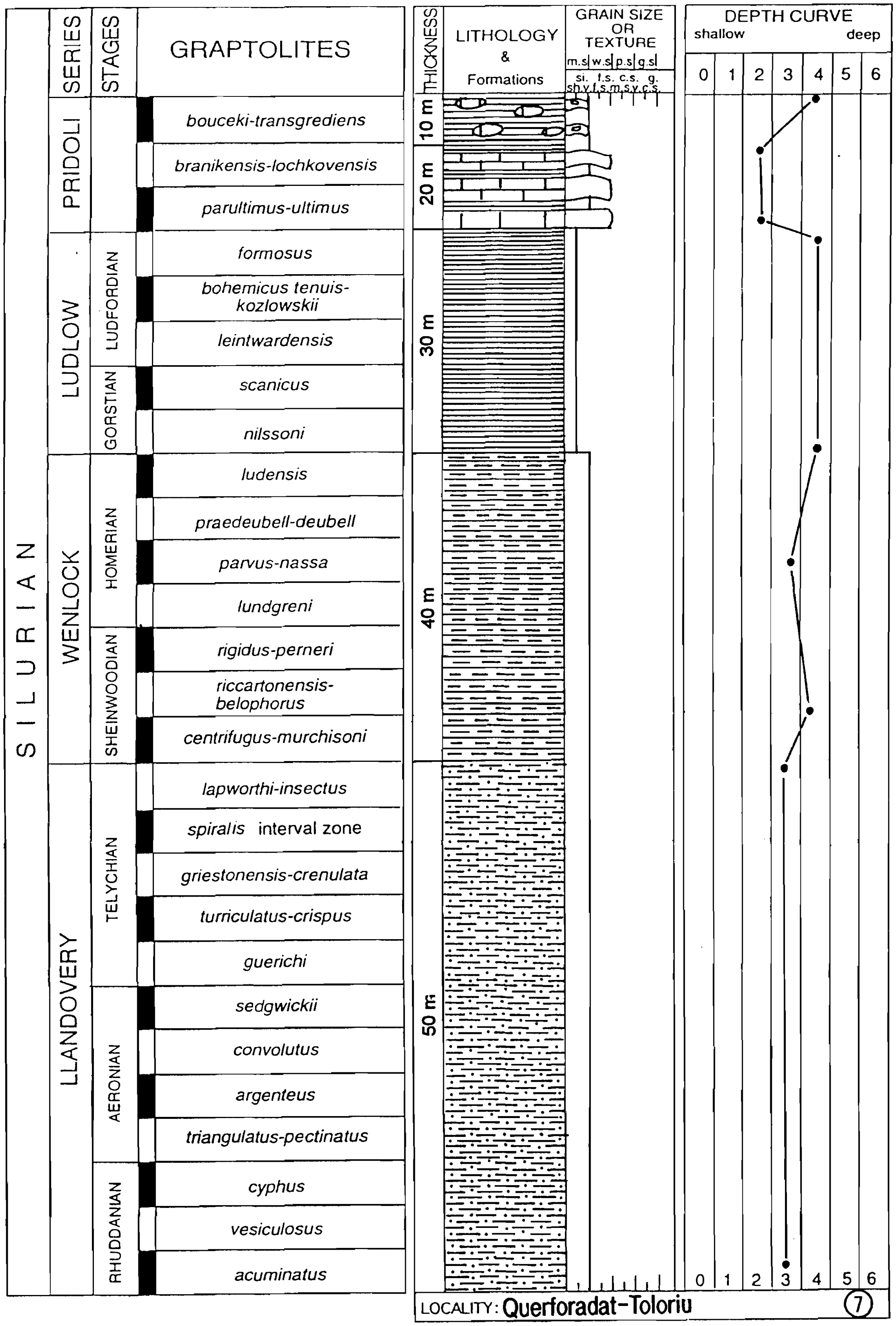

FIGURE 12 — Silurian of QuerforadatToloriu. For location, see Fig. 7 (locality 7).

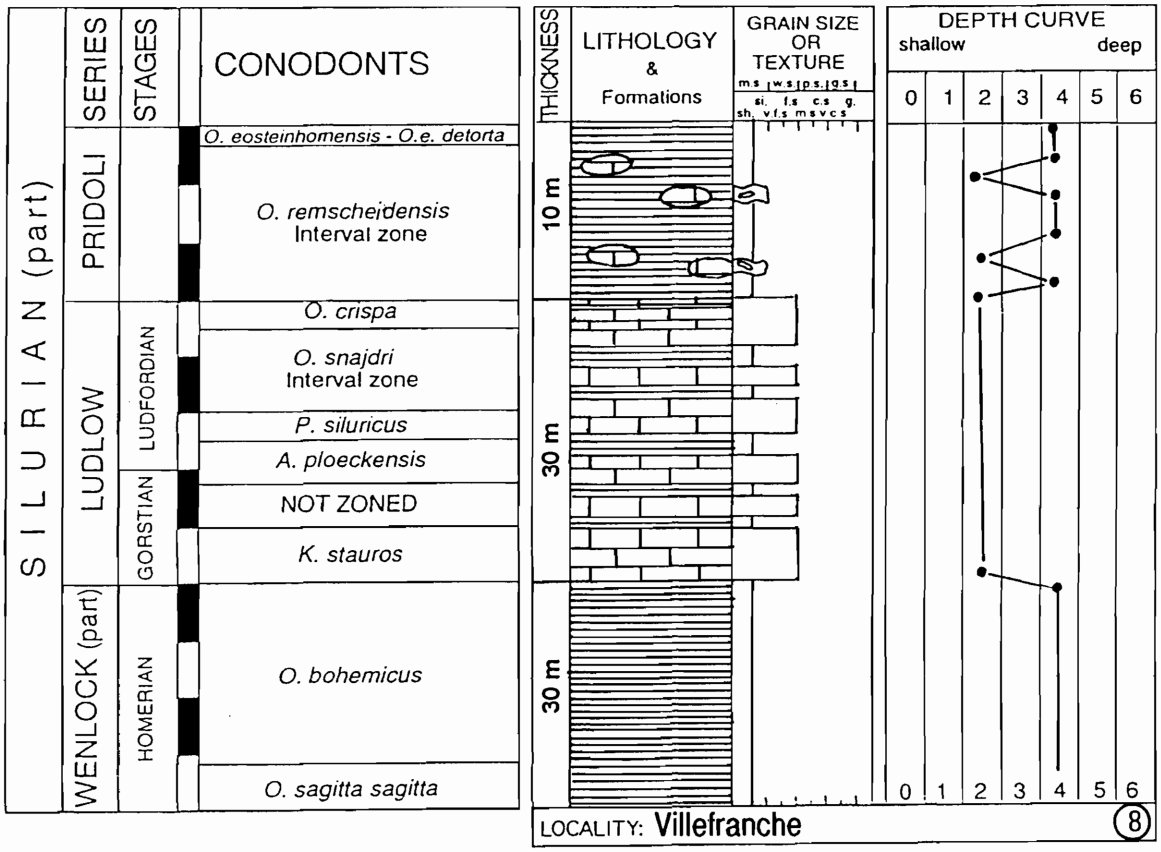

FIGURE 13 — Silurian of Villefranche. For location, see Fig. 7 (locality 8).

complex, with numerous small emergent areas in Brittany and the northwestern Iberian Peninsula, as suggested by Babin et al. (1980), who envisioned a number of islands which could have been the source of the coarser siliciclastics that are common in the Silurian of the Pyrenees.

Silurian of France

The main Silurian outcrops in France (Fig. 15) are in the Variscan massifs (i.e., Armorican Massif and Massif Central) and in the old Variscan cores of more recent orogens (i.e., the Pyrenees). More geographically restricted relics that are tentatively referred to the Silurian are in the Vosges, Alps, Maures Massif, and Corsica. Additional Silurian sequences are documented in the subsurface of the Aquitaine Basin and northern France (Artois and Boulonnais). All these Silurian strata are folded, and have suffered anchizonal to epizonal metamorphism, or even higher grades of metamorphism. The main features of these sedimentary and metamorphic rocks are given below, except those of the Pyrenees, which are documented above.

With the unique exception of the Artois sequence, which is regarded as part of the southern edge (eastern Avalonia) of the Baltica plate during the Silurian, the Silurian of France was deposited on the northern margin of Gondwana (Paris and Robardet, 1990). It shares, therefore, most of the basic sedimentological and climatic features of this huge paleogeographic province.

Northern France (Artois and Boulonnais) — Artois is located in northwestern France and lies in the French part of the Namur syncline, which extends into Belgium (Fig. 15). Additional Silurian subcrops (not documented herein) lie a few tens of kilometers west near the English Channel in the Boulonnais area (Fig. 15), where they are overlain disconformably by Middle Devonian conglomerates of the Caffiers Formation (Brice, 1988).

In the Artois area, several boreholes and coal pits expose the Upper Silurian part of the Angres Member of the Noulette Formation, as defined by Racheboeuf (1986). The formation, up to 200 meters thick, is part of the Pridoli and Lochkovian. It is a sequence of dark shales with local calcareous shales and/or limestone in the lower

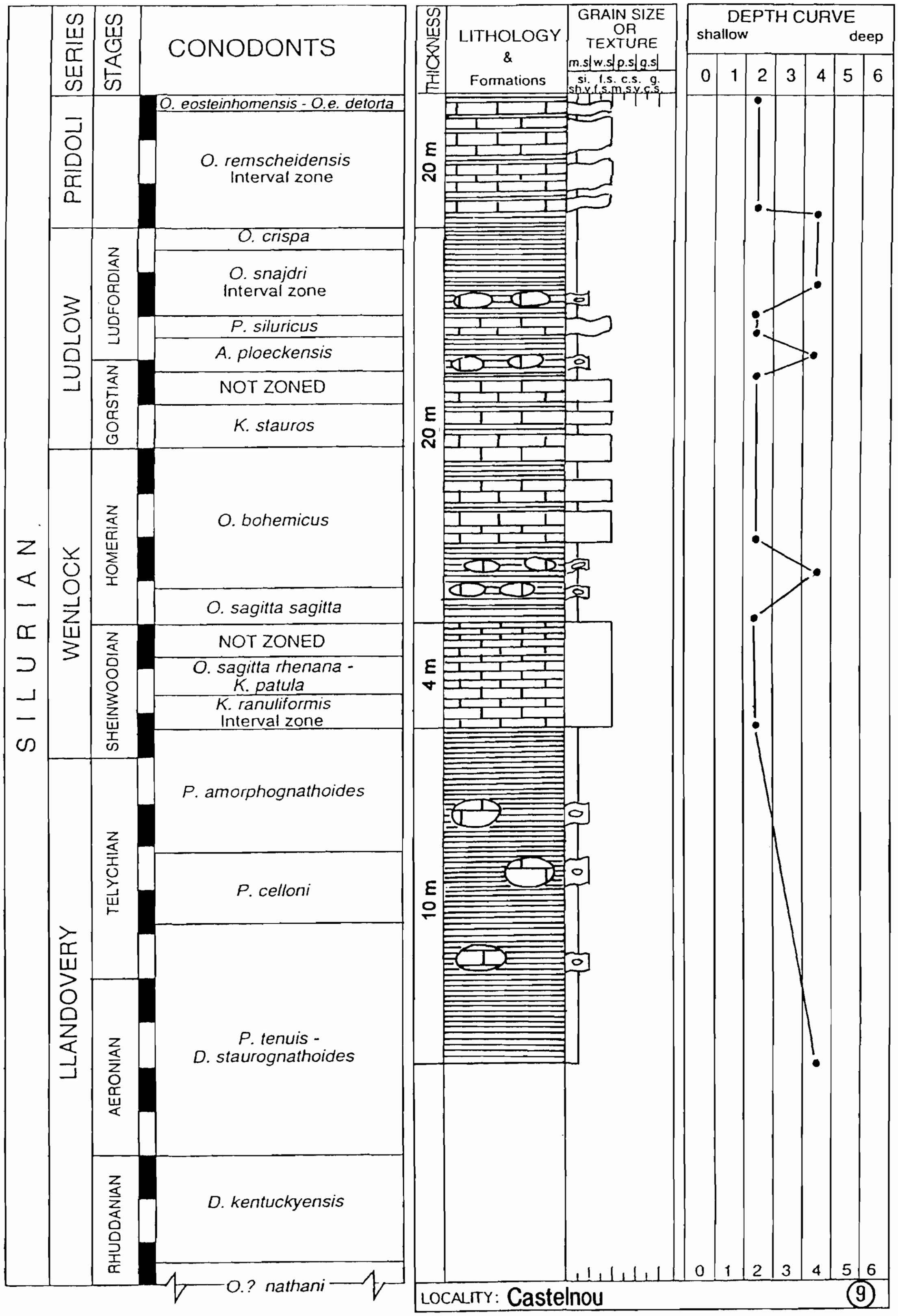

FIGURE 14 — Silurian of Castelnou. For location, see Fig. 7 (locality 9).

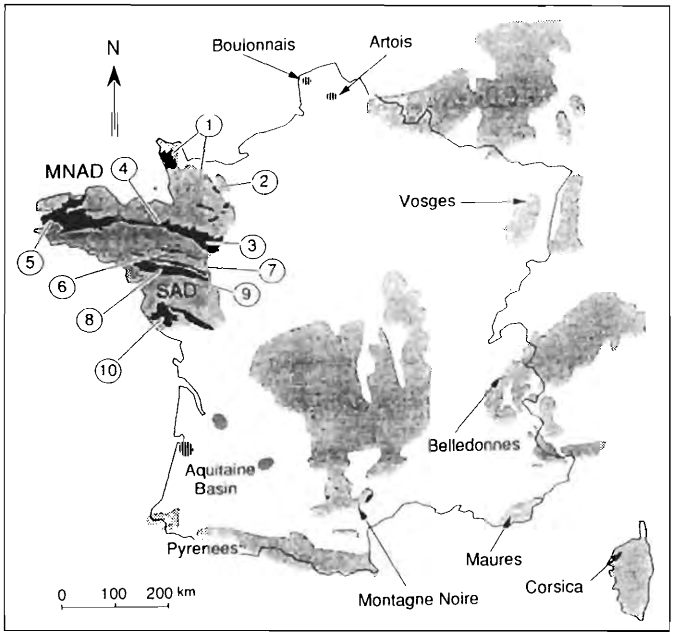

FIGURE 15 — Silurian of France. Variscan massifs (stippled); Silurian outcrops and subcrops (black); covered Silurian (vertical ruling). Armorican synclines and localities: *1*, Cotentin; *2*, Normandy; *3*, Laval; *4*, Ménez–Belair; *5*, Chateaulin; *6*, Martigné–Ferchaud; *7*, Angers; *8*, Candé and Saint-Georges-sur-Loire; *9*, Ancenis; 10, Vendée. Silurian of Mouthoumet and Pyrenées not indicated. MNAD is Medio–North Armorican Domain; SAD is South Armorican Domain.

Angres Member. The base of the formation is not known, and is always cut off by the Midi thrust fault (see Chalard, 1986).

The macrofauna of the Angres Member is fairly abundant and diverse. It includes numerous brachiopods, tentaculitids, bryozoans, bivalves, ostracodes, and trilobites. Microfossils, including conodonts, spores, acritarchs, and chitinozoans, also have been reported from the Silurian part of the Noulette Formation (see Racheboeuf, 1986).

The Artois and the Boulonnais areas are usually regarded as part of the southern edge of eastern Avalonia, and therefore belonged to Baltica in the Silurian. The Late Silurian map of Scotese and McKerrow (1990) locates the Artois and the Boulonnais at very low latitudes (ca. 15° S). However, no local paleomagnetic control is available.

No correlation chart is provided herein for the Artois area because its Silurian sequence is restricted to part of the Pridoli. The most diagnostic taxon is *Urnochitina urna*, a Pridoli chitinozoan reported from the lower and middle Angres Member, whereas *Eisenackitina bohemica*, a typical Lochkovian chitinozoan, has its lowest occurrence in the upper part of the member (Paris, 1986). This age is consistent with the occurrence of *Ozarkodina remscheidensis eosteinhornensis* Zone conodonts in the lower and middle Angres Member (Bultynck, 1986). No graptolites have been recorded in this succession.

Based on a detailed sedimentological study of the Noulette core drill (Pelhate, 1986), a shallowing trend is indicated from the base to the upper Noulette Formation, a unit which was deposited on the outer shelf. Evidence of numerous distal storm deposits is documented in the Angres Member. The Noulette Formation was deposited in a poorly oxygenated, but not anoxic, environment. The Noulette Formation has, especially in the upper Mericourt Member (Lochkovian), several condensed levels with phosphatic nodules.

The faunal associations are consistent with an outer-shelf environment for the Upper Silurian part of the Noulette Formation. This shelf was likely on the southern edge of the eastern Avalonian part of Baltica, as documented by paleobiogeographic affinities. Brachiopods and vertebrates, which are found mainly in the Lochkovian part of the Noulette, have close faunal relationships with Baltica, and especially with Podolia (Ukraine) (Racheboeuf and Babin, 1986). Less obvious paleobiogeographic affinities with Armorican, Iberian, and North African faunas that correspond to the northern Gondwana Province are consistent with the existence of a closing, but still fairly wide, Rheic Ocean in the Pridoli (Paris and Robardet, 1990).

Well-developed limestones in the Angres Member indicate a fairly low latitudinal location for northern France during the Pridoli. Such a location is consistent with a rather warm environment (F. Paris, unpublished data, 1998), and similar to the climate that prevailed in Podolia during the Pridoli However, no reefs developed in Artois.

ARMORICAN MASSIF — Silurian deposits are widely distributed in the Armorican Massif (Fig. 15), and likely extend under the Mesozoic cover of the Paris Basin and offshore to the west. Two main paleogeographic units are usually identified in the Armorican Massif; these are the Medio-North Armorican Domain (MNAD) and the South Armorican Domain (SAD) (sensu Paris and Robardet, 1994). The MNAD includes the Normandy, Cotentin, and the northern and central Brittany synclines (i.e., Chateaulin, Ménez-Bélair, Laval, and Martigné-Ferchaud synclines). The northern limb of the Angers syncline, located north of the northern branch of the South Armoican Shear Zone (i.e., Malestroit-Angers Fault) also belongs to the MNAD.

The SAD includes the southern Brittany (Candé anticline and Saint-Georges-sur-Loire syncline), as well as the Ligerian (Ancenis syncline) and Vendean (Vendée coast and Bas Bocage) terranes. The Medio-North and the South Armorican Domains correspond to northern

Gondwana terranes, but their original location on the northern margin of this paleocontinent is still unknown. However, obvious lithologic and geodynamic differences exist between the Silurian deposits of these two domains, which are discussed below.

Medio-North Armorican Domain (MNAD) — The Silurian of the MNAD is usually transgressive on latest Ordovician glacio-marine deposits formed with melting of the late Ashgillian ice cap that covered the North African part of Gondwana (see Paris et al., 1995). A hiatus of various local durations is documented above the Ordovician units (Figs. 16, 17). This gap is likely due to an absence of deposition, but Variscan tectonic disturbance may also be implicated, as Silurian black shales frequently acted as a décollement surface. The hiatus corresponds to part of the Llandovery (Rhuddanian to lower Telychian) in the Ménez-Bélair and Martigné-Ferchaud synclines (Fig. 15, localities 4, 6). It may extend into the early Wenlock in western Brittany and into the Laval syncline (Fig. 15, locality 3). However, local (central Chateaulin syncline and Cotentin; Fig. 15, locality 5) poorly preserved graptolites of early Rhuddanian affinities are found (Robardet, 1970; F. Paris, unpublished data, 1996). No gaps or disconformities have been documented at the Silurian–Devonian boundary in the MNAD.

With only local exceptions (e.g., the Wenlock Feuguerolles Limestone in Normandy and calcareous nodules or lenses in the Upper Silurian of the Chateaulin and Laval synclines), the MNAD Silurian is exclusively siliciclastic. It displays anoxic or strongly dysaerobic features that become less prominent in the Pridoli. Significant enrichment in such trace elements as vanadium (up to 5,600 ppm) are noted in black shales (Debard and Paris, 1986) that have TOC values ranging from ca. 10–40%.

At the base of the MNAD succession are pyritic and/or siliceous sandstones with interbedded graptolitic black shales. This type of sandy sequence is known from the Cotentin ("gres culminant" and lower Saint-Sauveur-le-Vicomte Formation, a few tens of meters thick), the Ménez-Bélair syncline (lower member of La Lande Murée Formation, up to 15 m), and the Martigné-Ferchand syncline (Poligné Formation with a prominent sandy lower member, or "gres culminant," ca. 80 m) (Fig. 15, localities 1, 4, 6). At other localities, similar sandstones are missing, either because of a sedimentary gap or tectonic disturbance (i.e., Crozon Peninsula in the Laval syncline and western Chateaulin syncline; Fig. 15, localities 3, 5). These sandstones are replaced by siltstones in eastern Normandy ("Schistes a Fucoides" Formation in the Caen area, ca. 40 m). Other examples of anoxic, condensed rocks developed in the Wenlock (e.g., laminated black shales of the middle member of La Lande Murée Formation or Veniec Member of La Tavelle Formation; a few meters thick); during the Ludlow and early Pridoli in the Chateaulin syncline (La Tavelle Formation, ca. 50 m, and Lostmarc'h Formation, 90 m); in the Menéz-Bélair and Laval synclines (upper member of the La Lande Murée Formation, up to 50 m); and in the Cotentin and Normandy areas (respectively, the Saint-Sauveur-le-Vicomte Formation, about 100 m, and Quesnay Formation, close to 200 m, but extending into the Pridoli). Several hiatuses probably occur within these sequences because only a few graptolite zones have been documented (Philippot, 1950; Jaeger et al., 1967; Paris et al., 1980).

By the early Pridoli or slightly later, depending upon the area (e.g., the northern Armorican Massif), the marine environment became normally oxygenated, as indicated by the litho- and biofacies. By that time, the subsidence rate increased drastically and allowed formation of thick sequences of alternating siltstones and sandy beds (Figs. 16, 17). These sequences correspond to the Plougastel Formation (at least 250 m of Pridoli) in the Chateaulin syncline and the Val Formation (ca. 200 m) and lower Gahard Formation in the Menéz-Bélair (ca. 250 m of uppermost Pridoli; F. Paris, unpublished data, 1998) and Laval synclines. In the Cotentin area, normal marine siliciclastic sedimentation is represented by the lower Saint-Germain-sur-Ay Formation, whereas in Normandy anoxic conditions persisted into the later Pridoli, which is the youngest Paleozoic in this area (Jaeger et al., 1967).

South Armorican Domain (SAD) — Anoxic or highly dysaerobic Silurian sequences of the SAD display fairly distal characteristics and have black cherts. In some areas, the cherts occur as olistoliths within a Carboniferous matrix (Dubreuil, 1986; Colchen and Poncet, 1989). At other localities, the Silurian includes volcaniclastics (e.g., in the Vendée; Peucat et al., 1986) or shales and siltstones interbedded with limestones, as in the Ancenis syncline (Cavet et al., 1971). A less common facies is the condensed black limestone and calcareous mudstone of the La Meignanne Formation (<20 m) that crops out on the northeast limb of the Candé anticline. These calcareous sedimentary rocks contrast sharply with the siliciclastic Silurian exposed nearby in the suburb of Angers, immediately north of the MNAD–SAD boundary (see Kříž and Paris, 1982).

Because of its tectonized setting (i.e., the different elements are not in stratigraphic continuity and are likely composed of exotic blocks), most of the SAD Silurian cannot be set into a stratigraphic succession. In addition, no formal formations have been designated, and sedimentologic data are not available.

No indisputable paleomagnetic data are available for the Silurian of the Armoricain Massif. However, on the Scotese and McKerrow (1990) maps, the Armorican Massif is regarded as a single terrane that shifted from 45° S to

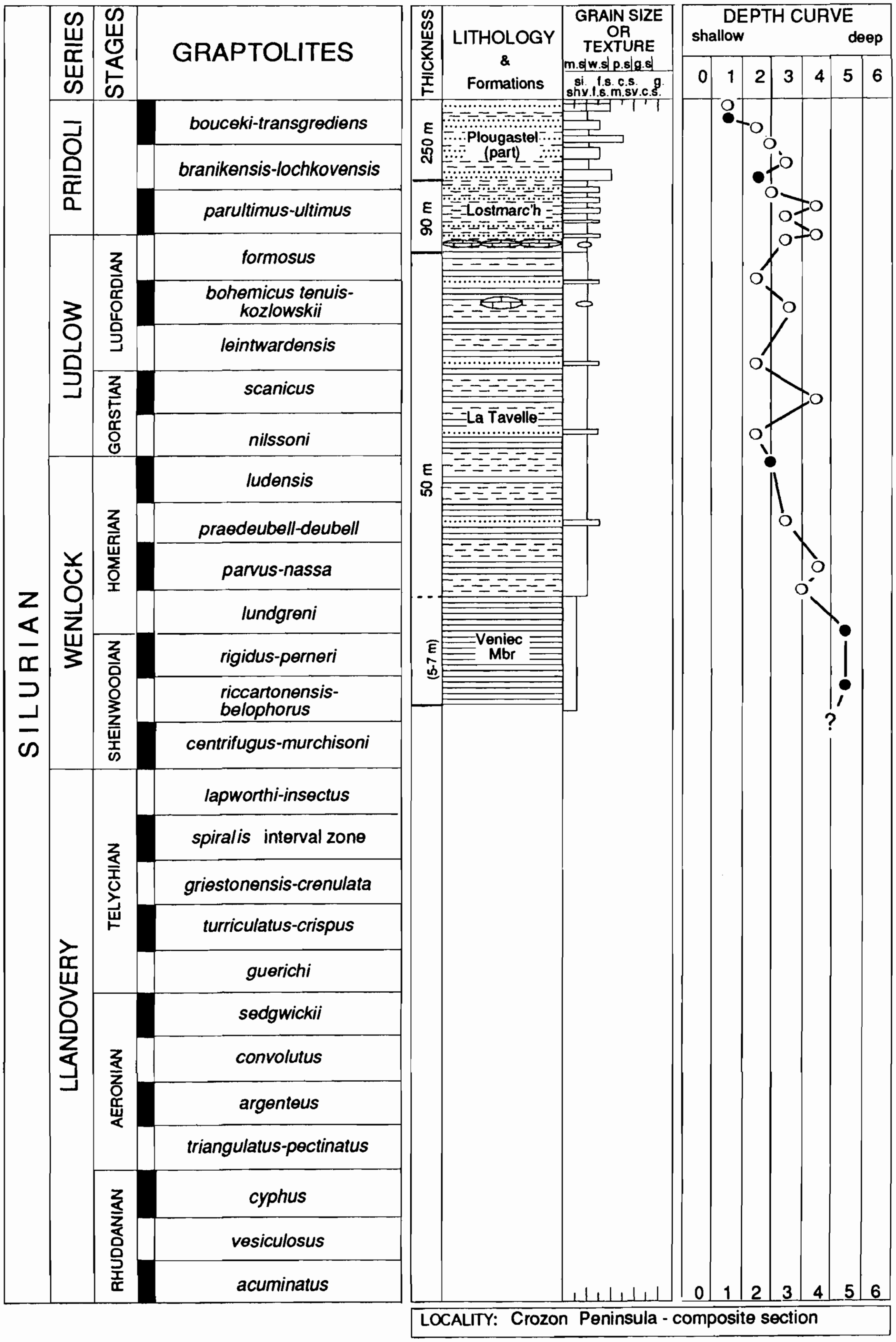

FIGURE 16 — Silurian of western Chauteaulin syncline (Crozon Peninsula, western Brittany), France.

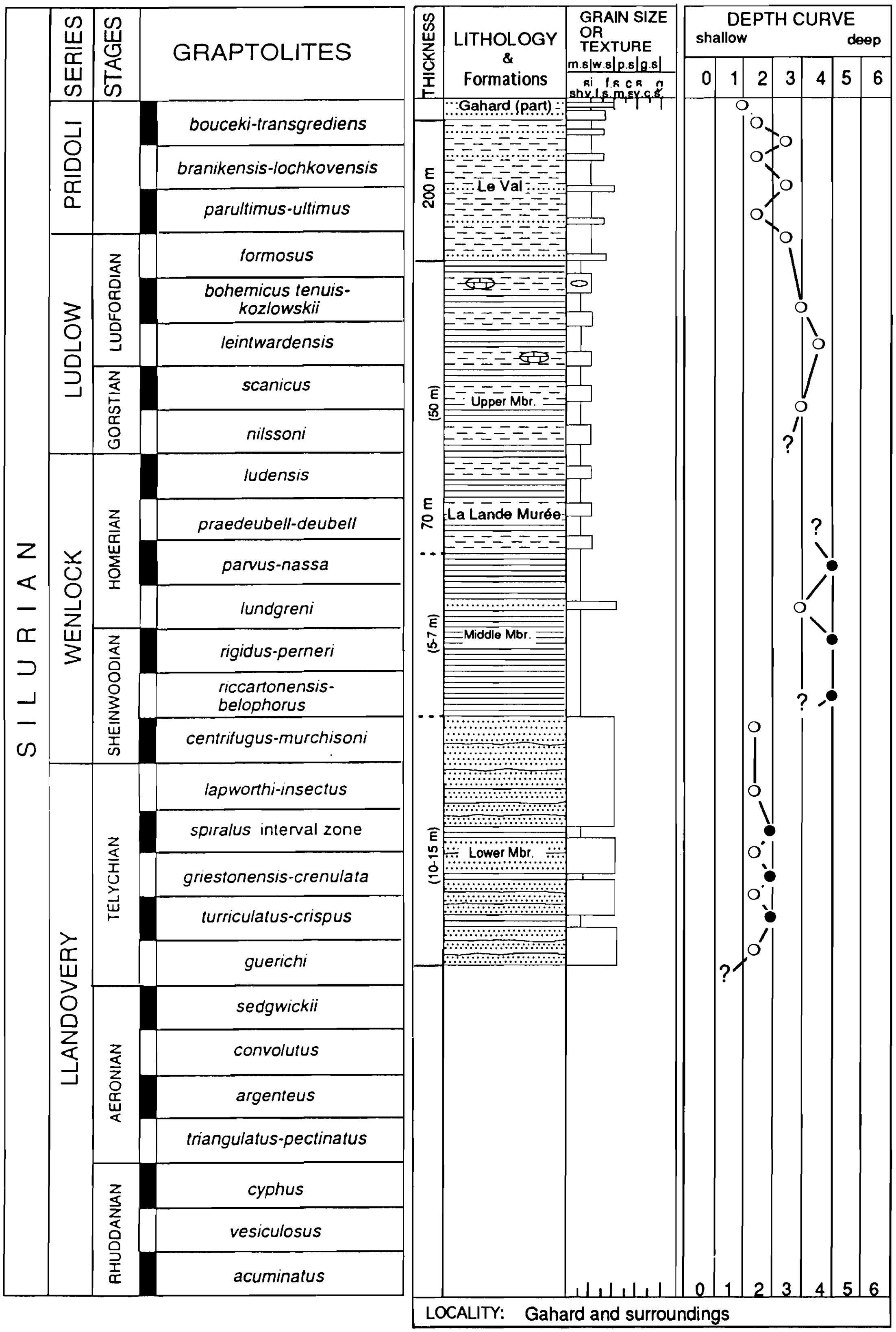

FIGURE 17 — Silurian of Gahard and surrounding area, Ménez–Belair syncline, France.

35° S between the Early and Late Silurian. Therefore, a temperate to fairly warm climate was probable in the area during the Silurian.

CORRELATION CHARTS — Two correlation charts are proposed for the MNAD. One deals with the Crozon Peninsula of western Brittany (Fig. 16), and the other with the central part of the median synclinorium (i.e., the Ménez-Bélair syncline in the area of Gahard; Fig. 17). Because of the discontinuous character of the succession in the olistolith-bearing sequences of the SAD, no correlation chart is given for this domain.

MEDIO-NORTH ARMORICAN DOMAIN (MNAD) CORRELATIONS — Limited biostratigraphic information is available because of the local occurrences of diagnostic fossils in the MNAD. Graptolites are the most useful fossils in the local Llandovery and Wenlock (Figs. 16, 17), but chitinozoans are also useful in the Ludlow and Pridoli.

The best-documented Telychian graptolite record is from the lower member of the Lande Murée Formation, where the *Spirograptus turriculatus-Monograptus crispus* and *Monoclimacis griestonensis-M. crenulata* Zones have been identified (Paris et al., 1980). The Sheinwoodian *Monograptus riccartonensis-M. belophorus* Zone is widely documented in the Chateaulin, Ménez-Bélair, and Laval synclines. The lower Homerian is probably present because *Cyrtograptus lundgreni* Zone graptolites have been reported from various localities (Philippot, 1950). Gorstian graptolites of the *Neodiversograptus nilssoni* and *Lobograptus scanicus* Zones have been identified in the Chateaulin and Laval synclines, where the Ludfordian *Bohemograptus bohemicus tenuis-Neocuculograptus kozlowskii* Zone is likely present in the upper member of the Lande Murée Formation (F. Paris, unpublished data, 1996). In the Cotentin and Normandy areas, dark shales of the upper Saint-Sauveur-le-Vicomte Formation and the Quesnay Formation yield lower Pridoli graptolites (*Monograptus parultimus-M. ultimus* Zone) (Jaeger et al., 1967; Roardet, 1970). Some or all of the chitinozoan zones defined recently in the Pridoli (Verniers et al., 1995) have also been identified in the Le Val and Gahard Formations (Paris, 1981; F. Paris, unpublished data, 1998), the lower Saint-Germain-sur-Ay Formation (Rauscher and Robardet, 1975), and the Lostmarc'h Formation (Paris, 1979). These chitinozoan data agree with a Pridoli age (*Ozarkodina remscheidensis eosteinhornensis* Zone) indicated by conodonts (Racheboeuf, 1979).

SOUTH ARMORICAN DOMAIN (SAD) CORRELATIONS — Diagnostic fossils are usually sparse in the SAD Silurian. At some localities (Candé anticline and Ancenis syncline; Fig. 15, localities 8, 9), black cherts have yield abundant lower Telychian graptolites (*Monograptus, Rastrites, Climacograptus, Petalograptus*) (Cavet et al., 1971; Lardeux and Cavet, 1994). This fauna, which needs taxonomic revision, likely corresponds to the lower *Spirograptus turriculatus-Monograptus crispus* Zone. In other localities in the Candé unit, late Telychian–earliest Sheinwoodian graptolites have been identified in Le Houx black shale (Cavet et al., 1986). Gorstian graptolites (*Neodiversograptus nilssoni* Zone) are reported from shales and limestones exposed near Chalonnes (Ancenis syncline) (Cavet et al., 1971; Lardeux and Cavet, 1994). The most precise biostratigraphic data on the SAD Silurian are given by the graptolites, bivalves, ostracodes, and chitinozoans recovered in the La Meignanne Formation (Kříž and Paris, 1982). This formation extends as high as the Lochkovian. The oldest known strata are ostracode-bearing Ludlow limestones with *Entomozoe (Richteria) migrans*, and they are overlain by black shale with numerous Pridoli bivalves (e.g., *Cheiopteria bridgei* and *Snoopia insolita* Communities with fairly abundant infaunal elements; Kříž and Paris, 1982). This age assignment is confirmed by the presence of *Urnochitina urna*, a chitinozoan which had an acme in the late Pridoli (Paris, 1981).

In the Vendée, modern paleontological data are absent for the Silurian. Poorly preserved chitinozoans, mazuelloids (phosphatic and organic-walled microfossils), and a few graptolites are known from black cherts and/or phosphatic nodules at localities along the Vendean coast (Ters, 1979; Le Hérissé et al., 1991) and from the Bas Bocage (Chalet et al., 1983). In the latter area, a U-Pb age of 405±5 m.y. was obtained (Peucat et al., 1986) on the Mareuil-sur-Lay Formation (ignimbrites and interbedded black cherts).

PATTERNS IN LITHOSTRATIGRAPHIC CHANGES — During the Silurian, sedimentation in the MNAD recorded a number of successive events: 1) eustatic rise resulting from melting of the northern Gondwanan ice cap, and Llandovery anoxia possibly related to this melting; 2) a phase of very restricted siliciclastic input (late Llandovery and Wenlock); 3) a slight increase in subsidence and a slow reduction in anoxia during the Ludlow; and 4) regional shallowing despite a rapid increase in subsidence (Figs. 16, 17).

The lowest Silurian sandy deposits represent high-energy environments, whereas the interbedded black shales correspond to very quiet, deeper environments (but probably not very deep, as suggested by the occurrence of eurypterids) (Figs. 16, 17). Sedimentologic evidence, with exception of a distal storm deposit observed in the Ménez-Bélair syncline (Fig. 17), cannot give a precise depth control for the Wenlock black shale environments. These laminated organic-rich shales were likely deposited in a fairly deep and quiet environment with only pelagic faunas and palynomorphs (Figs. 16, 17). Temporary breaks in sedimentation probably occurred in this condensed sequence. A shallowing trend took place during the Ludlow, and distal storm deposits and even ripples, indicative of normal wave-base, are recorded in the siltstones and shales of the upper Tavelle Formation (Fig. 16). This shallowing persisted during the Pridoli, as shown by a *Salopina-Clarkeia* Community and the occurrence of numerous ripples in the thin-bedded sandstones, siltstones, and dark shales of the Lostmarc'h Formation (Fig. 16). At the same time, a fairly similar environment, with a record of distal storm waves and ripples, prevailed in the east during deposition of the Val Formation (Fig. 17). Abundant endichnial and epichnial trace fossils in the Plougastel Formation show the end of an oxygen-deficient environment. This shallowing trend ended in the Lochkovian (Guillocheau and Rolet, 1982) with the advent of deltaic sedimentation at some localities (e.g., Landevennec Formation of western Brittany). Nearby emergent areas, possibly located northward in terms of modern geography (but different from the "Old Red Continent") provided land-derived palynomorphs (spores and tracheids), which are abundant close to the Pridoli–Lochkovian boundary (Deunff and Chateauneuf, 1976; F. Paris, unpublished data, 1998).

PATTERNS IN BIOFACIES CHANGES — The anoxic Llandovery and Wenlock rocks yield pelagic or epipelagic faunas, including graptolites (*Monograptus, Pristiograptus, Cyrtograptus, Petalograptus, Retiolites* as the most common genera; Philippot, 1950; Jaeger et al., 1967; Paris et al., 1980), a few thin-walled epifaunal brachiopods, and fairly abundant palynomorphs (chitinozoans, acritarchs, leiospheres). Eurypterid remains (e.g., Megalograplidae) must be stressed because these fossils usually indicate fairly shallow, near-shore environments of the Eurypterid Community. Anoxic conditions became progressively less severe during the Ludlow, and graptolites (e.g., *Saetograptus, Bohemograptus*), bivalves (Cardiolidae), myodocopid ostracodes (e.g., *Bolbozoe*), and orthocone cephalopods were typical parts of the fauna. They are frequently preserved in well-bedded nodules ("sphaeroids"). Later in the Pridoli, the normally oxygenated environments of the Lostmarc'h (uppermost part), Plougastel, Le Val, Gahard, and Saint-Germain-sur-Ay Formations supported brachiopods, crinoids, a few trilobites (Homalonotidae), ceratiocarids, bivalves, orthocone cephalopods, and very abundant and diverse palynomorphs (acritarchs, spores, a few tracheids), chitinozoans, and scolecodonts (Rauscher and Robardet, 1975; Deunff and Chateauneuf, 1976; Paris, 1981). The macrofossils and most of the palynomorphs are indicative of fairly shallow marine, near-shore environments.

Adequate sedimentological and faunal data for an accurate reconstruction of the SAD marine environment during the Silurian are mostly absent. However, the common record of mazuelloids and possible radiolarians in the black cherts of the Vendée (see references in LeHérissé et al., 1991) suggests a rather deep environment. On the other hand, the depositional environment at La Meignanne was deep enough that it was not obviously affected by the Pridoli–early Lochkovian regression.

CLIMATIC INDICATORS AND VARIATIONS — The occurrence of a cold-water *Clarkeia* fauna and the lack of important limestones suggest that MNAD was located at high to moderate latitudes (Paris and Robardet, 1990). Therefore, the successive latitudinal positions proposed by Scotese and McKerrow (1990) for the Armorican Massif for the Early–Late Silurian seem to be acceptable. However, the black micrites at La Meignanne in the SAD (see Kříž and Paris, 1982) suggest more moderate to low latitudinal locations of this area with regard to the MNAD regions during the Ludlow and Pridoli. Moreover, faunal affinities and lithologic features suggest that a closely related paleogeographic position of the La Meignanne area, Sardinia, Bohemia, Carnic Alps, and the Montagne Noire is likely (Kříž, 1996, 1999; Kříž and Serpagli, 1993; Paris, in press).

MONTAGNE NOIRE — With the exception of a local occurrence of black carbonates and shales in the south of the Albigeois, Silurian strata are only recorded in the southeast Montagne Noire (Fig. 15), in the Falgairas and Laurens areas of the Cabrières klippes (Feist and Echtler, 1994). In the rest of the Montagne Noire (or the "nappes" domain), the Lower Cambrian to Lower Ordovician are unconformably overlain by the lowermost Devonian. The complete Silurian sequence of the Cabrières klippes (Chaubet, 1937) is interpreted to be part of a huge olistolith in a Visean matrix (Engel et al., 1982). This tectonic association is mainly developed in the southeastern Montagne Noire.

From the late Llandovery–early Wenlock, the local sedimentary rocks include condensed black shales and subordinate limestone nodules ("Roquemaillère black shales," ca. 5,070 m; Fig. 18). These shales are overlain by Wenlock platy to nodular limestones with minor calcareous black shales (Fig. 18). The Ludlow is reduced to less than 5 m of black calcareous mudstones with calcareous nodules and a few bedded or nodular limestones (Fig. 18). The Pridoli is much thicker (Feist, 1977; De Bock, 1982). In the lower Pridoli, black shales are progressively replaced upward by dolostones and limestones. In the upper part of this black shale and pelagic carbonate sequence ("Falgairas shales and limestones," ca. 5,060 m), the sandy limestones pass up into calcareous sandstones and sandstones (Falgairas Sandstone, 9 m), with shallowing and increased siliciclastic input (Feist, 1977).

CORRELATION CHART OF MONTGNE NOIRE — A composite chart of the Roquemaillère and Falgairas Silurian is given in Fig. 18. Because a detailed sequence stratigraphy

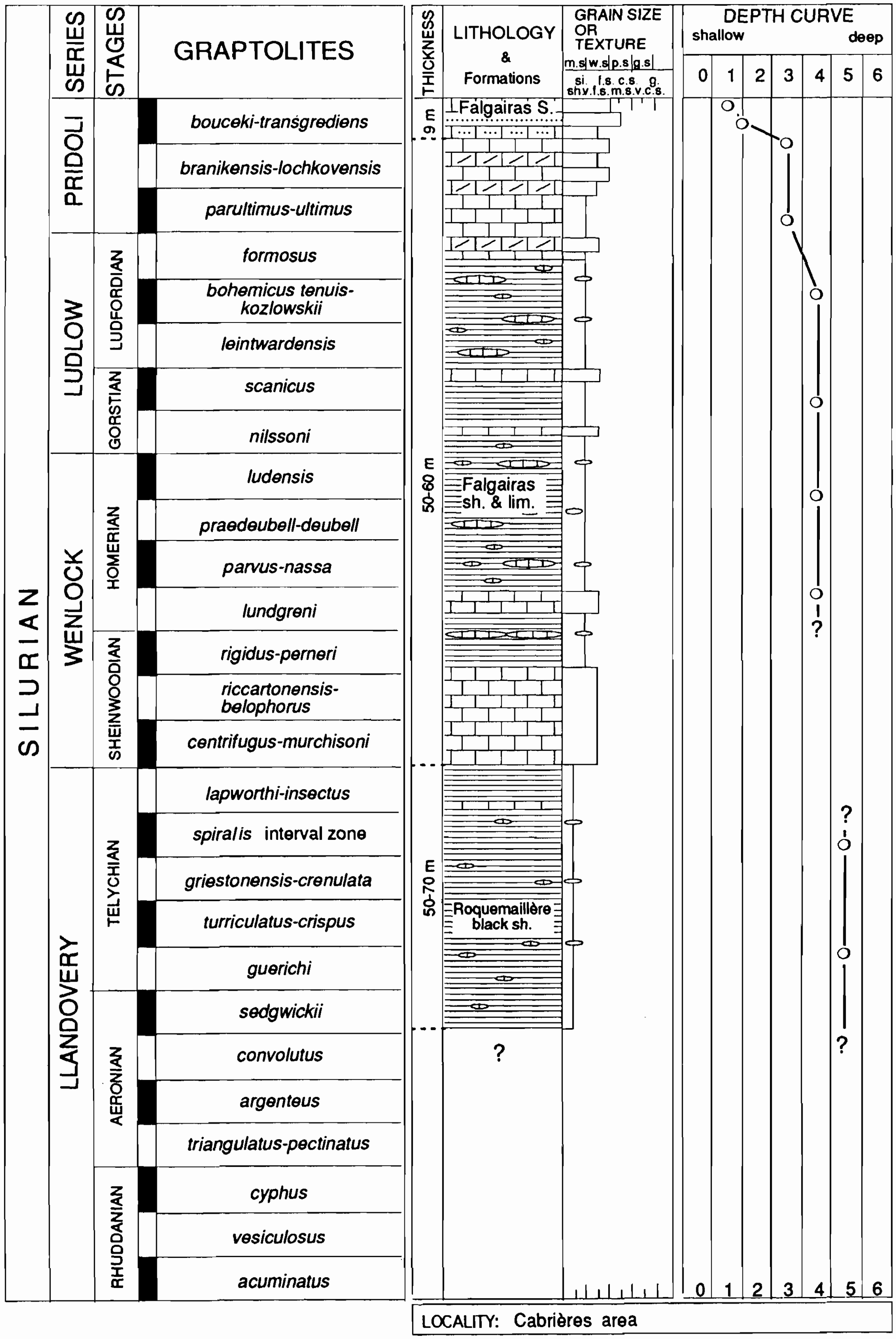

FIGURE 18 — Silurian of Cabrières area (southeast Montagne Noire), France.

is not available on these sequences, the proposed depth curve is tentative and generalized.

In addition to an abundant and diverse macrofauna with trilobites, brachiopods, bivalves, ostracodes, and crinoids (Feist, 1977), biostratigraphically diagnostic graptolites of the *Stimulograptus sedgwickii, Neodiversograptus nilssoni, Lobograptus scanicus, Bohemograptus bohemicus*, and *Monograptus ultimus* Zones have been recorded in this area (Centene and Sentou, 1975). The occurrence of a fairly diverse trilobite fauna in the Wenlock, and less diverse faunas in the Ludlow and Pridoli (Feist, 1977), has to be stressed because trilobites are virtually absent in other Silurian strata of France. Bivalves (see Kříž, 1996) represent an important group for detailed correlation with other regions of Gondwana and Perunica. Kříž (1996) described 41 bivalve species known from the Prague Basin from the Wenlock, Ludlow, and Pridoli of the Montagne Noire, and recognized five distinct bivalve-dominated communities also known from the Prague Basin, Sardinia, and the Carnic Alps (Austria, Italy). The bivalves show that the Silurian of the Montagne Noire has the closest faunal relationship with the Silurian of Sardinia. Three communities known from Sardinia occur in the Montagne Noire (Kříž and Serpagli, 1993). These are the *Cardiola figusi* (Wenlock–Ludlow boundary), *Cardiola donigala* (lower Ludlow), and *Cardiolinka sardiniana* (lower Přídolí) Communities. Diagnostic conodonts have also been recorded from the Telychian to the uppermost Pridoli in the Falgairas area (Feist and Schönlaub, 1974; Centene and Sentou, 1975). A number of conodont zones have been identified (e.g., *Pterospathodus celloni, P. amorphognathoides, Kockelella patula, Ozarkodina sagitta, Ancoradella ploeckensis, Polygnathoides siluricus, Ozarkodina snajdri, O. crispa, and O. remscheidensis eosteinhornensis* Zones; Centene and Sentou, 1975; Feist, 1977). Acritarchs and chitinozoans have been reported by Deflandre (1942) from the "Roquemaillre Limestone" (probably Wenlock). The distribution of *Urnochitina urna* and *Eisenackitina bohemica* — chitinozoans diagnostic, respectively, of the Pridoli and Lochkovian — allowed De Bock (1982) to locate the Silurian–Devonian boundary fairly precisely in the Falgairas and Laurens areas.

LITHOSTRATIGRAPHY AND BIOFACIES OF THE MONTAGNE NOIRE — No modern sedimentologic investigations have been done on the Silurian of the Cabrières klippes. A low-energy environment shown by pelagic and epipelagic faunas and the generally condensed succession prevailed from the late Llandovery to early Pridoli. Bivalve-dominated communities (Kříž, 1996) show that the environment was episodically ventilated by currents to depths similar to those in other northern Gondwanan terranes. The first evidence of a significant shallowing is given by the input of detrital quartz, and later by the deposition of sandstones in the latest Pridoli ("Falgairas Sandstone") in a high-energy environment (Fig. 18).

The faunas indicate this shallowing by the predominance of benthic taxa during the Pridoli. At that time, icriodontid conodonts, which are usually regarded as shallow-water, replaced the deeper water ozarkodinid species (Feist, 1977).

CLIMATE — The major development of carbonates, especially in the Pridoli, indicates a moderate to fairly low latitudinal location for the Silurian of the Cabrières klippen. In addition, obvious paleogeographic similarities exist with other "Mediterranean" regions (e.g., Sardinia, Carnic Alps, eastern Pyrenees, Mouthoumet, Aquitaine basin, and the South Armorican Domain [La Meignanne]) (Kříž, 1996, 1999). These similarities suggest a northern Gondwanan margin location for these areas in the Silurian, but with a lower latitudinal position than the MNAD (Robardet et al., 1994, Paris, in press).

AQUITAINE BASIN — In the Aquitaine Basin (Fig. 15), the Silurian has been encountered in only a few bore holes through the Mesozoic. Llandovery graptolitic black shales, probably of the *Spirograptus turriculatus-Monograptus crispus* or *Monoclimacis griestonensis-Monoclimacis crenulata* Zones; and Pridoli black shales with a few mazuelloids and poorly preserved chitinozoans, have been observed in the Castelsarrasin well (Cs. 102 of Paris and Le Pochat, 1994). Other Silurian rocks occur in the Saint-Martin-du-Bois borehole (SMB.1). Tuffaceous sedimentary rocks from SMB. 1 yield *Spinachitina fragilis?*, the index of the lowest Rhuddanian chitinozoan zone. Reworking cannot be excluded, but the sequence continues with black shale, and dolostone appearing only as minute cuttings. This anoxic or strongly dysaerobic succession ends with sandy limestones with *Eisenackitina bohemica* (Paris and Le Pochat, 1994). This Lochkovian chitinozoan was reported by De Bock (1982) from similar rocks in the Montagne Noire. Northwest and close to the Bay of Biscay shoreline, the slightly metamorphosed, black laminated shales of the Le Teich Formation are tentatively referred to the Silurian on the basis of limited graptolite evidence (Monograptidae) (Paris and le Pochat, 1994). No paleoenvironmental interpretation is possible due to the lack of suitable core samples.

POSSIBLE SILURIAN ELSEWHERE IN FRANCE — In several areas, rocks referred to the Silurian are strongly affected by folding, cleavage, and a fairly high-grade metamorphism. Original sedimentological features are not preserved, and fossils are generally lacking. These possible Silurian sequences are briefly listed below from north to south.

Part of the greenish to purple Steige Slate that crops out in the northern Vosges Massif (Fig. 15) has been referred to the Silurian based on chitinozoans (Doubinger, 1963). However, these palynomorphs were too poorly preserved to allow a firm generic identification, and the Steige Slate is not discussed further.

In the Paleozoic basement of the Alps, black and green schists crop out in the western Belledonne Massif (see Ménot et al., 1994). No firm evidence of Silurian fossils has been reported from this material. In southeastern France, lenses of crinoidal limestone and a black shale of Middle Silurian age are reported from the western Maures Massif (Crevola and Pupin, 1994; Fig. 15). These sedimentary rocks are not discussed further because Variscan epizonal metamorphism and tectonics obscure their lithology and relationship to surrounding formations.

In northern Corsica (Fig. 15), the Silurian is recorded by the Monte Martinu Formation, which has dark slates in the Campo Orbu Member that yield chitinozoans and acritarchs regarded as Silurian (Baudelot et al., 1976). However, additional paleontological work is necessary to document a more precise age for these slates and for the overlying sandstones and black cherts of the Capu Russellu Member.

SILURIAN OF SARDINIA

Silurian rocks are exposed only in southern Sardinia (Fig. 19). Two distinct outcrop belts occur in the southwest (Iglesiente and Sulcis sub-regions) and southeast (Gerrei and Sarrabus sub-regions) of the island. They resemble the Silurian of Bohemia and Thuringia, respectively. Their mutual relationship is unclear, and this justifies their separate treatment and the use of distinct correlation charts (Figs. 21–24). Although formal lithostratigraphic units have been proposed for southwest Sardinia (Gnoli et al., 1990), informal names adopted from Thuringia and used mostly as facies indicators are still used in southeast Sardinia.

Strong tectonism which affected Sardinia largely explains the fact that a complete Silurian section is unknown on the island. This is reflected also in the definition of lithostratigraphic units, which are largely based on several discontinuous outcrops. The use of graptolites and conodonts as zonal fossils is necessary because of the peculiar lithologies in the exposures. Corradini and Serpagli (1998) proposed a conodont zonation for the upper Llandovery–Pridoli of Sardinia. Their scheme has been adopted in the discussion of this report, but the figures used herein follow the standard conodont-graptolite zonation (Silurian Times, No. 3, 1995), and used in all the reports in this bulletin. For the relationship between the two zonations, see Fig. 20.

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY — The Silurian of Sardinia begins with graptolitic sandy to siliceous shale, which are organic-rich and pyritic, especially in southeast Sardinia ("alum slates" of Jaeger, 1977b) and are interbedded with chert in the lowest part of the succession. The Genna Muxerru Formation of southwest Sardinia (2,025 m; Fig. 21) has Llandovery graptolites (*Parakidograptus acuminatus*, *Cystograptus vesiculosus*, *Coronograptus cyphus*, *Demirastrites triangulatus-D. pectinatus*, *D. convolutus*, *Spirograptus turriculatus-Monograptus crispus*, and *Monoclimacis griestoniensis-M. crenulata* Zones) (Gnoli et al., 1990; Štorch and Serpagli, 1993). Nine more species were reported by Rickards et al. (1995) from these zones. Graptolites of the *Cyrtograptus lapworthi* Zone have been discovered by P. Štorch (unpublished data, 1998). The lower and upper boundaries of the Genna Muxerru Formation are not well exposed, but appear to be gradational.

The "lower graptolitic shales" of southeast Sardinia (3,040 m; Fig. 22) represent a long time interval, and range upward to the Wenlock and lower Ludlow. An undisturbed lower contact of the formation with the Ordovician is not known. Chert, otherwise rare in southwest Sardinia, is well developed in the Llandovery. They extend up to the *S. turriculatus-M. crispus* Zone and are generally thick-bedded, frequently radiolarian-rich, and have thin shale interbeds. Phosphorites are present in the middle–upper part of the formation from the *Cyrtograptus lundgreni–Neodiversograptus nilssoni* Zones, where they occur as nodules, lenses, or beds (Barca and Jaeger, 1990). Llandovery graptolites of the *Cystograptus vesiculosus*, *Demirastrites triangulatus-D. pectinatus*, *Monograptus argenteus (Coronograptus gregarius)*, *Demirastrites convolutus*, *Spirograptus turriculatus-Monograptus crispus*, *Monoclimacis griestoniensis-M. crenulata*, and *Oktavites spiralis* Zones are known. This formation extends into the lower Ludlow, at least as high as the *N. nilssoni* Zone (Barca and Jaeger, 1990). The base of the overlying calcareous formation has *Ozarkodina excavata hamata* Zone conodonts (Corradini et al., 1998), and seems to be equivalent to the lower *Lobograptus scanicus* (graptolite) Zone.

The Fluminimaggiore Formation (Fig. 23) overlies the Genna Muxerru Formation in southwest Sardinia and roughly corresponds to the "calcari a *Orthoceras*, *Cardiola*, *Monograptus*" of early reports (e.g., Meneghini, 1881). The approximate thickness of the Fluminimaggiore Formation can be estimated only indirectly due to strong tectonism, but is ca. 4,550 m. The formation is upper Llandovery–lowest Lochkovian. Black, calcareous nodule-like structures, generally ellipsoidal and up to 1 m in size (Fig. 23) alternate with dark, non-calcareous shales. Plastic deformation and cleavage have strongly altered the shales, while the limestone bodies have well-preserved fossils (Gnoli et al., 1980). The black color and the peculiar bituminous smell indicate high organic matter con-

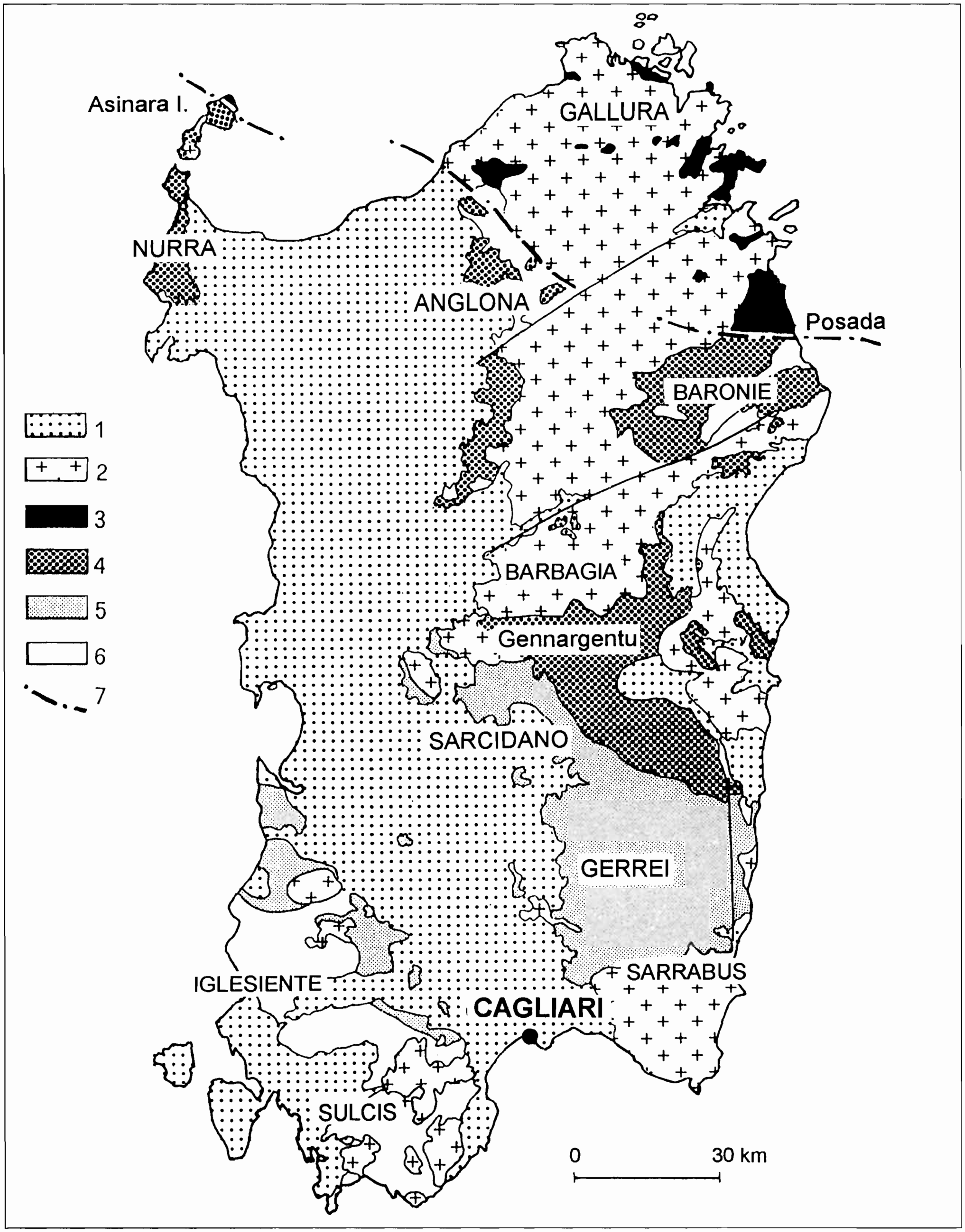

FIGURE 19 — Primary structural elements of Sardinia. *1*, post-Variscan cover; *2*, Variscan batholith; *3*, high-grade metamorphic complex; *4*, internal nappes; *5*, external nappes (including the Silurian); *6*, external zone; *7*, Posada–Asinara Line. After Carmignani et al. (1992).

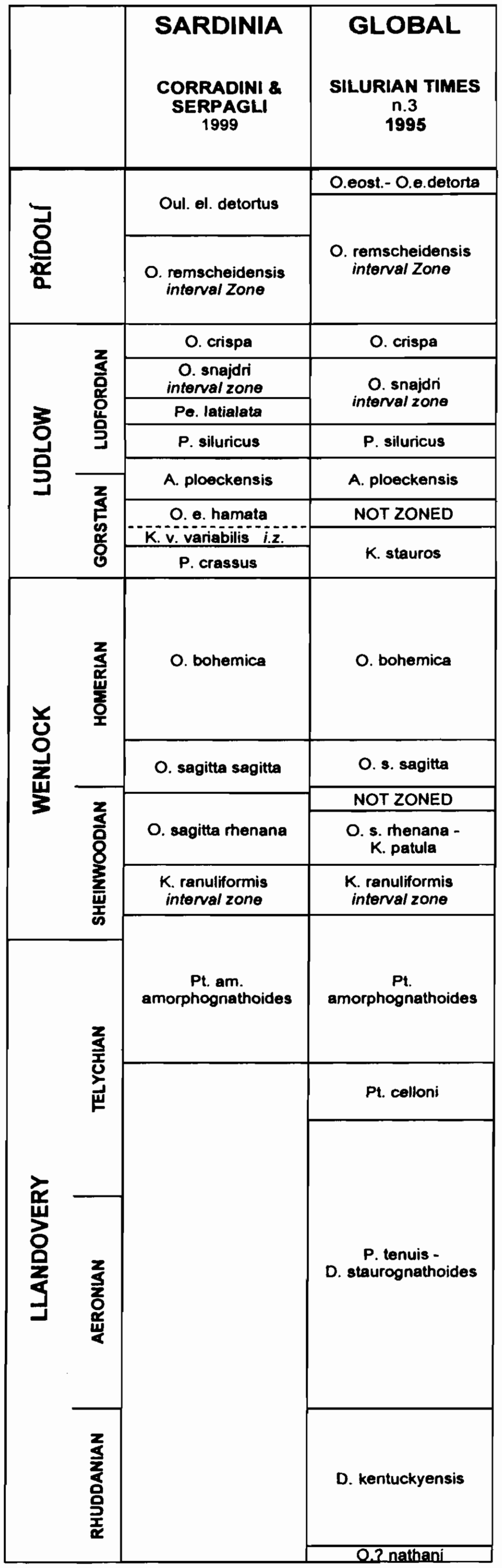

FIGURE 20 — Comparison between the the upper Llandovery–Pridoli conodont zonation of Sardinia and the standard zonation (after Corradini and Serpagli, in press).

tent. The fauna is dominated by cephalopods with associated bivalves, pelagic ostracodes, graptolites, conodonts, foraminiferans, chitinozoans, and muellerisphaerids. Gastropods, brachiopods, trilobites, and eurypterids are rare. Graptolites are frequently found packed together in the calcareous bodies. A single graptolite species, or a few at most, occurs in each calcareous body. Diverse *Saetograptus* or *Monograptus* species dominate the Wenlock–Ludlow and lower Pridoli associations (Ferretti and Serpagli, 1996b). Bivalves are the only important benthic forms, and almost no trilobites and brachiopods have been found. This suggests limited oxygenation, which could not be tolerated by these latter organisms. The carbonates are characterized by dominant fossiliferous wackestone–packstones that pass into sparse fossiliferous mudstones at the top of the formation. A crinoid packstone horizon with scyphocrinitids appears at the top of the Fluminimaggiore Formation. The *Pterospatodus amorphognathoides amorphognathoides, Kockelella ranuliformis, Ozarkodina sagitta sagitta, O. bohemica bohemica, Polygnathoides crassus, O. excavata hamata, Ancoradella ploeckensis, Polygnathoides siluricus, O. remscheidensis, Oulodus elegans detortus, Icriodius woschmidti woschmidti* and *I. woshmidti postwoshmidti* (conodont) Zones have been documented (Ferretti et al., 1998). Graptolites of the *Cyrtograptus lundgreni, Neodiversograptus nilssoni, Lobograptus scanicus, Saetograptus leintwardensis,* and *Monograptus parultimus-M. ultimus* Zones (H. Jaeger, personal commun., 1987; Rickards et al., 1995) have been reported from the limestones.

The limestones with cephalopods of the Fluminimaggiore Formation of southwest Sardinia have correlatives in southeast Sardinia (Gerrei sub-region) in the uppermost "lower graptolitic shales," in a calcareous unit ("Ockerkalk;" Fig. 24), and probably in the lowermost part of another shaly unit, the "upper graptolitic shales." The "Ockerkalk" (30 m), a blue-gray argillaceous limestone that weathers to an ochre color on which its name is based, is stylolitic. The fauna is composed largely of few nautiloids (Gnoli, 1993), with rare ostracodes, brachiopods, thin-shelled bivalves, trilobite fragments, gastropods, sponge spicules, phyllocarids (mainly mandibles), and crinoids. Trace fossils and very small solitary corals were reported from the "Ockerkalk" by Jaeger (1977b). These remains are scattered in a micritic matrix, and are locally concentrated in thin wackestone bands with disarticulated debris. Rich conodont faunas of the *Ozarkodina excavata hamata, Ancoradella ploeckensis, Polygnathoides siluricus, Pedavis latialata, Ozarkodina snajdri, O. crispa, O. remscheidensis,* and *Oulodus elegans detortus* Zones have been documented (Corradini et al., 1998). These conodont zones are in good agreement with the graptolite zones from the shales at the base and top of the

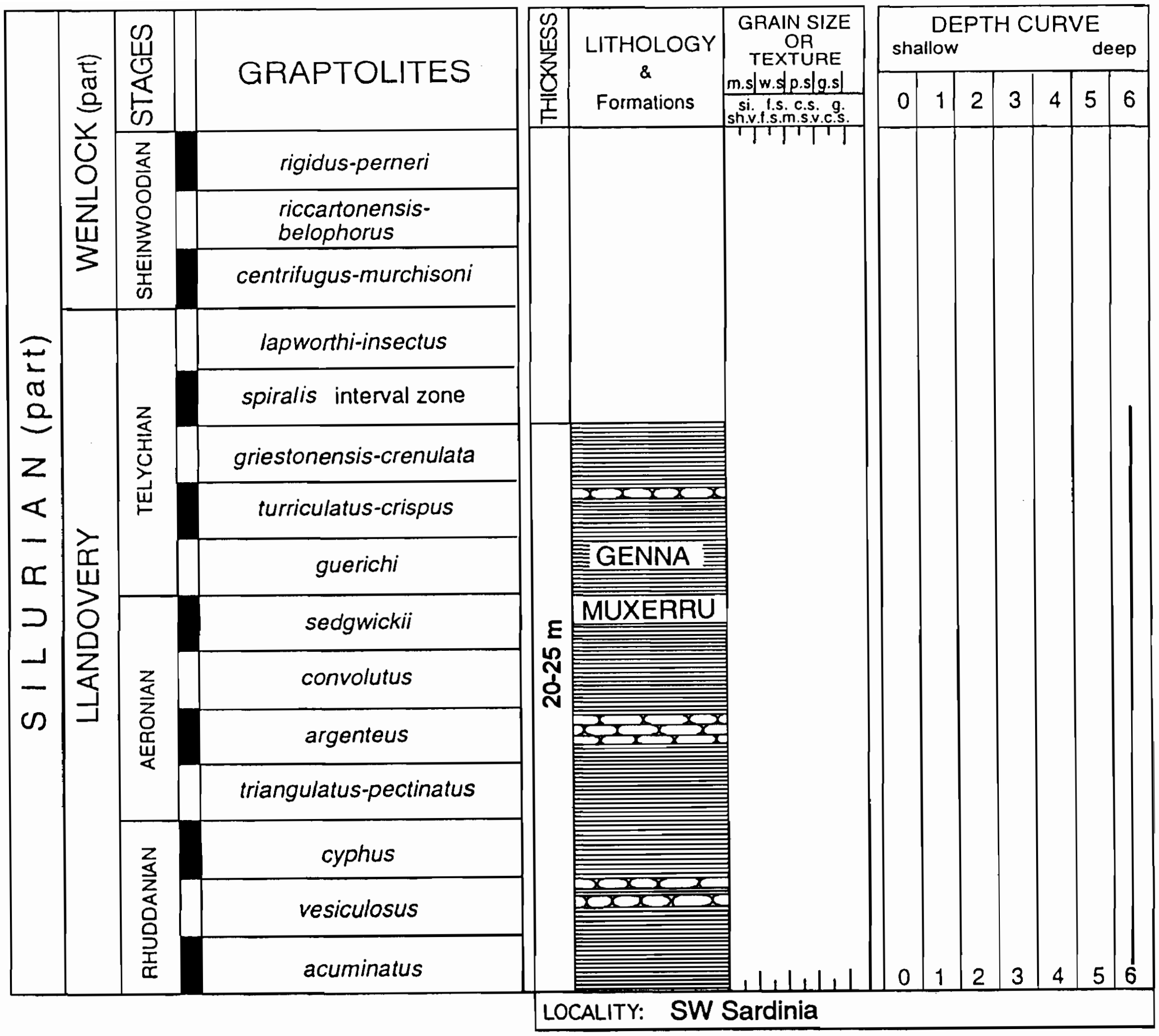

FIGURE 21 — Lower Silurian of southwest Sardinia.

unit (Jaeger, 1976). A lobolith horizon with the pelagic crinoid *Scyphocrinites* that is known along the northern Gondwana margin in the Silurian–Devonian boundary interval occurs in the Upper Silurian *Oulodus elegans detortus* Zone (Barca et al., 1995) of southeast Sardinia (Helmcke, 1973; Jaeger, 1976, 1977b; Barca and Jaeger, 1990). A similar lobolith horizon (*Camarocrinus?*) occurs in the uppermost Devonian of southwest Sardinia (Gnoli et al., 1988). In the Sarrabus sub-region of southeast Sardinia, Silurian and Devonian olistoliths and olistostromes are embedded in flysch-type rocks of probable Early Carboniferous age (Barca, 1991; Barca and Olivieri, 1991). Graptolitic black slates with interbedded cherts (Barca and Jaeger, 1990) and calcareous blocks (Barca et al., 1986; Barca and Olivieri, 1991) are part of this complex. This foredeep basin facies has many similarities to Culm-type, Hercynian flysches of southern Europe (Spalletta and Vai, 1982; Vai and Cocozza, 1986).

The Silurian–Devonian boundary in southwest Sardinia occurs in the calcareous Fluminimaggiore Formation (Gnoli et al., 1988). However, in southeast Sardinia (Gerrei sub-region, Fig. 24), it seems to be present at the base of the "upper graptolitic shales" and yields the index graptolite *Monograptus uniformis* (Jaeger, 1976). These "shales" are actually alum slates without chert or phosphorite. Based on the composite section of Barca and Jaeger (1990), *Scyphocrinites* also occurs in the lower part of this formation. These shales grade upward into fine-grained, nodular limestones of late Early to Middle Devonian age.

SILURIAN PALEOECOLOGY IN SARDINIA — The lowest Silurian of Sardinia consists of more or less uniform, dark

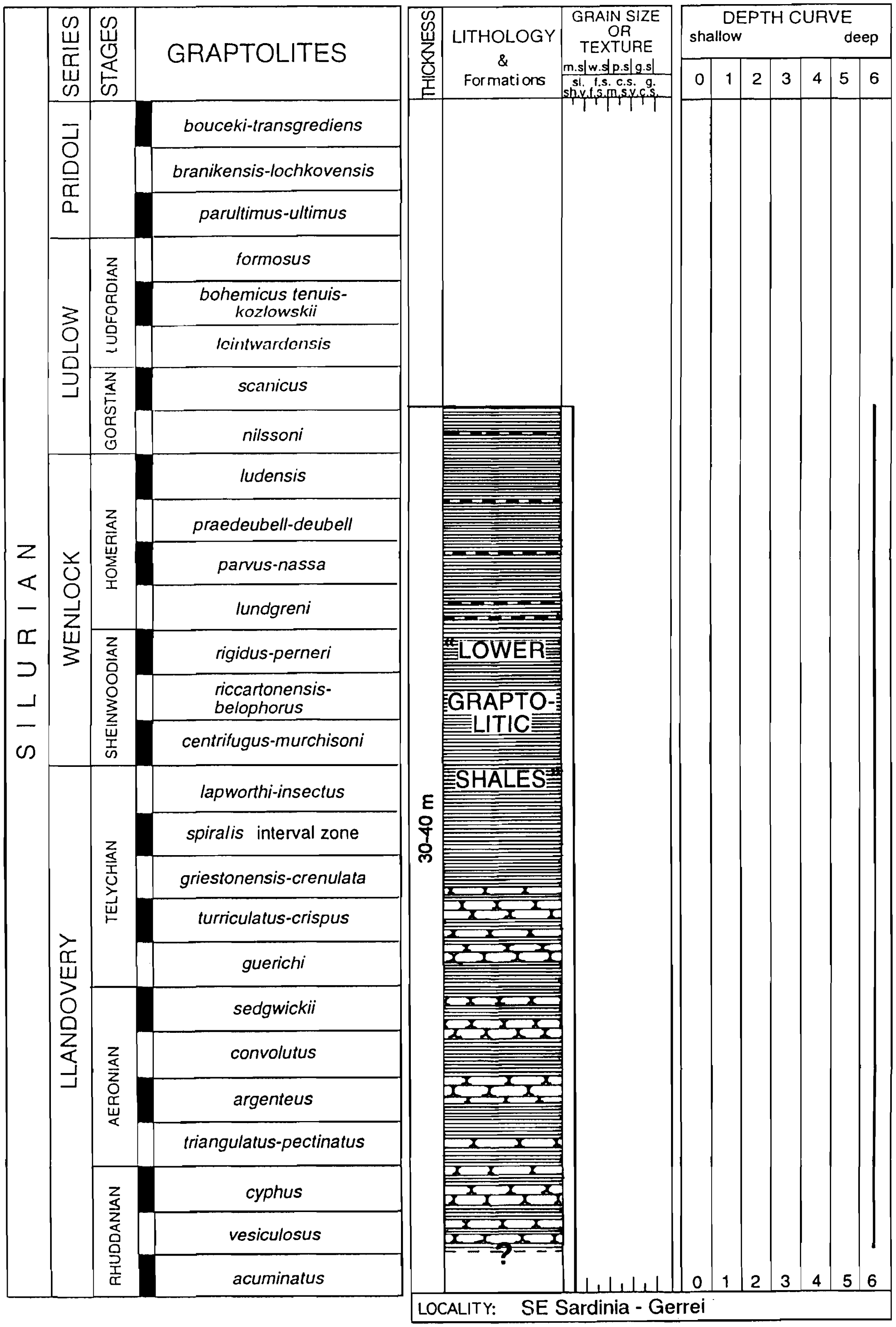

FIGURE 22 — Lower Silurian of Gerrei, southeast Sardinia.

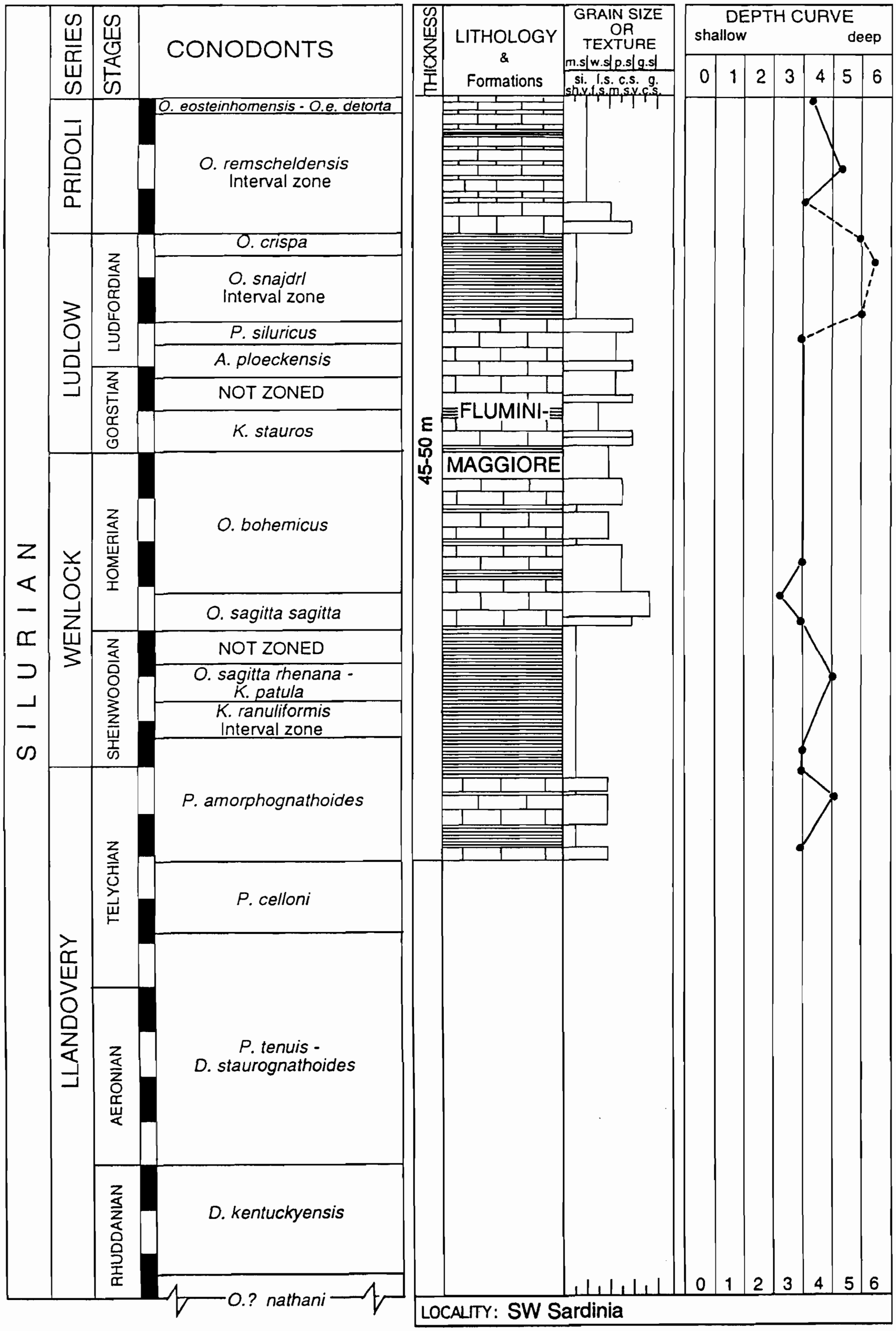

FIGURE 23 — Middle and Upper Silurian of Fluminimaggiore–Galemmu, southwest Sardinia.

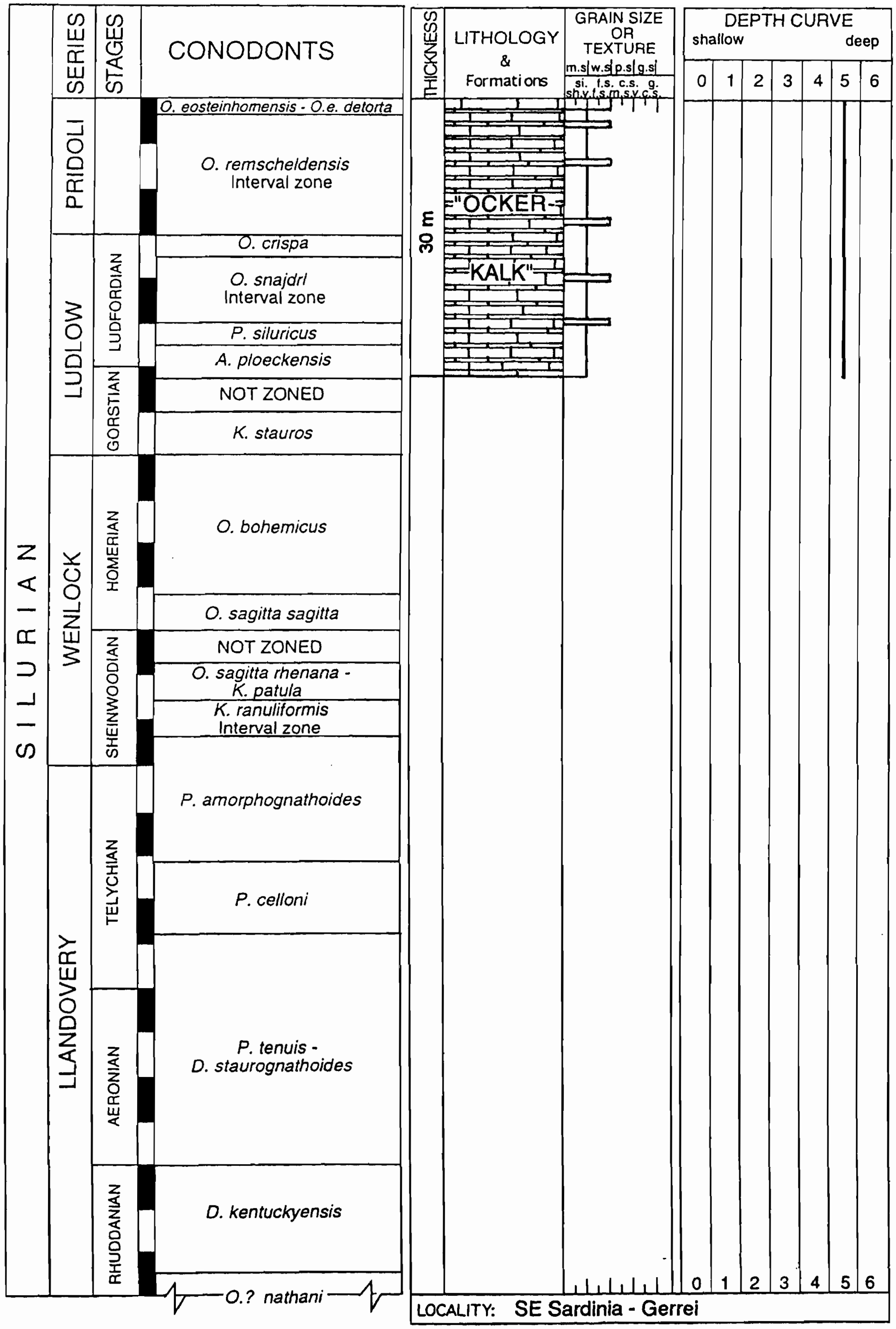

FIGURE 24 — Upper Silurian at Gerrei, southeast Sardinia.

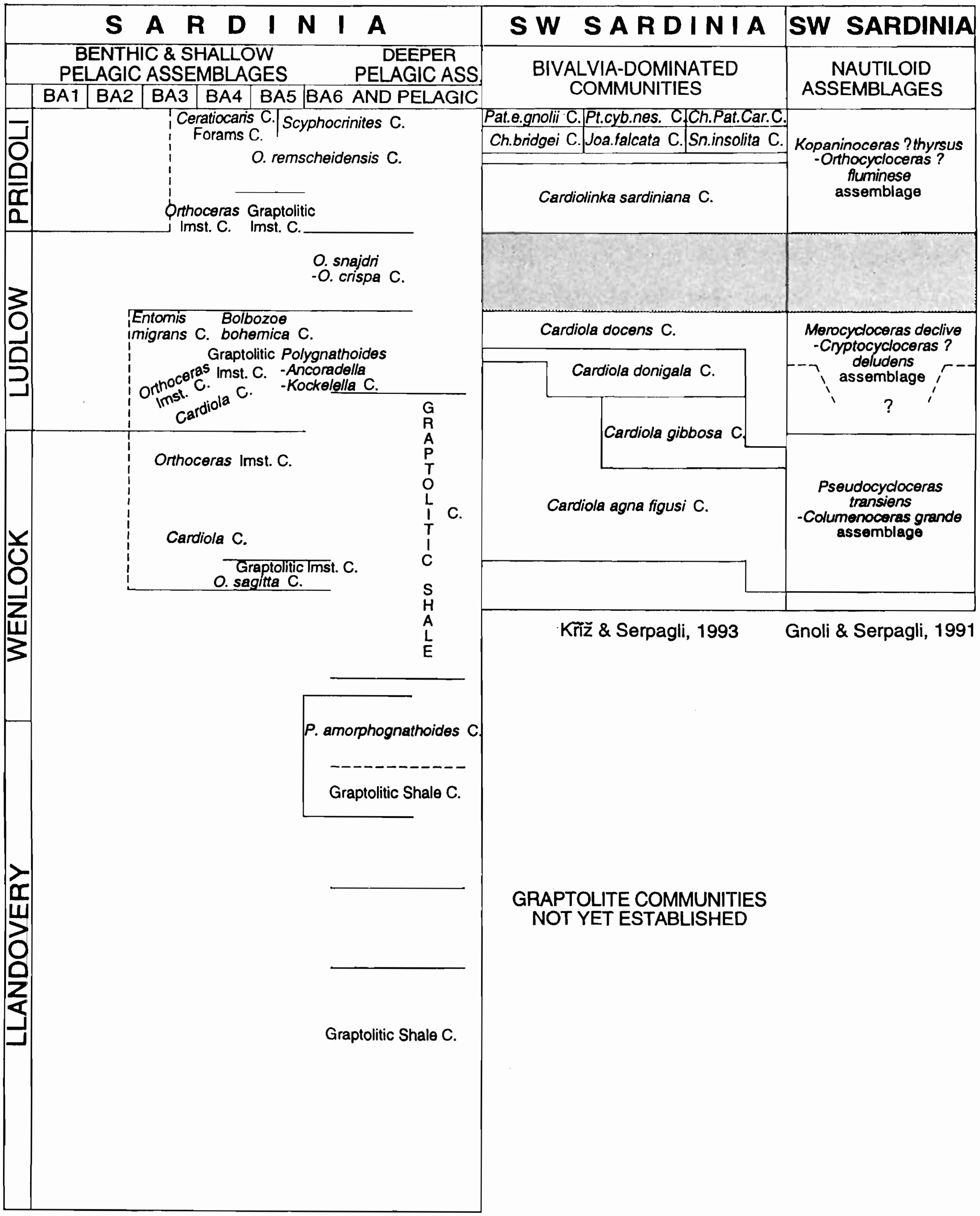

FIGURE 25 — Silurian communities in Sardinia.

shales that are rich in graptolites (Figs. 21–24). The same facies occur at the base of many other south European sequences, and indicate a common oceanographic domain along the northern margin of Gondwana after the marked provincialism of the Late Ordovician. Shale deposition occurred in an anoxic, sapropelitic basin (Jaeger, 1977b; Gnoli et al., 1990). Local calcareous deposition began diachronously in the late Llandovery, and became dominant in the Wenlock in southwest Sardinia and in the late Ludlow in southeast Sardinia (Fig. 25). Cephalopod-rich limestones from southwest Sardinia are lens-shaped beds intercalated with shales. They probably represent relatively short sedimentation events in the quiet depositional environments of the shales. Wave-oriented orthocones have been reported by Gnoli et al. (1980). A constant SSE-NNW conch orientation has been described from an upper Wenlock locality (Ferretti et al., 1995). This suggests uniformly oriented currents that carried cephalopod conchs along many parts of the northern Gondwana shelf. Local randomly oriented orthocones indicate that a current was not constantly active. Graptolitic limestones of southwest Sardinia are represented by centimeter-thick, graptolite-packed layers with a sharp base. These limestones are intercalated with fine-grained calcareous mudstones with sparse graptolite fragments and small cephalopods with abundant geopetal infills. Most of the cephalopod conchs still preserve body chambers, which are sometimes filled with graptolites. Both random- and current-oriented concentrations are present, even for graptolite rhabdosomes of the same genus and with similar hydrodynamic behaviour. These graptolitic concentrations represents discrete event horizons of individuals which were probably living in an environment adjacent to that of cephalopods (Ferretti and Serpagli, 1996a).

Five different microfacies have been recognized in the Wenlock–upper Ludlow limestones of southwest Sardinia. These include a shallow-water, high-energy depositional regime for the dominant peloid-cephalopod-ostracode packstone–wackestones (typical of the cephalopod limestone); the graptolitic packstones; and the coated-grain grainstone–packstones. Deposition below normal wave-base, but probably above storm wave-base, is indicated for the rare Ludlow mudstones with intercalated shell-lags and for the dark, laminated, fossiliferous mudstones. Pridoli sedimentation featured a shift to deeper waters, as shown by dark fossiliferous mudstones with occasional winnowed shell lags of disarticulated, thin-shelled, convex-up bivalves and ostracodes, small orthocones, and rare crinoid fragments (Ferretti, 1989).

The limestones from southeast Sardinia are largely represented by massive sequences of micritic limestone with millimeter-thick shell-lags of disarticulated debris. A quiet pelagic environment below wave-base has been suggested for these limestones (Barca et al., 1995).

As noted above, the transition into the Devonian takes place in a calcareous facies in southwest Sardinia, whereas in southeast Sardinia it appears to correspond to the lithologic change from the "Ockerkalk" limestone to the overlying "upper graptolitic shales."

SILURIAN COMMUNITIES IN SARDINIA — Ferretti and Serpagli (1996b) proposed a revised sketch of Silurian communities in Sardinia (Fig. 25, left side). Graptolite associations have been studied in the Lower Silurian of southwest Sardinia by Štorch and Serpagli (1993).

Twelve Silurian–lower Devonian bivalve-dominated benthic communities were recognized in the Fluminimaggiore Formation (Kříž and Serpagli, 1993; Fig. 25). The strong affinity between southwest Sardinia and Bohemia is shown by the common occurrence of 69 bivalve species. Four recurring, medium-diversity communities dominated by epifaunal forms were described within the *Cardiola* Community Group of latest Wenlock to late Ludlow age. Adaptation to conditions represented by the cephalopod limestone biofacies was achieved in these communities through an epibyssate life on a cephalopod shell substrate (Kříž, 1998). Pridoli communities that lived on soft micrite bottoms feature low-diversity infaunal and semi-infaunal taxa. Monospecific or very low-diversity communities developed in severe habitats (e.g., limited current activity and low oxygen content). More favorable habitats had communities with higher diversity and lower population density (Kříž and Serpagli, 1993).

Three nautiloid assemblages with potential stratigraphic value were recognized in the Middle to Upper Silurian of southwest Sardinia (Gnoli and Serpagli, 1991; Fig. 25). The *Pseudocycloceras transiens-Columenoceras grande* assemblage occurs in the *Ozarkodina sagitta sagitta–O. bohemica* Zones; the *Merocycloceras declive-Cryptocycloceras? deludens* assemblage is found in the *Ancoradella ploeckensis–Polygnathoides siluricus* Zones, and the *Kopaninoceras? thyrsus-Orthocycloceras? fluminese* assemblage extends from the *Ozarkodina remscheidensis* Zone to the *Icriodus woschmidti* Zone. Each nautiloid assemblage is widespread, and is similar to nautiloid assemblages from the Prague Basin, with several species in common (Gnoli, 1990; Gnoli and Serpagli, 1991).

THE SILURIAN OF AUSTRIA

In the Austrian Alps, fossiliferous Silurian strata are irregularly distributed (Fig. 26). They form a mosaic-like pattern of isolated units in the Alpine nappe system. Si-

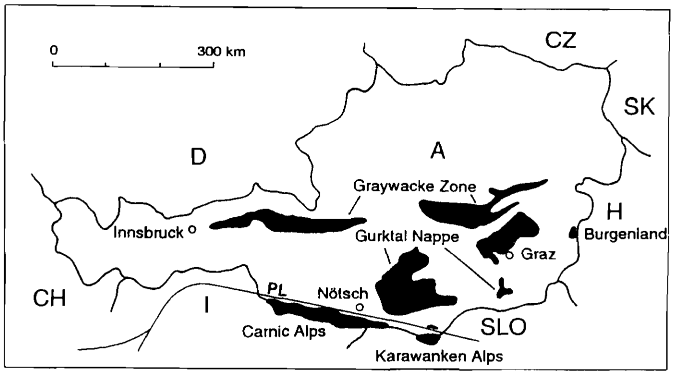

FIGURE 26 — Main regions of fossiliferous Paleozoic in the eastern and southern Alps. Abbreviations: A, Austria; CH, Switzerland; CZ, Czech Republic; D, Germany; H, Hungary; I, Italy; PL, Periadriatic Line; SLO, Slovenia; SK, Slovakia.

lurian outcrop areas include the Gurktal Nappe of middle Carinthia and southern Styria; the Graz region; and the Graywacke Zone of Styria, Salzburg, and Tyrol. Coeval rocks are exposed south of the Periadriatic Line along the northern margin of the southern Alps (i.e., in the Carnic and Karawanken Alps). In addition, some of the sedimentary precursors of quartz phyllites and even amphibolite-grade metamorphic rocks may also be Silurian, but it is not yet possible to correlate these non-fossiliferous units.

Since the discovery of Silurian fossils in the Alps by von Hauer (1847), the knowledge of Silurian rocks and organic remains has increased considerably. Microfossil research and field work by different working groups have elaborated a more detailed biostratigraphic framework and has documented the lithology of the Silurian.

Silurian deposits range from shallow-water carbonates to graptolitic shales. The thicknesses are regionally similar and generally do not exceed ca. 60 m. The main differences across the Periadriatic Line involve the distribution of fossils, facies patterns, rates of subsidence, supply areas, amounts of volcanism, and the spatial and temporal relationships of climate-sensitive rocks (Schönlaub, 1993).

The biostratigraphically important groups are primarily graptolites and conodonts. Other groups of almost equal importance in correlation are trilobites, bivalves, chitinozoans, and acritarchs. However, acritarchs are useful only in the Lower Silurian (upper Llandovery–lower Wenlock). Brachiopods, bivalves, and nautiloids provide further data and are useful in paleoecologic and paleogeographic syntheses.

The area north of the Periadriatic Line shares only a few lithologic features with the southern Alps. Shared features include thick siliciclastic sequences in the Ordovician–Devonian; local reefs during the Silurian and Devonian; and basic magmatism in the Ordovician, Early Silurian, and Middle Devonian. The increased input of siliciclastic material suggests proximity to a land area. On the other hand, intense volcanism may be related to crustal extension. However, this activity may also be responsible for the different facies which occurred in most areas north of the Periadriatic Line during the Silurian and part of the Devonian.

CARNIC AND KARAWANKEN ALPS — In the Carnic Alps, the Silurian transgression started in the earliest Llandovery *Akidograptus acuminatus* Chron. Due to the unconformity which separates the Ordovician and Silurian in the Carnic and Karawanken Alps, a varying thickness of sedimentary rocks is locally missing, which corresponds to several Llandovery and Wenlock conodont zones. Locally, the lowest Lochkovian rests disconformably on Upper Ordovician limestone (Schönlaub, 1971).

The Silurian is subdivided into four major facies belts that reflect different depth and energy conditions. The Plöcken facies represents a moderately deep-marine environment characterized, from bottom to top, by the pelagic Kok Formation, the Cardiola Formation, and the Alticola–Megaerella Limestones. The key section is the 60 m-thick Cellonetta profile (Fig. 27), well known for its classic Silurian conodont zonation (Walliser, 1964).

The Wolayer facies represents an apparently shallower environment. It is characterized by fossiliferous limestones with abundant orthoconic nautiloids, trilobites, bivalves, small brachiopods, gastropods, crinoids, and a few corals. Due to a hiatus at the base, this facies is represented by only 10–15 m of variegated limestones. The classical sections are located in the Lake Wolayer region of the central Carnic Alps (Von Gaertner, 1931; Schönlaub, 1971, 1980; Fig. 28).

The stagnant-water, graptolite facies is the Bischofalm facies. It is represented by 60–80 m of black siliceous shales, black cherty beds, and clayey alum shales (Fig. 29), which contain abundant graptolites. The graptolite succession has been clearly outlined (Jaeger, 1975; Flügel et al., 1977; Jaeger and Schönlaub, 1980, 1994; Schönlaub, 1985). According to Jaeger (1975), the Bischofalm facies can be subdivided into the lower, middle, and upper Bischofalm Shale.

The Findenig facies is intermediate between the shallow-water and the starved basinal environments. It comprises interbedded, black graptolitic shales, marls, and blackish limestone beds. At its base, a quartzose sandstone occurs locally (Fig. 30).

These four Silurian lithofacies reflect different rates of subsidence. Sediments from the Llandovery to the earliest Ludlow suggest steady basin subsidence and accompanying transgression. Subsidence and transgression apparently decreased and perhaps stopped during the Pridoli and led to balanced conditions with uniform lime-

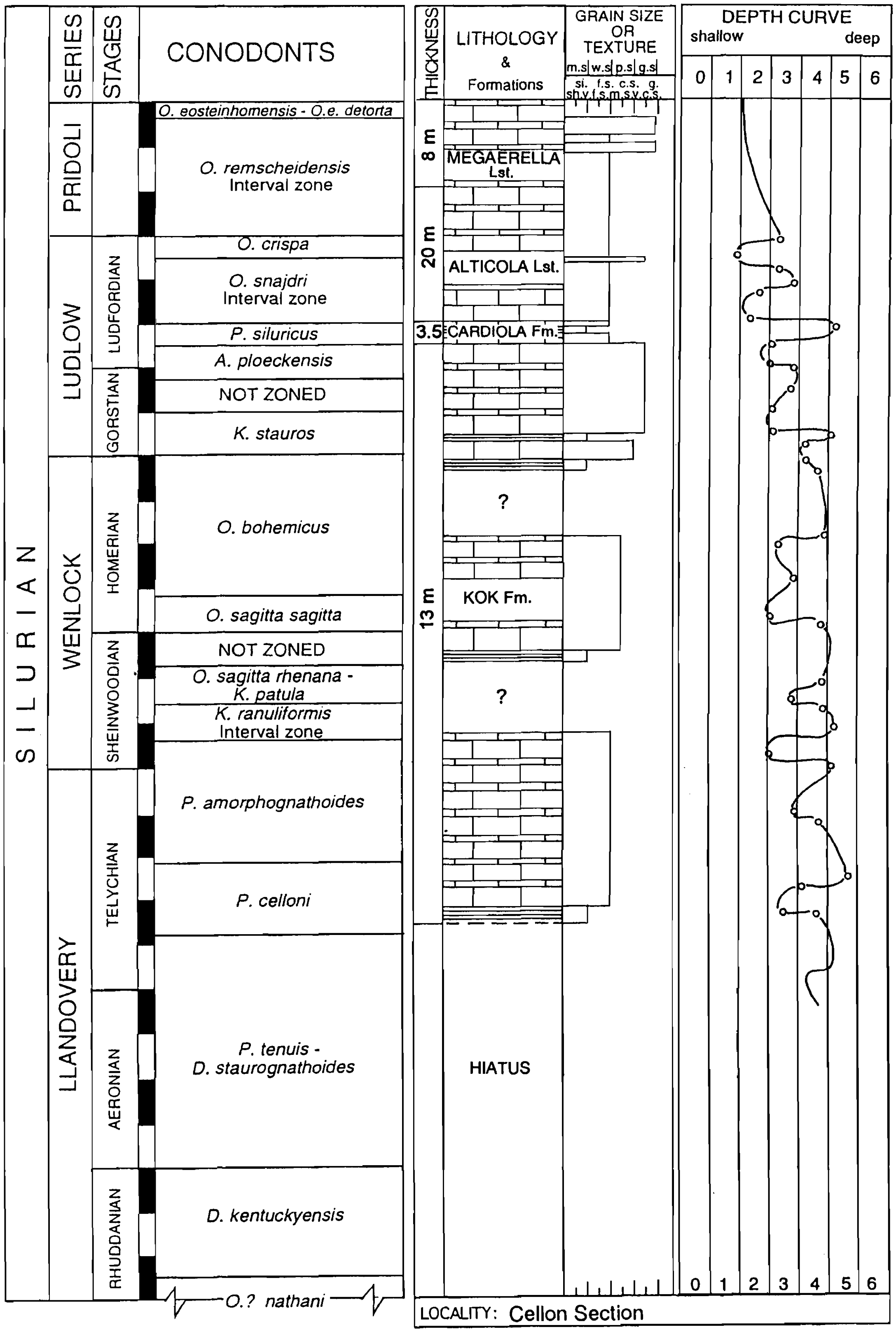

FIGURE 27 — Silurian at the Cellon section, Carnic Alps, Austria.

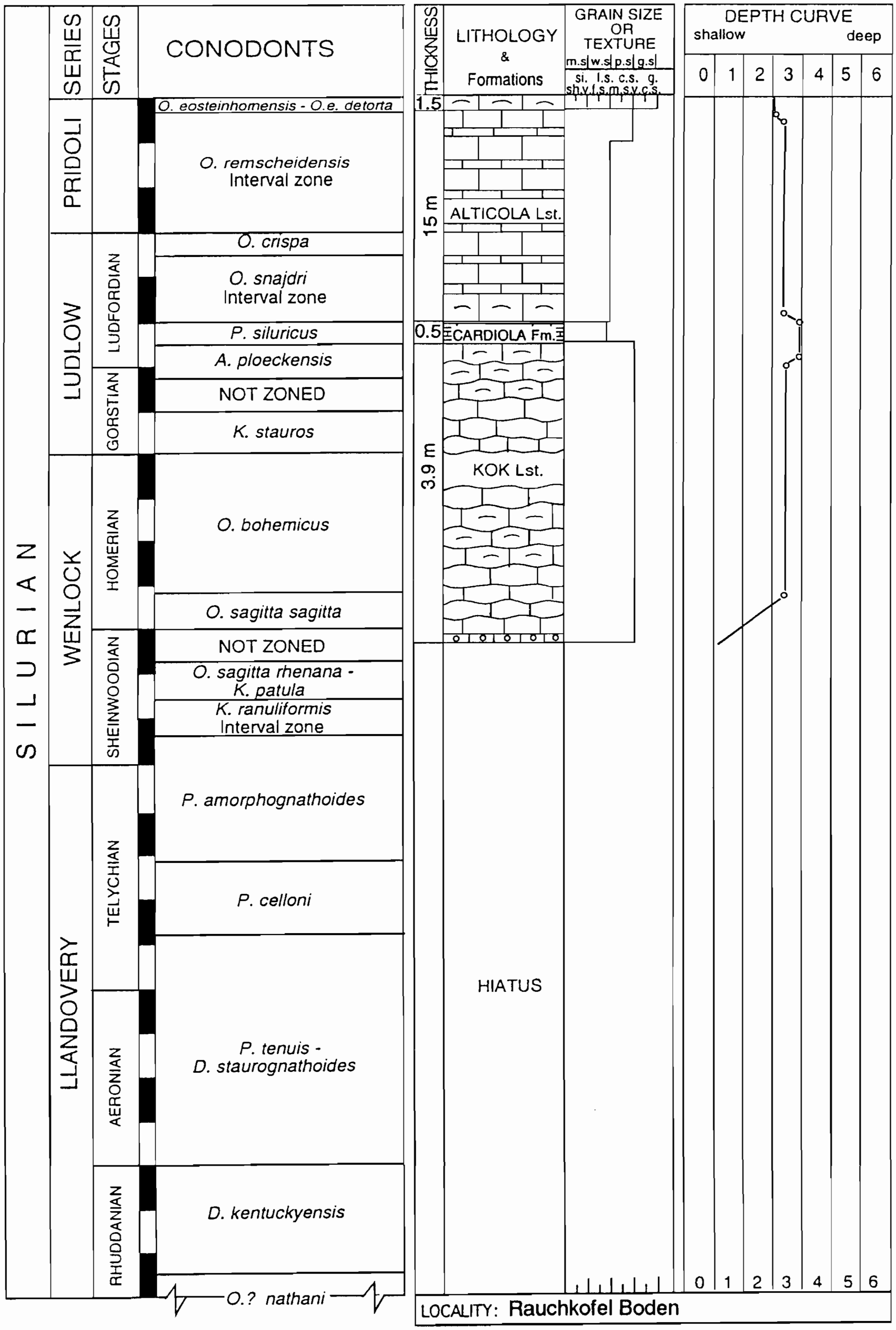

FIGURE 28 — Silurian at the Rauchkofel Boden section, Carnic Alps, Austria.

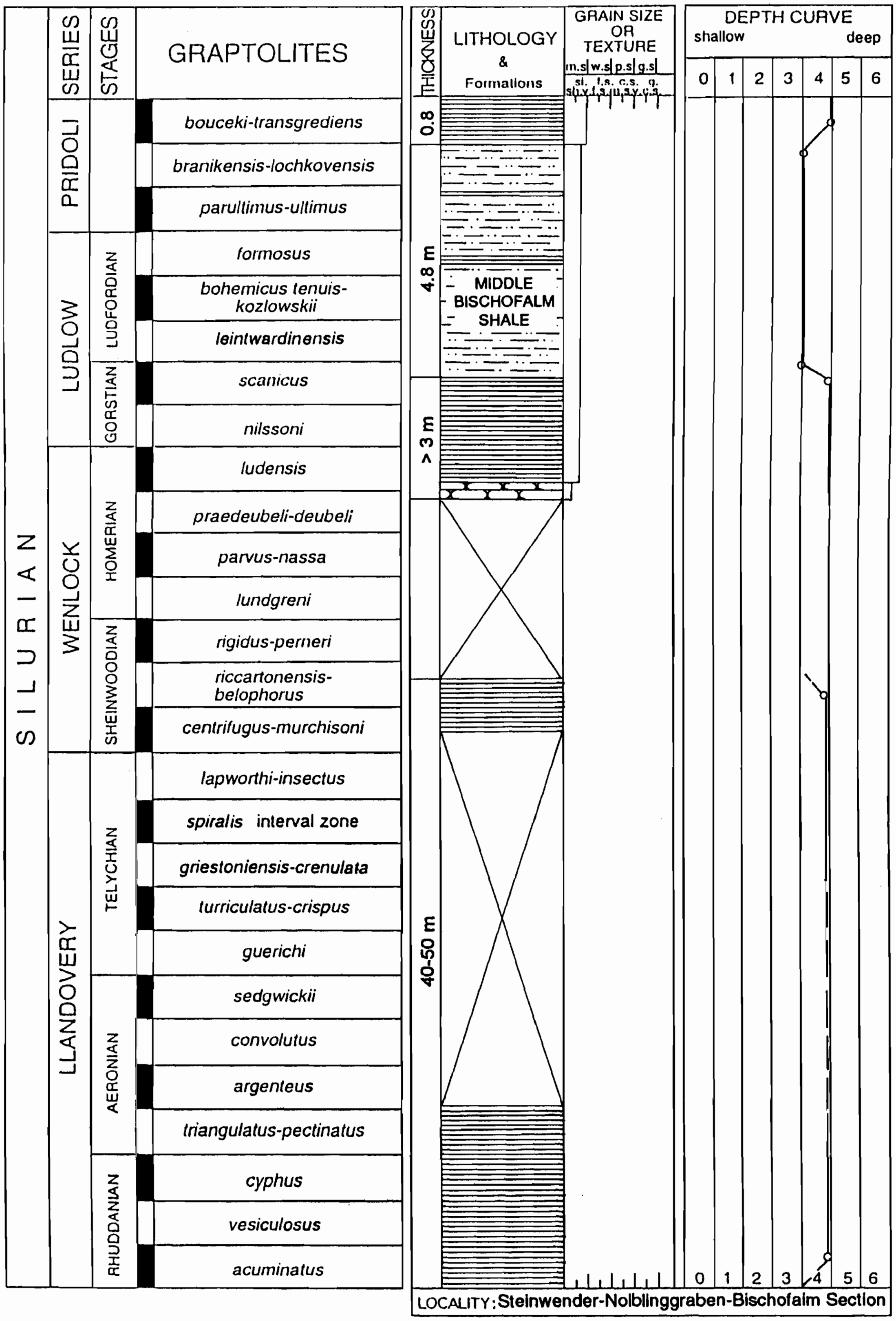

FIGURE 29 — Silurian at Steinwender–Nolblinggraben–Bischofalm section, Carnic Alps, Austria.

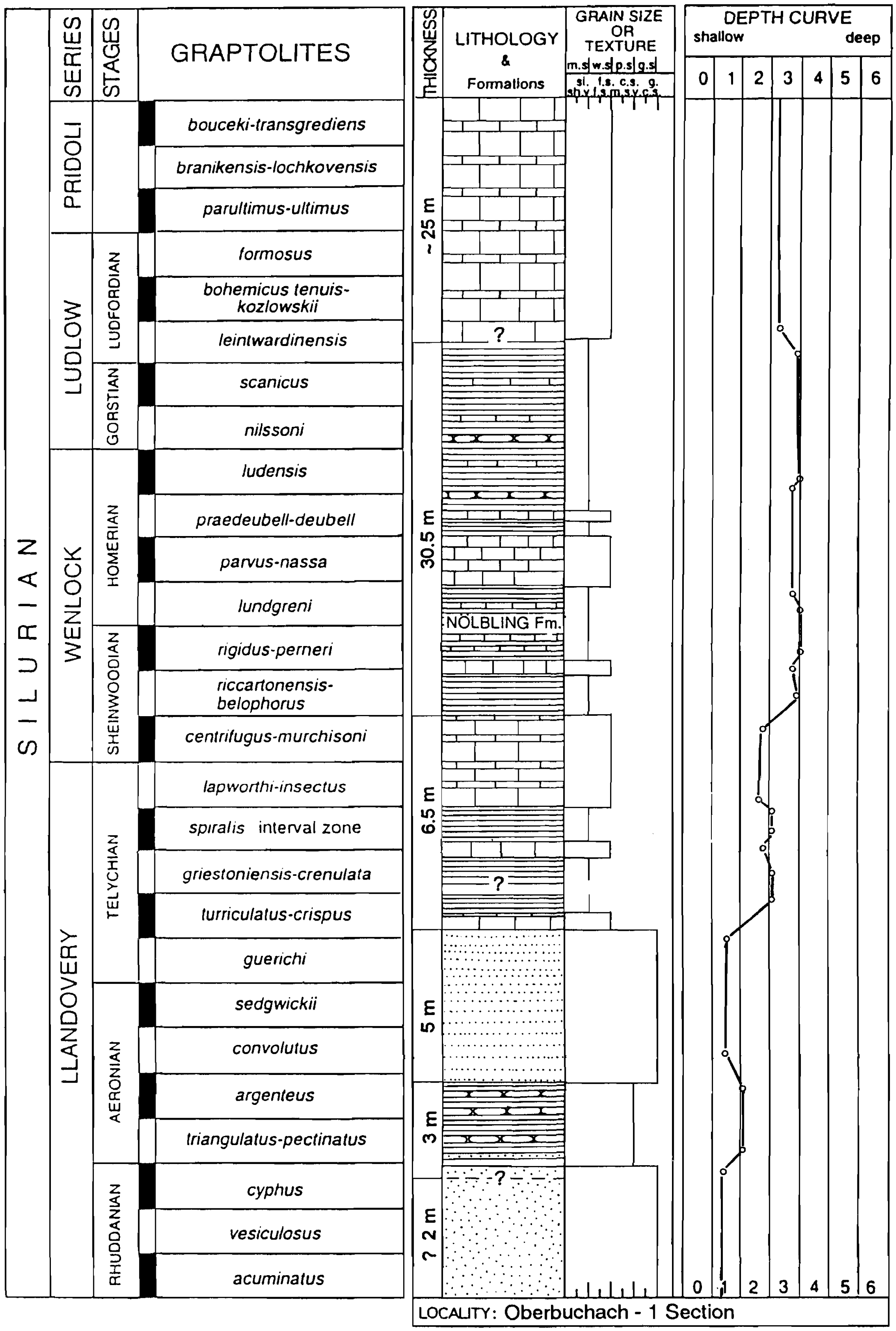

FIGURE 30 — Silurian at the Oberbuchach-1 section, Carnic Alps, Austria.

stone deposition. Simultaneously, the black graptolitic shale of the Bischofalm facies was replaced by green and gray shales called the middle Bischofalm Shale. At the base of the Devonian in the Bischofalm facies, the deep-water graptolitic environment reappeared and persisted until the end of the Lochkovian.

The Cellon section in the Carnic Alps (Fig. 27) has served since Walliser´s (1964) work as a standard for global conodont zonation that has been further refined and partly revised in other areas. In fact, this section represents the stratotype for the Silurian of the eastern and southern Alps (Schönlaub, 1994a). The conformable sequence suggests continuity from the Ordovician to the Devonian. However, in recent years, several small hiatuses have been recognized which reflect sea-level changes within an overall shallow to moderately deep environment. From top to base, the uppermost Ordovician–Silurian of the Cellon section is subdivided into the following formations: the Megaerella Limestone (gray, somewhat fossiliferous limestone, Pridoli, 8 m); Alticola Limestone (gray and pink, nautiloid-bearing limestone, Ludlow–Pridoli, 20 m), Cardiola Formation (alternating black limestone, marl, and shale, Ludlow, 3.5 m), Kok Formation (ferruginous nautiloid limestone with shale interbeds at the base, upper Llandovery–Wenlock, 13 m), and Plöcken Formation (calcareous sandstone, Ashgillian [Hirnantian], 4.8 m).

According to Schönlaub (1985, 1988), the Ordovician–Silurian boundary separates the Plöcken and Kok Formations. Conodonts and graptolites from the lower Kok Formation indicate that at least six graptolite and two conodont zones are missing in the Lower Silurian. Deposition began in the late Llandovery *Pterospathodus celloni* Chron.

The Llandovery–Wenlock boundary cannot be defined precisely in the Cellon section. Based on graptolites and conodonts, this boundary should be between Walliser's (1964) sample horizons 11 and 12. Consequently, the thickness of the Llandovery does not exceed ca. 3 m (Schönlaub, 1997).

The Wenlock–Ludlow boundary is drawn precisely between Walliser´s (1964) conodont samples 15B1 and 15B2. This level closely corresponds to the Wenlock–Ludlow boundary stratotype at Pitch Coppice quarry near Ludlow, England. The entire Wenlock at the Cellon section has an overall thickness of 5.0 m. By comparison with the Bohemian sections, strata equivalent to the range of the index conodont *Ozarkodina bohemica* are extremely condensed at Cellon, and this suggests that deposition occurred mainly during the early Homerian. As noted by Schönlaub (1994) on the underlying Sheinwoodian Stage, it may be inferred that the lowest Homerian is missing. *Cyrtograptus rigidus* Zone graptolites are found in the shale interbed between samples 12B and 12C, and indicate the upper Sheinwoodian.

Correlation with the Bohemian sequences and the occurrence of the basal Pridoli index graptolite *Monograptus parultimus* locate the Ludlow–Pridoli boundary a few centimeters (Walliser, 1964) above conodont sample number 32 (see H. P. Schönlaub *in* Kříž et al., 1986). The lowest Pridoli *Cardiolinka bohemica* Community appears just above sample horizon 32 (Kříž, 1999). This level is 8.0 m above the base of the Alticola Limestone, and suggests that the thickness of the Ludlow is about 16.45 m.

The Silurian–Devonian boundary at Cellon is placed at the bedding plane between Walliser´s (1964) samples 47A and 47B. At sample 47A, the lowest specimens of the index conodont *Icriodus woschmidti* occur. The lowest occurrence of diagnostic lowest Lochkovian graptolites is 1.5 m higher. However, Jaeger (1975) recorded the lowermost occurrences of *Monograptus uniformis*, *M. sp.* cf. *M. microdon*, and *Linograptus posthumus* in sample horizon 50. In total, the Pridoli at the Cellon section may reach 20 m. Data about the distribution of acritarchs, chitinozoans, brachiopods, bivalves, and taxonomically unrevised nautiloids and trilobites are included in the report edited by Schönlaub and Kreutzer (1994).

Two types of facies can be recognized in the Carnic Alps as early as the Late Ordovician. According to Dullo (1992), the Wolayer Limestone represents a near-shore, cystoid-rich facies, and the Uggwa Limestone is its offshore, basinal counterpart. Following a depositional gap at the base of the Silurian caused by glacially induced sea-level fall, renewed sedimentation started in a moderately shallow environment which may have lasted until the earliest Wenlock. This environmental interpretation is suggested by Walliser´s (1964) sample number 11, a bioturbated wackestone with algae and lumachelles (i.e., shell hash layers) that indicate a very shallow to intertidal environment. Later in the Wenlock, there was a progressive deepening. However, at the Wenlock–Ludlow boundary, a hiatus is present.

During deposition of the Cardiola Formation, a pelagic off-shore environment is indicated by radiolarian-bearing, black, marly interbeds and pelagic limestones with a diverse *Cardiola docens* Community and *Cardiola pectinata* Subcommunity (Kříž, 1999). The overlying Alticola Limestone reflects stable conditions in a pelagic setting that ended with a short regressive pulse recorded by Walliser´s (1964) sample 40 (a laminated grainstone with lumachelles). A further deepening trend can be assumed at the base of the Megaerella Limestone. More details are available in L.H. Kreutzer (*in* Schönlaub et al., 1994).

Graptolites have been known in the Alps since their discovery by Stache (1872). The pure graptolitic facies is best exposed in the so-called "Graptolithengraben" north of the Obere Bischofalm in the central Carnic Alps. The graptolite-bearing rocks form a monotonous sequence of interbedded radiolarian-bearing cherts and alum shales. The cherts dominate the Llandovery and Wenlock; the shales prevail in the upper part of the succession. The intermediate green and gray shales yield only a few graptolites in very thin layers (H. Jaeger *in* Flügel et al., 1977).

The thickness of the graptolite-bearing Silurian–Lochkovian ranges from 50–100 m. It is an extremely condensed sequence due to a very low, but nevertheless continuous, rate of sediment accumulation. This conclusion is supported by the very complete graptolite zonal succession. The environmental conditions were anoxic or strongly dysaerobic except for the short interval when the middle Bischofalm Shale was deposited.

Graptolites and a few conodonts on bedding planes are the only fossils to be found in the "Graptolithengraben" facies. The graptolites are common in many layers, both in the alum shales and in the cherts. Some intervals, however, are almost barren of graptolites.

Due to intense Variscan and Alpine tectonism, longer undisturbed sections are rare. By far the best exposed and least disturbed section is the "main section," or Hauptprofil (Fig. 29), which has been studied in great detail by H. Jaeger since 1965. This tectonic block is almost 20 m thick and covers the interval from the Wenlock *Pristiograptus ludensis* Zone to the Lower Devonian *Monograptus hercynicus* Zone. In the vicinity of the Hauptprofil, older strata are also well exposed, but they are in fault contact with the main section.

The main graptolite section is virtually undisturbed, except for a fault at the critical horizon between the *Monograptus uniformis* and the *M. transgrediens* Zones (i.e., at the Silurian–Devonian boundary). By comparison with other sections, it is concluded that there is no significant loss of strata at this fault (H. Jaeger *in* Flügel et al., 1977).

According to H. Jaeger (*in* Flügel et al., 1977), a number of important features are shown by the Hauptprofil. The Silurian–Devonian boundary is within a homogenous black shale facies. Obviously, there was no physical break at this boundary. A distinct change in facies from green and gray to black shales preceded the faunal change at the boundary by one graptolite zone. There is no evidence that the ranges of *Monograpatus transgrediens* and *M. uniformis* overlap. Finally, middle Bischofalm Shale occupies the same stratigraphic position as the non-graptolitic "Ockerkalk" of Thuringia and, presumably, Sardinia.

The intermediate facies between the shallow-water and basinal settings is best developed at the Oberbuchach section (Fig. 30). This facies is termed the "Findenig facies." The Silurian here is a mixed argillaceous–calcareous lithology referred to the Nölbling Formation. This almost 50 m-thick Llandovery–Ludlow unit is underlain by the Upper Ordovician Uggwa Limestone and the 10 m-thick siliciclastic Plöcken Formation of Hirnantian age. The latter formation is overlain by interbedded laminated pyritic sandstone, bedded black chert, and black shale with lower middle Llandovery *Coronograptus gregarius* Zone and *Demirastrites triangulatus* Subzone graptolites. It is not clear whether the lower Llandovery is missing, or whether this portion of the section is barren of fossils.

A second horizon of graphitic sandstones occurs in the upper Llandovery. Its age is inferred from diagnostic *Pterosphatodus celloni* Zone conodonts from limestones that overlie this siliciclastic interval. These limestones are followed by alternating dark argillaceous limestone, black graptolite shale, and chert that range through the Wenlock into the Ludlow. Conodonts in this interval are associated with uppermost Llandovery–Wenlock graptolites. In the overlying shales, graptolites occur at several levels, and include the Sheinwoodian *Monograptus riccartonensis* Zone and range up to the lowest Gorstian *Neodiversograptus nilssoni* Zone. The Wenlock–Ludlow boundary may thus be placed some 40 m above the base of the graptolite-bearing sequence.

At the Oberbuchach section, fossils other than graptolites and conodonts are very rare. Conodont assemblages are dominated by *Dapsilodus* and *Decoriconus.* Ramiform elements only occur in the lower part of the section (Schönlaub, 1980). Strata corresponding to the remaining part of the Ludlow and Pridoli are up to 20 m thick. This interval consists of lithologically distinct, gray, almost unfossiliferous, pyritiferous limestones with a characteristic weathered surface (Schönlaub, 1980).

The Rauchkofel Boden section (Fig. 28) represents the Silurian Wolayer facies. This facies is named after the Upper Ordovician, cystoid-bearing Wolayer Limestone and is overlain by highly fossiliferous Middle–Upper Silurian limestones. Strata corresponding to the Upper Ordovician Hirnantian Stage through the Lower Silurian are missing in this facies belt. The sedimentary gap may be ascribed to the glacially induced, terminal Ordovician eustatic fall.

The Wolayer Limestone is disconformably overlain by the gray, fossiliferous, cephalopod-bearing *Orthoceras* Limestone, a unit equivalent to the Kok Limestone at Cellon. Besides the dominant nautiloids, trilobites and bivalves are quite common (Gaertner, 1931; Ristedt, 1968; Kříž, 1979, 1999; Schönlaub, 1980). In addition, conodonts are fairly abundant and represent the Wenlock *Ozarkodina sagitta* Zone (basal Homerian). About 1.2 m above the Wenlock–Ludlow unconformity, the index conodont *Kockelella variabilis* appears, and this suggests the base of the Ludlow Series by comparison with Bohemia

(H. P. Schönlaub *in* Kříž et al., 1993). The overlying Cardiola Formation (Fig. 27) corresponds to the *Polygnathoides siluricus* Zone of the Cellon section. It is succeeded by pinkish and grayish limestones, which correspod to the Alticola and Megaerella Limestones at Cellon. However, no diagnostic conodonts have been found at Rauchkofel, except in the uppermost limestones with *Scyphocrinites* sp. This highest conodont fauna has common forms of the *Ozarkodina remscheidensis eosteinhornensis* Zone. Based on recent field data (J. Kříž, A. Ferretti, C. Histon, O. Bogolepova, and H. P. Schönlaub, unpublished data, 1997, and Kříž, 1999) bed number 331 is uppermost Pridoli (Schönlaub, 1980), and the Silurian–Devonian boundary lies just above, but below the beds with *Scyphocrinites. Antipleura bohemica* Community bivalves (Kříž, 1999) occur at the base of the Lochkovian, 40 cm above bed number 331 (Schönlaub, 1980).

Preliminary paleoecologic and paleogeographic analysis of the Wenlock–Pridoli at the Rauchkofel section (Fig. 28) indicate a shallow-water depositional environment dominated by the South Equatorial Current. This current may have been responsible for the exchange of faunas between such separated areas as northern Siberia, Perunica, the Carnic Alps, and Sardinia (Kříž and Bogolepova, 1995). Indeed, there is a SW–NE orientation of orthoconic cephalopod conchs in the Kok Limestone, and this changes to a NNE–SSW direction in the overlying Lochkovian (O. K. Bogolepova *in* Schönlaub and Kreutzer, 1994).

GURKTAL NAPPE — The Gurktal nappe (Fig. 26) is composed of several hundred meters of volcanic and siliciclastic rocks with intercalated limestones. The Silurian includes coral-bearing, fossil-fragment limestone lenses at the transition from the Llandovery to the Wenlock, and local 5–10 m-thick, Upper Silurian limestones and dolostones. Due to poor fossil control and exposure, it is not yet possible to reconstruct a composite Silurian section. However, the facies suggest a subdivision into a carbonate-dominated and a carbonate-poor facies (Buchroithner, 1979; Ebner et al., 1990; Schönlaub and Heinisch, 1994).

The Lower Paleozoic of the Gurktal nappe system is characterized by volcanic rocks. Volcanism occurred at different times, and was of varying intensity and of different geochemical character as a consequence of different paleotectonic settings (Loeschke and Heinisch, 1993).

GRAZ REGION — The Paleozoic of the Graz area (Fig. 31) is best displayed in the Rannach nappe, the uppermost nappe of the Graz thrust complex. The Silurian is dominated by alkaline mafic lavas and volcaniclastics, which suggest an initial rift stage. These volcani- and siliciclastics are succeeded by increased carbonate production during the Late Silurian and Devonian.

According to Fritz and Neubauer (1988) and Neubauer (1989), sedimentation of the Silurian Kehr Formation was controlled mainly by volcanism. During the early Ludlow, a more easterly area was characterized by a proximal shallow-water setting with lavas and coarse lapilli tuffs (Fig. 31), while the western distal facies shows intercalations of lapilli-rich beds, agglomerates, shales, and pelagic limestones. The interesting Kehr Agglomerate has 13% quartzite, dolostone, chert, and reworked limestone clasts.

During the Late Silurian, the volcanic centers were blanketed by fossiliferous carbonates that include approximately 4.0 m-thick bedded dolostones, with lenses of fossiliferous (crinoids, brachiopods, trilobites, nautiloids) dolomitic limestones interbedded with tuffs and tuffaceous shales. Based on conodonts, this sequence is Ludlow (Ludfordian) to Pridoli (Ebner, 1994).

Similar environmental conditions are suggested for the Upper Silurian of the other nappes of the Graz Paleozoic. In these nappes, pelagic, nodular limestones persisted from the Late Silurian to the Devonian.

The Silurian of the Graz area is best displayed in the Eggenfeld section (Fig. 31). In this area, the distribution of the Upper Silurian and Lower Devonian was controlled by Silurian volcanism. Despite poor outcrops, a well-constrained lithostratigraphic framework can be established (Ebner, 1994). Massive green basalts that interfinger with pinkish to greenish tuffs with the graptolite *Bohemograptus bohemicus tenuis* form the base of the succession. A overlying unit of dark dolostones (unit D/1) has common crinoids, brachiopods, nautiloids, and tabulate corals (*Favosites* sp.), and is succeeded by tuffs and tuffaceous shales. A second interval of dark dolostones (D/2 unit) has lens-like accumulations of crinoids, brachiopods, trilobites, nautiloids, and a few corals (e.g., *Syringaxon* sp.). The uppermost Silurian consists of tuffs and tuffaceous shales with intercalated dark dolostones (D/3 unit), with shell hash accumulations that include crinoids, brachiopods, trilobites, and nautiloids.

Biostratigraphically important macro- and microfossils at Egenfeld include conodonts and brachiopods. Conodonts are fairly abundant in all of the calcareous levels. Diagnostic species include *Polygnathoides siluricus*, *P. emarginatus*, and *Kockelella variabilis* in the dolostones immediately above the basalts. This indicates an end of basalt volcanism in the Ludfordian. *Bohemograptus bohemicus tenuis* from the lowest volcaniclastic layer (Hiden, 1996) is also Ludfordian. *Ozarkodina snajdri* has been identified in the second carbonate (D/2) and indicates the Ludfordian *O. snajdri* Zone. This index species is associated with *Ozarkodina remscheidensis eosteinhornensis*. In addition, the brachiopod *Septatrypa subsecreta* occurs in

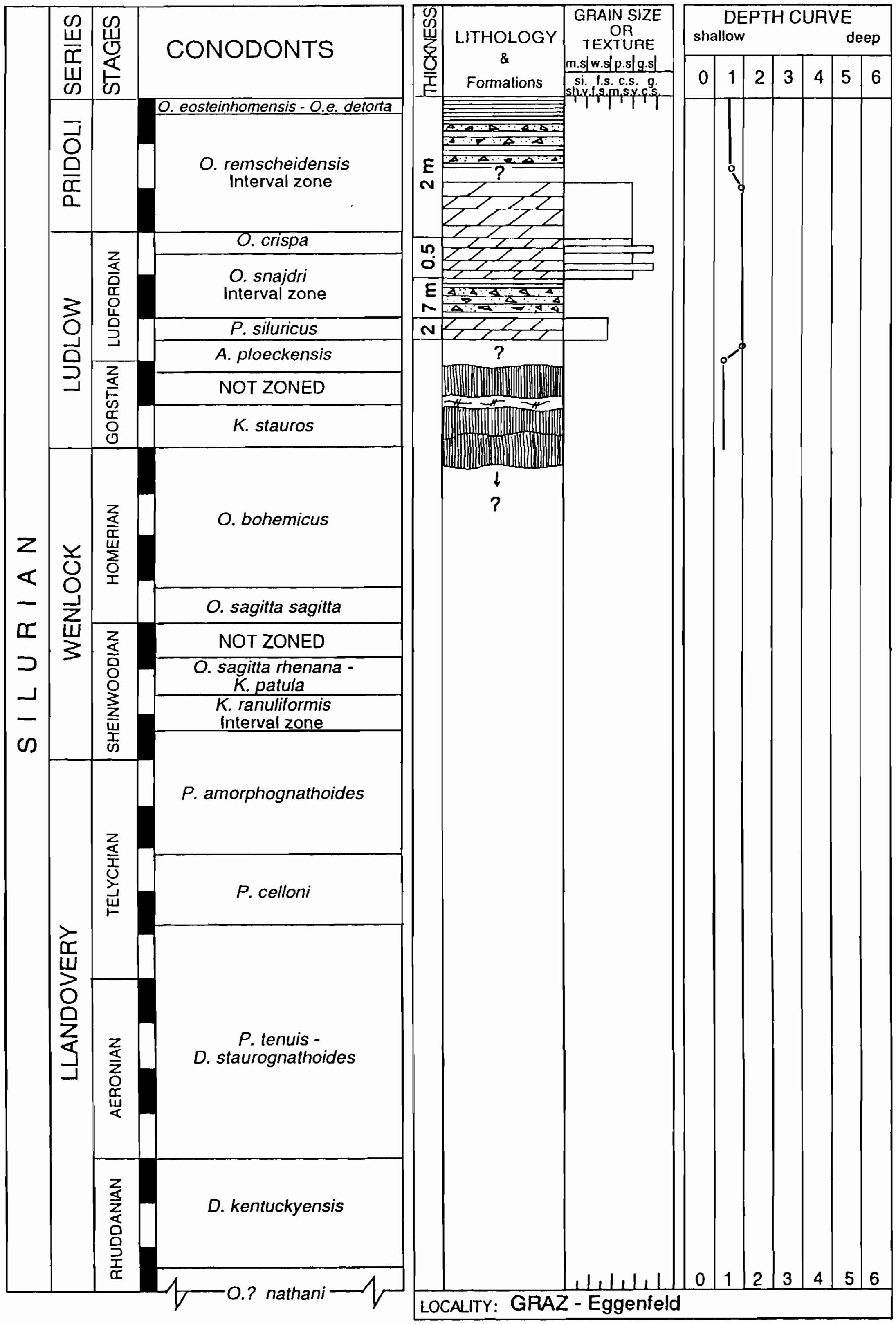

FIGURE 31 — Silurian at the Eggenfeld section near Graz, Austria.

thin carbonate beds in the overlying tuffaceous shales. Based on the occurrence of *Icriodus woschmidti*, *Septatrypa subsecreta* appears in the lowermost Lochkovian. However, index conodonts of the *Pedavis latialata* and *Ozarkodina crispa* Zones were not recovered.

The Eggenfeld section is of particular importance for dating Silurian volcanism in the eastern Alps. Based on its fossils, this section is an excellent example of a volcanic island surrounded and buried by fossiliferous carbonates during the Late Silurian. Carbonate production and volcanism increased later in the Devonian.

The Graywacke Zone — According to Schönlaub (1979) and Schönlaub and Heinisch (1994) the Silurian in the thick Lower Paleozoic of the Graywacke Zone of Styria (Fig. 26) shows vertically distinct facies that ranges from a lower 50 m of crinoid- and nautiloid-bearing limestone to overlying black graptolite shale. These two facies change laterally and vertically into interbedded limestone and shales overlain by pure limestones in the upper Ludlow and Pridoli. Local intercalations of Llandovery basic volcanics occur near the southern margin of the Graywacke Zone.

These facies changes also seem to be valid for the Tyrol and Salzburg segments of the Graywacke Zone. According to Heinisch (1988), two distinct facies can be distinguished within short distances. They are preserved in two nappes named the Wildseeloder and the Glemmtal Units. In the Silurian, the general facies range from black shale with local graptolites to chert; siliceous, pelagic limestone; condensed cephalopod limestone; and even dolomitic rock.

The Wildseeloder Unit in the western Graywacke Zone is characterized by the thick Upper Ordovician Blasseneck quartz porphyry, which is overlain by several meters of middle and upper Llandovery pelagic limestone. These limestones are overlain by the so-called "Dolomit–Kieselschiefer-Komplex" (Bedded Dolostone–Chert Formation). In the Late Silurian, a carbonate platform developed which lasted until the early Late Devonian.

The Glemmtal Unit in the western Graywacke Zone comprises more than 1,000 m of mainly siliciclastic units, which are known as the Wildschönau Group. Locally, up to 50 m of intercalated condensed pelagic limestone, marl, chert, siliceous shale, and basalt form the Klingler Kar Formation. Based on conodonts, the lower Klinger Kar Formation is Upper Silurian. This facies laterally grades into the turbiditic Löhnersbach Formation. In the latter formation, however, age determinations are not yet available.

With a few exceptions in Styria, Silurian conodonts and graptolites of the Graywacke Zone in Tyrol and Salzburg are fairly well known. However, no detailed biostratigraphic data are available on the exact position of the Ordovician–Silurian boundary (Schönlaub and Kreutzer, 1994).

The Spießnägel section south of Kirchberg, Tyrol, is one of the few sections in which the transition of presumably Upper Ordovician graywackes into the basal Silurian is exposed. According to Al-Hasani and Mostler (1969), the Silurian starts with 0.85 m of arenaceous and tuffaceous limestones with *Pterospathodus celloni* Zone conodonts. The lower part of these limestones feature bioturbated mudstones with varying amounts of siliciclastic and tuffaceous material. These mudstones grade into wackestones 0.7 m above the base of the Silurian. Of special interest is the occurrence of coated grains in the upper part of this bed. The nucleii of the grains are formed of crinoid ossicles or shell debris. This lowest part of the Silurian is succeeded by 1.1 m of limestone with interbedded shale and thin limestone lenses. This part consists of packstones with thin hash layers of bivalves, brachiopods, ostracodes, and echinoderms. These limestones are sharply overlain by grayish laminated dolostone assigned to the lower Wenlock *Kockelella patula* Zone.

The Spießnägel sequences correspond to the *Pterospathodus celloni–P. amorphognathoides* (conodont) Zone. They reflect late Llandovery–earliest Wenlock environments in this segment of the Graywacke Zone.

Another important Lower Silurian locality has long been known as the "Lachtal-Grundalm section" near the village of Fieberbrunn (Fig. 32). This classic graptolite-bearing sequence in the Graywacke Zone is a mixed shale–limestone succession known in the literature as "Lydit–Kieselkalk-Komplex." It is overlain by the 5.0 m-thick "Dolomit–Kieselschiefer-Komplex" (Mostler, 1966).

The basal cherty interval at Lachtal-Grundalm is formed of black, massive cherts known as "lydite" in the Alpine terminology. It is composed of radiolarian-bearing dolostones and reddish, cherty limestones that grade vertically into crinoidal limestones. The total thickness does not exceed 5.0 m. The accompanying microfauna consists of ostracodes, foraminiferans, brachiopods, radiolarians, conodonts, and echinoderms. In addition, bivalves, solitary corals, trilobites, and orthoconic nautiloids occur sparsely in the lower part of the 1.4 m-thick crinoidal limestone. The lower 2.1 m of the crinoidal limestones are assigned to the *Pterospathodus celloni* Zone; the upper part belongs to the *P. amorphognathoides* Zone.

According to Jaeger (1978), the only identifiable graptolites in the "Dolomit–Kieselschiefer-Komplex" occur in the upper Lachtal-Grundalm section. The lithology resembles the Silurian Nölbling Formation in the Carnic Alps. *Bohemograptus bohemicus* is most abundant in an upper horizon of the Dolomit–Kieselschiefer-Komplex.

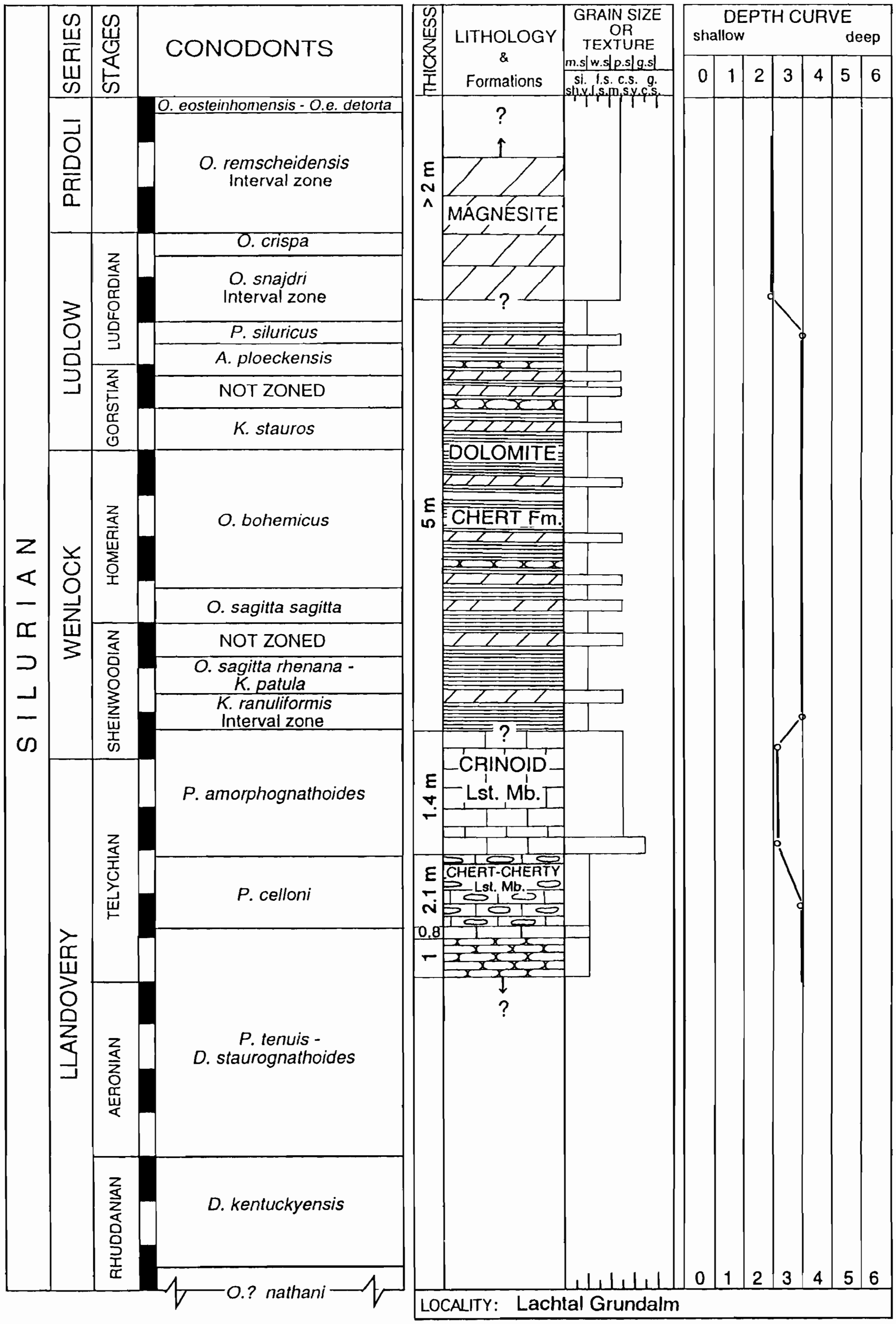

FIGURE 32 — Silurian at the Lachtal Grundalm section in the Graywacke Zone, Austria.

This species characterizes the basal Gorstian *Neodiversograptus nilssoni* Zone. Co-occurring conodonts are long-ranging forms that do not refine of this age assignment. Other graptolites include *Pristiograptus dubius* sp. cf. *P. frequens* and *Colonograptus* sp. cf. *C. colonus.*

In the Graywacke Zone in Tyrol, the "Dolomit–Kieselschiefer-Komplex" is overlain by dolomitic rocks and magnesite. According to Mostler (1966), the base of these carbonates can be assigned to the *Polygnathoides crassa* Zone or to the base of the overlying *Ancoradella ploeckensis* Zone at the Gorstian–Ludfordian boundary.

In summary, the data from the Lachtal-Grundalm section show that it represents a composite succession that extends through most of the Silurian. Biostratigraphically dated rocks start in the middle Llandovery and can be followed through the Wenlock to the middle Ludlow. In the Graywacke Zone in Tyrol, no record of the Pridoli is known, although the Pridoli may be represented by recrystallized dolostones.

SILURIAN FAUNAS AND CLIMATE IN AUSTRIA — The Alpine Silurian is characterized by a wide range of lithofacies. The strata are locally fossiliferous and feature distinct faunal assemblages that include nautiloids, trilobites, bivalves, brachiopods, graptolites, conodonts, foraminiferans, acritarchs, chitinozoans, and ostracodes. During the last few decades, most of these groups have been revised. Available data suggest a complete but condensed succession in the carbonate-dominated facies and a continuous record in the graptolite-bearing sequences. This is particularly true in the Carnic and Karawanken Alps. In other areas, stratigraphic continuity has yet not been demonstrated, and this may be due to poor preservation, original lack of fossils, and metamorphic overprints.

Silurian faunas after the terminal Ordovician mass extinction are generally regarded as cosmopolitan, and provide little evidence to reconstruct the paleolatitudinal position of individual areas (Schönaub, 1992). Lithologic data and improved information on fossil assemblages may improve this situation.

Conodonts from the Alpine Silurian have a close affinity with coeval faunas from central, southern, and southwestern Europe. Avalonian Britain and Baltic Gotland occupied a more equatorial position; consequently, the conodonts are more diverse (Bergström, 1990; Aldridge and Schönlaub, 1989).

The distribution of acritarchs suggests an intermediate position of the Alpine Silurian between the high latitude *N. carminae* and the tropical *Domasia-Deunffia* biofacies. Chitinozoans show close relationships with those from Bohemia, a connection most strongly shown in the upper Ludlow–lower Lochkovian (Paris and Kříž, 1984; Kříž et al., 1986; Dufka, 1992; Kříž, 1992).

Silurian trilobites from the Carnic Alps are closely related to those from Bohemia and other central European regions (Alberti, 1970). Affinities with Moroccan trilobites exist, but the trilobites are not yet studied in detail (Alberti, 1970).

According to Berry and Boucot (1967), Silurian graptolites show little endemism, and this suggests intercontinental dispersal. Their distribution may have been controlled mainly by surface oceanic currents that flowed between Silurian continents and volcanic islands. As noted by Jaeger (1976), essentially uniform Ludlow and Pridoli graptolite faunas developed in Europe. The changing environment of this time as seen in coeval changes in African and Baltic lithofacies includes a characteristic vertical change from black graptolitic shale to limestone and back to shale. Sea-level rises and falls are considered to have been responsible for these changes.

In the late Llandovery, nautiloids became the predominant organisms in Alpine carbonate facies. The abundant Wenlock and Ludlow orthoceratids then decreased in the Pridoli (Ristedt, 1968, 1969). These diverse Alpine faunas seem closely related to those of Bohemia, the Montagne Noire, and Sardinia (Kříž and Serpagli, 1993; Kříž, 1996, 1998, 1999). Ongoing studies show that Silurian cephalopod biofacies even reflect close links to northern Siberia. Supposedly, this relationship resulted from activity of a South Equatorial Current that flowed along the southern margin of Siberia and Laurussia (Kříž and Bogolepova, 1995).

The distribution of other mollusks, particularly bivalves, generally resembles that of nautiloids. According to Kříž (1979), Silurian cardiolids from the Carnic Alps and the western Graywacke Zone inhabited a warm equatorial belt or were dispersed by surface currents. Kříž (1999) recognized the oldest Silurian bivalve-dominated community of the *Cardiola* Community Group from the Carnic Alps (Kříž and Serpagli, 1993; Kříž, 1996, in press). This is the *Carnalpia nivosa* Community in the *Cyrtograptus rigidus* Zone (Wenlock). Other recurring communities of the *Cardiola* Community Group are also known from the Bohemian Prague Basin and other regions in Europe. In the Wenlock (*Cyrtograptus lundgreni* Zone), the *Cardiola agna* Community and its *Slava pelerina-Isiola zila* Subcommunity occur at the Rauchkofel Boden section. The lower Ludlow is characterized by the *Cardiola consanguis* Community, which is also known from the Prague Basin. The Cardiola Formation is characterized by the *Cardiola docens* and *Cardiola alata* Communities, which are known from the Prague Basin (Bohemia), Sardinia, eastern Serbia, the Montagne Noire, Spain, and Morocco. In the lowest Pridoli (*Monograptus parultimus* Zone), the *Cardiolinka bohemica* Community occurs at the Cellon section; at Nagelschmieddpalfen near Dienten in the

Graywacke Zone; in the Prague Basin (Bohemia); and at Elbersreuth, Frankenwald (Germany). In the uppermost Pridoli, a *Patrocardia-Dualina* Community (*Dualina nigra-Patrocardia* Subcommunity) occurs at the Rauchkofel Boden section. This latter community is related to the *Patrocardia evolvens evolvens* Community of the *Patrocardia* Community Group in the Lower Devonian (Lochkovian) of the Prague Basin, Sardinia, and South Armorican Domain (La Meignanne) in France. The *Cardiola* Community Group is characterized by epibyssate bivalves, which were adapted to the cephalopod limestone biofacies and indicate episodically ventilated, relatively shallow bottom conditions (=Boucot's [1975] Benthic Assemblage 2–3). The *Patrocardia* Community Group is characterized by epibyssate *Patrocardia* with infaunal and reclining Dualina, and also lived in the cephalopod limestone facies.

Silurian corals from the Alps were prominent constituents of a probable shallow-water facies in the tropical belt. During the Early Silurian, only weak indications of provincialism are seen among tabulate and rugose corals at the generic level. However, long-living teleplanic larvae might also have been transported by ocean currents over great distances (Kaljo and Klaamann, 1973; Pickett, 1975; McLean, 1985; Pedder and Oliver, 1990). Rugose and tabulate corals occur in the upper Llandovery of middle Carinthia and the Upper Silurian (Ludlow) near Graz. However, they are very rare in the shallow-water and local coated-grain-bearing limestones in the upper Llandovery of the Tyrolean Graywacke Zone (Schönlaub, 1994b).

Lithologic and faunal data from the Alps can be used to infer Silurian climates and to provide insights into such parameters as light, temperature, salinity, water agitation, and other factors that controlled organism distribution. During the Silurian, the Alpine facies belts shifted from higher to lower latitudes. Paleomagnetic data from Gondwana seem to support rather rapid northward plate movements (Schönlaub, 1992). Based on the evidence presented above, we estimate that Alpine Silurian deposition was at ca. 30–40° S. During the Silurian, close faunal relations existed with northern Europe, but minor links existed with other southern Europe regions.

Silurian of Germany

In Germany, the Silurian occurs in the Rhenohercynian Zone and, in particular, in the Saxothuringian–Lugian Zone of the Variscan orogen. Other occurrences are known from the intermediate area of the Mid-German Crystalline Rise and from the German part of the southern Variscan orogen, or Moldanubian Zone (Fig. 33).

Rhenohercynian Zone — In the Rhenohercynian Zone, the Silurian is exposed in the Harz Mountains (Fig. 32A) and Rhenish Slate Mountains (Figs. 32B). Wells drilled between Flechtingen and Rosslau also encounter the Silurian. These core rocks are similar to the Silurian of the Harz Mountains.

Harz Mountains — The Silurian crops out in the area of Bad Lauterberg, Hasselfelde, and Harzgerode in the lower Harz Mountains, and there are a few more northern occurrences in the central Harz Mountains. The Lower Silurian in the Harz Mountains consists of black to green shales, and the Upper Silurian is predominantly calcareous shales and dark limestones. Cherts, black carbon-rich alum shales, and phosphatic nodules, which are typical of the "lower graptolitic shales" in the Saxothuringian Zone (discussed below), are nearly absent. The occurrence of a fossil-rich limestone lentil in the bed of the Wieda River near Zorge (Fig. 33) is remarkable, but this lentil is no longer accessible. At least the lower part of it is Pridoli (Maronde, 1968). The fauna consists mainly of brachiopods, trilobites, bivalves, and orthoconic nautiloids; the lithology is that of a typical cephalopod limestone (Heritsch, 1930).

Graptolite studies by Jaeger (1991a) show that the base and top of the Silurian are exposed, as are sections through the upper Llandovery, Wenlock, most of the Ludlow, and the Ludlow–Pridoli transition (Maletz, 1996). It is probable that a complete Silurian succession occurs in the Harz Mountains. The thickness of the entire Silurian is between 50 m and 100 meters (Jaeger, 1991a). Nearly all Silurian occurrences in the Harz Mountains occur in Late Devonian and Early Carboniferous olistostromes.

Rhenish Slate Mountains — In the Rhenish Slate Mountains, the easternmost occurrences of Silurian graptolitic shales are known from the northeast Lahn syncline (Marburg region) and from south of the Kellerwald in an area north of Gilserberg (Figure 33). These Steinhorner Schichten are allochthonous and probably span the upper Llandovery–Pridoli.

In contrast, the Silurian of the Ebbe anticline (Herscheid area, Sauerland region) is characterized by dark shale, marly shale, and interbedded ochre-weathering limestone. These Köbbinghäuser Schichten are about 110 m thick. They are transitional into the overlying Ockrige Kalke—black shale interbedded with nodular, bluish-gray carbonate rocks. Bio- and lithofacies of the Köbbinghäuser Schichten indicate a shallowing of the basin during the Late Silurian (Timm, 1981a). In the Remscheid-Altenaer anticline (Bergisches Land), a Silurian succession comparable to that of the Ebbe anticline was deposited. The only difference may be the lesser thick-

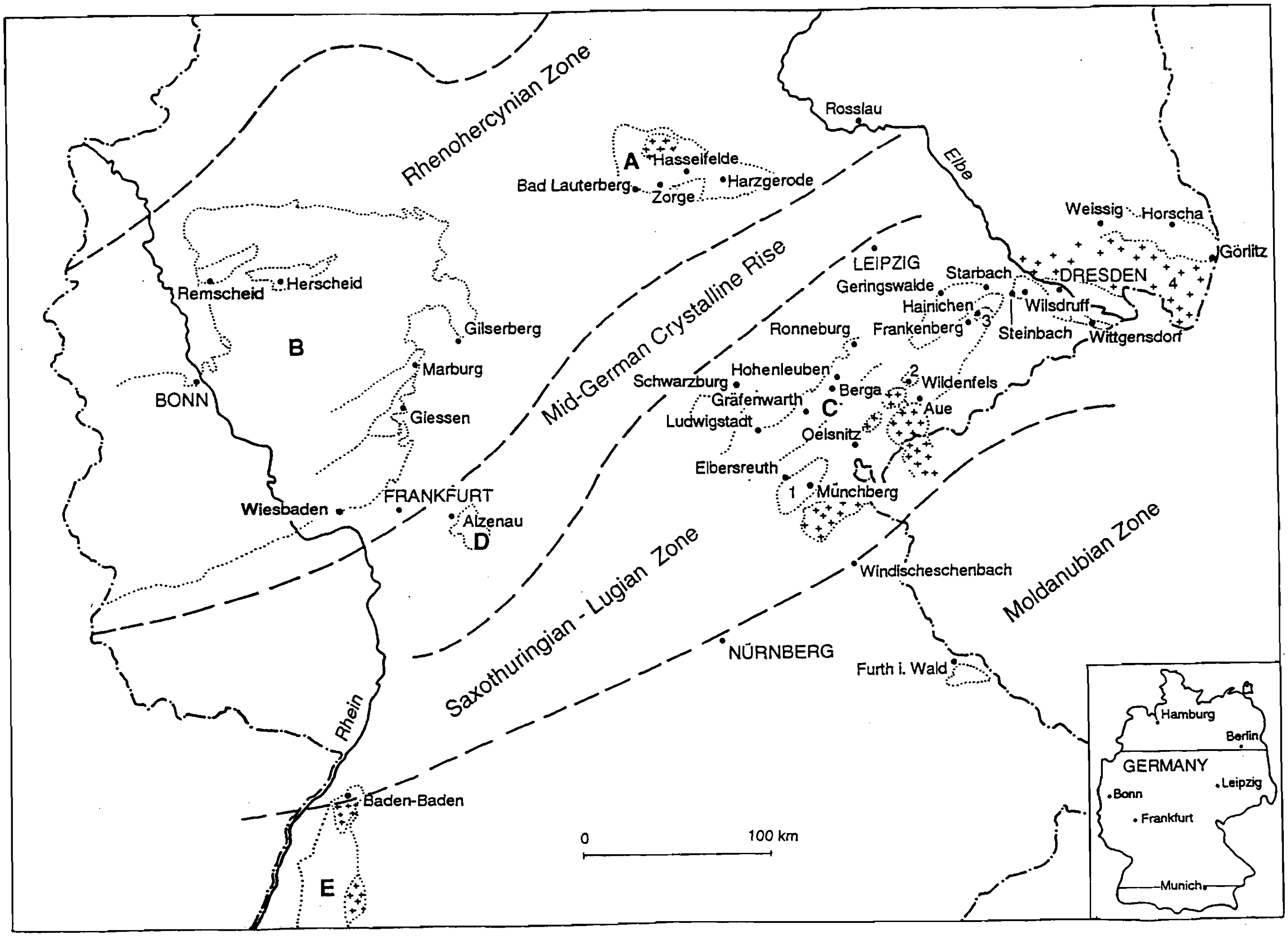

FIGURE 33 — Major tectonic units and Silurian outcrops of the Variscan orogen in Germany. Crosses indicate major Variscan and pre-Variscan plutons. A, Harz Mountains; B, Rhenish Slate Mountains; C, Thuringian–Vogtlandian Mountains; D, Spessart; E, Black Forest. 1, Munchberg gneiss massif; 2, 3, Betwixt Mountains (crystalline "Zwischengebirge") of Wildenfels and Frankenberg–Hainichen; 4, Lusatian granodiorite massif.

ness. Conodonts (Ziegler, 1960) and trilobites (Timm, 1981b) suggest the Silurian–Devonian boundary is in the lowermost Ockrige Kalke.

Silurian beds are also exposed south of Giessen in the Lindener Mark region. The upper Wenlock–Pridoli (*Monograptus transgrediens* Zone) thickness is only about 12 m. The succession starts with an ostracode-rich limestone that is thick-bedded in its lower part and interbedded with cherts in its upper part. The overlying beds consist of a 5 m-thick cephalopod limestone. According to Bahlburg (1985), a calcareous silty shale with limestone nodules in its lower part has orthoconic nautiloids, brachiopods, bivalves, and graptolites. The ostracode- and cephalopod-bearing limestones were deposited on the lower and upper subtidal shelf, respectively.

In contrast to Saxothuringia (discussed below), the Ordovician–Silurian boundary in the Rhenish Slate Mountains is marked by a distinct hiatus. Generally, most of the Llandovery is absent. The reason for this stratigraphic gap is still uncertain, but it is doubtful that it represents a real break in sedimentation. The Silurian–Devonian boundary in the Rhenish Slate Mountains lacks or has only minor breaks, and can be interpreted to represent nearly continuous shelf sedimentation at various local depths.

The southeastern margin of the Rhenish Slate Mountains abuts the metamorphic zone of the South Taunus Mountains near Wiesbaden (Fig. 33). In this area, probable Silurian rocks are metamorphosed to greenschist (Thews, 1996).

SAXOTHURINGIAN ZONE — The Lugian Zone east of the Elbe River and the Saxothuringian Zone were areas of continuous marine deposition throughout the Silurian. Northern Bavaria, Thuringia, and west Saxonia are part of the Saxothuringian Zone. East Saxonia lies in the German part of the Lugian Zone (Fig. 33). The thickness of the Silurian is no more than about 100 meters. Traditionally in the Saxothuringian–Lugian Zone, the Silurian and lowermost Devonian are regarded as a single depositional cycle composed of lower and upper graptolitic shales. These shales are separated by a widely persistent carbonate, the "Ockerkalk" (ochre limestone) (Fig. 34). Because of the mainly early Lochkovian age of the "lower graptolitic shales," these beds are not discussed herein.

Silurian strata are well documented in all of these areas. Tentaculitids (e.g., *Tentaculites scalaris*) in phyllites of the Traischbach "Series" of the Baden-Baden-Zone in the northern Black Forest (Fig. 33, region A) are probably Late Silurian (Mehl, 1989). This area is also considered to be part of the Saxothuringian Zone, while the southern Black Forest is part of the Moldanubian Zone.

Because of its relatively widespread distribution, the litho- and biofacies of the Silurian in the Saxothuringian Zone must be further detailed. The typical Silurian of the Saxothuringian Zone is graptolitic shale in the Lower Silurian, and the "Ockerkalk" is a peculiar limestone in the Upper Silurian.

"Lower graptolitic shale" — This sequence consists of black alum shales with high carbon and pyrite contents, and has black, thick-bedded cherts. The two rock types intergrade. Normally, chert is dominant in the Llandovery, and interbedded shale and chert or alum shale are predominant in the Wenlock and lower Ludlow. Because of weathering, the Llandovery cherts are often bleached.

The upper alum shales of the Ludlow are interbedded with thin layers of argillaceous shale that is transitional into the "Ockerkalk" of the Thuringian facies (Fig. 34), or into gray-green shales in the Bavarian facies. In the interval from the *Pristiograptus dubius parvus-Gothograptus nassa* Zone to the *Neodiversograptus nilssoni* Zone, but particularly in the *N. nilssoni* Zone, there is a striking concentration of phosphatic nodules. Such layers are also known in the *Stimulograptus sedgwickii* and *Cystograptus vesiculosus* Zones (Schauer, 1971). In a few Thuringian sequences, black dolomitic layers up to 0.5 m thick are intercalated in the shale. The "lower graptolitic shales" are pure sapropelites. An enrichment in such elements as vanadium, molybdenum, selenium, uranium, and sulfur is typical for its anoxic or strongly dysaerobic depositional environment.

Strongly deformed graptolites are the only abundant fossils found throughout this condensed sequence. Rare conodonts are present on bedding planes at Gümbelit. White quartz lenses in the cherts contain radiolarians. In Thuringia, beds transitional into the "Ockerkalk" have very rare bivalves (e.g., *Cardiola*), indeterminable brachiopods, very rare myodocopid ostracodes, orthoconic nautiloids, and eurypterids (Jaeger, 1959; Schauer, 1971). The absence of any typical benthic organisms suggests extremely unfavorable bottom conditions.

For the most part in the Saxothuringian–Lugian Zone, the "lower graptolitic shales" are not completely exposed. In the Görlitz Slate Mountains, Llandovery cherts occur at the Pansberg near Horscha and at the Eichberg near Weissig (Fig. 33). In the NossenWilsdruff Slate Mountains (Fig. 33), the "lower graptolitic shales" are a 25 m sequence of Llandovery chert and Wenlock alum shale. In the Slate Hills of the Elbe valley, only a few meters of graptolitic shales are exposed, as at the Sandberg near Wittgensdorf.

A few outcrops of slaty "lower graptolitic shales" are known at the margin of the granulite massif in Saxony. These include a locality at Geringswalde in the northern part, from an area north of Frankenberg (Zschopau Valley), and from Hainichen in the southern part. Upper Llandovery chert and alum shale are exposed in abandoned quarries north of Aue and between Loßnitz, Zwönitz, and Affalter in the Erzgebirge Mountains (Fig. 33). Llandovery cherts also occur near Altmannsgrün and at the Engelspöhl, southeast of Oelsnitz (Fig. 33).

"Lower graptolitic shales" are known from numerous outcrops along the flanks of the anticlines in the Thuringian and Vogtlandian Slate Mountains. Representative localities include an outcrop near Gräfenwarth (Jaeger, 1991b) and the Lichtenberg quarry near Ronneburg (Fig. 33). According to bore hole data, their thickness may be more than 50 m locally. At almost all outcrops in the slate mountains, these shales were affected by strong tectonism (Jaeger, 1959).

The "lower graptolitic shales" in the Frankenwald region of northeast Bavaria were described by Stein (1965), Greiling (1966), and Zitzmann (1968). The ca. 3,040 m-thick sequence consists predominantly of chert and alum shale. The *Stimulograptus sedgwickii* Zone (Llandovery) and, in particular, the Wenlock have tuffs and basalts.

"Ockerkalk" — The term "Ockerkalk" was introduced by Gümbel (1863). The name was later used for all calcareous beds that separate the "lower" and "upper graptolitic shales" in Thuringia and Saxony. However, there are at least minor local differences between the same limestones in different areas.

The typical "Ockerkalk" lithology occurs in the western Thuringian Slate Mountains (Schwarzburg anticline), and is a thick-bedded (up to nearly 3 m), dense, bluish-gray to grayish-black limestone with irregular nodular texture. Pyrite often occurs as framboidal aggregates up

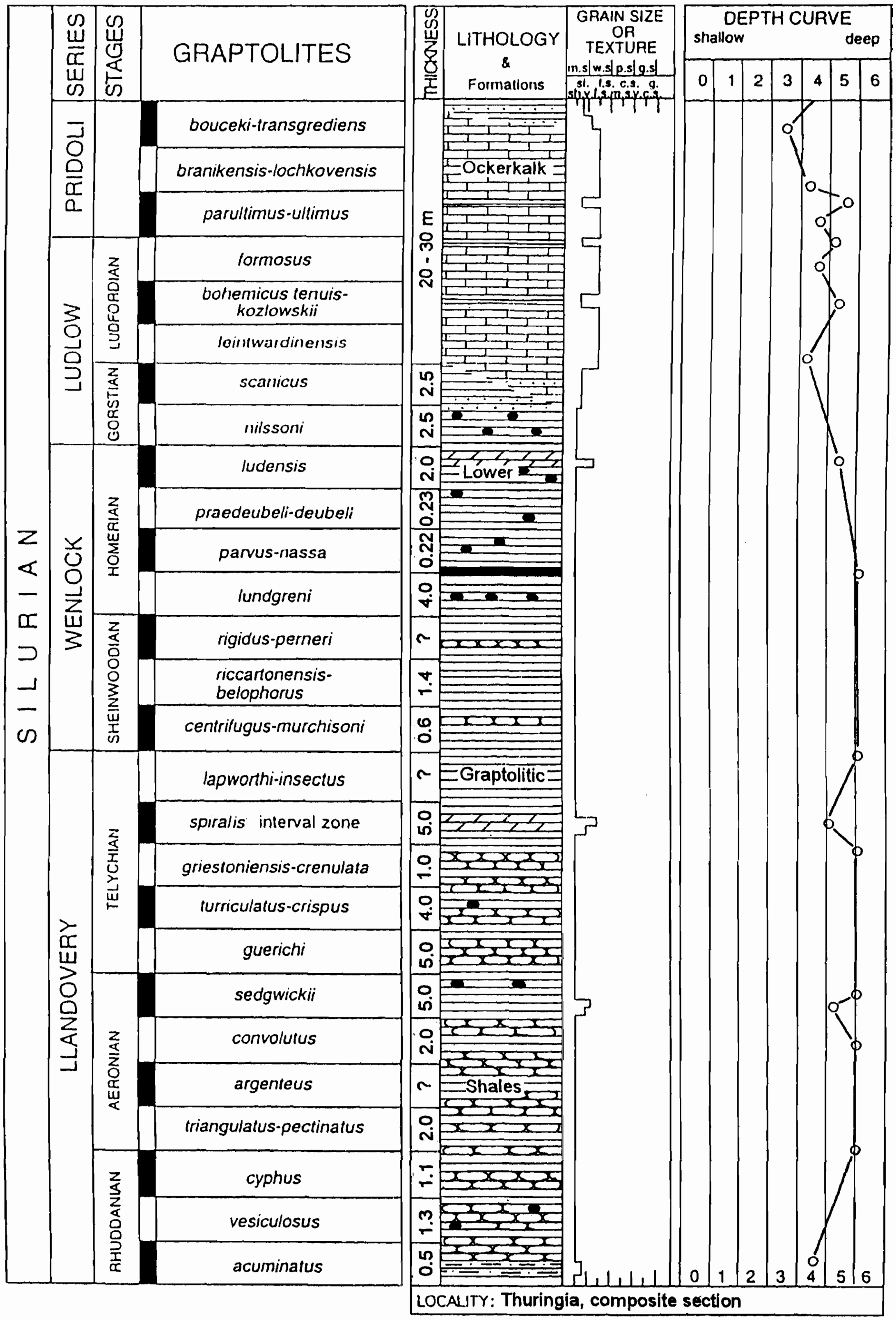

FIGURE 34 — Silurian of Thuringia, Germany. After Jaeger (1959, 1977b, 1988, 1991b), Schauer (1971), Hansch (1993a, and unpublished data).

to 3 cm in size. Insoluble residue normally forms 5–35% of the limestone and consists of quartz, pyrite, mica, and chlorite. The typical "Ockerkalk" weathers ochre. Because of the higher content of iron in the dolomite, the nodular bands discolor first with weathering.

The "Ockerkalk" differs slightly elsewhere in the Saxothuringian Zone. The limestones may not locally show the nodular texture, and the typical weathering features may be missing. Locally, there is a transition from the "Ockerkalk" to gray or light-colored marls, or the pure limestones are replaced by dolostones. Following intrusion of Late Devonian mafic rocks, the "Ockerkalk" is metamorphosed at a few outcrops in the Berga anticline.

The "Ockerkalk" contains interbeds of black alum shale, locally with phosphatic nodules, and interbeds of grayish-green argillaceous or arenaceous shale. For this reason, it has been called the Ockerkalk Group (Jaguar, 1959). The siliciclastic layers range from a few centimeters to ca. 1 m in thickness. At least two thick shale intercalations can be used to correlate between eastern and western Thuringia (W. Hansch, unpublished data, 1997). The shale content of the "Ockerkalk" varies between 10–30%. Generally, the "Ockerkalk" can be considered a typical condensed pelagic limestone.

In contrast to the "lower graptolitic shales," the fossils of the "Ockerkalk" are more diverse. On the other hand, the fauna is imperfectly known because of the lack of modern descriptions. Apparently, the fauna is impoverished. The following fossils are known: rare articulate brachiopods, a few trilobites that were blind or had highly reduced eyes, very rare conulariids and gastropods (preserved as molds), very small solitary corals, orthoconic nautiloids, indeterminable bryozoans, a few foraminiferans (*Psammosphaera* sp., *Tolypammina* sp., hyperamminoid-type forms), and very rare bivalves (Cardiola sp.). Jaeger (1977b) listed the trilobites *Ampyx* sp. cf. *A. rouaulti*, *Harpes* sp. cf. *H. ungulata*, *Cheirurus* sp. cf. *C. propinquus*, *Denckmannites caecus*, and *Scutellum* sp. The shale interbeds yield some graptolites and very rare eurypterids. Crinoid fragments are common and abundant in the upper "Ockerkalk" (*Scyphocrinites* horizon). Conodonts occur, but are not yet described. The most diverse faunal component is silicified ostracodes characterized by thin-shelled, smooth, or sometimes spine-bearing podocopes and a few paleocopes. At least 40 species of ostracodes are known (Hansch, 1993a). The ostracodes apparently lived in a relatively low-energy, open-marine environment. These ostracodes and conodonts may have the best biostratigraphic potential in the Thuringia Slate Mountains. Most "Ockerkalk" fossils are known from the western Thuringian Slate Mountains. This may reflect better habitats associated with the more carbonate-rich facies in the west.

The "Ockerkalk" has been used for ochre and building stone since the seventeenth century, particularly in Thuringia. After World War II, the "Ockerkalk" received study in East Germany because of its uranium content. The average content was between 1.5–5 gm/metric ton, and its mining ended in 1990.

The thickness of the "Ockerkalk" in the Saxothuringian Zone probably decreases from west to east. According to bore hole data, the greatest thickness (ca. 40 m) occurs in the Schwarzburg anticline. At present, the only well-exposed outcrop of "Ockerkalk" (24.5 m) occurs in the completely exposed Silurian in the Lichtenberg quarry near Ronneburg. In the eastern part of the Saxothuringian Zone, the "Ockerkalk" is 12 m thick in the Triebisch Valley near Steinbach in the Nossen–Wilsdruff Mountains. In northeast Bavaria, a 15-m thick "Ockerkalk" section was completely exposed at the Löhmar mill northwest of Münchberg. Additional outcrops are known in this region in the Ludwigstadt area.

OTHER SILURIAN LOCALITIES IN GERMANY — According to Reitz (1987), well-preserved Upper Silurian (probably Ludlow) spores are found in a quartzite mica schist unit in the northwest Spessart near Alzenau. The Spessart is part of the Mid-German Crystalline Rise. Pflug and Reitz (1987) described Silurian spores in the Moldanubian Zone from the southeast Hoher Bogen Mountains near Furth in Wald, eastern Bavaria. According to Pflug and Prösl (1989), Late Silurian microfossils and plant fragments were also found in paragneiss of the KTB research bore hole near Windischeschenbach, northeast Bavaria.

GERMAN SILURIAN FACIES — In the Cambrian–Lower Carboniferous of the Saxothuringian–Lugian Zone, two types of facies are distinguished. These are the Thuringian and Bavarian facies. The boundary between these facies is not sharp, and there are transitional zones (Gandl, 1992).

THURINGIAN FACIES — The "lower graptolitic shale"–"Ockerkalk" succession is the typical Silurian sequence in the Thuringian facies. This succession is widely distributed in the Saxothuringian Zone from northeast Bavaria to west Saxony. In the Lužice Zone (Fig. 35) east of the Elbe River, limited Silurian occurrences are known from the southern margin of this zone in Bohemia.

The Thuringian facies is a monotonous basinal facies with only moderate lateral changes, particularly in the Upper Silurian. In the Lower Silurian, its thinness, predominance of planktic organisms, absence of current-orientated fossils, and lithology indicate a calm, open-marine environment for the deposition of the "lower graptolitic shales." This anoxic or highly dysaerobic environment changed during deposition of the Ockerkalk

Group. Thick-bedded carbonates, intercalations of thin quartzite beds which may be distal turbidites, monospecific concentrations of ostracode valves that indicate weak bottom currents, and the greater diversity of fossil groups show a distinct upward change to a better-oxygenated environment. On the other hand, varying thicknesses of the limestones and intercalated shales, the differing abundances of fossils, and changing carbon and pyrite content in the limestones indicate that sedimentation and the environment were not uniform across the Saxothuringian Basin. However, at no time during deposition of the Ockerkalk Group were habitats suitable for a rich benthos.

BAVARIAN FACIES — As in the Thuringian facies, typical Bavarian facies rocks are interbedded black graptolitic cherts and alum shales. The main difference with the Thuringian facies is merely that carbonate sedimentation did not take place even during the Ludlow and Pridoli. Thin, gray-green shales, with a thickness of ca. 5.0 m (according to bore hole data in the Görlitz hills), correlate with the "Ockerkalk." In contrast to the wide distribution of the Thuringian facies in the Saxothuringian Zone, the Bavarian facies forms a discontinuous belt confined to narrow strips on either side of the Münchberg gneiss massif and the Betwixt Mountains near Wildenfels and Frankenberg (Jaeger, 1988). East of the Elbe River, the Bavarian facies is known from small outcrops and bore holes north of the Lusatian granodiorite massif (Fig. 33). Outside of these areas, the Bavarian facies only comprises 20m of Silurian shale in the slate mountains of the Elbe valley southeast of Dresden and in the Nossen-Wilsdruff Slate Mountains near Starbach.

A cephalopod limestone ("Elbersreuther Orthoceratenkalk") is limited to an area between Elbersreuth and Wildenstein-Triebenreuth and in the Steinbachtal, west of Münchberg (northeast Bavaria). This limestone represents Bavarian facies sedimentation on a submarine rise. It is a light gray-red, dense, fossil-rich limestone with interbedded arenaceous layers and a thickness of ca. 10 m. Its deposition probably began in the early Ludlow, following volcanic activity in this region in the late Llandovery and Wenlock.

Since the 1920s, the contrast between these facies has led to the idea that the Bavarian facies, the Münchberg gneiss massif, and the crystalline complex of the Betwixt Mountains are elements of large nappes that were transported from the Moldanubian Zone (Franke, 1984, 1995). If the Silurian alone is examined, its facies patterns can be explained without the assumption of nappes (Gandl, 1992).

In the Rhenohercynian Zone, similar facies are obvious. The Silurian of the Harz Mountains, the Kellerwald region, and part of the Giessen area show relatively close relationships with northern Gondwana facies. However, the facies in the Ebbe and Remscheid–Altenaer anticlines are probably more closely linked to the outer shelf facies of Baltica (discussed below).

In summary, the German Silurian on the northern Gondwana margin is characterized by stratigraphically continuous, widely distributed, anoxic or highly dysaerobic black shale with planktic organisms in the Early Silurian. Laterally discontinuous carbonate sedimentation took place in the Late Silurian, during which time the habitats apparently became more oxygenated.

The Late Silurian featured bathymetrically and hydrodynamically different sedimentation areas. Areas with carbonate sedimentation in shallower and well-oxygenated water allowed development of a rich shelly fauna. This facies is represented by the "Elbersreuther Ortoceratenkalk" in northeast Bavaria, by the Pridolicephalopod limestone of the Giessen area (Lindener Mark), and by the Wieda stream lens (Harz Mountains). Other areas have discontinuous, rarely thick-bedded carbonates or nodular limestones interbedded with shale or marl. This deposition took place in poorly oxygenated deeper water, which supported an impoverished benthic fauna but allowed the appearance of more diverse planktic and pseudoplanktic faunas with graptolites, crinoids, ostracodes, conodonts, orthoconic nautiloids, and foraminiferans. This facies is best represented by the Ockerkalk Group of Saxothuringia and northeast Bavaria and, possibly, the Ostracodenkalk of the Giessen area. The latter areas featured condensed but continuous shale deposition in deeper, probably poorly oxygenated water, with planktic graptolites and orthoconic nautiloids. This facies comprises gray-green shales found in small outcrops and bore holes close to the Münchberg gneiss massif near the Görlitz hills.

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY — The Silurian Thuringian facies in the Saxothuringian Zone is compiled in Fig. 34. The Ordovician–Silurian boundary in the Thuringian facies is represented by a transition from buff-weathering, black siliciclastic mudstone with a high mica content in the uppermost Ordovician (Lederschiefer) to lowest Silurian cherts and alum shales. The Ordovician–Silurian boundary is known in the Lichtenberg quarry near Ronneburg, in the Weinberg area near Hohenleuben south of Gera, and from the Engelspöhl southeast of Oelsnitz. On the other hand, the boundary in the Bavarian facies is represented by a transition from nearly black Upper Ordovician quartzitic sandstone with shale interbeds (Döbrasandstein) into Silurian shales. It is best exposed along the northwest side of the Münchberg gneiss massif, at Döbra and at Starbach in the Nossen–Wilsdruff Mountains.

The Silurian–Devonian boundary is located within the first meter above the Ockerkalk Group. There is no distinct lithological change at the boundary, which is exposed at various outcrops in Thuringia (Jaeger, 1977a, 1977b). The most accessible section through the boundary is in the Lichtenberg quarry near Ronneburg. At the base of the "lower graptolitic shales" is the "lower shelly bed horizon" with abundant bivalves (*Pterinea* sp.), gastropods (*Platyceras* sp.), crinoids (*Camarocrinus?* sp.), and rare spiriferids.

The Llandovery–Wenlock boundary is not well defined in the Saxothuringian–Lugian Zone. Only the basal Wenlock graptolite zone is known in the Ronneburg area. The best known section of the Wenlock–Ludlow boundary is located at the Wetterberg, near Gräfenwarth (see Jaeger, 1991b). The Ludlow–Pridoli boundary is not established in the Saxothuringian–Lugian Zone because of the lack of index graptolites.

The succession and thickness of graptolite zones in the Thuringian facies column (Fig. 34) are based on Jaeger (1959), Schauer (1971), and unpublished data. According to Jaeger (1959), deposition of the "Ockerkalk" always started within the *Lobograptus scanicus* Chron. Because of the remarkable differences in thickness within the Saxothuringian Zone, it is possible that the onset of "Ockerkalk" sedimentation was diachronous. This was suggested by Stein (1965) for northeast Bavaria.

The "Ockerkalk" has lower and upper ostracode associations separated by a thick shale interval. The ostracodes of the upper "Ockerkalk" show some similarity to the Lower Devonian ostracode fauna of the tentaculite nodular limestone (Tentakulitenknollenkalk) (Hansch, 1993a).

GERMAN SILURIAN PALEOBIOGEOGRAPHY — Comparable information on Silurian fossils is not available for the Rhenohercynian and Saxothuringian Zones. Few of the Silurian rocks, with exception of the cephalopod limestones, are rich in benthic fossils. Even the planktic faunas are impoverished or badly preserved. Only the graptolites, based largely on nearly 40 years of work by H. Jaeger, L. Greiling, V. Stein, M. Schauer, and others, are relatively well studied. Nearly all other fossil groups lack modern systematic descriptions. Nevertheless, relationships between the Silurian faunas of the Rhenohercynian and Saxothuringian Zones and other regions can be briefly summarized.

Though not recently studied (see Schindewolf, 1924; Heritsch, 1930; Kegel, 1953), Upper Silurian cephalopod limestone faunas from the Elbersreuth area (northeast Bavaria), Harz Mountains, and Lindener Mark south of Giessen are obviously comparable with other cephalopod limestone faunas of the Prague Basin, Carnic Alps, and southwest Sardinia.

As noted by Jaeger (1976) and Barca and Jaeger (1990), Silurian and lowest Devonian lithologies and faunas (graptolites) are comparable in Thuringia and southeast Sardinia. The ostracodes also confirm this similarity (W. Hansch, unpublished data, 1996). Additionally, all three major Upper Silurian facies of the Saxothuringian Zone exist in Sardinia.

Genus-level relationships exist between the ostracode fauna of the upper Wenlock–Ludlow limestone near Giessen (Lindener Mark) and faunas of Baltoscandia and the Saxothuringian "Ockerkalk." A few ostracodes of Bohemian character also occur in the Lindener Mark (Hansch, 1993a, b, 1995). The Giessen ostracode limestone only has species typical of a low-energy environment (offshore to open marine, below wave-base; Schallreuter, 1991, 1995; Hansch, 1994, 1995). The Saxothuringian Ockerkalk ostracodes are distinct at the species level from the other faunas. However, the genera have a distinct affinity with the Prague Basin fauna.

The index graptolite *Monograptus transgrediens* and a crinoidal limestone (*Scyphocrinites* horizon) in the upper Pridoli are typical for the Rhenohercynian and Saxothuringian Zones (e.g., in the Harz Mountains, Kellerwald region, and most areas of Saxothuringia). This points to a common paleogeography. The Silurian of the Ebbe and Remscheid–Altenaer anticlines is characterized by a lack of graptolites, a relatively great thickness (>110 m; Timm, 1981a), and trilobites with Baltoscandian affinity.

Lithologic and faunal data indicate that the Silurian of the Saxothuringian–Lugian zone, Harz Mountains, and Kellerwald region was deposited on the north Gondwanan shelf, whereas the Silurian of the Sauerland and Bergisches Land regions (Rheinish Slate Mountains) is more closely related to the east Avalonia–Baltica margin. The position of the Giessen area is more problematic because of its faunal similarity to Baltica and to the margin of Gondwana. Oczlon (1994) considered the Harz Mountains, Kellerwald region, and Giessen area to be parts of a Harz terrane on the outer Gondwanan shelf, and discussed the possible tectonic implications. However, Franke and Oncken (1995) assumed that the Silurian carbonates of the Harz Mountains, Kellerwald, and Giessen area were deposited on the northern margin of the Armorica microcontinent, and were separated from Avalonia by the Rheic Ocean. They considered that Armorica collided with Avalonia in the Early Devonian. Renewed rifting split the intervening Silurian arc and left Armorican sedimentary rocks stranded on the Avalonian side. The result of this tectonic history is that Silurian carbonates with Armorican (Gondwanan) fossils occur as olistoliths and displaced blocks in mud flows (olistostromes) in the Rhenohercynian Zone (e.g., Harz, Geissen area) in the Devonian and Lower Carboniferous.

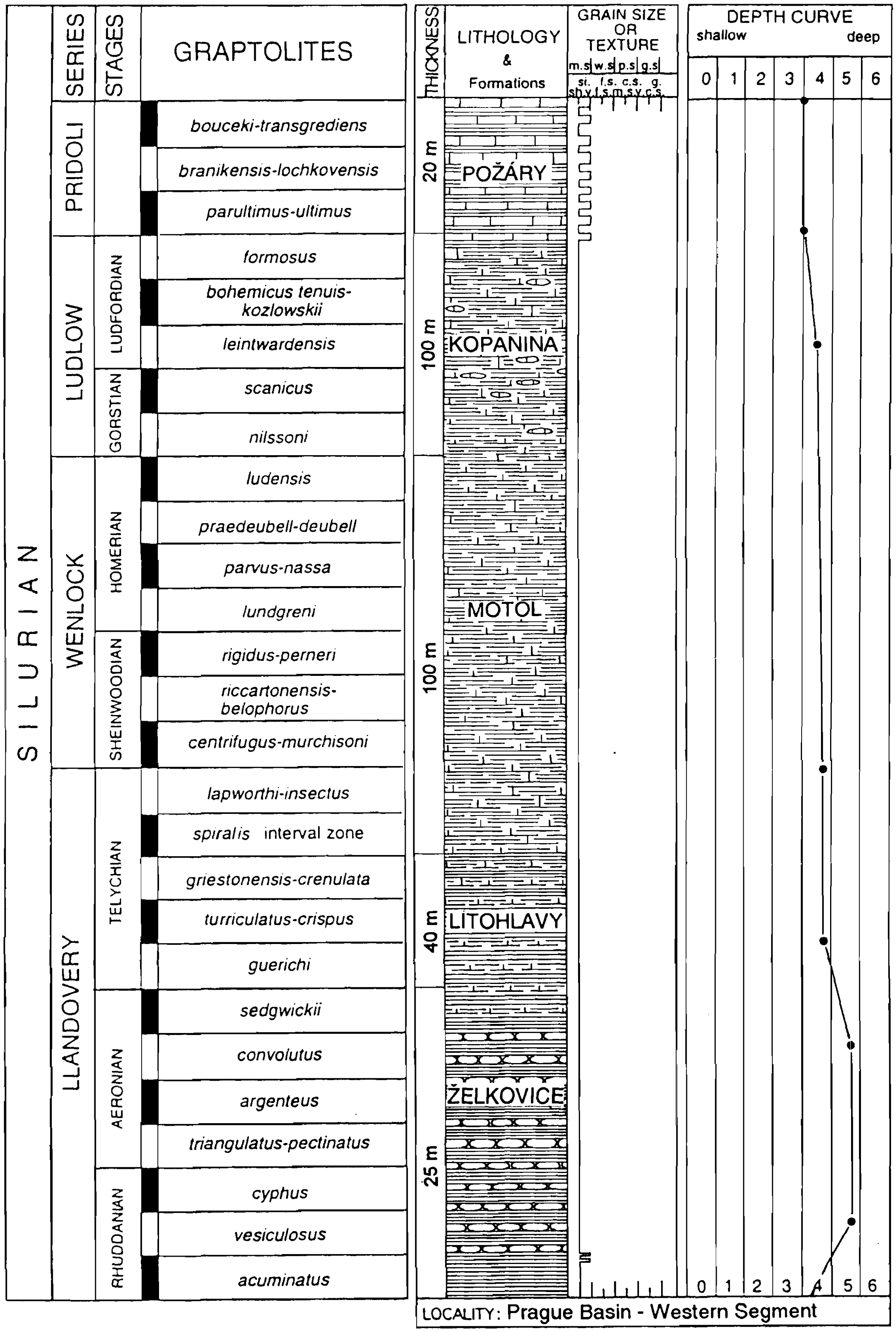

FIGURE 36 — Silurian of western Prague Basin, Bohemia. For detail, see Fig. 38.

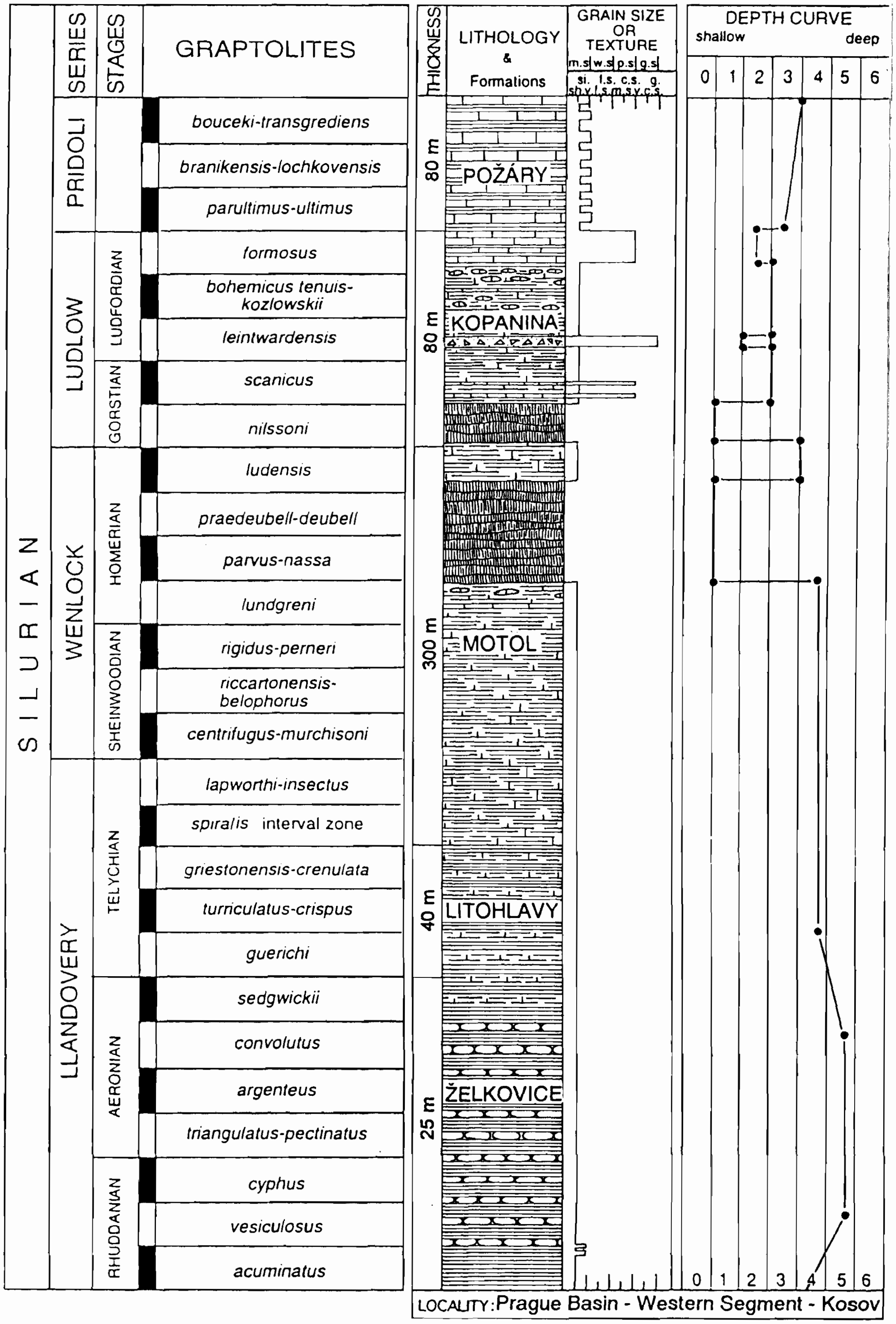

FIGURE 37 — Silurian of Kosov volcanic center, western Prague Basin, Bohemia. For detail, see Fig. 38.

The Prague Basin Silurian is more than 580 m thick in regions of igneous activity and maximum deposition of volcaniclastics (Kosov and Svatý Jan volcanic centers). In other areas, the dominant shale facies reach a maximum of 250–450 m in the southwestern Prague Basin (Fig. 39). Thicknesses are much less in the north- and southeastern basin.

The Silurian near Stínava in Drahany (Moravo–Silesian region) is similar to that of the Prague Basin. The lower part of the sequence is dominated by black graptolitic shale (upper Telychian) and the upper part by calcareous shale with limestone nodules that contain cephalopods, bivalves (*Cardiola*, *Patrocardia*, and *Dualina*), and crinoids in the lower Gorstian. The entire thickness is about 100 m (Bouček, 1935).

BIOSTRATIGRAPHY — The Bohemian Silurian is dated mostly by graptolites (Kříž, 1994). In the west Sudetes at Poniklá, northwest of Jilemnice, Horný (1964) described Wenlock graptolites from phosphatic concretions in graphitic phyllites. Ockerkalk Group limestones in the upper part of this sequence have columnals and stems of *Scyphocrinites* (Chlupáč, 1993). Graptolitic shales occur in the Telychian–Sheinwoodian boundary interval, the Homerian, and the lower Gorstian in the Rožmitál area in the Bohemian "Islet Zone" (Havlíček, 1977). In the Mirovice metamorphic "islet," the Homerian is dated by graptolites (Štorch et al., 1984). Two horizons in the Sedlčany–Krásná Hora metamorphic "islet" in the middle Llandovery and at the Llandovery–Wenlock boundary are dated by graptolites (Chlupáč, 1986). In the Železné Hory Mountains, Silurian phyllitic black shales have uppermost Telychian, Sheinwoodian, Homerian, and lowest Gorstian graptolites and *Scyphocrinites* in Pridoli black limestones (Svoboda and Prantl, 1950). In the Hlinsko region, Aeronian and Telychian graptolite zones are recognized (Horný, 1956).

The Prague Basin Silurian is well dated paleontologically (e.g., Horný, 1955, 1962, and Kříž, 1991, 1992). Štorch (1994) developed a zonal scheme for the Llandovery and Wenlock in the Prague Basin and recognized 27 graptolite zones. In the Ludlow, Štorch (1995) recognized eight graptolite zones, and H. Jaeger (*in* Kříž et al., 1986) recognized six in the Pridoli. The conodont biostratigraphy has been documented by Walliser (1964), who defined eleven successive zones in the carbonate facies. The Wenlock–Ludlow boundary interval was studied by H. P. Schönlaub (*in* Kříž et al., 1993). The lowest *Pristiograptus dubius parvus-Gothograptus nassa* Zone–*Colonograptus colonus* Zone is characterized by the acme of *Ozarkodina bohemica*, which has three morphotypes below and above the Wenlock–Ludlow boundary. Bed-by-bed conodont biostratigraphy has been done through the Ludlow–Pridoli boundary (H. P. Schönlaub *in* Kříž et al., 1986). In particular, the *Polygnathoides siluricus, Ozarkodina snajdri, O. crispa* and *O. eosteinhornensis* Zones are well documented. Chitinozoan biostratigraphy through the Ludlow–Pridoli boundary has been detailed by F. Paris (*in* Kříž et al., 1986). Dufka (1992, 1995; P. Dufka *in* Kříž et al., 1993; Dufka et al., 1995) applied the global Lower Silurian chitinozoan zonation to the Prague Basin (see Verniers et al., 1995). An ostracode biostratigraphy has been developed by Bouček and Přibyl (1955) and Hansch (1993b).

Other fossil groups allow correlation of the Prague Basin Silurian with that of other Gondwana basins. Bivalves allow correlations with western Macedonia (Bouček et al., 1968), the Moesian Platform of Romania (Kříž and Iordan, 1975), eastern Serbia (Kříž and Veselinovič, 1975), Sardinia (Kříž and Serpagli, 1993), the Taimyr Peninsula of Russia (Kříž and Bogolepova, 1995), the Montagne Noire and Mouthoumet Massif in France (Kříž, 1996), and the Carnic Alps (Kříž, 1999).

The Silurian near Stínava, in Drahany, (Moravo–Silesian region; Fig. 34) is correlated with the *Oktavites spiralis* and *Stomatograptus grandis* Zones (upper Telychian) and the *Cyrtograptus lundgreni* and *Pristiograptus ludensis* Zones (Homerian). Graptolites and bivalves (*Cardiola*, *Patrocardia*, and *Dualina*) allow correlation of the uppermost Silurian with the lower Gorstian *Colonograptus colonus* Zone (Bouček, 1935).

BOHEMIAN SILURIAN COMMUNITIES — Graptolite associations were studied through the Lower Silurian and into the upper Pridoli by H. Jaeger (*in* Kříž et al., 1986) and Štorch (1994, 1995) in the Prague Basin. Benthos-dominated communities are known in the Bohemian Massif only from the Prague Basin facies and from the shelly-fauna facies near Stínava, Drahany. In the Prague Basin, Chlupáč (1987) recognized one trilobite-dominated assemblage in the upper Aeronian, one in the middle Sheinwoodian, five in the Homerian, nine in the Ludlow, and three in the Pridoli. Kříž et al. (1993) revised the Wenlock–Ludlow boundary trilobites. These trilobite assemblages are lithofacies-related. Most assemblages occur in light-colored, fossiliferous limestones which were deposited in the shallow subtidal to intertidal zones (i.e., Boucot's [1975] Benthic Assemblages 2–3). Four assemblages occur in the dark grey and bituminous limestone facies (Benthic Assemblage 4), which was deposited in oxygen-deficient environments (Ludlow–Pridoli). Seven trilobite assemblages are associated with the Wenlock–Ludlow volcanic archipelago, and occur in tuffaceous, calcareous shales and carbonates (Benthic Assemblages 4–6).

Havlíček and Štorch (1990) analyzed the brachiopod-dominated benthic communities from the Prague Basin. One community was described from the Aeronian, three

from the Sheinwoodian, six from the Homerian, two from the Gorstian, five from the Ludfordian, and four from the Pridoli. These communities are strictly related to lithofacies and depth. Most communities occur in shallow environments (Benthic Assemblages 2–4) around the Wenlock–Ludlow volcanic archipelago. Few communities occur in calcareous shale (Benthic Assemblages 5–6). Kříž et al. (1993) described two additional communities from the upper Homerian and Homerian–Gorstian boundary. Havlíček (1995) distinguished five brachiopod biofacies in the Homerian and lowermost Gorstian. Brachiopods are the dominant benthic elements in the non-strophic-brachiopod-dominated biofacies, the pyroclastic biofacies, and the crinoid-stromatoporoid-coral biofacies. They are uncommon in the deeper-water trilobite-dominated biofacies, and very rare in the graptolitic shale biofacies. In general, the brachiopods of the Homerian differ from those of the lower Gorstian, but they are not useful in defining the Wenlock–Ludlow boundary. Kříž et al. (1993) showed that several species cross this boundary without any change in shell morphology.

Silurian communities dominated by bivalves were grouped by Kříž (1997a, in press) into four community groups that are related to lithofacies and depth of deposition (Fig. 40). The shallow-water *Cardiola* Community Group is represented in the Prague Basin by six *Cardiola*- and one *Cardiolinka*-dominated community that characterize the episodically oxygenated cephalopod limestones (Benthic Assemblage 2–3). The *Cheiopteria* Community Group is represented by two recurrent *Cheiopte-*

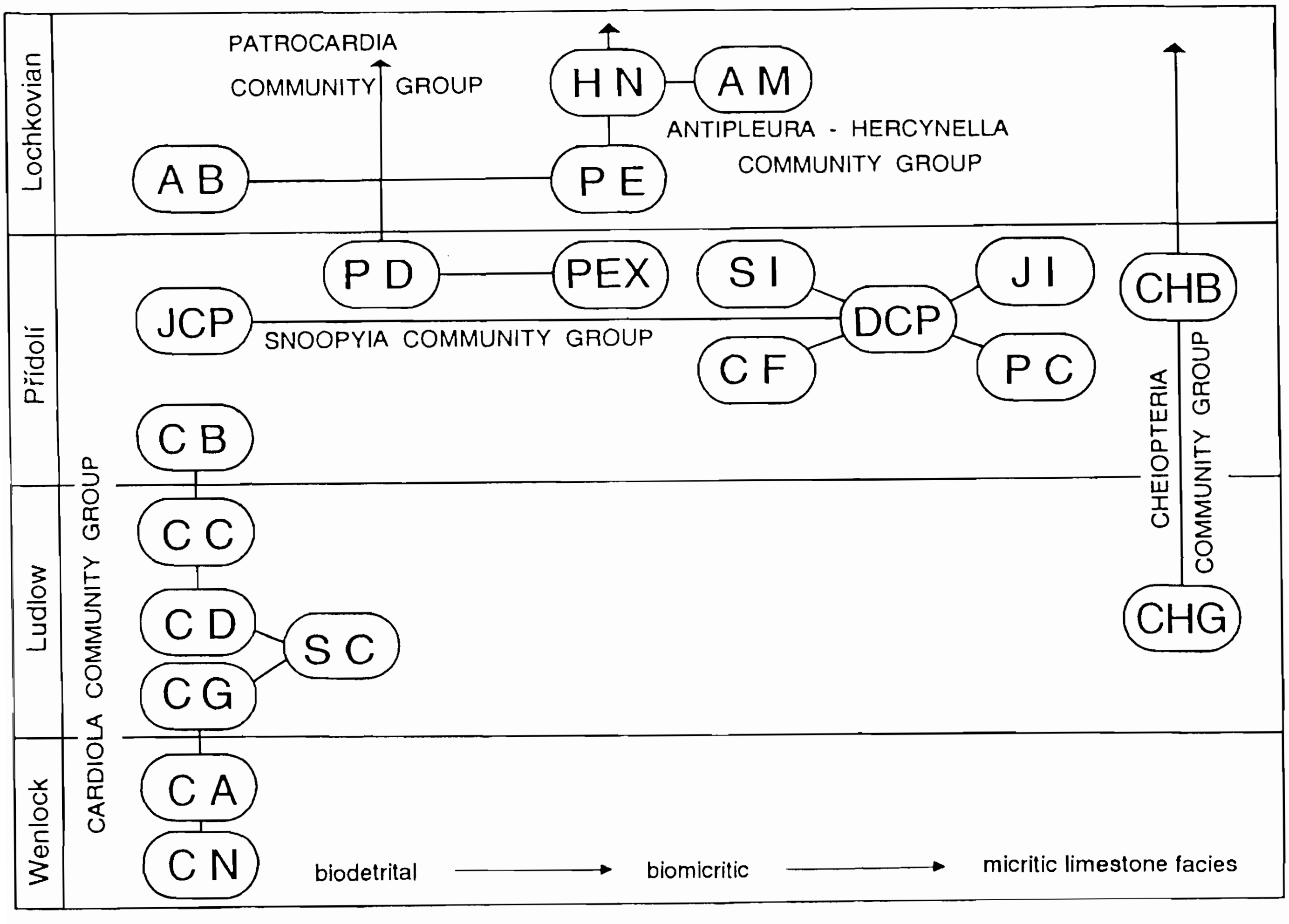

FIGURE 40 — Bivalve-dominated community succession and environmental framework for the Silurian and Lower Devonian carbonates of North African Gondwanan and Perunican facies. AB, *Antipleura bohemica* Community (=Cm.); AM, *Actinopteria migrans migrans* Cm.; CA, *Cardiola agna* Cm.; CB, *Cardiolinka bohemica* Cm.; CC, *Cardiola conformis* Cm.; CF, *Cardiolinka fortis* Cm.; CG, *Cardiola gibbosa* Cm.; CN, *Carnalpia nivosa* Cm.; CHB, *Cheiopteria bridgei* Cm.; CHG, *Cheiopteria glabra* Cm.; DCP, *Dualina-Cardiolinka-Paracardium* Cm.; HN, *Hercynella-Neklania* Cm.; JCP, *Joachymia-Cardiolinka-Pygolfia* Cm.; JI, *Joachymia impatiens* Cm.; PD, *Patrocardia-Dualina* Cm.; PE, *Patrocardia evolvens* Cm.; PEX, *Patrocardia excellens* Cm.; PC, *Pterinopecten (P.) cybele cybele* Cm.; SC, *Slava cubicula-Cardiola donigala* Cm.; SI, *Snoopyia insolita* Cm. (after Kříž, in press).

ria-dominated communities; it occurs in a micritic limestone facies that was less oxygenated and deeper than the cephalopod limestone biofacies (Benthic Assemblages 3–4). The *Snoopyia* Community Group is represented by six commonly monospecific communities; it occurs in the Pridoli in deeper water micritic limestones (Benthic Assemblages 3–4), where it occupied less favorable habitats with limited current activity and low oxygen. The *Patrocardia* Community Group is represented in the Silurian of the Prague Basin by three communities dominated by *Patrocardia;* it occurs in somewhat shallower, better-ventilated environments than the *Snoopyia* Community Group. Wacke- to packstones of the *Patrocardia* Community Group facies correspond to Benthic Assemblages 2–3.

The *Cardiola* Community Group is very closely related to the cephalopod limestone biofacies. Limestones with cephalopods occur in the Prague Basin Silurian at eleven horizons (Kříž 1997b, 1998, in press). Each horizon has characteristic cephalopods (Š. Manda, personal commun., 1997) and indicates a period when the sea floor below wave-base was ventilated by surface-water currents.

BOHEMIAN SILURIAN PALEOGEOGRAPHY — The Bohemian Massif (=Perunica microcontinent of Havlíček et al., 1994) drifted from high southern to low northern latitudes in the latest Paleozoic. Data for this supposition were summarized by Krs et al. (1986, 1987) from Bohemia. Lower Middle Cambrian graywacke from the Příbram–Jince Basin shows a paleolatitude of ca. 39° S, and Upper Cambrian andesite records a paleolatitude of ca. 29° S. Lower Ordovician chert and tuffaceous rock in the Prague Basin record a paleolatitude 28° S, and Lower Devonian micrites show a 5–9° S paleolatitude. The Upper Carboniferous in northern Bohemia was deposited approximately on the equator. Younger Permian rocks show a paleolatutude of 6–10° N, and the Triassic preserves a 14–18° N paleolatitude. The insular development of Perunica is suggested by its probable rotation, as shown by changes in paleomagnetic directions from ca. 65° in the Middle Cambrian, 90° in the Late Cambrian, and to 127–132° in the Early Ordovician (Krs et al., 1986). These data support Burrett's (1983) interpretation of apparent polar wander path and his suggestion that the Bohemian Massif moved independently of Armorica during the Early Paleozoic. Havlíček et al. (1994) supported the existence of an insular Perunica in the Ordovician by the analysis of microplate and plate separations based on brachiopod and trilobite assemblages.

Early Silurian sedimentation on the Bohemian Massif was influenced by Late Ordovician glaciation. Seawater temperature was probably relatively low in the post-glacial period. In the Early Silurian, the sea was anoxic or strongly dysaerobic, and black graptolitic shale was deposited in all Bohemian Massif basins. During the Telychian, the temperature slowly increased and better circulation commenced. In the Prague Basin, this is related to the deposition of calcareous shales. Wenlock and especially Homerian limestone facies in the Prague Basin indicate a further increase in temperature, which reached a maximum in the Pridoli. Climate in the latter interval corresponded to a 20–30° latitudinal separation of Perunica from the equator (Krs et al., 1986).

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PART III: EASTERN GONDWANA AND RELATED TERRANES

Silurian of Australia and New Guinea: Biostratigraphic Correlations and Paleogeography

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ABSTRACT — The Silurian of Australia, illustrated herein by 33 stratigraphic columns, is correlated by conodonts and graptolites. The consequences of such correlations in understanding transgression–regression patterns and changes in paleogeography are synthesized. Biostratigraphic correlations allow a dozen new conclusions on the Australian Silurian. 1) The Silurian of eastern Victoria (earlier regarded as Late Silurian) includes almost all of the upper Llandovery (*Pterospathodus celloni* Zone)–Pridoli. 2) The Sardine Conglomerate of eastern Victoria probably reflects earliest Devonian synorogenic sedimentation. 3) The Yass sequence seems to extend down into the upper Wenlock. 4) The Canberra succession is likely mainly Ludlow, rather than Wenlock. 5) The Silurian of the northern Capertee High extends well down into the Wenlock. 6) The two major units of the northern Molong Arch (Nandillyan and Narragal Limestones) are appreciably different in age (early Wenlock and late Ludlow, respectively). 7) Late Ludlow–Pridoli siliciclastics at the base of the Winduck Formation extend west into the Darling Basin and Bancannia Trough, western New South Wales. 8) Silurian clasts appear at six localities in the Tamworth Belt, northern New South Wales; this implies substantial platform carbonates to the west. 9) The Pridoli–Lochkovian is absent in the Calliope Arc of east-central Queensland except at Craigilee. 10) Sequences from the Camel Creek Province of northern Queensland yield mixed Ashgillian and Llandovery–earliest Wenlock conodonts that show cannibalization of shelf platform carbonates to the west and/or southwest, a region where autochthonous pre-*Pterospathodus amorphognathoides* Zone (Llandovery) carbonates are unknown in the Broken River and Camel Creek Provinces. 11) An apparently unbroken sequence exists from the *Pterospathodus celloni* Zone (and possibly older horizons) through the Lower Devonian in the Hodgkinson Province of northeast Queensland. 12) Improved biostratigraphic control is possible on evaporites in the Carnarvon (Llandovery–Pridoli) and Canning basins (middle Llandovery and older). Improved biostratigraphic precision, particularly for the latter two areas have resulted in revised timing of the transgression–regression pattern.

The new data, especially from the Indi and upper Buchan River areas of eastern Victoria, are consistent with the main events of the Benambran cycle (arguably the most profound Phanerozoic orogenic cycle in Australia). This orogeny occurred in the late(?) Ashgillian to early or middle Llandovery, rather than in a generalized Llandovery–Wenlock interval, as earlier assumed.

Introduction

The Silurian is arguably the most neglected interval in Australian geology, despite large tracts of Silurian igneous rocks and locally highly fossiliferous sedimentary rocks in the eastern states. Biostratigraphically significant Silurian rocks in Australia are restricted to the meridional Tasman Fold belt of the east. With two exceptions — one in the Canning Basin (Barbwire Terrace, Llandovery; Nicoll et al., 1994) and one from the offshore part of the Carnarvon Basin (Dirk Hartog Formation, upper Ludlow; Gorter et al., 1994) — Silurian correlation in the western 80% or more of the continent is speculative. Some cratonic sequences (evaporitic and siliciclastic) in the Carnarvon, Canning, and Bonaparte Basins of Western Australia, and the Amadeus and Arafura Basins of central and offshore northern Australia, may be partly or entirely Silurian (Fig. 1).

There are relatively abundant carbonates with potential for conodont and chitinozoan recovery. Many of these units have produced poor yields of these fossils, and they seem to represent shoal water environments inimical to conodonts and chitinozoans. Many carbonates have yet to be examined biostratigraphically. In Australia, grapto-

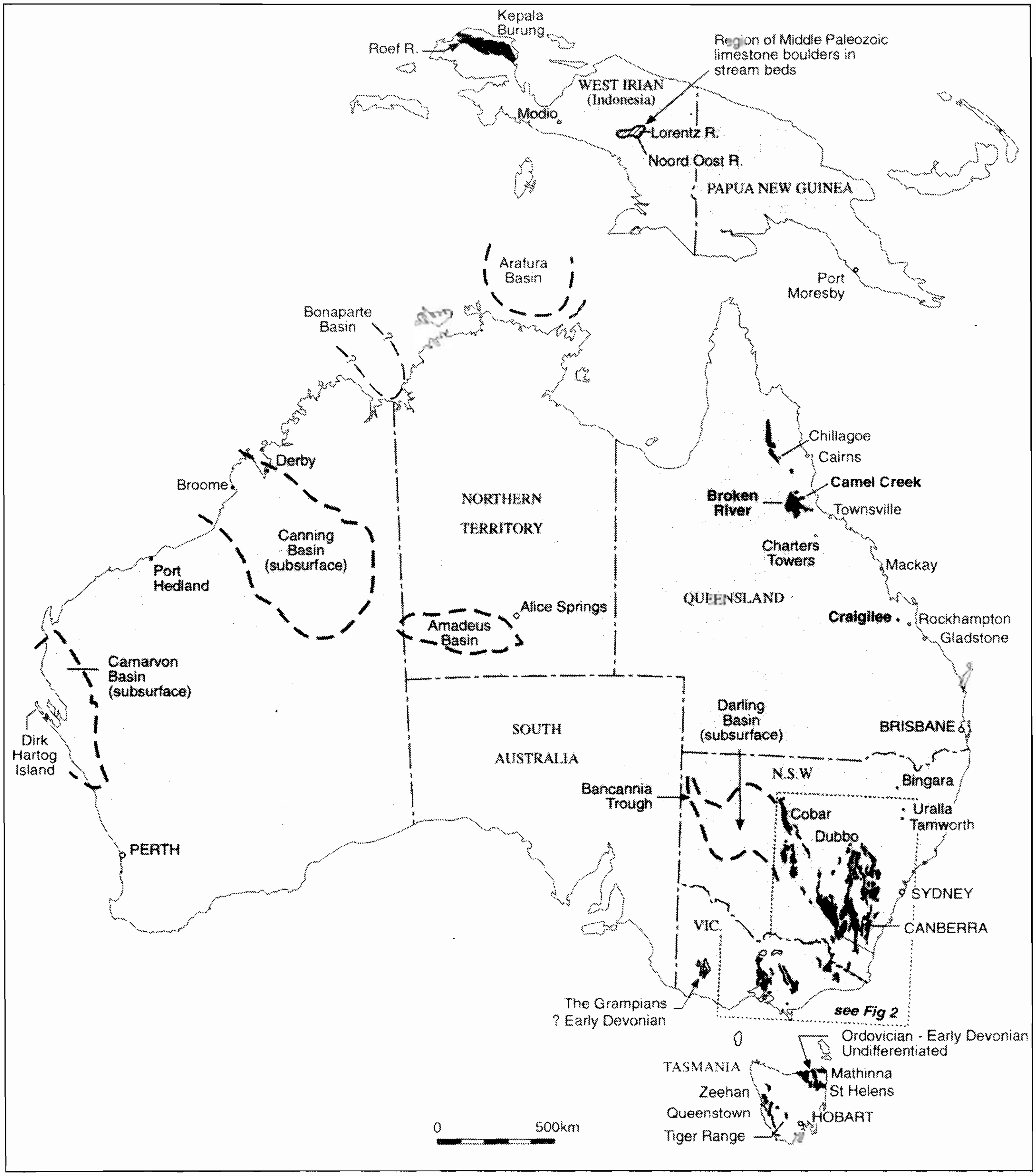

FIGURE 1 — Distribution of the Silurian in Australia and New Guinea with locations mentioned in the text.

lites have a long history of use in correlation, but they have been useful principally in central–western New South Wales; the Yass synclinorium, central Victoria; and, in a minor way, far eastern Victoria and the Broken River region of northeastern Queensland (Figs. 2, 3). There remain large tracts of potentially suitable Silurian facies,

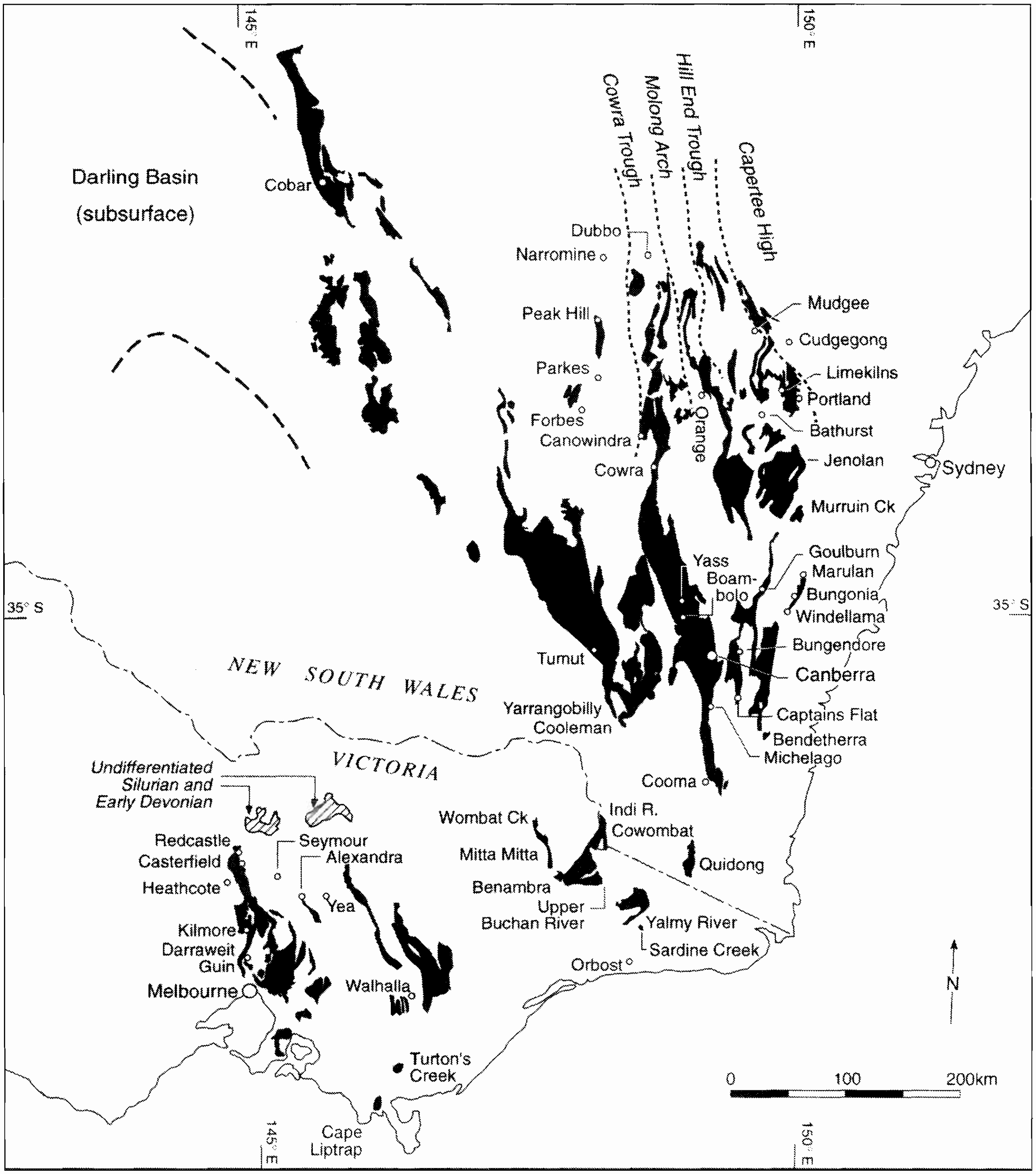

FIGURE 2 — Distribution of the Silurian in southeastern Australia with principal locations mentioned in the text.

particularly in New South Wales, to be examined for conodonts, graptolites, and chitinozoans.

Recent conodont work, especially in New South Wales, eastern Victoria, and northern Queensland, has substantially changed long-accepted biostratigraphic correlations (e.g., Simpson et al., 1993; Simpson and Talent,

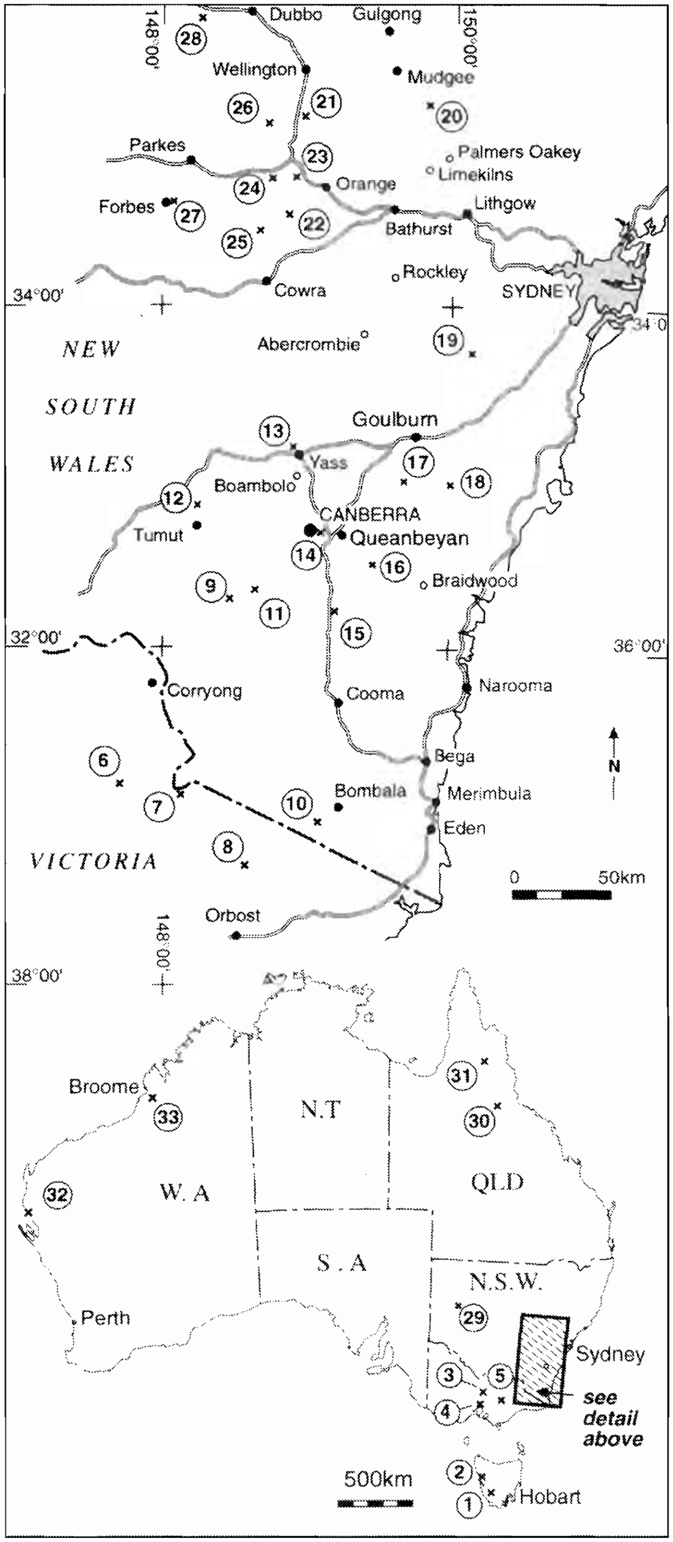

FIGURE 3 — Location of the biostratigraphically best-constrained Silurian sequences in Australia, see Figs. 4–6.

1995; Sloan et al., 1995; Simpson, 1995a, 1995b, 1999; Cockle, 1999). New graptolite data, especially from New South Wales and the mainly flyschoid associations of the Melbourne Terrane, central Victoria, have improved biostratigraphic precision (e.g., Sherwin, 1979; Rickards et al., 1993, 1995, 1998; Rickards and Wright, 1997a, 1997b; Rickards and Sandford, 1998). Improved correlation brings greater precision in determining the timing of sedimentary, volcanic, and tectonic events in the Late Ordovician–Early Devonian.

Autochthonous Silurian limestones are widespread in eastern Australia, and there are many occurrences that are allochthonous. These include debris-flow megabreccias and isolated olistoliths (sometimes with a largely coherent stratigraphy). Noteworthy among these occurrences, some reflecting synorogenic sedimentation, are major tracts in the Hodgkinson and Camel Creek Provinces, northeast Australia (Bultitude et al., 1993; Sloan et al., 1995; Simpson, 1999); on the flanks of the Hill End Trough on the Molong Arch, New South Wales (Talent and Mawson, 1999); and some but not all of the Silurian limestones of the Mitta Mitta, Gibbo, and Indi (upper Murray) Rivers and Sardine Creek, eastern Victoria (Simpson and Talent, 1995).

Older references can be found in recent (Walley et al., 1990) and earlier biostratigraphic syntheses (Talent et al., 1975; Cooper and Grindley, 1982; Pickett, 1982a; Garratt and Wright, 1989; Jell and Talent, 1989). Corals, especially rugose corals, and to a lesser extent brachiopods and trilobites were prominent in earlier studies (see Hill, 1978, for coral bibliography). These biostratigraphic re-evaluations include advances in graptolite and conodont biostratigraphy (Rickards et al., 1995, 1998; Simpson and Talent, 1995; Simpson, 1995b, 1999; Rickards and Wright, 1997a, 1997b, 1999; Rickards and Sandford, 1998; Talent and Mawson, 1999; Cockle, 1999). The primary aim of this report is to review conodont and graptolite data from the past decade that are relevant to the definition of series, stage, and zonal boundaries in the Australian Silurian. These correlations have implications for the Silurian transgression–regression pattern outlined by Talent (1989). For location of the important stratigraphic sequences, see Figs. 1–3. The correlation tables are arranged from south to north (Figs. 4–6) and are discussed in this order to build on Talent et al. (1975) and the Devonian correlation chart of eastern Australia (Mawson and Talent, 2000, In press). A relevant survey of the Silurian time scale, with special reference to Australia, has been given by Strusz (1989) and Young and Laurie (1996).

Tectonic Setting

Middle Paleozoic sedimentary and volcanic rocks of Australia and New Guinea occur in two tectonic settings. These include intra-cratonic basins with relatively simple tectonics—the Amadeus, Carnarvon, Canning, Bonaparte Gulf, and Arafura Basins (Fig. 1) — and the tectonically complex Lachlan Fold Belt, with its extension to the north and northeast as the Thompson Fold Belt and the Northern New Guinea Fold Belt. The latter belts represent almost one-third of the modern Australian continent. Collins and Vernon (1994) proposed a rift–drift–delamination model for the tectonic development of eastern Australia. By this model, the Lachlan Fold Belt and its northern extension, the Thompson Fold Belt, represent a tectonic collage. This collage developed as an amalgam of Proterozoic continental and Cambrian oceanic fragments that formed by rift, drift, and subsequent convergent tectonism that welded the fragments to the leading edge of eastern Gondwana. Oceanward dispersal of detritus across this belt produced an overlap assemblage of quartzose Ordovician turbidites. Subsequently, a new subduction zone developed along the eastern margin of modern Australia in a tectonic setting similar to that of the modern Philippines Plate of the western Pacific. Back-arc Silurian–Devonian convergent tectonics over a 60-m.y. interval converted what had been a 1,700–2,000 km-wide continental margin into a 750 km-wide, thin-skinned, fold–magmatic belt.

There was earlier acceptance that three orogenic maxima occurred during the Silurian–earliest Devonian of eastern Australia. These are the Benambran, Quidongan, and Bowning events, which were originally based on important unconformities. However, new biostratigraphic data lead to questions about the timing, extent, and possible diachronism in these and lesser events (Talent, 1988; Cockle, 1999). The episodic nature of orogeny in eastern Australia was challenged by Bucher et al. (1996) and Gray et al. (1997) for being outmoded, as it does not embrace the possibility of localized or regionally diachronous deformation. This concern was not earlier in question (Talent, 1988), but perhaps was not well elucidated. Gray et al. (1997) noted that extensive areas of the Lachlan Fold Belt are dominated by Cambrian–Early Devonian submarine fan deposits that were thrust over Cambrian oceanic crust. $^{40}Ar/^{39}Ar$ dates from metamorphic mica within a number of thrust zones further suggest that deformation was dominated by an east-vergent fold-and-thrust system that developed diachronously from west to east from the latest Ordovician into the Late Devonian (Gray et al., 1997). We suggest the possibility of a more complex history that includes the possibility of diachronous deformation and sedimentation along the Lachlan Fold Belt, for which there appears to be evidence during the Devonian (Talent 1985, 1988). This model takes into account the possibility of large-scale translocations of some of the crustal blocks that make up eastern Australia (Talent, 1985, 1988; Packham, 1987). Sinistral strike-slip displacement on the order of 500 km during the Silurian, possibly preceded by dextral motion, has been suggested (Packham, 1987). Relevant radiometric, paleontologic, and paleomagnetic data to clarify these questions need to be expanded.

Tasmania

The Silurian sedimentary sequences of Tasmania (Figs 1, 3, 4, columns 1, 2) consist exclusively of siliciclastics, lack carbonates, and, being predominantly turbiditic, resemble the Silurian–Lower Devonian of central Victoria. Age-constraining data are sparse.

Tiger Range–Florentine Valley — The Ordovician–Silurian boundary in the Florentine Valley in the Tiger Range (Fig. 4, column 1) may lie high in the uppermost unit of the largely Ordovician Gordon Group. This unit is the Arndell (= Westfield) Sandstone, and the boundary lies above a horizon with a rich fauna with *Akidograptus?*, *Atavograptus*, *Climacograptus normalis*, and *Normalograptus persculptus*, the latter of which is the diagnostic species of the uppermost Ashgillian. Overlying the Arndell Sandstone with apparent unconformity is the Gell Quartzite, the basal unit of the Tiger Range Group (Baillie, 1989), a siliciclastic sequence interpreted as a probable sandy tidal-flat facies (Baillie, 1989). The overlying Richea Siltstone has three fossiliferous intervals. The lower two have low-diversity trilobite faunas (Holloway and Sandford, 1993), brachiopods (Laurie, 1991), echinoderms, and the graptolites *Monograptus priodon*, *M.* sp. cf. *M. rickardsi*, *M? parapriodon*, *Pristiograptus nudus*, *P. denemarhae*, and *Monoclimacis* sp. (Baillie et al., 1978). These graptolites suggest the upper Llandovery *Monoclimacis crenulata* or *M. griestoniensis* Zones. A higher horizon has *P.* sp. cf. *P. dubius*, which suggests the middle or upper Wenlock. The overlying Currawong Quartzite is poorly fossiliferous; its shelly fauna appears to be Late Silurian (Baillie et al., 1978; Baillie, 1979, 1989; Sheehan and Baillie, 1981).

Zeehan and Queenstown areas — The Eldon Group of western Tasmania, initially documented from the Queenstown and Zeehan areas (Gill, 1948, 1950; Gill and Banks, 1950; Baillie, 1989), spans much of the Silurian. The oldest unit, the Crotty Quartzite, is probably unconformable on the Gordon Group (Fig. 4, column 2). Shelly faunas from the Crotty Quartzite and the overlying

Amber Slate are poorly preserved and not particularly age-diagnostic (Baillie, 1989). Although its base may be uppermost Ludlow or Pridoli, the youngest unit of the Eldon Group, the Bell Shale, appears to be largely Lower Devonian. This is supported by strata correlated with the Bell Shale in the lower Gordon River that have *Ozarkodina remscheidensis remscheidensis*, a conodont consistent with a generalized Early Devonian age (Baillie, 1989, p. 22). The underlying Florence Quartzite is thought to span the latest Wenlock and most of the Ludlow, but could extend into the Pridoli or even the Lochkovian. There is no tight biostratigraphic control on any of the units of the Eldon Group.

NORTHEAST TASMANIA — A column for this area is not illustrated. Until recently, fossils unequivocally of Silurian age have not been reported from the turbiditic Mathinna Group of northeastern Tasmania (Banks and Baillie, 1989; Powell and Baillie, 1992; Powell et al., 1993), although the group has long been assumed to include Silurian rocks (Baillie, 1989). There are reports of Lower Ordovician graptolites from Turquoise Bluff near Mathinna (Fig. 1). Other fossils include the middle Lower Devonian (Pragian) graptolite *Monograptus thomasi* Jaeger near Scamander (Rickards and Banks, 1979), Pragian dacryoconarids (G. K. B. Alberti, personal commun., 1995), as well as occurrences of vascular plants assumed to be Lower Devonian (Banks and Baillie, 1989). A locality on Golden Ridge 20 km WSW of St. Helens (Fig. 1) yielded *Monograptus dalejensis, M. insignitus, M. podoliensis australis, Bohemogratus bohemicus tenuis, B. bohemicus minutus*, and *Linograptus posthumus*, an assemblage referred to the upper Ludlow *Bohemograptus kozlowskii* Zone (Rickards et al., 1993) The Mathinna Group closely resembles the Silurian–Early Devonian turbiditic sequences of central Victoria.

VICTORIA

THE GRAMPIANS — No column for this area is illustrated herein. The precise age of the predominantly non-marine, siliciclastic Grampians Group of western Victoria (Spencer-Jones, 1965) is uncertain. A low-diversity fauna of linguloid brachiopods, ostracodes, and thelodont and poracanthodid scales occurs in its middle unit, the Silverband Formation (Talent and Spencer-Jones, 1963; Turner, 1986; Burrow, 1997). *Poracanthodes* sp. cf. *P. qujingensis* in this fauna indicates a Ludlow age (*Ozarkodina crispa* [conodont] Zone), but isotopic data (Simpson and Woodfull, 1994; Mawson and Talent, 2000; compare Gradstein and Ogg, 1996) suggest an age somewhere between early Pragian and early Emsian for the overlying Rocklands Rhyolite. This determination does not rule out a possible latest Silurian age for at least part of the very thick (ca. 5.3 km) Grampians Group.

HEATHCOTE–REDCASTLE — Correlations of the entirely siliciclastic Silurian of the Heathcote–Redcastle–Costerfield area (Figs. 2, 4, column 3) diverge from those proposed earlier (Talent, 1964; Holloway and Neill, 1982; VandenBerg, 1988; Jell and Talent, 1989). This is a consequence of recent redefinitions both of the stratigraphy in central Victoria (Fig. 4, column 5) and of the base of the Wapentake Formation (Rickards and Sandford, 1998). The Dargile Formation has been raised to group level, and now consists of the Yan Yean Formation (units I–2b of the "Dargile Formation") and overlying Melbourne Formation (units 2c–4 of the "Dargile") in the Heathcote–Redcastle–Costerfield area (Rickards and Sandford, 1998). These formations were based originally on sections in the Melbourne–Kilmore area. Despite relatively diverse shelly faunas in the overlying thick McIvor Sandstone–Mt Ida Formation sequence (Talent, 1964), these faunas are rarely adequately preserved for species-level identification. Moreover, there is an absence of graptolites or other taxa useful in precise biostratigraphy. The boundary between the Melbourne Formation and the McIvor Sandstone is so lithologically and faunally abrupt that a significant unconformity may be involved. Questions therefore remain as to the position of the Silurian–Devonian boundary in the Heathcote–Redcastle–Costerfield area. A biofacies pattern (Talent, 1964, and unpublished data) is apparent for the McIvor Sandstone and Mt Ida Formation, in which western bivalve faunas are replaced by eastern faunas dominated by rhynchonellids, retziidines, and occasional *"Howellella"* and then by faunas dominated by strophomenidines with a few larger spiriferidines. Farther east in the Seymour–Seymour East and Yea–Alexandra areas (Fig. 2), stratigraphic equivalents of the McIvor Sandstone and Mt Ida Formation are turbiditic sequences with graptolites, land plants, and, very rarely, debris flows with generally comminuted shelly material.

MELBOURNE–KILMORE — Continued study of the stratigraphy and graptolites of the purely siliciclastic Silurian rocks of central Victoria (Rickards and Sandford, 1998, and references therein) has improved correlation in this extensive, lithologically monotonous, and largely turbiditic region (Fig. 2). No evidence has been found to support earlier suggestions of a west–east diachronism in sedimentation. As in the Heathcote–Redcastle area, two groups are now recognized. These are the Keilor Group (Llandovery), based on sequences in Deep Creek and its tributaries west of the meridian of Melbourne, and the Dargile Group (Wenlock–lower Ludlow), which derives its name from part of the Silurian sequence in the Heathcote–Redcastle area farther north (Fig. 4, column 4). Graptolites of the uppermost Ordovician *Normalograptus persculptus* Zone and lower Llandovery *Parakidograptus acuminatus* and *Demirastrites triangulatus* Zones indicate that the Ordovician–Silurian boundary is low in the Deep Creek Siltstone at Darraweit Guim (VandenBerg et al., 1984). The definition of the overlying Springfield Sandstone and Chintin Formation follows VandenBerg (1992), with minor modification herein of its correlation based on new graptolite data. Strata earlier referred to the "Kilmore Siltstone" (VandenBerg, 1992) have been referred to the Springfield Sandstone and other units (Rickards and Sandford, 1998). Graptolites indicative of the lower *Monograptus nilssoni* Zone high in the Yan Yean Formation (i.e., units 2a and 2b of the "Dargile Formation") provide a useful lower bracket for the Melbourne Formation (i.e. unit 2c of the "Dargile Formation" and higher). The boundary between the Melbourne Formation and overlying Clonbinane Member of the Humevale Formation is so lithologically and faunally abrupt that a significant unconformity may be involved, as may be the case for the boundary between the Melbourne Formation and McIvor Sandstone of the Heathcote–Redcastle area farther north.

The Upper Silurian–Lower Devonian of the Goulburn Valley north of Melbourne in the Seymour–Yea–Alexandra–Eildon area (Fig. 2) displays fold repetition of the Humevale Formation and, more rarely, of the underlying Melbourne Formation. Noteworthy in this area are plant- and graptolite-bearing beds, the highest of which is Lower Devonian. The lowest of these beds (i.e., "Lower Plant–Graptolite Horizon" of Couper, 1965) has the upper Ludlow–Pridoli graptolites *Bohemograptus bohemicus* and *Monograptus* sp. aff. *M. uncinatus* (Garratt, 1978, 1983b; Garratt et al., 1984; Garratt and Rickards, 1984, 1987; Rickards and Garratt, 1990).

UPPER YARRA, THOMSON AND GOULBURN RIVERS — The correlations in Fig. 4, column 5, follow VandenBerg (1975, 1977, 1988), who clarified stratigraphic relationships in this mountainous area (Fig. 2). *Monograptus exiguus, M. pandus, M. spiralis permensus, M. turriculatus, M. priodon, Stomatograptus australis,* and *Rastrites?* sp. confirm a late Llandovery age for the McAdam Sandstone, the lowest unit of the Jordan River Group. An isolated occurrence of McAdam Sandstone with *Monograptus marri* is known on the south coast of Victoria near Cape Liptrap (Douglas and Paton, 1972; Fig. 2).

Pristiograptus sp. cf. *P. dubius* indicates a generalized Wenlock–early Ludlow age for the overlying Bullung Siltstone, a unit that also has an undescribed small, rather nondescript, shelly fauna with the trilobite *Encrinurus* sp. (VandenBerg, 1988). *Bohemograptus bohemicus* and *Monograptus* sp. cf. *M. uncinatus* from shales in the overlying Sinclair Valley Siltstone at Bonnie Doon, Telbit Crossing, and Upper Thomson provide better biostratigraphic control (VandenBerg, 1988). The Whitelaw Siltstone appears to be unfossiliferous, but is assumed to span the Silurian–Devonian boundary as it passes, apparently gradationally, up into the lower Pragian–lower Emsian Boola and Coopers Creek Formations (Mawson and Talent, 1994).

MOUNT USEFUL SLATE BELT — There is little chronostratigraphic control for this extensive and structurally complex region in the mountainous country east of the Walhalla synclinorium (Fig. 2; no column illustrated). Two Silurian units have long been recognized: the Serpentine Creek Sandstone and the Donnellys Creek Siltstone (Baragwanath, 1925; VandenBerg, 1977, 1988). The former has graptolites of generalized Silurian age, but no identifications have been published; it may correlate with the McAdam Sandstone west of the Walhalla synclinorium (see Fig. 4, column 5). No fossils have been reported from the Donnellys Creek Siltstone, but it is assumed to equate with the Whitelaw Siltstone west of the Walhalla Synclinorium. The Donnellys Creek Siltstone passes up into the conglomeratic Wurutwun Formation, which includes probable early Pragian olistolith limestones in Marble Creek and Deep Creek east and southeast of Walhalla (Mawson and Talent, 1994).

It has been customary to assume shallow marine shelf conditions east of the Mount Useful Slate Belt in the Silurian–Early Devonian (e.g., Garratt, 1983a), but there is no evidence for this. A zone on the eastern flank of the Melbourne Trough with shallow marine-shelf environments, if it ever existed, may have been severed from it by large scale, pre-Late Devonian movements along the Mount Wellington–Dolodrook Fault Zone.

MITTA MITTA RIVER, WOMBAT CREEK, GIBBO RIVER — Whether or not limestones in this region (Fig. 2, 4, column 6) are allochthonous (VandenBerg, 1998) or autochthonous (Whitelaw, 1954; Bolger, 1982) is of pivotal importance in dating associated strata and determining the time span of the Benambran orogenic cycle at its type locality here. Our principal data come from a cliff section on the flanks of the Mitta Mitta River about 260–320 m upstream from its junction with Wombat Creek (Whitelaw, 1954, fig. 2E). The elegant exposures show a gradual change from lower conglomerates through overlying arenites to thin-bedded and then massive limestone. There is very only gradual change from the massive limestones into overlying bedded limestones, and then into interbedded crinoidal limestones and mudstones, with an overall upward decrease in calcareous content of the latter. The gradual change in lithologies is not what would be anticipated if this succession is within an olistolith. We interpret the gradual change to mudstones up-section as reflective of deepening.

Conodonts in preliminary samples of these carbon-

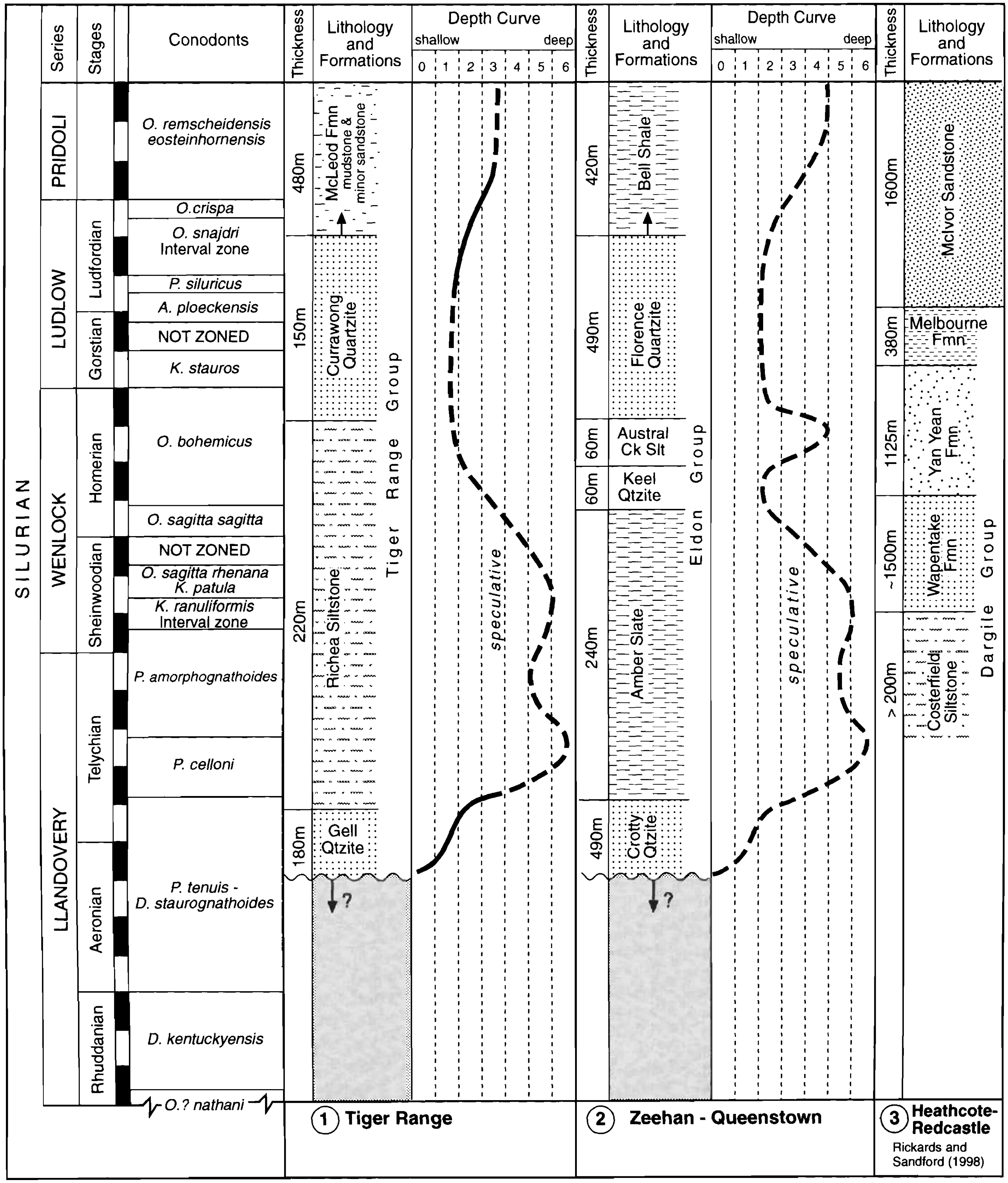

FIGURE 4 — Correlation of the Silurian in Tasmania and southeastern mainland Australia. Location of stratigraphic columns on Fig. 3. No attempt has been made to provide depth curves for stratigraphic columns where correlations are conjectural.

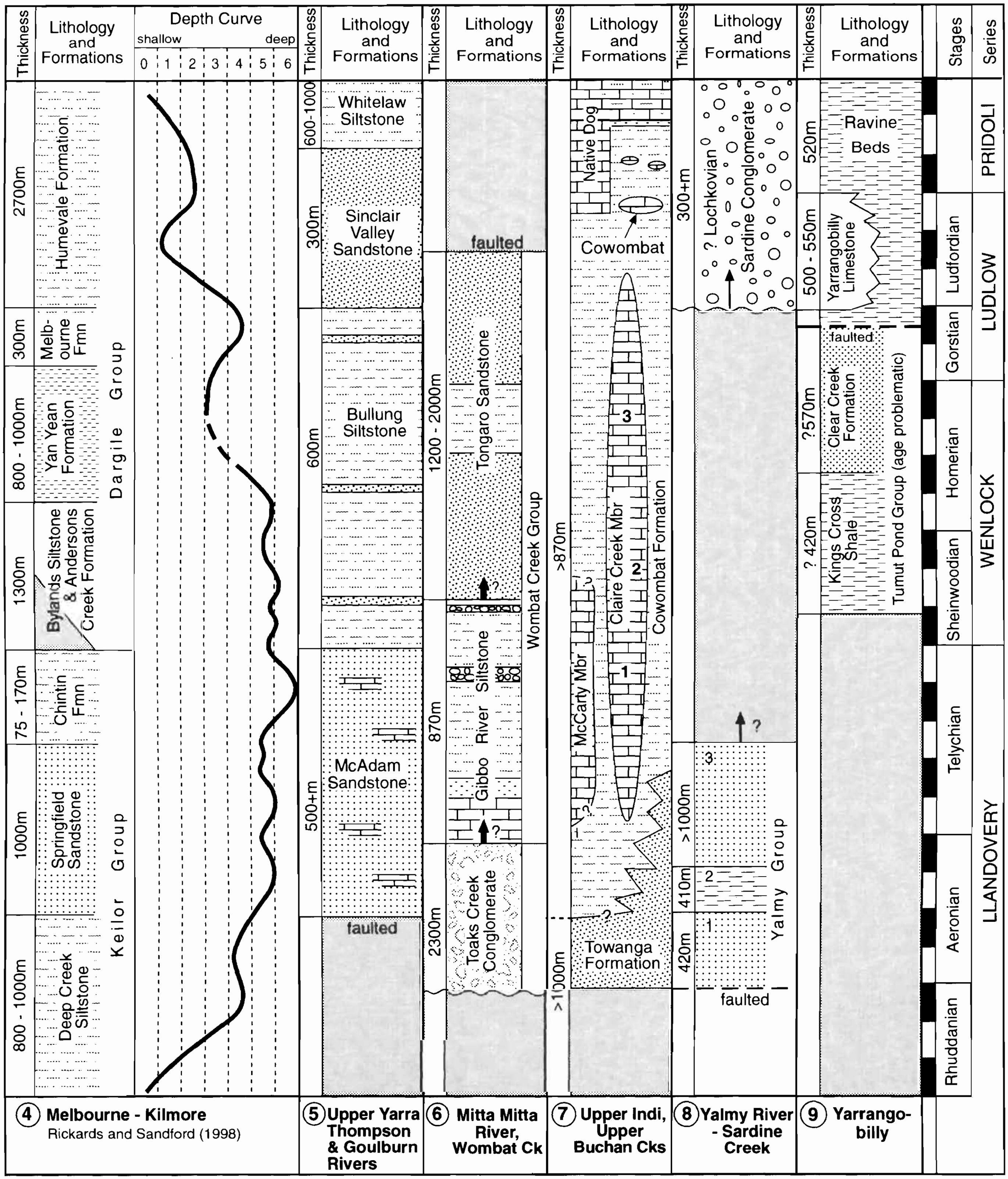

FIGURE 4 *continued.*

ates do not allow a precise correlation. The faunas are dominated by *Panderodus* elements, but also include the long-ranging taxon *Ozarkodina excavata excavata* and other elements comparable with those of *O. confluens.* A generalized Wenlock or younger Silurian age is inferred. Some *Panderodus* elements are similar to those of a new but unnamed species recovered from the late Llandovery Quinton Formation and the Wenlock of the Jack Formation in north Queensland (Simpson, 1999). A more accurate age determination will require further samples.

Pyle's deposit (Whitelaw, 1954, fig. 3D) is a tiny, century-old quarry with metamorphosed calcareous siltstones and arenites with minor, generally thin limestone bands. Despite poor exposures, we have no reason to suspect that this occurrence is allochthonous. It has been extensively sampled and has produced a small, poorly preserved conodont fauna with *Ozarkodina excavata excavata, O. remscheidensis remscheidensis, Panderodus?* ssp. and *Dvorakia?* sp. (Talent et al., unpublished data). Despite taxonomic problems with the *"remscheidensis"* plexus, the Pa element recovered here is within the range of the morphotype group described by Klapper and Murphy (1975). This Pa element is usually associated with the Lower Devonian, but may extend down into the upper Pridoli.

HEADWATERS OF INDI, BUCHAN AND TAMBO RIVERS — The region about the headwaters of the Indi (upper Murray), Tambo, and Buchan Rivers (Fig. 2) consists of complexly deformed, Ordovician–Early Devonian sedimentary, volcanic, intrusive, and metamorphic rocks. The oldest Silurian unit appears to be the unfossiliferous Seldom Seen Conglomerate (about 2.6 km of conglomerate, sandstone, and minor mudstone), which apparently passes up into the widespread Towanga Formation (Fig. 4, column 7). The Towanga Formation is primarily sandstone interbedded with siltstone intervals, and locally seems to be overlain disconformably by the Thorkidaan Volcanics. As no paleontologic data are available to constrain the ages of the Seldom Seen Conglomerate, Towanga Formation, or the Thorkidaan Volcanics, they are not shown in Fig. 4, column 7, although a generalized Llandovery age has been suggested from regional stratigraphy (VandenBerg, 1988).

All the carbonates in this region have been given a generalized Late Silurian age (VandenBerg, 1988; Walley et al., 1990). Most of the carbonates have been sampled for conodonts, and a broad range of Silurian ages has been determined (Simpson and Talent, 1995).

Samples from low in the prominent McCarty Member limestone lens yield a poor, lower Llandovery, *Distomodus combinatus* Zone fauna (sensu Simpson, 1995b). The ranges of some of the associated conodonts are poorly understood at present, and a slightly younger age that predates the late Llandovery *Pterospathodus celloni* Chron is possible (Simpson and Talent, 1995). In the middle and upper McCarty Member limestone lens, poor upper Llandovery *Pterospathodus celloni* Zone to upper Llandovery–lower Wenlock *Pterospathodus amorphognathoides* Zone faunas have been recovered (Simpson and Talent, 1995). Biostratigraphic control of the lens is very poor, but it is possible that the lens extends into the Wenlock (Simpson and Talent, 1995).

The Lobelia and Farquar Limestone lenses of the Reedy Creek area in the south of the region yield poor but distinctive faunas. These range from the upper Llandovery to lower Wenlock *Pterospathodus celloni* and *P. amorphognathoides* Zones (Simpson and Talent, 1995).

The Claire Creek–Stoney Creek outcrop belt in the central part of the region consists of two main limestone units separated by a pelitic sequence with subordinate carbonates. The upper and lower limestone units have conodonts. Despite poor yields of conodont elements, particularly for the lower unit, the faunas indicate deposition through much of the Silurian (Simpson and Talent, 1995). Despite metamorphism to lower greenschist facies, the higher intervals of the lower limestone unit yield poor faunas of the upper Llandovery–lower Wenlock *Pterospathodus celloni* and *P. amorphognathoides* Zones. The lowest samples in the lower limestone unit yielded a specimen identified as *Ozarkodina? aldridge,* which suggests an earliest possible age of middle Aeronian (Simpson and Talent, 1995). Equivocal fragmentary specimens from near the top of the lower unit suggest that, as the McCartys Member limestone lens, it may extend into the Wenlock of the "post-*P. amorphognathoides*" Zone.

The lower intervals of the upper limestone unit have generalized Wenlock taxa. Higher in the unit, typical European index species of the Wenlock *Ozarkodina sagitta* Zone appear, with overlying strata recording the lowest appearance of cosmopolitan index species of the Ludlow *Ancoradella ploeckensis* Zone. Ludlow-aspect faunas above this level appear high in the upper limestone unit of the Claire Creek Member. The intervening pelitic sequence between the two limestone units is inferred to be Wenlock because it is constrained by data from the overlying and underlying Claire Creek limestone units. Intermittent carbonates in the predominantly siliciclastic sequence overlying the upper Claire Creek limestone lens also yield faunas regarded as Ludlow (Simpson and Talent, 1995). Despite the lack of index species, this highest sequence is thought to extend well into the upper Ludlow.

An isolated limestone lens at Cowombat Flat has upper Ludlow *Ozarkodina crispa* Zone conodonts (Simpson et al., 1993). The interval of fine siliciclastics above this lens is, therefore, probably Pridoli.

Conodonts from limestones associated with siliciclastics at Native Dog Plain are generalized Late Silurian forms, but faunas of the Pridoli *Ozarkodina eosteinhornensis* Zone occur high in the sequence. It is well established, however, that the lowest occurrence of *O. eosteinhornensis* is below the Ludlow–Pridoli boundary in many parts of the world (Aldridge and Schönlaub, 1989). However that may be, the occurrence at Native Dog Plain is most likely the youngest preserved Silurian horizon in the headwaters of the Indi, Buchan, and Tambo Rivers. The Native Dog Plain section has the typical Late Silurian acanthodian *Poracanthodes* sp. cf. *P. qujingensis* (Parkes and Simpson, 1997).

VandenBerg (1998) suggested that Silurian limestones in the headwaters of the Indi, Buchan, and Tambo Rivers may be allochthonous. This would mean that paleontological data derived from them by Simpson and Talent (1995) may not be pertinent to dating the associated, "matrix" strata.

In our sampling, we gave particular attention to the sequence exposed in Claire and Stoney Creeks, approximately 45 km east of Benambra (Fig. 2) because of the exceptional length and thickness of the section, the excellent exposures, and the intricate sedimentology. The sequence includes a wide spectrum of generally thin-bedded arenites, siltstones, and mudstones, and a broad spectrum of pure and impure limestones earlier figured by Whitelaw (1954, fig. ID, Section A). Such lithologically diverse and generally thin-bedded sequences characterized by a broad spectrum of contrasting lithologies with abundant mudrocks are prone to disintegration with downslope movement. This section was sampled (367 samples) through a thickness of 530 m over a distance of 1.4 km (Simpson and Talent, 1995, figs. 2, 6, 7, tables 2–5); it displays no inconsistencies in conodont biostratigraphy. We accordingly accept this pivotal sequence as being fundamentally untectonized and autochthonous. The conodonts indicate accumulation from the middle Aeronian (middle Llandovery) through the upper Gorstian (lower Ludlow) *Ancoradella ploeckensis* Zone and, probably, into the Ludfordian (upper Ludlow). Sequences at Cowombat and Native Dog Plain extend the sequence almost to the end of the Silurian. These are shaley sequences with minor limestones, not the sort of sequences that resist major downslope movement. Discussion of what we believe to be allochthonous and possibly allochthonous limestone occurrences is outside the scope of thisreport.

YALMY RIVER AND SARDINE CREEK — The Yalmy Group (VandenBerg, 1988; VandenBerg et al., 1990) is a much-faulted, ca. 2.5 km sequence of Lower Silurian flyschoid sandstones, mudstones, and quartzites that crop out extensively in County Croajingolong (Fig. 4, column 8), east of the Snowy River in eastern Victoria (Fig. 2). Three informal units have been recognized (Fig. 4, column 8): a lower sandstone with minor mudstones (unit 1, 480 m); middle laminated and massive mudstones (unit 2, 410 m); and upper sandstones and orthoquartzites (unit 3, 1,600 m). Graptolites (i.e., *Glyptogratus* sp., *Petalograptus* sp., *Retiolites geinitzianus perlatus, Monograptus convolutus,* and *M. triangulatus*) 200 m below the top of unit 2 indicate the middle Llandovery *Monograptus convolutus* Zone. Discovery of *M. turriculatus* in unit 3 indicates that the Yalmy Group extends into the upper Llandovery. Unit 3 of the Yalmy Group thus correlates generally with the Tombong Beds of the Quidong area farther east (Fig. 4, column 9) and with the Tawonga Formation in the headwaters of the Indi and Buchan Rivers farther west (Fig. 4, column 7). VandenBerg (1988, p. 131) suggested that the Seldom Seen Conglomerate, which apparently underlies the Towanga Sandstone, equates with unit 1 of the Yalmy Group.

A fault-bounded(?) area of Silurian occurs about 32 km north-northeast of Orbost. This succession was earlier referred to as the Sardine Beds (Talent et al., 1975). It is now regarded as consisting of two units, a lower Sardine Conglomerate (a fan deposit) and an overlying Wibenduck Limestone (VandenBerg, 1988; VandenBerg et al., 1990). There are no exposures of the contact between these units. However, the distribution of limestone and silicilastics in the vicinity of the Martins Creek Saddle where the Wibenduck Limestone crops out (J.A. Talent and R. Mawson, unpublished data) leads us to believe that the Wibenduck Limestone consists of clasts and megaclasts of various carbonate lithologies. Thus the "Wibenduck Limestone" is viewed as an integral part of this spectacular Sardine Conglomerate fan deposit. Conodonts from the "Wibenduck Limestone," which was lithified before cannibalization and incorporation into the fan deposit, are viewed as probably providing a maximum age for the Sardine Conglomerate. Conodonts reported but not documented from the "Windenduck Limestone" include *Kockelella variabilis, K. ranuliformis, Ozarkodina confluens, O. excavata, Belodella anomalis,* and *Coryssognathus dubius* (recorded as *Pelekysgnathus dubius* by VandenBerg (1988, p. 131). *Kockelella ranuliformis* suggests a generalized Wenlock age, but may extend into the *Polygnathoides siluricus* Zone of the lower upper Ludlow. *Kockelella variabilis* suggests the *Ancoradella ploeckensis* and *Polygnathoides siluricus* Zones, and *C. dubius* suggests the Ludlow. The fauna is thus consistent with a generalized Ludlow (Gorstian–early Ludfordian) age of the "Wibenduck Limestone." The fan, of which these limestone clasts and megaclasts are interpreted to form a part, is therefore regarded as probably late Ludlow–Pridoli or, based on regional relationships in which the highest Cowombat Group is Pridoli at Native Dog Plain, is earliest Devonian and reflective of synorogenic sedimentation.

Southeastern New South Wales and Australian Capital Territory

Extensive areas of Silurian and presumed Silurian sedimentary rocks occur in southeastern and east-central New South Wales (Figs. 2, 3). In these areas, numerous apparently unfossiliferous units have been referred to the Silurian on the basis of inferred stratigraphic relationships or reconnaissance determinations of macrofossils. Information on these units is available in the useful synthesis presented by Pickett (1982a). We limit discussion herein to the most important sequences, especially those for which there are recent conodont and/or graptolite data.

Yarrangobilly — The lowest 50 m of section at Yarrangobilly (Fig. 3) consist of interbedded limestone and mudstone overlain by 450–500 m of massive limestone. The latter are overlain, in turn, by about 100 m of bedded limestone with some mudstone and siltstone (Fig. 4, column 9). Cooper (1977) inferred a middle Ludlow *Polygnathoides siluricus* Zone correlation for a horizon in the massive limestone interval that lies about 190 m above the base of the Yarrangobilly Limestone. He assumed a comparable age for the lower part of the formation, an interval virtually barren of conodonts. This conclusion was based on the occurrence of *Ozarkodina confluens* with *Kockelella variabilis* (Cooper, 1977, sample Y7). Subsequent work has shown that both taxa have longer ranges than assumed by Cooper (1977). The lower 190 m of the section may be partly or entirely older than the *P. siluricus* Zone, and may be referable to the *Ancoradella ploeckensis* Zone. An upper Ludlow *Ozarkodina crispa* Zone fauna was obtained from two localities at the top of the upper 100 m interval of bedded limestones with subordinate mudstones and siltstones (Cooper, 1977). The Yarrangobilly Limestone may thus represent most of the Ludlow. Its top correlates biostratigraphically with several other Upper Silurian units in southeastern Australia (see Fig. 4). These include the limestone in the Cowombat Formation at Cowombat (Simpson et al., 1993). The Yarrangobilly Limestone passes gradationally up into the thick (ca. 520 m), virtually unfossiliferous mudstone and graywacke of the Ravine Beds. The Ludlow–Pridoli boundary may approximate the base of the Ravine Beds, or may be somewhere within them.

Quidong — The Silurian at Quidong (Fig. 2), which is preserved in a faulted synclinal structure, consists of four main stratigraphic units (Crook et al., 1973; Talent et al., 1975; Pickett, 1982a; Trounson, 1982; R. Parkes in Talent and Mawson, 1997, p. 9–11). The lowest unit (Fig. 5, column 10), the Tombong Beds, is a proximal flysch that rests unconformably on the Late Ordovician Bombala Beds and passes upwards abruptly with a decrease in arenites into the more distal flysch of the Merriangah Siltstone. Graptolites from the latter unit include *Retiolites geinitzianus angustidens, Monograptus* sp. cf. *M. auduncus*, and *M.* sp. cf. *M. priodon*, and indicate an upper Llandovery correlation between the *M. crenulatus* to *M. crispus* Zones (G. H. Packham *in* Talent et al., 1975). The Merriangah Siltstone is overlain unconformably by the fossiliferous Quidong Limestone, a correlative of the Silverdale Formation in the Yass succession (e.g., Talent et al., 1975; Pickett, 1982a). However, there has been so little study of its diverse biota (stromatoporoids, algae, tabulate and rugose corals, brachiopods, trilobites, mollusks, conodonts, chitinozoans) that this correlation may reflect similar biofacies rather than temporal equivalence. The presence of *Coryssognathus* high in the Quidong Limestone suggests that at least some of the Quidong is Ludlow. The Quidong Limestone is overlain by mudstone with occasional siltstone, sandstone, and rare limestone that form the Delegate River Mudstone. Macrofaunas are diverse, well preserved, and abundant in the lower and upper Delegate River Mudstone, but are comparatively unstudied, and good biostratigraphic control has not been established. Conodonts from the highest beds include panderodids that do not provide high biostratigraphic resolution. The positions of the Wenlock–Ludlow and Ludlow–Pridoli boundaries in the Quidong sequence are still problematic (R. Parkes, personal commun., 1998).

Long Plain and Cooleman — Relationships between the various units in this area (Figs. 2, 3) are complex (Owen and Wyborn, 1979). The lithologies and lateral variation of the Peppercorn Beds (Fig. 5, column 11) are documented by Pickett (1982a). Hill (1954) identified corals from the Peppercorn Beds, and suggested they were Wenlock to Ludlow. Pickett (1982a) reported a shelly fauna from fine siliciclastic intervals in the Peppercorn Beds; the age indicated is broadly similar to that suggested by Hill (1954). The best biostratigraphic control from the lower parts of the sequence comes from a limestone "lens" in the Peppercorn Beds near Cooinbil Hut. Conodonts from this locality were interpreted as upper Llandovery by Nicoll and Rexroad (1974); the fauna consists of taxa common to the upper Llandovery–lowest Wenlock *Pterospathodus amorphognathoides* Zone. The Peppercorn Beds are thought to be laterally equivalent to the lower Pocket Beds, and to be conformably overlain by, and partly equivalent to, the Cooleman Limestone (Pickett, 1982a).

The Cooleman Limestone is thick, lithologically variable, and generally recrystallized and dolomitized. Pickett (1982a) listed corals and brachiopods suggestive of a generalized late Wenlock–Ludlow age. A small collection of conodonts 147 m above the base of the Cooleman Limestone type section includes *"Spathognathodus sagitta"* (Pickett, 1982a). This could be one of a plexus of elements that indicates a generalized Wenlock age. *Ozarkodina* sp. cf. *O. remscheidensis* was reported (Pickett, 1982a) only 32 m above this level; this species is upper Ludlow–Devonian and may indicate a hiatus within the Peppercorn Beds. If such a hiatus is present, it may be located between the bedded dark-grey limestones and the overlying, poorly exposed, massive limestone at the top of the Cooleman Limestone. This interpretation may, however, be reading too much into sparse, undocumented conodont faunas.

The Blue Water Hole Beds are a lithologically variable unit that interfingers with the Cooleman Limestone in some areas and discomformably overlies the Cooleman Limestone in others (Pickett, 1982a). A local discomformable relationship with the underlying Cooleman Limestone is suggested by a karst developed on the top of the Cooleman Limestone. Allochthonous carbonate blocks in the Blue Water Hole Beds are inferred to have been derived from the Cooleman Limestone (Owen and Wyborn, 1979; Pickett, 1982a). Corals, trilobites, and brachiopods from siliceous siltstones in the Blue Water Hole Beds have been assigned a general Wenlock or Ludlow age (Pickett, 1982a). Conodonts from allochthonous blocks high in the Blue Water Hole Beds includes *Ozarkodina remscheidensis* (Pickett, 1982a). A generalized late Ludlow–Early Devonian age is indicated, which is similar to that indicated for the uppermost Cooleman Limestone, and from whence the block is assumed to have originated.

The Micalong Creek Beds about 50 km north of Cooleman Plains are considered middle Ludlow (R. Nicoll in Owen and Wyborn, 1979). This correlation is based on conodonts, but no faunal list has been given.

TUMUT–COOLAC — Since Pickett's (1982a) synthesis, there have been important advances in understanding of the stratigraphy and tectonics of the Tumut Trough. This feature runs approximately 50 km north of Tumut (Basden, 1990; Stuart-Smith, 1990; Stuart-Smith et al., 1992; Warner et al., 1992; Dadd, 1998a) in the Tumut–Coolac region (Fig. 2). However, there are unresolved questions regarding the precise ages of the various stratigraphic units in this approximately 2.5 km-thick complex of Silurian volcanics and subordinate flysch. This succession was folded along a N–S axis during the earliest Devonian Bowning orogeny. Column 12 (Fig. 5) is based on the synthesis presented by Warner et al. (1992). A generalized Ludlow but possibly older age was suggested for the Blowering Beds on the basis of a small, nondescript conodont fauna that occurs with halysitid corals identified as *Halysites pycnoblastoides* (Ashley et al., 1972). This collection was presumably from a limestone clast. A late Llandovery–early Wenlock age is suggested for the Wyangle Formation, based on conodonts from limestone clasts in a debris flow (J. D. Lightner *in* Stuart-Smith et al., 1992). Neither conodont fauna has been documented.

Blocks of Late Ordovician siltstone and Silurian limestones of diverse lithologies, some with pentamerid brachiopods or tabulate corals such as *Halysites chillagoensis*, occur in the Goobarragandra Volcanics in the vicinity of "Talmo" and "Glenrock," 25–40 km west-southwest of Yass (Fig. 2). Nineteen limestone occurrences were recorded by Dadd (1998b), who interpreted them as olistoliths and clasts in a caldera-fill megabreccia with a dacitic ignimbrite matrix. The few fossil identifications suggest a generalized Middle Silurian age.

YASS SYNCLINORIUM — The Silurian in the Yass synclinorium in the vicinity of Yass (Cramsie et al., 1975; Fig. 2) commences with the Mundoonen Sandstone, which rests unconformably on the Late Ordovician "Jerrawa Beds" (Fig. 5, column 13). A small shelly fauna high in the Mundoonen Sandstone was noted by Crook et al. (1973, p. 125). Of potential biostratigraphic importance is a fauna with *Monograptus dubius* from a mudstone lens low in the overlying Hawkins Volcanics (Crook et al., 1973, p. 126). A seemingly nondescript coral fauna has also been listed from limestone of the stratigraphically higher Bango Member of the Hawkins Volcanics (Brown, 1941; see Pickett, 1982a). None of these potentially biostratigraphically important faunas has been illustrated.

Carbonates from the Yass Basin were the basis of the first detailed investigation into Silurian conodont biostratigraphy in Australia (Link and Druce, 1972). A local biostratigraphy was proposed that principally involved a modification of Walliser's (1964) work. Biostratigraphic resolution has not improved for most units examined by Link and Druce (1972), but alternative correlations for lower parts of the sequence have been suggested (Simpson, 1995b).

The lowest carbonate unit in the sequence is the Bango Member of the Hawkins Volcanics. The Bango has a generally nondescript coral fauna, and has failed to yield conodonts (Pickett, 1982a); it is therefore poorly constrained biostratigraphically. It is separated from other fossiliferous units by a considerable thickness of volcanics. The lowest Yass Basin sequence is therefore biostratigraphically unconstrained.

Above the Hawkins Volcanics is the Yass Formation or Subgroup. It consists of two units (not discriminated on Fig. 5, column 13): the O'Briens Creek Sandstone and the Cliftonwood Limestone. The latter produced Link and Druce's (1972) oldest conodont assemblage, which was interpreted as lower Ludlow. The same conodont

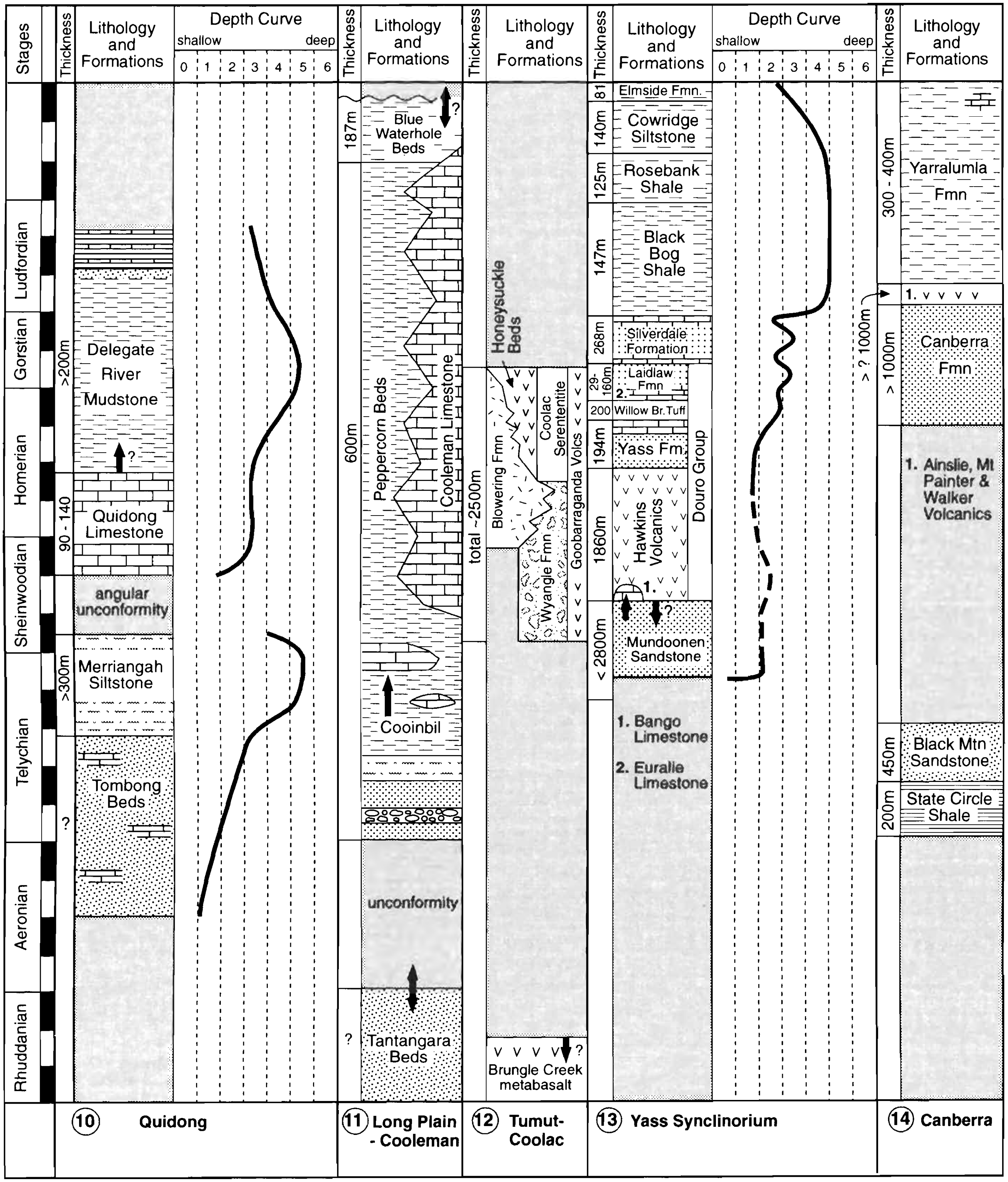

FIGURE 5 — Correlation of the Silurian in eastern New South Wales. Location of stratigraphic columns on Fig. 3. Depth curves speculative. No attempt has been made to provide depth curves for stratigraphic columns where correlations are conjectural.

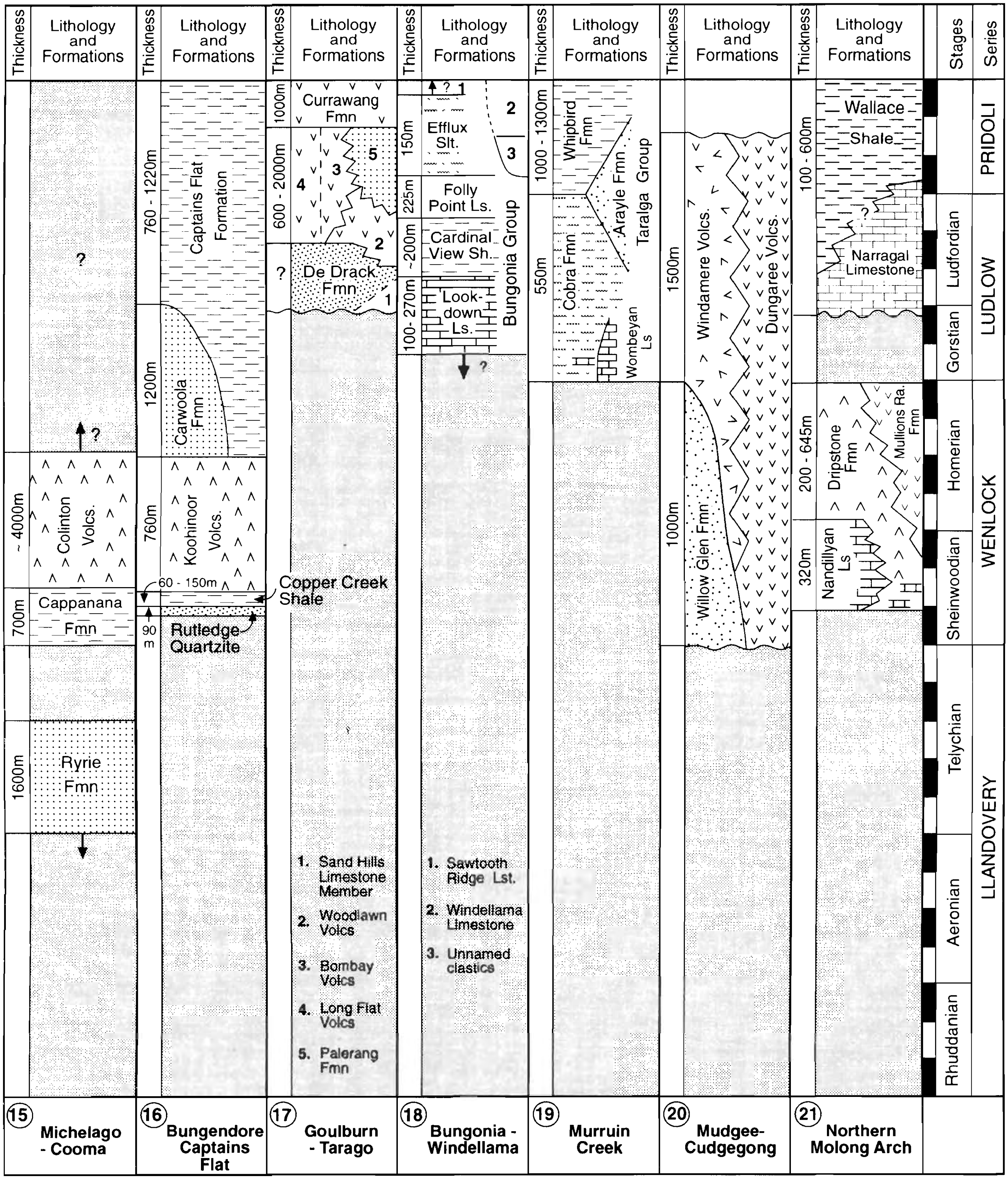

FIGURE 5 *continued.*

association extends into the basal beds of the overlying Euralie Limestone of the lowest Laidlaw Formation. Simpson (1995b) indicated that this assemblage of long-ranging taxa is possibly Wenlock.

The Yass Formation Formation is overlain by the Willow Bridge Tuff. The Willow Bridge is overlain, in turn, by carbonates, siliciclastics, and volcanics of the Laidlaw Formation. The lowest unit of the Laidlaw Formation is the Euralie Limestone Member, as noted above. This unit, excluding its lowest horizons, was the basis for the lower part of Link and Druce's (1972) second-oldest conodont assemblage zone. The Laidlaw Formation is overlain by carbonates and siliciclastics of the Silverdale Formation. This formation consists of three successive limestone members (not discriminated on Fig. 5, column 13): the Gums Road, Bowspring, and Hume Limestones, in ascending order, which are separated by sandstone of the Tullerah Member and shale of the Barrandella Members. The lower 8 m of the Bowspring Limestone predates the first appearance of the eponymous species of the *Ancoradella ploeckensis* Zone, which has its lowest appearance in the lowest Ludlow *Neodiversograptus nilssoni* graptolite Zone (Jeppsson, 1994). Link and Druce's (1972) second-oldest conodont zone extends from the upper Euralie Limestone through the lower Bowspring Limestone, and may therefore extend down into the Wenlock.

The *Ancoradella ploeckensis* Zone extends from the upper Bowspring Limestone through the lower Hume Limestone. The upper Hume is referable to the upper Ludlow *Polygnathoides siluricus* Zone (Link and Druce, 1972). Comments on the taxonomy of the Yass Basin conodonts have been made by Cooper (1980) and Simpson (1995b).

The Silverdale Formation is overlain successively by the predominantly pelitic Black Bog and Rosebank Shales. Link and Druce (1972) recovered a small, low-diversity conodont fauna from the Yarwood Siltstone (a member of the Black Bog Shale) some 18–25 m above the Hume Limestone. This was termed "Fauna A" by Link and Druce (1972), and was tentatively correlated with Walliser's (1964) upper Ludlow *Kockelella latialata* Zone. Above this level in upper Ludlow and Pridoli horizons, the primarily siliciclastic lithologies have yielded a few graptolites, which have been summarized by Jell and Talent (1989) and documented by Rickards and Wright (1999).

Low numbers of uppermost Pridoli–lower Lochkovian *Icriodus woschmidti* Zone conodont elements have been recovered from shallow-water algal limestone lenses in an unnamed mudstone member above the basal sandstones of the Elmside Shale (Link and Druce, 1972). The uppermost Elmside Formation and overlying Sharpeningstone Conglomerate are thus inferred to be Lochkovian.

Despite the lack of a clearly defined interval with Pridoli conodonts, the Yass Basin has several graptolite horizons, three of which are Pridoli. Rickards and Wright (1999) appreciably expanded the knowledge of graptolites from the Yass sequence. A graptolite fauna 12 m below the top of the Black Bog Shale, which is dominated by *Bohemograptus bohemicus tennis* and *B. praecornutus*, indicates the *Bohemograptus praecornutus* Zone. *Bohemograptus* sp. nov. from the top of the Black Bog Shale indicates the *Bohemograptus cornutus* Zone (Rickards and Wright, 1999). The presence of Jaeger's (1967) *Monograptus formosus* fauna (a species earlier known as *M. salweyi*) low in the Rosebank Shale refers this unit to the lowest Pridoli *Paramonoclimacis parultimus* Zone. *Monograptus bouceki* and *M. transgrediens* in the Cowridge Siltstone at Barambogie Creek (Jaeger, 1967) indicate the middle Pridoli *Monograptus bouceki* Zone. Latest Pridoli graptolites are documented from the lowest 20 m of the Elmside Shale (Jenkins, 1982).

BOAMBOLO — A thick sequence of presumed Yass Formation (Pickett, 1982a, p. 92–95), which is much thicker than the typical Yass Formation in the Yass synclinorium, occurs as an inlier beneath Silurian volcanics in the Murrumbidgee River valley near Boambolo. This locality is 12 km south of Yass and 36 km northwest of Canberra (Fig. 2; no stratigraphic column included herein). The lower 500 m or more of shallow-marine to estuarine(?) siltstones with subordinate sandstone and limestone beds, is the Glen Bower Formation. The Glen Bower is overlain by 80–110 m of calcareous quartz arenites with a less diverse fauna than that from the Boambolo Formation. The Boambolo Formation is overlain, in turn, by volcanics thought to equate with the Willow Bridge Tuff of the Yass synclinorium. Despite numerous horizons with relatively high-diversity, well-preserved fossils, the area has been neglected paleontologically. The Silurian at Boambolo is inferred to be entirely Wenlock on the assumption that it correlates with the Yass Formation, but conodonts so far obtained by the authors do not allow a precise correlation of these Wenlock–Ludlow boundary strata.

CANBERRA — The sedimentary intervals in the Silurian about Canberra occur in a polygonal fault-delineated area of about 250 km^2. Despite generally poor outcrops and frequent faults, careful mapping (e.g., Henderson and Mutveev, 1980; Abell, 1982, 1991) that utilized temporary excavations has led to a much-improved understanding of the stratigraphy (Fig. 5, column 14). Correlation with the graptolite and conodont zonal schemes, however, remains generally problematic. The salient exception is the State Circle Shale. It rests unconformably on Ordovician "Pittman Formation," is overlain conformably by the unfossiliferous Black Mountain

Sandstone, and has *Monograptus exiguus, M. turriculatus, M. spiralis,* and *Retiolites* sp., which indicate the upper Llandovery *Spirograptus turriculatus* Zone.

Higher units in the Canberra region are not biostratigraphically tightly constrained. Unconformably overlying the Black Mountain Sandstone is the Canberra Formation, a unit considered entirely Wenlock (Henderson and Mutveev, 1980; Abell, 1982, 1991; Strusz, 1985). The report (Link, 1970) of *Kockelella variabilis* and the biostratigraphically less significant *Ozarkodina excavata* (as "*Spathognathodus inclinatus inclinatus*") from limestone in the "Riverside Formation," an abandoned term for an interval now regarded as high in the Canberra Formation, indicates a Ludlow, rather than Wenlock, correlation for the upper Canberra Formation. Unfortunately, the material studied by Link has been lost; sampling undertaken by Abell and Nicoll (*in* Abell, 1991, p. 19) failed to produce additional specimens.

The Canberra Formation is interpreted as overlain by the Ainslie Volcanics, and seemingly also by the Walker and Mount Painter Volcanics. However, faulted and tectonically problematical contacts have led to equivocal interpretations. The Mount Painter Volcanics are overlain by the Yarralumla Formation, which has been correlated with the Yass Formation of the Yass synclinorium (Strusz, 1984). This correlation is puzzling, as the uppermost Yass Formation is upper, not uppermost, Wenlock (Fig. 5, column 13), while on the basis of the conodont data noted above, the upper Canberra Formation is upper lower Ludlow.

If the Walker Volcanics correlate in a general way with the Ainslie and Mount Painter Volcanics, then we cannot accept that there is "firm stratigraphic evidence for a Wenlock age" for the Walker Volcanics (Strusz, 1982, p. 108). This interpretation seems to have arisen from placing undue reliance on an undocumented identification of the Wenlock–Ludlow pentameridine brachiopod *Rhipidium.* This genus was reported from the Camp Hill Sandstone by Öpik (1958) as an exclusively Wenlock form. Conodonts from the Yarralumla Formation and from limestones in the Walker Volcanics (e.g., near The Pinnacle and along the Molonglo River downstream from Coppins Crossing) are long-ranging forms and not helpful in precise correlations in the Wenlock–Ludlow (R. S. Nicoll *in* Abell, 1991; Simpson, Mawson, and Talent, unpublished data).

Column 14 (Fig. 5) follows Abell (1991), but extends the Canberra Formation into the lower Ludlow and adjusts the correlation of overlying units generally correlated with the Laidlaw–Elmside Formations of the Yass synclinorium (Fig. 5, column 13). The Deakin Volcanics, which overlie the Yarralumla Formation and are regarded as lower Ludlow, could be appreciably younger.

MICHELAGO–COOMA — The tract of grabens and folds with Silurian volcanic and sedimentary rocks extends southwards from Canberra for about 120 km through Michelago and Bredbo to "Dangalong," southeast of Cooma (Figs. 2, 3). Despite highly fossiliferous horizons in this ca. 800 km^2 area, surprisingly little paleontologic data, apart from reconnaissance identifications at the generic level (see Pickett, 1982a, p. 61, 62), are available. Poorly preserved Llandovery graptolites (i.e., *Glyptograptus incertus, Orthograptus* sp., *Climacograptus* sp., *Rastrites* sp., *Petalograptus* sp., *Monograptus communis communis, M. exiguus, M. gregarius, M. lobiferus, M. runcinatus, M.* sp. cf. *M. barrandei, M.* sp. cf. *M. intermedius, M.* sp. cf. *M. marri, M.* sp. cf. *M. nodifer, M.* sp. cf. *M. priodon* and *M.?* sp. cf. *M. regularis*) are reported from eight localities in proximal flysch of the Ryrie Formation (Richardson and Sherwin, 1975; Fig. 7, column 15). Shelly faunas reported from the Cappanana Formation and the Colinton Volcanics include taxa well known in the Yass district. They indicate a generalized Ludlow or latest Wenlock age, but there is an obvious need for systematic collecting and study, as well as sampling for conodonts in the scores of limestone lenses in this belt, especially on the Michelago 1:100,000 sheet (Richardson, 1979).

BUNGENDORE–CAPTAINS FLAT SYNCLINORIAL ZONE — There seems to have been no significant advance in the stratigraphy or paleontology of the Bungendore–Captains Flat area (Figs. 2, 3) over the past 25 years. The stratigraphic column (Fig. 5, column 16) is therefore much the same as in earlier reports (Talent et al., 1975; Pickett, 1982a; Walley et al.; 1990). Paleontologic data provide little constraint on this column.

GOULBURN–TARAGO SYNCLINORIAL ZONE — The stratigraphy for Goulburn–Tarago (Fig. 2, 5, column 17) follows Felton and Huleatt (1977). It is poorly constrained biostratigraphically. *Bohemograptus bohemicus* in the De Drack Formation and the rugose corals *Circophyllum, Zenophila,* and *Phaulactis* in its Sandhills Creek Limestone Member (Pickett, 1982a) are consistent with a generalized Ludlow correlation, and further specify the upper Ludlow (i.e., *Bohemograptus bohemicus–B. kozlowskii* Zone). The overlying Woodlawn Volcanics, Currawang Volcanics, and Covan Creek Formation are therefore assumed to be latest Ludlow, but possibly extend through the Pridoli and into the Lochkovian.

Four substantial but poorly known limestones are associated with volcanics south of Tarago. None are well constrained biostratigraphically. The largest and most southerly, the Bendithera Limestone, 55 km south of Braidwood (Fig. 2), is said to be a few hundred meters thick. *Tryplasma, Pycnostylus,* and *Kirkidium*-like brachiopods (Talent et al., 1975, p. 38) suggest a generalized Ludlow age.

BUNGONIA–WINDELLAMA — Mapping of the Bungonia area (Carr et al., 1980, 1981; Jones et al., 1981, 1984, 1986; Figs. 2, 3) led to recognition of five stratigraphic units. These include three limestone intervals separated by fine-grained siliciclastic units (Fig. 6, column 18). All five units have been formally named (Bauer, 1994, 1998). An exhaustive list of older fossil identifications is in Bauer and Bauer (1998). The stromatoporoids and tabulate and rugose corals of the limestone units are diverse, but have not been described, with the exception of the rugosans *Bungoniella clarkei* and *Hedstroemophyllum* sp. indet. from the Lookdown Limestone (Wright and Bauer, 1995). Because of the lack of biostratigraphically useful fossils, the position of the Wenlock–Ludlow boundary within or below the Lookdown Limestone is problematic. *Bohemograptus bohemicus tenuis* in the Cardinal View Shale indicates a late Ludlow age (Carr et al., 1980). More information has come from the Efflux Siltstone and the Sawtooth Ridge Limestone. A brachiopod-trilobite fauna from the Efflux Siltstone was thought to be Lochkovian (Carr et al., 1980; Jones et al., 1981). However, a conodont identified as *Spathognathodus* sp. cf. *S. remscheidensis* from limey nodules 5 m above the brachiopod-trilobite fauna is believed to be *Ozarkodina remscheidensis eosteinhornensis*, and likely Pridoli (Mawson, 1986). We suggest that the Efflux Siltstone may equate generally with the lower dolomitic interval of the Windellama Limestone, from which *O. remscheidensis eosteinhornensis* was reported (Mawson, 1986), and possibly with the unnamed siltstone unit beneath it at Windellama (Mawson, 1975). The Sawtooth Ridge Limestone would then correlate approximately with the thinly bedded dark limestones in the middle Windellama Limestone, and with lithologically similar limestones at the old Jacqua lime kiln, approximately 15 km to the north (Talent and Mawson, 1997; Mawson and Talent, 1999). This agrees with a report of the brachiopod *Cyrtina* from the Sawtooth Ridge Limestone (Bauer, unpublished data). The richly fossiliferous Windellama Limestone (Mawson, 1975, 1986; Mawson and Talent, 1999) spans at least five conodont zones (i.e., *Ozarkodina eosteinhornensis–Pedavis pesavis* Zones), and is Pridoli–upper Lochkovian (Mawson, 1986).

TARALGA AND MURRUIN CREEK — There have been no significant advances in understanding of the biostratigraphy of the Taralga Group (Fig. 5, column 19) over the past 25 years. A review of available data was summarized by E. Scheibner (in Pickett, 1982a, p. 186–190). Of special importance are the occurrences of the Ludlow graptolites *Monograptus crinitus*, *M. dubius*, and *Bohemograptus bohemicus* in the Cobra Formation, a siltstone unit with interbedded, richly fossiliferous limestones. However, the faunas have not been described or illustrated.

JENOLAN, ABERCROMBIE RIVER, OBERON, AND ROCKLEY — This large area south of the Bathurst batholith (Fig. 2, no stratigraphic column included herein) has been remapped (Pogson and Watkins, 1998; Raymond et al., 1998) and has several substantial outcrop belts of Silurian. However, there has been little advance in understanding the biostratigraphy of the region since Pickett's (1982a) summary. The most significant change has been nomenclatural. The Silurian sedimentary units have been united into the Campbell Formation (formerly Group), and assigned to the Mumbil Group. The Mumbil Group was based originally on sequences north and northwest of the Bathurst batholith. In addition, the former Kildrummie Group (a limestone-rich unit) has been reduced to a member of the Campbell Formation. An understanding of the stratigraphy and biostratigraphy has been hampered by structural complications, metamorphism, a dearth of paleontologic data, and the suspicion that many of the major limestones in the Kildrummie Member are olistoliths. Conodonts from an area of Kildrummie Member outcrop south of Rockley were thought by De Decker (1976) to indicate the upper Ludlow post–*Polygnathoides siluricus* Zone. Subsequently, they have been reinterpreted to be upper Wenlock–early Ludlow (Simpson, 1995b, p. 340). This accords with a correlation based on tabulate corals (J. G. Byrnes in De Dekker, 1976).

One of the most important areas of Silurian limestone is at Jenolan, where two belts of lithologically different limestones have been mapped and interpreted as a synclinal repetition of the Jenolan Caves Limestone. There is still uncertainty about this interpretation. The western limestone tract extends at least up to the upper Ludlow *Ozarkodina crispa* Zone (P. D. Molloy, personal commun., 1972).

CENTRAL AND WESTERN NEW SOUTH WALES

Few biostratigraphically useful conodonts and graptolites are known in the Lower Devonian north of the Bathurst batholith. These localities are on the eastern and western flanks of the Hill End Trough and in northern parts of the Capertee High (Colquhoun, 1995; Talent and Mawson, 1999; Mawson and Talent, 2000; Fig. 2). Unfortunately, there is a dearth of biostratigraphic data for the Silurian of this region, apart from the Mudgee–Cudgegong area (Fig. 2).

SOFALA–PORTLAND–PALMERS OAKEY — Conodonts have been listed from various localities in the vicinity of Palmers Oakey (Bischoff and Fergusson, 1982), about 12 km east and northeast of Limekilns. However, their stratigraphic position was problematical in the Hill End Trough until Pickett et al. (1996) concluded that many of the samples were collected from the Tanwarra Shale (Silurian). There are other limestone olistoliths farther east in the Portland area. These indicate formerly more extensive areas of a "now-lost" carbonate platform of generalized Silurian age along the Capertee High southeast of the western flank of the Hill End Trough.

MUDGEE–CUDGEGONG — Information on the Mudgee–Cudgegong area (Figs. 2, 3) is based on mapping over the past two decades (Pemberton, 1980, 1989; Cook, 1990; Pemberton et al., 1994; Wright et al., 1994; Colquhoun et al., 1997), and includes radiometric and graptolite data (Rickards et al., 1998). The Tannabutta Group consists of six stratigraphic units, but generalized biostratigraphic control is possible on only three of them (Fig. 5, column 20). The oldest, or Willow Glen Formation, is a suite of shallow-water arenite, conglomerate, and carbonate, that rest unconformably on the Upper Ordovician Sofala (=Cudgegong) Volcanics. The Willow Glen is overlain by, and is partially laterally equivalent to, the lower Windamere Volcanics, a thick sequence of subaerial dacitic lavas and breccias that grade laterally into shallow marine volcaniclastics with debris flows (Toolamanang Formation, not on Fig. 5, column 20). The Windamere Volcanics are transitional upwards into shallow-marine breccias, limestones, and dacitic conglomerates (Millsville Formation, not shown on Fig. 5, column 20). To the east in the vicinity of Kandos, the Moonbucca Formation, a succession of conglomerate, arenite, limestone breccia, and mudstone, correlates broadly with the above assemblage. Another generally correlative unit is the Dungeree Volcanics, a unit of dacitic and rhyolitic lavas, volcaniclastics, mudstones, and rare limestones to the northeast of Cudgegong in the Lue district. Biostratigraphic control for the upper part of the Windamere Volcanics is based on the recovery of *Monograptus parultimus minutus* (Rickards et al., 1998), which indicates a probable early Pridoli age. This fossil, coupled with five radiometric dates from the Windamere and Dungeree Volcanics, leads to the conclusion that the Willow Glen Formation is Wenlock, rather than Ludlow, as previously believed (Rickards et al., 1998).

NORTHEAST FLANK OF MOLONG ARCH — Silurian limestones occur as isolated clasts and carbonate-rich channel and/or fan deposits in the Upper Silurian and Lower Devonian on the Molong Arch (Fig. 2; no stratigraphic column herein), as well as along its eastern and western flanks. Talent and Mawson (1999) discussed carbonate fan deposits in the Wallace Shale in the "Canobla" area southwest of Stuart Town, where conodonts indicate derivation of material (including clasts) from the lower Wenlock Nandillyan Limestone. They also documented a large olistolith with an age close to that of the Ludlow–Pridoli boundary. The olistolith occurs within the Emsian of the Nubrigyn Member of the Cunningham Formation in the Cooper Creek area (northwest of Euchareena). This Silurian interval is not known to be preserved along platform sequences west on the Molong Arch.

NORTHERN MOLONG ARCH — Recent mapping and conodont investigations have brought about substantial changes in the stratigraphy of the northern Molong Arch (Walley et al., 1990, column 32). There are remaining questions of interrelationships, including degree of diachroneity, between the various units. The thick Nandillyan and Narragal Limestones (Fig. 5, column 21) are lower Wenlock (at least in part) and lower–middle Ludlow, respectively, rather than correlative (Percival, 1998). The Nandillyan Limestone seems to correlate with, rather than postdate, the Dripstone Formation (formerly "Group;" Vandyke and Byrnes, 1976) and its southerly extension, the Mullion Range Formation. The Narragal Limestone is, in part, diachronous with the Wallace Shale (formerly Barnby Hills Shale [Richards and Wright, 1997a] but now rejected as a junior synonym of the Wallace Shale) (Cockle, 1999; Talent and Mawson, 1999). Not shown on Fig. 5, column 21, are the Molong Limestone (Adrian, 1971), for which there is limited biostratigraphic control, and the thick (>380 m) Camelford Limestone that crops out prominently in two principal areas in the northern Molong Arch. These localities are at The Gap, between Molong and Cumnock, and at Neurea, south of Wellington. Though reported to have conodonts of the *Ozarkodina eosteinhornensis* and *Icriodus woschmidti* Zones (G. C. O. Bischoff *in* Chatterton et al., 1979), Farrell (in Talent, 1995) demonstrated the Camelford Limestone to be entirely Lochkovian.

SOUTHERN MOLONG ARCH — Pivotal to an understanding of the stratigraphy of the important Silurian sequences west of Orange (Fig. 2) has been the work of G. H. Packham and his students, particularly C. J. Jenkins. The Silurian stratigraphy has been reviewed (Talent et al., 1975; Jenkins et al., 1986; Jell and Talent, 1989; Simpson, 1995b). The stratigraphic framework and associated graptolite and conodont faunas (Stevens and Packham, 1952; Packham, 1969; Jenkins, 1977; Rickards et al., 1995) was used to infer a series of tectono-stratigraphic events (Packham, 1969; Talent, 1989). Other useful contributions to Llandovery–Wenlock conodonts of the region include reports by Bischoff (1987), Simpson (1995b), and Cockle (1999).

The Four Mile Creek and Quarry Creek area of central New South Wales (Fig. 6, column 22) contains the best Lower Silurian conodont faunas known in Australia. The oldest carbonate units are separated from the underlying Angullong Tuff by the Cobblers Creek hiatus. Jenkins (1977) considered that the Angullong Tuff correlates with the two uppermost Ordovician graptolite zones and probably extends into the Silurian. From the overlying sequence, Bischoff (1987) discriminated four Llandovery "assemblage zones" below the upper Llandovery *Pterospathodus celloni* Zone. Simpson (1995b) reinterpreted the data and proposed retention of three of these zones, but with minor modifications and redefinition. The oldest interval, the *Distomodus combinatus* Zone (Simpson, 1995b), is represented in the carbonates of the lower Bagdad Formation, which includes the Wire Gully Limestone. Bischoff (1987) correlated this member with the upper lower Llandovery (Rhuddanian) *Monograptus cyphus* (graptolite) Zone, but considered the lower boundary of the member could extend down into the older *Cystograptus vesiculosus* (graptolite) Zone. The oldest carbonates were barren of conodonts.

The *Distomodus combinatus* Zone is succeeded by the *D. pseudopesavis* Zone of Simpson (1995b). This latter zone occurs within the Bridge Creek Limestone and the upper Bagdad Formation. It is equated with the upper lower–lower middle Llandovery (Rhuddanian–Aeronian) *Monograptus cyphus* and *Demirastrites triangulatus* (graptolite) Zones (Bischoff, 1987). The top of the *D. triangulatus* Zone cannot be recognized with certainty due to the Panuara hiatus, which, based on graptolite data (Jenkins, 1977), is believed to encompass both the Aeronian *Monograptus convolutus* and *M. sedgwicki* Zones.

Deposition of a suite of varied lithologies that comprise the Waugoola Group followed the Panuara Hiatus. The oldest carbonate unit in the Waugoola Group is the Cobbler's Creek Limestone, which has been correlated with the late Aeronian *Monograptus turriculatus* Zone but extends into younger strata in some areas. Bischoff (1987) inferred a number of "pre-*Pterospathodus celloni*" conodont assemblage zones, and equated the Cobbler's Creek Limestone with the *Aulacognathus antiquus–Distomodus staurognathoides* morphotype A Assemblage Zone. Simpson (1995b) discussed some conceptual difficulties with this, and recommended abandoning this assemblage zone for the more globally applicable *D. staurognathoides* Zone.

Bischoff (1987) also proposed an overlying *Astropentognathus irregularis–Pterospathodus pennatus* Assemblage Zone. Most or all of this zone can probably be equated with the *Pterospathodus celloni* Zone (Simpson, 1995b). The *P. celloni* Zone has been identified in the Liscombe Pools Limestone, Quarry Creek Limestone, Glendalough Formation (Burly Jack Sandstone), and lower Boree Creek Formation. Simpson (1995b) suggested that the base of the *P. celloni* Zone is probably close to the base of the Liscombe Pools Limestone and Boree Creek Formation.

Many of the carbonates of the Waugoola Group and other more-or-less coeval limestones from mid-western New South Wales correlate with the *Pterospathodus celloni* Zone and extend into the upper Llandovery–lowest Wenlock *P. amorphognathoides* Zone. Simpson (1995b) suggested the lower boundary of the latter zone could be identified in the lowest Quarry Creek Limestone and within the Boree Creek Formation (Fig. 6, column 23).

Although the base of the *Pterospathodus amorphognathoides* Zone can be discriminated with reasonable confidence, recognition of the top of the zone in the Four Mile Creek–Quarry Creek area is more difficult. This problem is exacerbated by the definitions of the overlying *Kockelella ranuliformis* Zone and the pattern of intense faunal turnover associated with the Irevikin Event (Jeppsson, 1998). Simpson (1995b) noted differences in the nature of the lower boundary of the overlying *K. ranuliformis* Zone in the Boree Creek Formation and the Quarry Creek Limestone. However, he overlooked the fact that the uppermost unit sampled at Quarry Creek (Bischoff, 1987, sample Qr 17), and the only unit referable to the *K. ranuliformis* Zone, is allochthonous. The apparent sharpness of the boundary at the base of this zone, in comparison with that of the Boree Creek Formation, is therefore illusory. Bischoff (1987) noted the lack of taxa commonly associated with the upper *P. amorphognathoides* Zone at Quarry Creek and suggested the uppermost beds of this limestone did not extend to the top of this zone. The sequential turnover of conodont taxa, commonly associated with the Irevikin Event, is not seen in Quarry Creek (though this may be an artifact of sampling). As the base of the Irevikin Event is close to the Llandovery–Wenlock boundary (Aldridge et al., 1993), it is feasible that the entire Quarry Creek Limestone is upper Llandovery. This possibility requires testing by more intensive sampling.

Rickards et al. (1995) recorded five Wenlock and Ludlow graptolite zones in Panuara Formation siliciclastics above the Quarry Creek Limestone. The oldest is the *Monograptus riccartonensis* Zone. This implies a possible hiatus of up to two lower Wenlock graptolite zones between the Quarry Creek Limestone and the Panuara Formation.

Boree Creek — Column 23 (Fig. 6) follows Cockle (1999). The Boree Creek Formation (Sherwin, 1971) extends from the *Pterospathodus amorphognathoides* Zone into the *Kockellela ranuliformis* Zone. Thus, the Boree Creek Formation embraces the Ireviken Event, for which isotope data have been presented (Talent et al., 1993). The Borenore Limestone, which overlies the Boree Creek Formation with a distinct unconformity, commences somewhere within the *K. ranuliformis* Zone. The Quarry Creek hiatus may have been shorter in the Boree Creek area than in the Four Mile Creek–Quarry Creek area; it is in the *K. ranuliformis* Zone in the Boree Creek area.

Manildra–western flank of Molong Arch — The oldest Silurian unit of this area (Fig. 2), the Greengrove

Formation, crops out in the core of the Cudal anticline as poorly exposed shale and limestone with minor conglomeratic horizons (Fig. 6, column 24). The lowest exposed horizons yield *Glyptograptus tamariscus, Pseudoclimacograptus (Metacimacograptus) hughesi, Monograptus argutus, M.* sp. cf. *M. revolutus*, and *Rastrites* sp. cf. *R. approximatus* that indicate the middle Llandovery *M. gregarius* Zone. Graptolites high in the sequence are probably Wenlock (Savage, 1968; Sherwin, 1974). The limestone bodies of the Greengrove Formation are 30–50 m in thickness, and crop out better than the siliciclastic parts of the succession. The local distribution and shoal-water faunas of the limestones contrast with the surrounding graptolitic facies; this suggests the limestones are probably allochthonous. The undescribed faunas include pentamerid and trimerellid brachiopods (Talent and Mawson, unpublished data).

The Kurrajong Park Formation has *Monograptus flemingi, M. dubius*, and *M. testis* (Savage, 1968), which indicate that at least part of it is upper Wenlock. The Mackeys Creek Shale has *M. colonus?, M. flemingi?*, and *Bohemograptus bohemicus?*, which indicate a Ludlow, but possibly not latest Ludlow, age. The overlying Goonigal Group (=Fairview Formation of Savage, 1968) is a lithologically diverse interval of greywacke and tuffaceous sandstone with siltstone, tuff, and shale. Though four units have been discriminated elsewhere in the Goonigal Group (Pickett, 1982a), only the Wansey Formation is recognized in the Manildra area (Pogson and Watkins, 1998; Raymond et al., 1998). In the vicinity of "Fairhill," there are eight richly fossiliferous rudaceous limestone lenses composed of allochthonous materials. Four of these lenses are shown on Savage's (1968) map. We have sampled these, but all eight lenses were barren of conodonts. Diverse brachiopods from the uppermost Wansey Formation (Savage, 1968, 1974) and monographed from the basal Maradana Shale (Savage, 1974) at Manildra appear to be Pridoli–early Lochkovian, but these faunas are not tightly constrained relative to the Silurian–Devonian boundary.

CANOWINDRA — Information for column 25 (Fig. 6) is derived from Pickett (1982a). The Millarnbri Formation conformably overlies the Upper Ordovician Rockdale Formation. *Glyptograptus tamariscus, Pseudoclimacograptus (Metaclimacograptus) hughesi, P. (M.) undulatus, P. (M.) retroversus*, and *Monograptus jonesi* indicate the middle Llandovery, probably what is known as the Monograptus gregarius Zone. Conodonts from the lowest Liscombe Pools Limestone (Percival, 1976; Pickett, 1978) are thought to be close to the base of the *Pterospathodus celloni* (conodont) Zone, and correlate with the *Monoclimacis griestoniensis* (graptolite) Zone. *Monograptus priodon, M.* sp. aff. *M. vomerinus, Retiolites geinitzianus*, and, possibly, *M. spiralis* indicate a late Llandovery–early Wenlock age for the overlying Gospel Oak Formation. Nondescript shelly faunas and *Monograptus.* sp. cf. *M. dubius* from the Avoca Valley Shale, and *M. dubius* and equally nondescript shelly faunas from the overlying Ghost Hill Formation, are consistent with a Wenlock age of the Avoca Valley Shale and a Wenlock–early Ludlow age of the Ghost Hill Formation (Pickett, 1982a). *Bohemograptus* sp. aff. *B. bohemicus* in the Belubula Shale (Pickett, 1982a) indicates a late Ludlow age for at least part of the Belubula Shale. In the absence of fossils, the position of the Silurian–Devonian boundary in the Tenandra Formation is problematic.

CUMNOCK–YEOVAL — The Loombah, Bournewood, and Yullundry Formations are not well constrained biostratigraphically (Fig. 6, column 26). The Cary Formation has an upper Homerian graptolite association (*Monograptus testis, M. irfonensis, M. dubius, M. flumendosae* and *M.* sp. cf. *M. uncinatus*), which suggests that the formation extends into the Gorstian (L. Sherwin *in* Pickett, 1982a, p. 79). *Monograptus colonus, M. salweyi, Bohemograptus bohemicus*, and *Spinograptus spinosus?* high in the Sourges Shale (L. Sherwin *in* Pickett, 1982a, p. 86) suggest that the Sourges extends through the Gorstian into the Ludfordian. However, the Sourges Shale is also reported to be conformable on Late Ordovician limestone (Meakin and Morgan, 1999, p. 53–55), and this suggests a remarkably long time span. The highly fossiliferous Burrawong Limestone is not well constrained biostratigraphically, but a generalized middle–late Ludlow age is inferred (Pickett, 1982a, p. 76). The overlying Buckinbah Volcanics and Myrangle Formation of the Goonigal Group are therefore suggested to be Pridolian–earliest Lochkovian.

TOONGI–COWRA TROUGH AXIS — The almost entirely siliciclastic Toongi Group crops out 15–35 km south of Dubbo (Colquhoun et al., 1997; see Fig. 2; no stratigraphic column included herein). It consists of fourteen formations with a composite thickness of ca. 8,370 m (Packham, 1969; Pickett, 1982a, pp. 124–132). Paleontologic control is lacking for any of these poorly exposed units. Only the highest unit, the Glengeera Formation, has produced poorly preserved conodonts (Pickett, 1982a). Early Devonian corals have also been identified from the formation (I. G. Percival, personal commun., 1997).

WEST FLANK OF COWRA TROUGH — Latest Silurian–Early Devonian sequences west of the Cowra Trough (Fig. 2) are predominantly siliciclastic, with marine influence apparently decreasing west over 500 km into the Darling Basin and Bancannia Trough. A single stratigraphic column is not presented herein, but two columns, Forbes (Fig. 6, column 27) and Narromine (Fig. 6, column 28), are included. Lochkovian and higher units in western New South Wales have been reviewed (Sherwin, 1997;

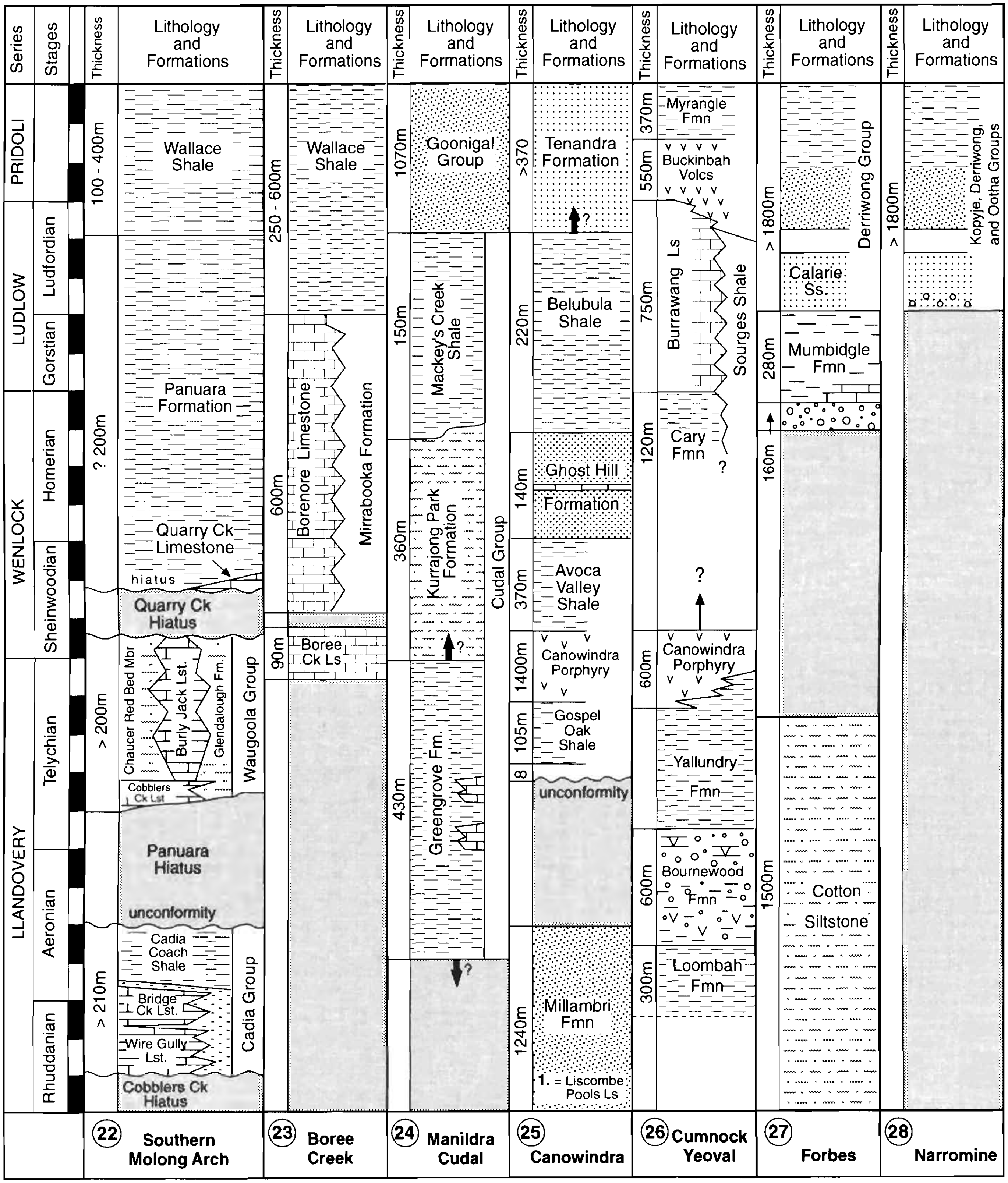

FIGURE 6 — Correlation of the Silurian of east-central New South Wales, Queensland, and Western Australia. Location of stratigraphic columns on Fig. 3. Depth curves speculative. No attempt has been made to provide depth curves for stratigraphic columns where correlations are conjectural.

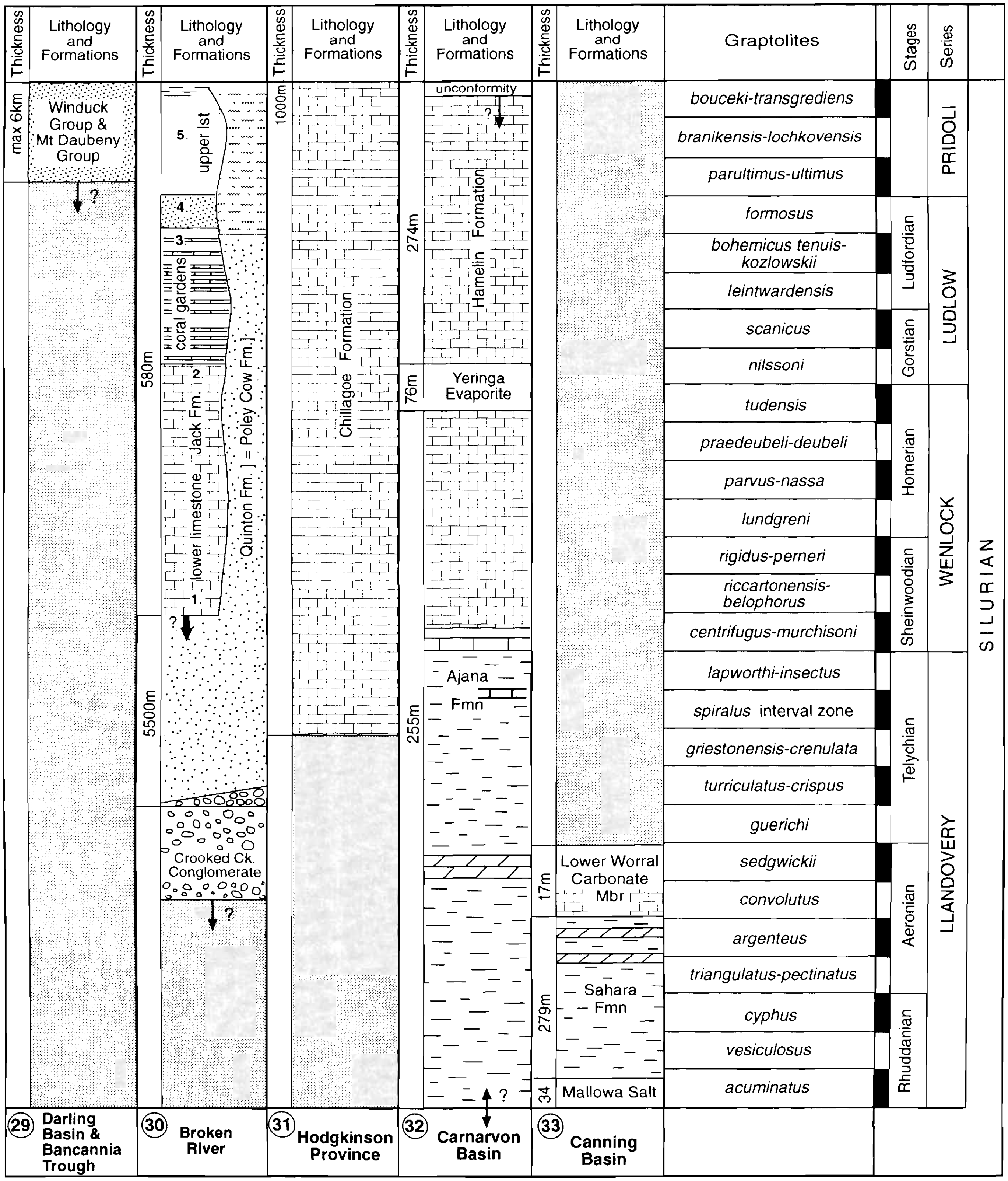

FIGURE 6 *continued.*

Mawson and Talent, 2000). The Silurian is almost entirely siliciclastic, generally poorly exposed, often deeply weathered, and, with a dearth of paleontologic data, not always readily differentiable from the lithically similar Lower Devonian. Conodonts from about a dozen minor limestone occurrences (Pickett, 1980, 1982b, 1984, 1986), and typically poorly preserved brachiopods from a few localities (Sherwin, 1990, 1995), provide some biostratigraphic data.

FORBES — The Cotton Siltstone at Forbes (Fig. 6, column 27) appears to be continuous from the Upper Ordovician into the Llandovery. The highest graptolite assemblage indicates the *Monograptus turriculatus* Zone (Sherwin, 1973, 1974; Sherwin et al., 1987; Loydell, 1990). Limestones from the lowest Mumbidgle Formation have a Wenlock–early Ludlow corals, with *Monograptus dubius* and *M. sherrardae* higher in the Mumbidgle. The correlation of the Calarie Sandstone is speculative; it is based on an uppermost Ludlow–lowest Lochkovian correlation (*Ozarkodina crispa–Icriodus woschmidti* Zones) of the Cookeys Plains Formation of the Derriwong Group in the Narromine area (Fig. 6, column 28).

NARROMINE — Three major Pridoli–Lochkovian units have been discriminated in the Tullamore and Murda synclines on the Narromine 1:250,000 sheet (Sherwin, 1980, 1992, 1994, 1997). The Kopyje and Derriwong Groups and the incompletely documented, partly turbiditic Ootha Group occur at Narromine (Fig. 6, column 28). These groups may be parts of a single sedimentary–volcanic sequence, but intervening tracts of older units and large areas of poor outcrop inhibit local correlations. All three groups are characterized by a coarse, 50–500 m-thick interval of conglomerates and sandstones (Mount Susannah Conglomerate, Edols Conglomerate, Calarie Sandstone). This siliciclastic-dominated basal interval is interbedded with or overlain by about 200–300 m (but perhaps as much as 1,650 m) of acid or intermediate volcanics (Mineral Hill, Meloola, Byong, Fermor, and Milpose Volcanics). These volcanics are overlain by up to 1.6 km of sandstone with subordinate siltstone (Talingaboolba, Cookeys Plains, and Yarrabandai Formations). Lenticular limestones in the Cookeys Plains Formation of the Derriwong Group have conodonts of the *Ozarkodina crispa*, *O. eosteinhornensis*, and *Icriodus woschmidti* Zones (Pickett and Ingpen, 1990; Pickett and McClatchie, 1991; Pickett, 1992). Horizons below the *Ozarkodina crispa* Zone may be present in such older units as the Calarie Sandstone.

DARLING BASIN AND BANCANNIA TROUGH — West and south of Cobar (Fig. 1) are major outcrop belts of the Winduck Group (Fig. 6, column 29), an interval divided into three formations that equate with some, or all, of the generally more arenaceous Amphitheatre Group farther east (Baker et al., 1975; Glen, 1979). Three formations have been defined: the Gundaroo Sandstone, Buckambool Sandstone, and Sawmill Tank Siltstone, which have generally been assumed to be Lower Devonian (e.g., McRae, 1989a, 1989b). However, Sherwin (1995) used macrofaunas from The Meadows area, 50 km southwest of Cobar, to argue that the Winduck Group extends down into the Pridoli, although it has Lochkovian and, possibly, Pragian shelly faunas, as well (Sherwin, 1995). Winduck Group siliciclastics with very rare carbonate intervals have been penetrated in several bore holes in the Darling Basin farther west. The carbonates tend to be reddish and oolitic/oncolitic; some are white and lack macrofossils; others may represent calcretes. The Silurian–Devonian boundary in the Winduck Group, and presumably coeval intervals penetrated in bores in the Darling Basin, remain problematic (Bembrick, 1997).

Very shallow-marine conditions appear to have extended northwest from the Darling Basin into the generally fluviatile Bancannia Trough (Neef et al., 1996), where the Mount Daubeny Formation is a thick, terminal fan-deposited suite of siliciclastics considered Upper Silurian–Lochkovian. Walley et al. (1990) indicated that the Mount Daubeny Formation was approximately 2,000 m thick adjacent to the Koonenberry fault, and thinner elsewhere. The type section of the formation (Neef et al., 1989) is considered to be more than 6 km thick. The Mount Daubeny Formation is extensive, overlies Proterozoic or Lower Paleozoic, and is overlain by Mulga Downs Group siliciclastics, a unit regarded as Emsian–Eifelian (Neef et al., 1996).

A Late Silurian–Lochkovian age is indicated by a *Baragwanathia* flora ca. 4 km above the base of the formation at the type section. Neef et al. (1996, p. 11) cited a lack of typical Pragian plant material to support a Late Silurian–Lochkovian age. Trace fossils resembling the nonmarine *Scoyenia* ichnofauna are found and used to support this correlation (Neef et al., 1996). It is therefore possible that part or all of the Mount Daubeny Formation is Silurian.

NEW ENGLAND (TAMWORTH BELT AND TABLELANDS COMPLEX) — Autochthonous Silurian sedimentary rocks are not known from northeast New South Wales. However, conodonts and shelly fossils, principally corals, occur in limestone olistoliths occur in the Tamworth belt, also known as the Gamilaroi terrane (Aitchison, 1988; Blake and Murchey, 1988; Aitchison et al., 1992, 1994), and in the Tablelands complex east of the Peel fault (Leitch, 1974; Cawood, 1980; Ishiga and Leitch, 1988) in northeast New South Wales. The Tamworth belt extends 300 km north-northwest from 200 km northeast of Sydney through Tamworth to Bingara (Fig. 1). Six occurrences of well-dated fossils are documented, one from chert and five from limestone clasts in debris-flow sequences. These debris flows indicate a "now-lost" carbonate platform or platforms to the west, or a derivation from sea mounts. One of the fossil assemblages, thought to be Late Silurian rather than earliest Devonian, is from a clast associated with Ordovician and lower Emsian clasts in the lower Emsian Drik Drik Formation (Furey-Greig, 1995) at grid reference 161441, Dungowan 1:25,000 topographic sheet. This clast produced *Ozarkodina* sp. nov., presently being described by A. J. Simpson on the basis of material from the Broken River region, northeastern Queensland. This form ranges from the middle Ludlow *Polygnathoides siluricus* Zone to the lower Lochkovian *Icriodus woschmidti* Zone (A. J. Simpson, personal commun. 1998). The other occurrences are from the Tablelands complex. Two of these, both east of Manilla, have late Llandovery corals in allochthonous limestone blocks from the Wisemans Arm Formation (Chappell, 1961; Hall, 1978; Leitch and Cawood, 1980). The limestone blocks at grid reference 065866, Klori 1:25,000 topographic sheet, have produced a single element of *Kockelella variabilis* Walliser (Pickett, 1982a). This suggests that at least one of the blocks is Ludlow. A locality at Uralba, about 5 km further north along strike, has Ludlow corals, as well as the conodonts *Aspidognathus* sp. cf. *A. tuberculatus* and *Distomodus* sp. (Furey-Greig, 2000). Both localities have Late Ordovician (Eastonian–Ea3) and Llandovery olistoliths. A fourth locality, 6 km farther east in the Wisemans Arm Formation, yielded a fragment of *Aspidognathus?* (Furey-Greig, 1999, Pl. 4, fig. 18, identified as Gen. et sp. indet. A). The fifth locality, an allochthonous limestone block in the "Woolomin Beds" about 3 km east of Bingara, has such corals as *Favosites gothlandicus*, *F.* sp. cf. *F. libratus*, and *Entelophyllum* sp. (Lusk, 1964, locality L5; identifications by G. M. Philip). The block failed to produce conodonts, but a generalized Wenlock–Ludlow age is indicated.

Ishiga et al. (1988) reported *Ozarkodina eosteinhornensis* and *Walliserodus* sp. from bedded cherts associated with thrust slices of oceanic basalts, cherts, and fine-grained sedimentary rocks of the Woolomin Formation (sensu Leitch and Cawood, 1980) at grid reference 179494, Dungowan 1:25,000 topographic sheet. A Pridoli age is indicated.

The above allochthonous occurrences are in facies that ranges from abyssal (Woolomin Formation) to probable outer forearc basin (Wisemans Arm Formation), and extends over 250 km along strike. The faunas indicate a Silurian source or sources that range from late Llandovery through Pridoli and, possibly, into the Lochkovian. An absence of clasts older than the *Pterospathodus amorphognathoides* Zone is consistent with the limited record of early and middle Llandovery sedimentary rocks in eastern Australia. This rarity is apparently connected with the Benambran orogenic cycle, an event with profound impact on sedimentation throughout eastern Australia.

Queensland

Conodonts obtained from extensive limestone sampling in the Silverwood Formation near the Queensland–New South Wales border (K. Van Noord and R. Mawson, unpublished data) have failed to demonstrate the Silurian. All of the limestones appear to be allochthonous and derived from a "now-lost" Early Devonian (Lochkovian–Pragian) carbonate platform.

Rockhampton hinterland — Druce (1970) reported *Icriodus woschmidti* from Marmor and the Mount Holly Beds of central coastal Queensland (Murray, 1975; Fig. 3; no stratigraphic column included herein). It was presumed from this report that some and perhaps much of the Middle Paleozoic that crops out from Mt Etna north of Rockhampton (Fig. 1) and extends south to Gladstone could straddle the Silurian–Devonian boundary and could include substantial Silurian intervals. However, the major limestones in the region are primarily upper Emsian (Mawson et al., 1995; Mawson and Talent, unpublished data), and the matrix in which they occur, the Mount Alma Formation, is Middle–Upper Devonian (P. Blake *in* Yarrol Project Team, 1997). By contrast with the Devonian of central Queensland (Mawson and Talent, 2000), marine conditions seem not to have extended as far west as the Anakie High.

Telford (1969) documented *Ozarkodina remscheidensis eosteinhornensis* and, thus, the uppermost Ludlow–Pridoli in the Craigilee Beds. This predominantly volcanic and volcaniclastic sequence with minor limestone and siltstone (Talent et al., 1975) crops out along the core of the Craigilee anticline about 55 km west of Rockhampton (Fig. 1). A small brachiopod–trilobite fauna (McKellar, 1969a) accords with this age assignment. Recent mapping and conodont work have improved knowledge of the Craigilee Beds and have shown that they are Pridoli–Lower Devonian (G. Simpson *in* Yarrol Project Team, 1997).

Broken River and Gray Creek — Silurian rocks in the Broken River–Gray Creek region (Figs. 1, 3, locality 30) consist of allochthonous and autochthonous siliciclastic, argillaceous, and carbonate sequences (Fig. 6, column 30). The predominantly siliciclastic and areally extensive Quinton Formation crops out in the northern Gray Creek area, where it overlies the Crooked Creek Conglomerate. The latter has carbonate clasts that include large limestone olistoliths. Withnall (1985, p. 42) noted that conodonts from limestone olistoliths in this unit consist of long-ranging forms that could be Late Ordovician or Early Silurian. No faunal list is available from this earlier work.

Age is better constrained for the overlying Quinton Formation, particularly in the Gray Creek area, where conodonts and graptolites have been obtained from several localities. These include Early Silurian (Telychian) graptolites at Gray Creek (Jell et al., 1993, p. 240). A small unit of allochthonous carbonates near Top Hut, while not yielding zonal species, includes a rich upper Llandovery *Pterospathodus celloni* Zone conodont fauna (Lane and Thomas, 1978; Simpson, 1995a; Simpson, 1999). Another limestone interval, apparently a carbonate fan with limestone clasts at Tomcat Creek, has a poorer upper Llandovery–lowest Wenlock fauna of the *Pterospathodus celloni–P. amorphognathoides* Zones (Simpson, 1999).

The "Magpie Creek Limestone," formerly assigned to the upper Quinton Formation, was assumed to be Silurian and was equated in some way with the Jack Formation farther south. It has been interpreted as allochthonous (Munson, 1987; Munson and Jell, 1999) or autochthonous (Withnall et al., 1993). Recent mapping (Talent and Mawson, unpublished data) demonstrates that the "Magpie Creek Limestone" is a mappable unit up to about 1 km thick, and is distinct from the Jack and Shield Creek Formations. It also differs from the Quinton Formation, except in outcrop areas north of the typical "Magpie Creek" occurrences. The "Magpie Creek Limestone" is a complex of limestone-rich debris flows, submarine fans, and discrete olistoliths in a predominantly mudrock matrix that was emplaced during the middle Lochkovian (post-*Icriodus woschmidti*–pre-*Pedavis pesavis* Zone). Conodonts now available from the "Magpie Creek Limestone" are sparse and imply a generalized late Llandovery–Pridoli age for the source materials (Simpson, 1995a). However, conodonts from clasts in the "Magpie Creek Limestone" in the Jessey Creek area include *Icriodus* sp. cf. *I. woschmidti* (considered by Simpson, unpublished data, as an *Icriodella–Icriodus* transitional form). This form demonstrates derivation by cannibalization of the upper Jack Limestone (Sloan et al., 1995), a unit which extends into the *Icriodus woschmidti* Zone near Broken River Crossing (Simpson, 2000).

To the south near Broken River, siliciclastics referred to as the Poley Cow Formation (Withnall et al., 1993) are regarded as a proximal equivalent of the Quinton Formation (Simpson, 1999). Graptolites from several localities in the Poley Cow Formation (Jell et al., 1993) indicate a general late Llandovery (Telychian) age. A carbonate lens from the vicinity of the Broken River Crossing has *Pterospathodus celloni* Zone conodonts. Lack of any discernible age differences between conodonts from the allochthonous carbonates and graptolites of the enclosing siliciclastics indicates nearly contemporaneous lithification, erosion, and redeposition of these carbonates.

The Quinton Formation is overlain by the Jack Formation. The latter formation, as presently interpreted, is extensive and lithologically heterogeneous. Limestones crop out in two principal belts. One extends northeastward from the Jack Hills Gorge to the Diggers Creek–Jessey Creek area; the other crops out near Broken River Crossing. In the type section at Jack Hills Gorge, the Jack Formation consists of two limestones separated by a siliciclastic interval. Sparse conodonts indicate that the base of the lower limestone is lower Wenlock (Simpson, 1995a); age control on higher levels is imprecise. The intervening siliciclastics pass up into a distinctive interbedded carbonate and siliciclastic sequence termed the "coral gardens;" the Ludfordian *Polygnathoides siluricus* Zone occurs within this interval (A. Simpson, unpublished data). The upper unit of the Jack Formation is a thick sequence of massive carbonates. Conodonts indicate that the uppermost horizons in the type area of the Jack Formation are still Ludlow (Simpson, 1995a).

There is no equivalent of the lower limestone unit of the Jack Formation in the Broken River Crossing area, where the "coral gardens" directly overlie Quinton Formation siliciclastics. The Gorstian–Ludfordian *Ancoradella ploeckensis* and *Polygnathoides siluricus* Zones are recognized in the "coral gardens" sequence, and the uppermost Jack Formation has uppermost Silurian–lowest Devonian *Icriodus woschmidti* Zone conodonts (Simpson, 2000). There is no obvious break in the sequence; the Jack Formation extends through a generalized Pridoli interval into the Lochkovian. In other areas, Jack Formation carbonates are not younger than Ludlow (Simpson, 1995a).

The Jack Formation is overlain paraconformably to unconformably, locally with a deeply incised erosion surface, by the Ralph Flint Formation, a mudstone–shale–debris flow sequence with a maximum thickness of almost 1 km. Many limestone clasts and olistoliths in this unit are obviously derived from the Jack Formation (Sloan et al., 1995), but others have dimensions up to 0.5 km. These large blocks occur in the Turtle Creek watershed, 1.5–3 km southwest of where the Wando Vale road crosses the "Magpie Creek Limestone" and around the Wade anticline (notably at Arch Creek Gorge). These massive, often sheared, rarely fossiliferous, pale grey blocks resemble the large Upper Ordovician–Llandovery limestone olistoliths in the Crooked Creek Conglomerate and the Carriers Well and Perry Creek formations (Talent and Mawson, unpublished data).

CAMEL CREEK SUBPROVINCE — This area (Fig. 1; stratigraphic column not included herein) was mapped by Withnall (1993). Limestones, which include large olistoliths, have been sampled for conodonts at 22 localities in the flyschoid Perry Creek and Kangaroo Hills Formations in the Camel Creek oroclinal area. All of the limestones are allochthonous. Some of the olistoliths are highly sheared and recrystallized; others have well-preserved but unstudied macrofaunas. Conodonts indicate source rocks as young as the uppermost Llandovery–lowest Wenlock *Pterspathodus amorphognathoides* Zone. A salient section through the Perry Creek Formation along Thatch Creek displays a sequence with Late Ordovician–Llandovery limestone clasts argued to document cannibalization of a "now-lost"carbonate platform (Sloan et al., 1995), which was possibly located upslope in the Broken River–Gray Creek area. Clasts from the Kangaroo Hills Formation indicate possible late Pridoli, Lochkovian–Pragian, and younger Devonian source rocks (Sloan et al., 1995).

HODGKINSON PROVINCE — The Chillagoe Formation, up to 1.2 km of shallow-water limestone with subordinate radiolarian and sponge-spicule chert, submarine mafic volcanics, and siliciclastics, has latest Llandovery to, at least, early Emsian conodonts. The conodonts have been pivotal in recognition of fault repetitions of the sequence. The Chillagoe Formation accumulated in a rift context (Bultitude et al., 1987, 1993, 1995). Overlying the Chillagoe Formation (Fig. 6, column 31), and apparently a lateral facies equivalent with at least part of it, is the Hodgkinson Formation (not shown on Fig. 6, column 31). Monotonous flyschoid arenite and mudstone with minor channel and slope-apron conglomerate, chert, basalt, and rare limestone comprise the Hodgkinson.. The limestones blocks, which were mostly if not entirely cannibalized from "now-lost" platform carbonates (Green, 1990), include Late Silurian, Devonian, and Early Carboniferous limestones. The Hodgkinson Formation crops out along strike for about 400 km and across strike for as much as 130 km.

CRATONIC AUSTRALIA

AMADEUS BASIN — The Mereenie Sandstone of the Amadeus Basin (Fig. 1; no stratigraphic column included herein) is a thick siliciclastic unit that is predominantly aeolian and fluviatile. It is relatively constant in thickness (500–1,000 m) over large areas (Lindsay and Korsch, 1991). A shallow-marine environment has been inferred on the basis of trace fossils and *Skolithos*-like tubes low in the formation (Talent et al., 1975). Other authors (e.g., Walley et al., 1990, 1991) consider the Mereenie Sandstone to be entirely aeolian, with the exception of fluvial material in its upper, possibly Devonian, part. Evidence for the Silurian is indirect. It is based on conodonts, trilobites, and vertebrates from underlying Ordovician units and Devonian vertebrates in overlying units. Fish remains high in the Mereenie Sandstone indicate that part of it may be Devonian (Talent et al., 1975). Available information has been summarized by Shergold et al. (1991). R. Nicoll (*in* Shergold et al., 1991) disaggregated Mereenie Sandstone samples, but he found only reworked Ordovician conodonts. Substantial periods of Silurian erosion are also inferred for this unit.

CARNARVON BASIN — The Silurian is known only from the subsurface in the southern Carnarvon Basin (Figs. 1, 7). Some authors (e.g., Hocking et al., 1987) suggest that substantial thicknesses of rock in the Carnarvon Basin may be Silurian. These include a thick, siliciclastic unit that crops out extensively and which was known as the Tumblagooda Sandstone (now Tumblagooda Group; Hocking, 1991; McNamara and Trewin, 1993; Trewin and McNamara, 1994; Mory and Iasky, 1996). The Tumblagooda is overlain by the mixed siliciclastics, carbonates, and evaporites of the Dirk Hartog Limestone (now Dirk Hartog Group, which consists of the Ajana Formation, Yeringa Evaporite, and Hamelin Formation; see Fig. 6, column 32) and the overlying Kopke Sandstone.

Gorter et al. (1994) discussed previous attempts to evaluate the age of the Tumblagooda Group; these used radiometric, paleomagnetic, and sparse trace fossil and macrofossil data. These attempts produced a spectrum of Early Ordovician to Middle Silurian ages. Gorter et al. (1994) reported rare elements of the conodont *Teridontus*, and suggested a Late Cambrian–Early Ordovician age.

The first report of Silurian conodonts from Australia was based on material from a bore hole on Dirk Hartog Island (Fig. 1) that penetrated the Dirk Hartog Group. Glenister and Glenister (1957) reported mostly stratigraphically long-ranging forms but, on the basis of several "*Paltodus*"species, proposed an Early or Middle Silurian age. Their faunal list included the Wenlock conodont "*Ctenognathus.*" Philip's (1969) discussion of Glenister and Glenister's (1957) collection did not list "*Ctenognathus.*" On the basis of elements he identified as *Ozarkodina ziegleri tenuiramae,* Philip (1969) assigned the fauna to the middle Ludlow *Ancoradella ploeckensis* Zone.

The Dirk Hartog Group has been subdivided into several units (Gorter et al., 1994; see Fig. 6, column 32), the lowest of which, the Ajana Formation, has been further subdivided into three informal units. The lower Ajana Formation consists principally of fine-grained siliciclastics, possibly of lacustrine origin, with palynofloras that indicate a generalized Ordovician or, possibly, Early Silurian age. The middle Ajana Formation has a larger carbonate component and a more diverse microflora with the Silurian form *Ambitisporites.* Massive carbonates of the upper Ajana Formation yield such conodonts as

Ozarkodina confluens, from which a generalized Wenlock–Ludlow age is inferred (Gorter et al., 1994).

The Ajana Formation is overlain by the Yaringa Evaporite, which has a microflora and fragmentary conodonts indicative of a generalized Wenlock or Ludlow age (Gorter et al., 1994). The Yaringa Evaporite is overlain by the Hamelin Formation. Resampling of carbonates in the latter formation has yielded a poor conodont fauna comparable to that documented by Philip (1969). On the basis of new and older conodont data, Gorter et al. (1994) concluded that the Hamelin Formation represents a shallow-marine environment of Ludlow age (*Kockelella crassa–Polygnathoides siluricus* Zones).

Higher intervals of the Hamelin Formation have enigmatic conodonts with forms referred to Amydrotaxis n. sp. A. A latest Ludlow–early Pridoli age (*Ozarkodina crispa–O. eosteinhornensis* Zones) is suggested (Gorter et al., 1994). The Pa elements of this taxon may be the same as those identified as *"Ctenognathus"* by Glenister and Glenister (1957). Units overlying the Dirk Hartog Group are not considered to be Silurian (Gorter et al., 1994).

CANNING BASIN — The Carribuddy Group (formerly Carribuddy Formation) is a sequence of siliciclastics, carbonate, and evaporites identified in the subsurface of the southern and central Canning Basin (Fig. 1) It consists of the Mallowa Salt and overlying Sahara Formation, and underlies siliciclastics and carbonates of the Worral Formation. These two units were thought to be separated by an Early Devonian hiatus of uncertain duration. For many years, some authors, primarily on stratigraphic grounds, considered the Carribuddy to be Early Devonian, with older parts of the succession possibly extending into the Upper Silurian (Playford et al., 1975; Lehmann, 1984). Nicoll (1984) summarized the conodont data of the Canning Basin.

The first useful biostratigraphic data from the Carribuddy Group was a palynoflora from the Mallowa Salt, an evaporitic unit that underlies the Sahara Formation. This palynoflora includes *Tetrahedraletes medinensis,* a Late Ordovician or Early Silurian indicator (Foster and Williams, 1991). Recovery of sparse early–middle Llandovery (Rhuddanian–Aeronian) conodonts from the lower carbonate unit of the Worral Formation (Nicoll et al., 1994) provided more precision. The fauna includes *Oulodus* sp., *Icriodella?* sp., and *Ozarkodina hassi,* and is very different from coeval faunas from the Tasman fold belt of eastern Australia (Simpson, 1995b). A major hiatus spanning most of the Silurian seems to be located somewhere in the Worral between its lowermost carbonate unit and its overlying divisions in the Canning Basin. This relationship has been demonstrated only for the Barbwire Terrace; more continuous sequences may occur elsewhere in the Canning Basin. Some authors, using sequence stratigraphic correlations into the Carnarvon Basin, suggest that this hiatus is time transgressive (e.g., Gorter et al., 1994, fig. 14). However, others, because of the lack of evidence for diachroneity, depict the hiatus as a non-diachronous boundary in the Lower Silurian (Warris, 1993, fig. 2; King, 1998, fig. 4).

BONAPARTE GULF BASIN — Salt assumed to be Silurian–Early Devonian intrudes Upper Devonian and younger rocks in the Petrel Sub-Basin and Lacrosse Terrace (Mory, 1990) of the Bonaparte Gulf Basin (Fig. 1; stratigraphic column not included herein). Evaporitic environments may have been coeval in the Silurian of the Canning and Bonaparte Gulf Basins (Fig. 1).

ARAFURA BASIN — The Arafura Basin (Fig. 1; stratigraphic column not included herein) has extensive Paleozoic rocks. The geographic limits of the basin are not clearly defined; it may extend beneath younger northern Australian cover sequences into Papua New Guinea and Irian Jaya. An interval of Silurian siliciclastic and volcanics was reported in the subsurface at Money Shoal I (Balke et al., 1973); this was included in the review by Walley et al. (1990, column 5). Based on a reinterpretation of palynological data, Bradshaw et al. (1990) concluded that a Devonian age is more probable for this interval, with the Silurian as a period of non-deposition and significant erosion in the basin. No column is therefore provided in this report.

NEW GUINEA

The tectonically complex New Guinea fold belt (Figs. 1, 7; stratigraphic column not included herein) has Paleozoic sequences, and is considered part of the Australian continental margin (Pieters et al., 1983). There is no Silurian in Papua New Guinea. However, there are three occurrences of Silurian rocks in Irian Jaya, which are all possibly connected in some way with the Arafura Basin. Recent mapping (e.g., Dow et al., 1986) shows many undifferentiated Paleozoic areas, which could include Silurian units.

KEPALA BURUNG (VOGELKOP) REGION OF IRIAN JAYA — Several thousand meters of low-grade Silurian metasediments (i.e., thin pelitic lithologies with subordinate coarser siliciclastics typical of distal turbidites) have long been known in the "Bird's Head," northwestern Irian Jaya (Visser and Hermes, 1962; Pieters et al., 1983; see Fig. 1). This sequence, the Kemum (formerly Kemoen) Formation, has rare intercalations of probable allodapic limestones. *Monograptus turriculatus* and *M. marri* from outcrops in the Roef River indicate that some of the sequence is upper Llandovery (*Monograptus turriculatus* Zone). Early Ordovician graptolites from similar litholo-

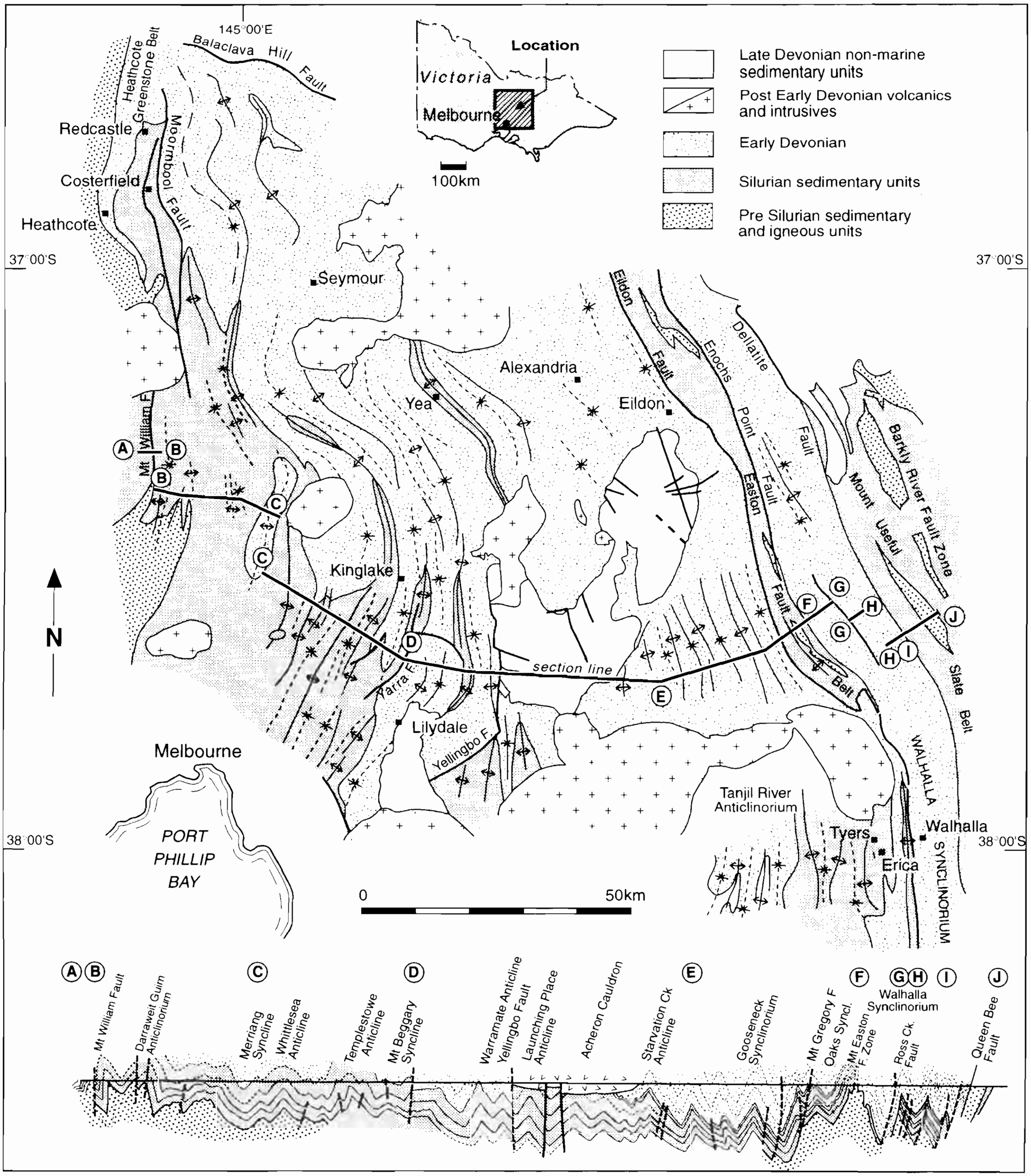

FIGURE 7 — Outcrop-tracts and structure of Upper Ordovician, Silurian, and Lower Devonian in central and east-central Victoria (after VandenBerg, 1988).

gies in the Heluk River area (Fortey and Cocks, 1986); Devonian ostracodes (Pieters et al., 1983); and a scale of *Thelodus trilobatus*, a middle Ludlow–middle Pridoli species (Turner et al., 1995), indicate that these deep-water rocks may represent much of the Paleozoic.

WAGHETE — The Modio Dolostone, a thick (ca. 2 km) sequence of dolomitic limestone on the Waghete 1:250,000 sheet (Fig. 7; stratigraphic column not included herein), was considered to be Carboniferous (Dow and Sukarnto, 1984). However, two samples from the westernmost outcrop of this formation near Modio in the Charles Louis Range yielded generically indeterminate conodonts. Associated specifically indeterminable elements of *Panderodus* and *Scolopodus* are similar to Silurian and Early Devonian(?) forms. *Panderodus* ranges from the early Middle Ordovician (Sweet, 1988) into the Middle Devonian (Barrick and Klapper, 1976; Barrick, 1977). Based on similarities of the *Panderodus* elements to the form-species *"P. simplex,"* a Llandovery–Ludlow form, a Silurian age has been suggested (Nicoll and Bladon, 1991). The species-level taxonomy of Silurian *Panderodus* is poorly understood. The morphology of *P.* sp. cf. *P. simplex* (sensu Nicoll and Bladon, 1991) appears closest to the slender graciliform elements of Sansom et al. (1994). The poorly preserved element of Scolopodus sp. (see Nicoll and Bladon, 1991, fig. 5.2) resembles coniform symmetry transition elements of the Distomodontidae, a family restricted to the Silurian. If this interpretation is accepted, it supports Nicoll and Bladon's (1991) tentative age assignment.

LORENTZ AND NOORD OOST RIVERS — Boulders with Middle Paleozoic corals have long been known from watercourses on the southern flank of the Central Highlands, West Irian Jaya (Teichert, 1928; Musper, 1938; Kruizinga, 1957; Visser and Hermes, 1962; Oliver et al., 1995), especially from the Noord Oost and Lorentz Rivers (Fig. 1). These boulders have predominantly Devonian faunas (Oliver et al., 1995), but others are of unequivocal to possible Silurian age, although there is no known in situ Silurian. The boulders are reported to have the Silurian tabulate coral *Halysites wallichi?* (Teichert, 1928). Abundant Silurian conodonts were obtained from a boulder collected in the Lorentz River early in this century. The fauna consists of *Coryssognathus dubius, Ozarkodina confluens*, and *O. crispa*, and indicates the upper Ludlow *Ozarkodina crispa* Zone (Van den Boogaart, 1990). Turner et al. (1995) obtained a thelodont scale of *Turinia* sp., acanthodian scales comparable to *Gomphoncus sandalensis* and *Nostolepis striata*, and a fragment of a possible acanthodian fin spine from this sample. There are seven reports of fossiliferous limestone boulders in the same general region along an east–west tract of ca. 90 km (see Oliver et al., 1995, fig. 1, for localities). Most of the boulders are Devonian, but those with *"Apyhyllum"* sp. and tabulate corals could be Silurian.

PALEOGEOGRAPHY

Earlier paleogeographic reconstructions for the Silurian of the Australia–New Guinea region have been based on the areal extent of sequences known or thought to be of a particular age, and have considered regional changes in lithofacies thickness. This has led to reconstructions that are consistent with tectono-sedimentary models (Talent et al., 1975; Pickett, 1982a; Walley et al., 1990). It is obvious that the actual shorelines may have lain well outside the preserved outcrop belts. Because tectonism may have transported some of the Australian–New Guinea Silurian for hundreds of kilometers, we provide reconstructions only for the late Llandovery (Telychian) and Pridoli (Fig. 8), and have not attempted palinspastic reconstructions. A. M. Walley (*in* Walley et al., 1990) has reviewed the Silurian tectono-sedimentary history of Australia. Even if the stratigraphic correlations presented herein (Figs. 4–6) are taken into account, Walley's review remains a very useful synthesis.

No shoreline is hypothesized for the early Llandovery (Rhuddanian) because of the mere half-dozen sequences with a marine record for this interval. Only Darraweit Guim (Fig. 2) is of particular interest; its graptolites show that the Deep Creek Siltstone spans the Ordovician–Silurian boundary (Fig. 4, column 4). Limestone clasts from debris flows in the Broken River–Camel Creek Province, northern Queensland, have Ashgillian to Llandovery–lowest Wenlock conodonts. These conodonts demonstrate cannibalization of a "now-lost" shelf area to the west and/or southwest; this is the Georgetown carbonate platform of Talent et al. (2000).

Evaporitic conditions extended over large areas of the Carnarvon and Canning Basins (Fig. 6, columns 32 and 33) during the early Llandovery, and extended at least 650 km southeast into the Canning Basin. Available data suggest that evaporitic conditions may have ceased in the Canning Basin before the late Llandovery. There is no evidence for Llandovery sedimentation in the Amadeus Basin of central Australia, although it may have had aeolian conditions during that time.

For 23 (perhaps 25) of the 33 stratigraphic columns presented herein (Figs. 4–6), the early and most of the middle Llandovery correspond to a hiatus. Indeed, in most cases, an angular unconformity emphasizes a profound contrast in tectonic style with underlying stratigraphic units. This unconformity is the primary basis for proposal of a major orogenic cycle, the Benambran orogeny, which was possibly the most profound Phanero-

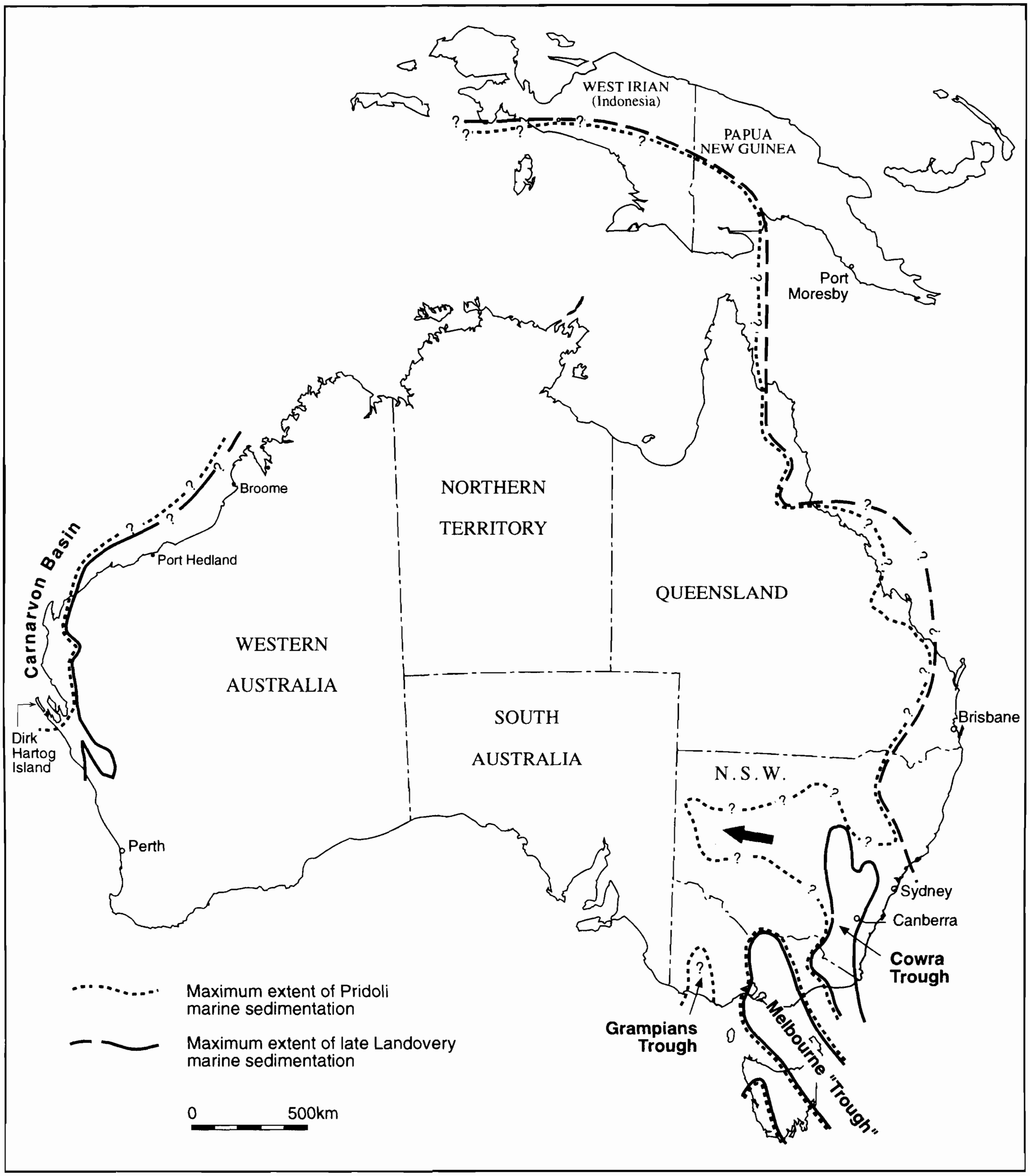

FIGURE 8 — Extent of marine transgressions during the late Llandovery (Telychian) and Pridoli in Australia and New Guinea. Paleogeographic map does not take into account long-distance tectonic movements.

zoic orogeny in Australia. Data from the Indi and upper Buchan River regions of eastern Victoria (Simpson and Talent, 1995) are consistent with development of the main events of the Benambran cycle in the Ashgillian (late?) to early–middle Llandovery, rather than in a generalized Llandovery–Wenlock interval, as earlier assumed. However, questions of west–east and meridional diachronism need further clarification (see Tectonic Setting, above).

With notable exceptions in the Yass and much-faulted Canberra sequences, there is a general dearth of adequate correlations within the voluminous Wenlock, Ludlow, and Lochkovian volcanics of the Tumut–Cooma–Canberra–Yass–Goulburn areas (Fig. 5, columns 12–17). How these might relate to the hypothesized Quidongan orogeny, which is based on a possible local unconformity at Quidong, is problematic. We have, therefore, been reluctant to propose shorelines for the Wenlock or Ludlow. Evaporitic conditions seem to have continued throughout this interval in the Carnarvon Basin, Western Australia (Fig. 6, column 32). However, noteworthy late Ludlow–Lochkovian marine transgression involved the Darling Basin and, perhaps, briefly covered the Bancannia trough of western New South Wales. The latter was a region of rapid siliciclastic sedimentation (Fig. 8) that persisted well into the Devonian.

The onset of the Bowning orogenic cycle bracketed biostratigraphically by conodonts from pre-orogenic units in three widely spaced regions of eastern Australia (Fig. 4, column 6, Fig. 5, column 13, Fig. 6, column 30). This orogeny commenced in the early Lochkovian, but began after the earliest Lochkovian *Icriodus woschmidti* Chron in eastern Victoria (Tongaro Siltstone), the Yass Synclinorium (Elmside Formation), and northeastern Queensland (Jack Formation).

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SILURIAN OF THE INDIAN SUBCONTINENT AND ADJACENT REGIONS

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ABSTRACT — An overview of the Silurian of the Indian subcontinent and adjacent regions, illustrated by ten stratigraphic columns, underscores uncertainties in precise correlation. Silurian sequences with some biostratigraphic control are documented from the Dasht-e-Nawar region, Afghanistan; the Peshawar Basin, Pakistan; central Nepal; southernmost Tibet; and Shan State, Myanmar (Burma). Near-shore, primarily siliciclastic sequences with reef developments in the western Himalaya, that range from Late Ordovician to Middle and, conceivably, younger Silurian, are not tightly constrained biostratigraphically. Similar sequences in the central Himalaya, often referred to as the Muth Quartzite and regarded as broadly Devonian, are largely Silurian, but need renewed study. The Paleozoic of Bhutan, assumed by some to include Silurian rocks, appears to have a Middle Cambrian–Late Devonian hiatus. There is an urgent need for a multi-pronged re-investigation of these Silurian and associated Late Ordovician and Devonian sequences, particularly if transgression–regression patterns and changes in paleogeography are to be documented adequately.

INTRODUCTION

Silurian fossil-based correlation of sequences on the Indian subcontinent and adjacent regions (Figs. 1, 8) is best established in the Dasht-e-Nawar tract, southeast Afghanistan; the Nowshera area, northwestern Pakistan; the northwest flank of Annapurna/Nilgiri, central Nepal; the Nyalam area, southernmost Tibet; and Shan State, Myanmar [Burma] (Fig. 8). The first two sequences are correlated by conodonts, the last three by graptolites.

Unfortunately, the Silurian database for much of the western and central Himalaya from Kashmir to Nepal has been corrupted by false paleontologic reports. These form part of a general pattern of mis-information that spans the Phanerozoic in this region. This problem has been documented (Talent et al., 1988, 1991; Talent, 1989, 1990a, 1990b, 1995; Ahluwalia et al., 1989; Brock et al., 1991; Webster, 1991; Webster et al., 1993; Shanker et al., 1993). Excision of this spurious material has left a modest paleontologic database. This database was reassessed by Talent et al. (1988, p. 31, 32) on the basis of reports largely from the early 1900s and earlier, although inclusive of a few more recent studies (Mehrotra et al., 1982; Khanna and Sah, 1983; Goel et al., 1987). Recent 1:50,000 mapping in Spiti has been coupled with closer study of sedimentary facies, cyclicity, and breaks in sedimentation. With this as a base, a re-evaluation has been made of earlier observations on the Kashmir and Kumaon sequences (Bhargava and Bassi, 1996, In press).

AFGHANISTAN

Information on the Silurian of Afghanistan has advanced little since investigations in the 1960s, which focused on the Dasht-e-Nawar region of east-central Afghanistan (Fig. 1, locality 1), about 100 km southwest of Kabul (Fig. 2, column 1). The Logar Formation in this area has Ordovician–Silurian fossils, with conodonts from one horizon that include *Kockelella variabilis*, *K.* sp. cf. *K. variabilis*, *Ozarkodina confluens*, *O. excavata excavata*, and *O.* sp. cf. *O. sagitta* (Fesefeldt, 1964; Dürkoop, 1970; Weippert et al., 1970; Schreiber et al., 1972; Wolfart and Wittekind, 1980). These show a Ludlow, but not latest Ludlow, age. A convenient summary of fossils known from central and southern Afghanistan is in Weippert et al. (1970).

PAKISTAN

There are several units in northwest Pakistan which, despite a lack of fossils, have been assigned to the Silurian. One of these units that crops out in the mountains west of Peshawar is the Landi Kotal Formation (Stauffer,

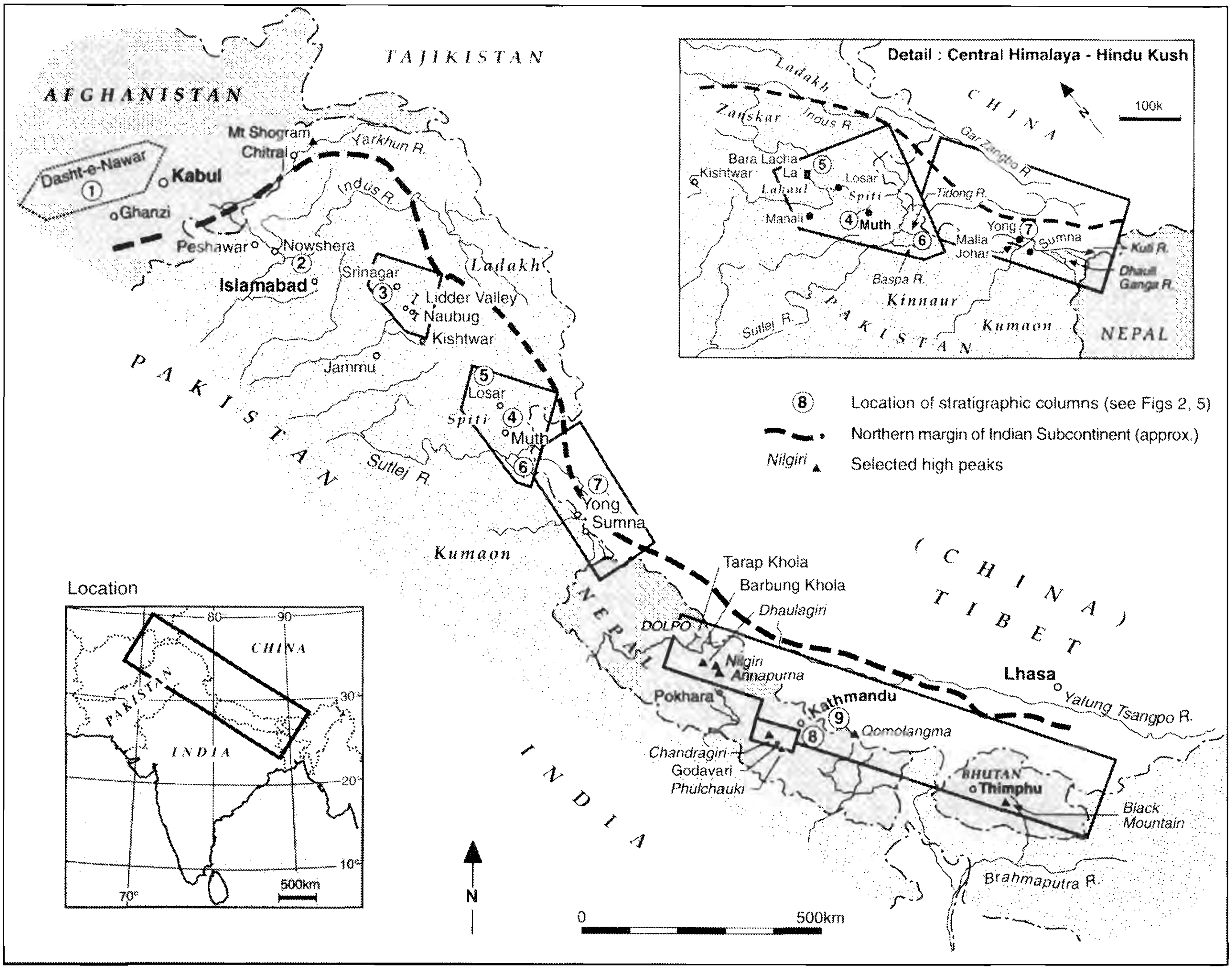

FIGURE 1 — Himalaya–Hindu Kush region showing location of stratigraphic columns (Fig. 2, 5) and principal Silurian localities.

1968a; Shah, 1977; Shah et al., 1980; ?=Lala China Slatey Shales of Tahirkheli et al., 1975), an assemblage of slate and phyllite with subordinate quartzite, lenticular limestone, and dolostone. The other is the Shagai Limestone (Stauffer, 1968a; Jan and Kempe, 1970; Shah, 1977; Shah et al., 1980; Molloy et al., 1997). Their age is problematic, and could be Proterozoic. Possible easterly correlatives of these units in the Attock-Cherat Range are suggested to be Silurian (e.g., the Hissartang Formation; Haneef et al., 1989; compare Hussain et al., 1990). Silurian intervals have been thought to occur in the Ordovician–?Silurian of the Karakorum Block in northernmost Pakistan (Gaetani, 1997), but these have yet to be unequivocally dated.

PESHAWAR PLAIN — The Misri Banda Quartzite (?=Chamla and Swabi Quartzites of Martin et al., 1962), consists of ca. 600 m of calcareous and dolomitic quartzite (Fig. 2, column 2). This sequence was thought to be Devonian (Stauffer, 1968b) until discovery of the ichnofossil *Cruziana rugosa* from a locality east of Misri Banda village that is close to the type section (Pogue and Hussain, 1986; Pogue et al., 1992). The trace fossil was thought to indicate the Lower–Middle Ordovician, but according to B. D. Webby (personal commun., 1996), *C. rugosa* does not supply any more compelling evidence for an Ordovician than for a Silurian age. If the Misri Banda Quartzite is older than the Kandar Formation, then it could be the source of the Llandovery–early Wenlock conodonts reported (discussed below) in the lower Kandar Formation. The unit could, moreover, correlate with the Caradocian–Llandovery–?Wenlock transgression–regression interval. This interval elsewhere in the northern Indian Subcontinent has an upper arenaceous interval that was earlier assigned to the Muth Quartzite of Spiti, Kashmir, and elsewhere in the Indian High Himalayas. This interval is now referred to the Takche Formation (discussed below; Fig. 2, columns 4, 5).

The Kandar Formation (Kandar Phyllite of Stauffer, 1968b, and Ali and Anwar, 1969; =Swabi Shale of Martin et al., 1962, and Panjpir Formation of Hussain et al., 1990, and Pogue et al., 1992) crops out around the southeast and east margin of the Peshawar basin (Fig. 1, locality 2). The Kandar Formation consists of more than 600 m of argillite, phyllite, and meta-siltstone with a basal interval of calcareous quartzite and discontinuous conglomerate. Thin limestones interbeds are common. Conodonts have been recovered from eight localities, including a section that exposes the basal calcareous quartzites near Misri Banda, about 12 km east of Nowshera. Ludlow–Pridoli conodonts from at least the *Polygnathoides siluricus–Ozarkodina remscheidensis eosteinhornensis* Zones are known (A. G. Harris *in* Pogue et al., 1992). Biostratigraphically important forms identified from this locality are *Kockelella variabilis, P. siluricus, Ozarkodina crispa*, and *O. remscheidensis eosteinhornensis.* Llandovery–Wenlock faunas have also been reported from low in the Kandar Formation in the Chingalal synclinorium about 50 km northeast of Nowshera, and northeast of Misri Banda (Pogue et al., 1992). More data are needed to confirm that these conodonts are not reworked from older horizons. Thin crinoidal limestones in what may be the youngest horizons of the Kandar Formation crop out on the Nowshera–Risalpur Road 3 km north of Nowshera. These limestones have the Pridoli conodonts (Barnett et al., 1966; Molloy, personal commun., 1980) *O. remscheidensis eosteinhornensis, O. excavata excavata*, and *Panderodus unicostatus.* The Kandar Formation is overlain, apparently gradationally, by the Nowshera Limestone. The latter is interpreted as a reef complex (Teichert and Stauffer, 1965; Stauffer, 1968b; Ali and Anwar, 1969), with conodonts and brachiopods consistent with the Lower Devonian (Lochkovian) (Barnett et al., 1966; Molloy personal commun., 1980). A sequence at Gundhai Sar west of Peshawar, once thought to be Upper Silurian–Lower Devonian (Khan, 1969), is now reported (Pogue et al., 1992) to have upper Famennian conodonts.

CHITRAL — Slates with Silurian graptolites (*Monograptus priodon;* H. Jaeger, personal commun., 1976) and undescribed trilobites occur on the crest of Mt. Shogram, 50 km northwest of Chitral Township (Fig. 1). This crustal block is separated from the Indian Subcontinent sensu stricto by the main Karakoram thrust fault (Searle, 1991; Tahirkheli, 1996; Searle and Khan, 1996) and its southwest extension, the Drosh–Yasin ophiolite belt. This terrane could have been detached from the northern margin of the Gondwana supercontinent, with a long history of motion independent from the Indian subcontinent, before becoming sutured to the Asia. Accretion of this terrane to Asia may have taken place well before the suturing of the Indian subcontinent with the Asia (compare Klootwijk et al., 1994).

INDIA

There has been little modern biostratigraphic work on the Ordovician–Silurian of India, apart from spot-sampling during reconnaissance structural and stratigraphic investigations. The fossils, as presently known, are generally sheared and otherwise poorly preserved.

KASHMIR — Pivotal in the literature on the Paleozoic stratigraphy of the Kashmir synclinorium (Fig. 1, locality 3; Fig. 3) are the classic works of Lydekker (1876, 1883) and Middlemiss (1910), with additional data in Shah (1978) and, for the Lidder Valley, in Srikantia and Bhargava (1983). Middlemiss (1910) appears to have been the first to collect extensively from the area east of Anantnag (=Islamabad). The faunas are predominantly brachiopods that occur in ca. 30 m of rusty-weathering, patchily calcareous, sandy shale (Middlemiss, 1910). These shales underlie what was formerly referred to as the Muth Quartzite (discussed below). This fossiliferous unit is now placed within the Rishkobal Formation of Srikantia and Bhargava (1983). Minor coral-algal buildups occur in this sequence (Fig. 2, column 3; Bhargava, 1997). The faunas were illustrated in a monograph by Reed (1912b), and considered Early Silurian (Llandovery). Boucot and Gauri (1968), on the basis of collections from several localities in the same interval near Gugaldar and Gudramer east of Anantnag, concluded that the faunas were likely Ashgillian–Llandovery. Although these faunas are not well preserved or useful in high-precision biostratigraphy, they need modern documentation. Thin-section study might establish the generic and specific identifications needed to establish the biogeographic affinities of these brachiopods and corals.

The Rishkobal Formation is unconformably overlain by the Muth Quartzite sensu stricto, and the latter, in turn, conformably overlain by the "*Syringothyris* Limestone" (Aishmuqam Formation of Srikantia and Bhargava, 1983). However, an unconformity is reported from the base of the "*Syringothyris* Limestone" (Wadia, 1934, p. 143). The stratigraphic nomenclature of Fig. 2, column 3, is that of Srikantia and Bhargava (1983) for the Lidder valley (Fig. 4).

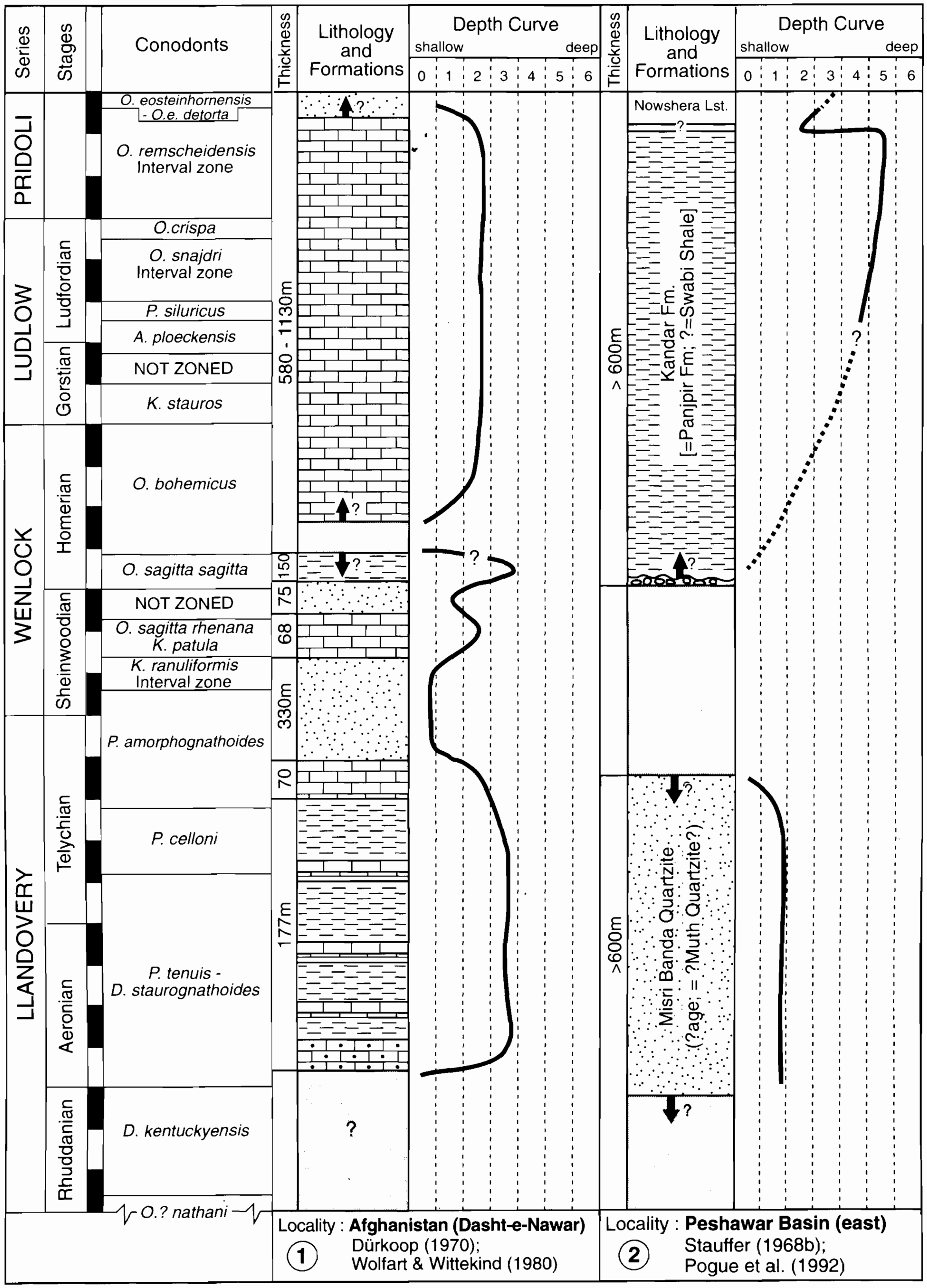

FIGURE 2 — Stratigraphic columns for Afghanistan, Pakistan, and India.

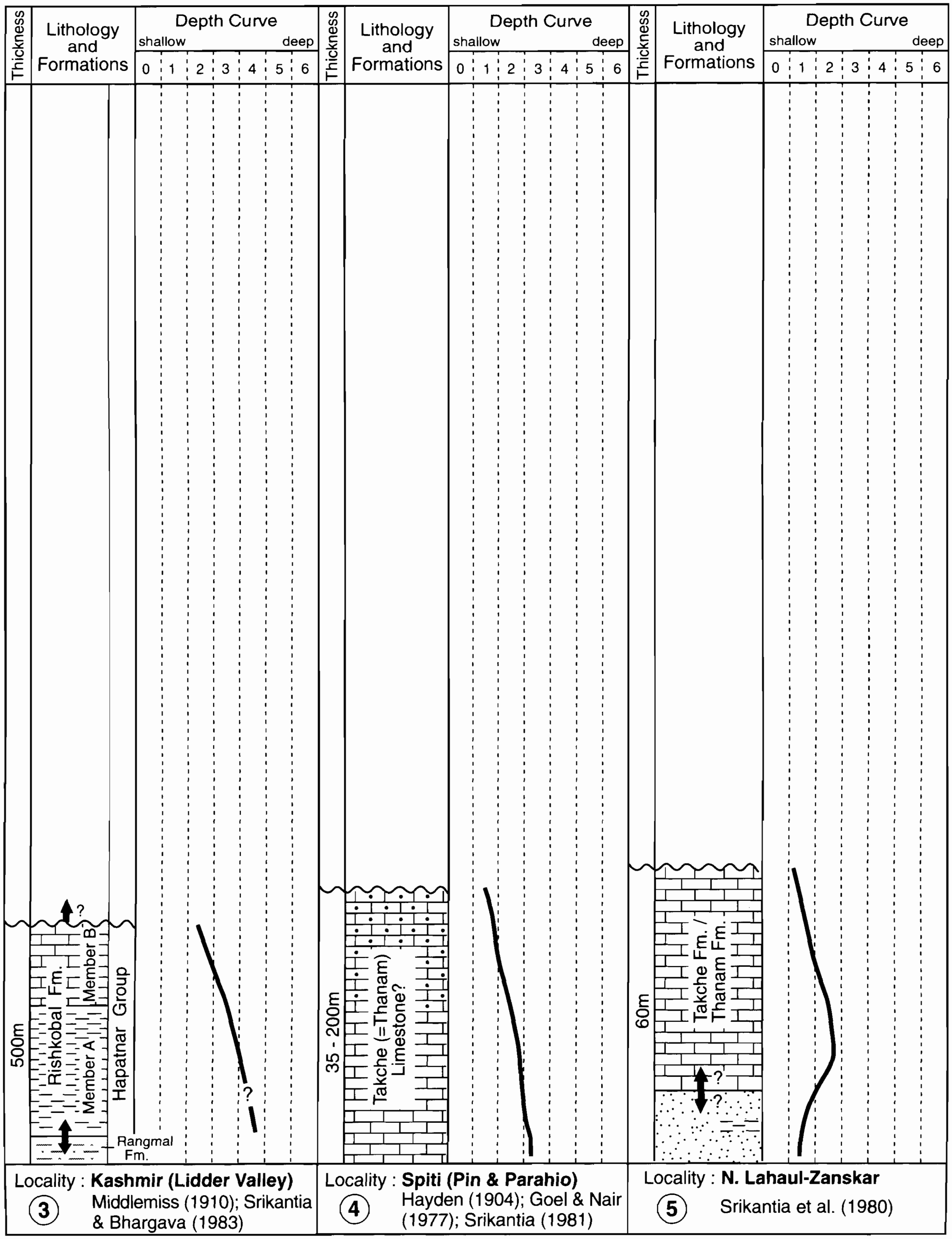

FIGURE 2 *continued.*

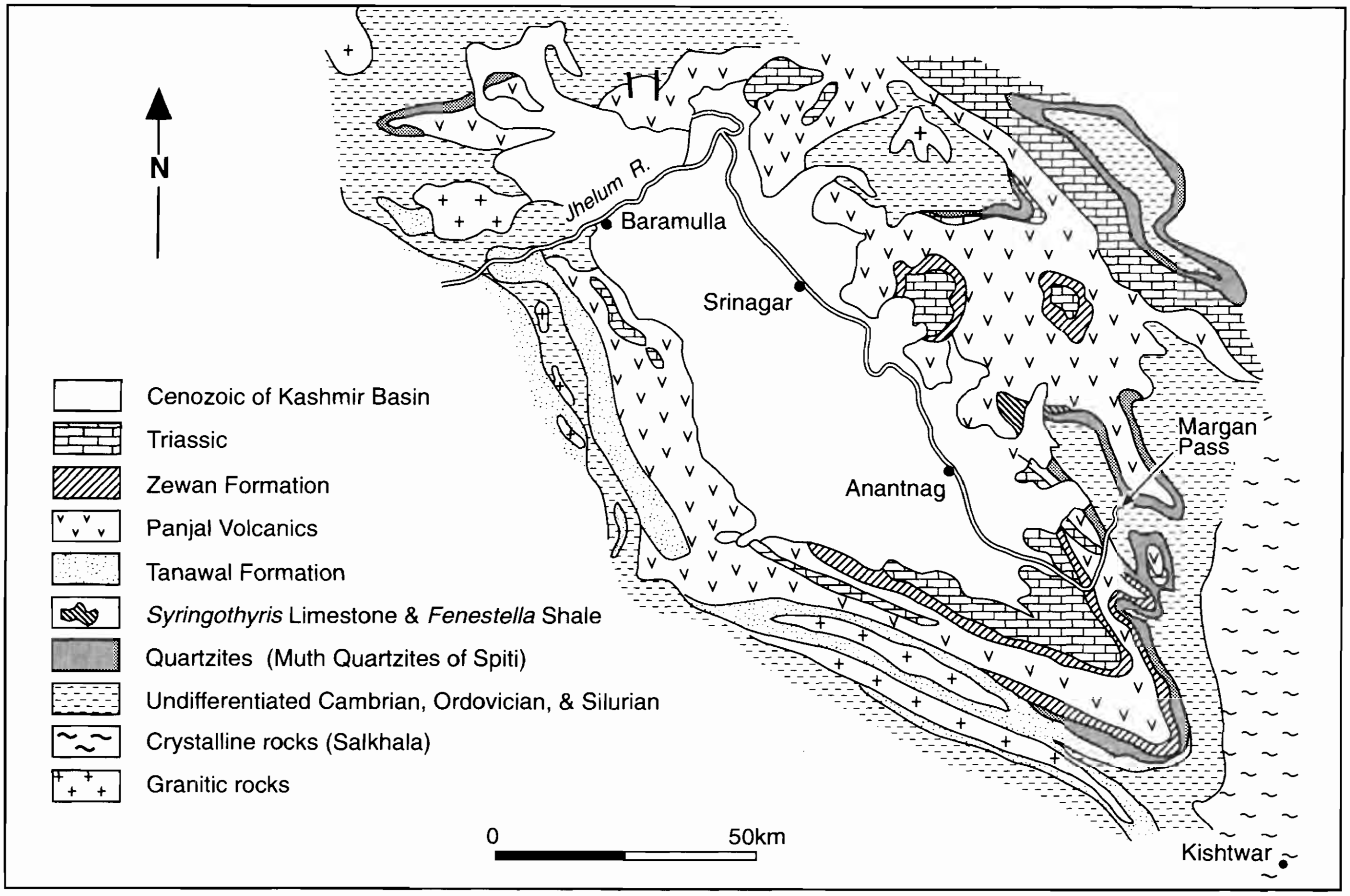

FIGURE 3 — The Kashmir synclinorium delineated by its Paleozoic–Triassic stratigraphy (after Shah, 1978).

SPITI, LAHAUL, AND ZANSKAR — The successions of the Spiti, Lahaul, and Zanskar regions (Fig. 1, localities 4, 5; Fig. 2, columns 4, 5) were previously considered separately, though the classic Spiti Paleozoic–Mesozoic sequence was known to be identifiable, at least in a general way, in Lahaul and Zanskar. The three regions were long believed to lie in single synclinorium. Regional 1:50,000 mapping has demonstrated that the stratigraphic nomenclature developed for the Spiti valley sequences can be applied in all three regions (Bhargava and Bassi, In press).

The "Muth Series" was introduced by Stoliczka (1866a, 1866b), but without evidence of the age of this often boldly outcropping unit. The name was later modified to "Muth Quartzite." Hayden (1904, p. 27) provided more information and stressed the gradational relationship between underlying "hard, light grey, siliceous limestones that gradually become less and less calcareous and pass through calcareous quartzite into the reddish and brown quartzites which form the lowest beds of the Muth Quartzite." This underlying calcareous sequence was named the "Pin Dolomite" by Goel and Nair (1977).

The Takche Formation, proposed by Srikantia (1981) with a type section in Spiti, includes dolostone, shale, siltstone, and sandstone. It equates generally with the interval for which Goel and Nair (1977) had used the terms "Thanam Limestone," "Pin Dolomite," and "Pin Limestone." Relationships in the Spiti Valley reveal that the well-washed arenites of the Muth Quartzite sensu stricto are distinct from the underlying matrix-rich Takche Formation. A distinct unconformity has been identified between the two units (Bhargava and Bassi, 1996), for at least part of the region, but its temporal significance is problematic.

The Takche Formation is well exposed in the original type section for the Muth Quartzite sensu lato. This section is less than 1 km south-southwest of Muth in the Pin Valley of Spiti (Hayden, 1904; Goel and Nair, 1977; Jain et

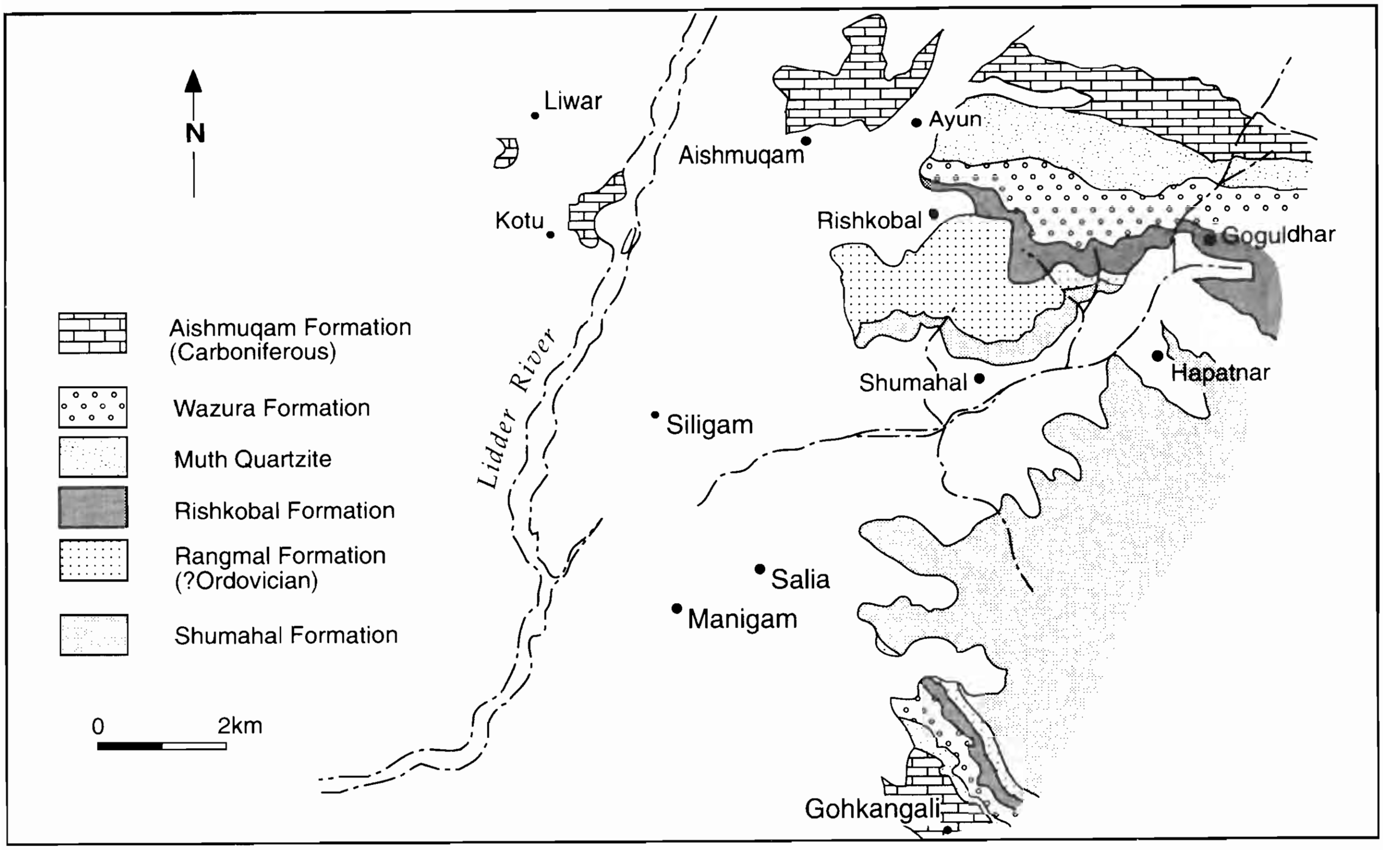

FIGURE 4 — Geology of the Lidder Valley, Kashmir (after Srikantia and Bhargava, 1983). Note difference in stratigraphic nomenclature for this part of the Kashmir synclinorium with earlier nomenclature (compare Fig. 3).

al., 1980; Srikantia, 1981; Talent, 1982; Fuchs, 1982; Ranga Rao et al., 1983; Bhargava et al., 1991) and in the Parahio Valley. The Takche Formation in this section records the regressive phase of an Ordovician–Early Silurian transgressive–regressive cycle.

The Takche Formation is best developed in the southeastern part of the Spiti synclinorium in the Pin and Parahio Valleys. It becomes more sandy towards the northwest in Lahaul and Zanskar, where its thickness is highly variable (Srikantia et al., 1980). This thickness variation is related to varying accumulation rates of Muth, as well as varying amounts of erosion of sub-Muth stratigraphic units. Because of this, various workers (e.g., Kanwar and Ahluwalia, 1978) failed to distinguish the Takche Formation, and grouped it either with the Thango arenites or with the Muth Quartzite. Several prograding cycles can be discriminated in the sandstone, shale, and carbonate sequence of the Takche Formation. Environments represented in the sequence range from about the middle of the lower shore-face to upper shore-face and were affected by low-intensity storms. The carbonate component is prominent in areas east of the Ratang River. Small carbonate build-ups in these carbonate beds (Bhargava and Bassi, 1986) have *Plasmoporella*, *Favosites*, *Halysites*, *Thamnopora*, *Heliolites*, *Tentaculites*, and brachiopods.

Collections made by H. H. Hayden, C. L. Griesbach, and earlier workers were illustrated in a monograph by Reed (1912a) and evaluated by Talent et al. (1988, p. 31, 32), but need restudy, preferably with larger collections. Mehrotra et al. (1982) reported poorly preserved chitinozoans, and Khanna and Sah (1983) recovered acritarchs from horizons now assigned to the Takche Formation. Hayden (1904, p. 27) noted numerous casts of what he took to be *Pentamerus oblongus* Sowerby in siliceous limestones of the Pin Dolostone at Gaichund in the Parahio Valley. These fossils in rocks now referred to the Takche Formation are similar to the large pentamerids in the "Muth Quartzite" east of Sumna in the Kumaon Himalaya (discussed below). Goel et al. (1987) reported

Caradocian algae low in the Takche Formation close to Muth. The biostratigraphic significance of the poorly preserved cephalopod and indeterminate coral reported by Ameta and Gaur (1980) from a loose block of quartzite, said to have been from the "Muth Quartzite" near Muth, is uncertain.

The Muth Quartzite sensu strictu grades into the Lipak Formation. An Early Carboniferous (Tournaisian) age was assigned to the Lipak Formation (Vannay, 1993), but Middle Devonian (Givetian) to middle Famennian conodonts occur in it near Muth and Mikim in the Pin Valley (Draganits et al., in press). The Lipak Formation has clean sandstones in its lower part which are identical to those of the Muth Quartzite sensu stricto. Body fossils are unknown in the Muth Formation in the Spiti area; earlier reported body fossils are from horizons that equate with the Takche Formation. Xiphosuran trackways have been reported from the Muth Quartzite sensu stricto by Bhargava and Bassi (1988). Superbly preserved giant arthropod trackways (eurypterid and myriapod) up to 42 cm across and desiccation features are reported from it in the Pin Valley east of Mikkim (Draganits et al., 1998).

KINNAUR–KUMAON (HIMACHAL PRADESH) — The High Himalaya in the Kinnaur–Tibet border region on the left flank of the Sutlej River (Fig. 1, locality 6; Fig. 6) was assumed to be composed of igneous and metamorphosed Precambrian rocks (Gansser, 1964). However, a Paleozoic and Mesozoic succession was discovered at the western nose of the Kumaon synclinorium in the upper Tidong and Gyamthing Valleys (Bassi and Chopra, 1978, 1982; Bassi et al., 1983; Bassi, 1989). New stratigraphic units were recognized (Bassi et al., 1983), but it was realized that the nomenclature applied in Spiti could be used satisfactorily in eastern Kinnaur (Fig. 5, column 6; Bhargava et al., 1984).

An orthoid brachiopod has been reported from the Muth Quartzite in this area (Bassi, 1988), which indicates that at least part of this succession is marine. The unconformably underlying Takche Formation (=Manchap Formation of Bassi et al., 1983) is predominantly calcareous, with coral-algal build-ups and several identifiable prograding cycles. Diverse fossils are associated with the carbonate build-ups (i.e., tabulate and rugose corals, stromatoporoids, and calcareous algae; Bhargava and Bassi, 1986, 1987). The age of these faunas is assumed to be the same as in the Spiti area, namely Llandovery and perhaps Wenlock. The sedimentary environments range from subtidal to intertidal with well-developed back reef areas with build-ups, local upper slope facies, and protected low-energy environments (Bhargava and Bassi, in press).

Griesbach (1891) laid the foundations for the stratigraphic synthesis of the central High Himalaya in the Kumaon region (Fig. 1, locality 7). His maps and sections are useful more than a century after they were published. Ordovician–Llandovery faunas collected by C. L. Griesbach and earlier workers were described by Salter and Blanford (1865) and Reed (1912a). These faunas need restudy, preferably with more extensive collections. The Ordovician–Silurian of Kumaon (Fig. 5, column 7) consists of the Shiala and Variegated Formations (Kumar et al., 1977), with the Yong Limestone (Khanna et al., 1983) laterally equivalent to the Shiala Formation. Members A and B of the Muth Formation (indicated on the right side of Fig. 5, column 7 in upper-case letters), as described by Kumar et al. (1977), are believed to be Silurian. Various areas of the Kumaon Himalaya have been the focus of studies since C. L. Griesbach's work (Heim and Gansser, 1939; Shah and Sinha, 1975; Kumar et al., 1977; Khanna et al., 1983; Goel et al., 1987; Sinha, 1989; Sinha et al., 1996). However, two publications on the geology of the headwaters of the Kali River in northeastern Kumaon contain spurious paleontologic reports (i.e., Valdiya et al., 1971; Valdiya and Gupta, 1972).

It was believed until recently that the Ordovician–Silurian boundary approximated the contact of the Shiala Formation and overlying Yong Limestone. However, Sinha et al. (1996) suggested that acritarchs show the that boundary is in the Shiala Formation. Collections made by R. K. Goel and colleagues in 1978 and 1985 from the Muth Quartzite at the 4,325 km marker on the Sumna–Rewalibagar (Laptal) track 4 km east of Sumna (Goel et al., 1987) have numerous large specimens of the Silurian pentameridine *Pentamerifera* sp. This horizon is overlain by about 100 m of coarse-grained quartz arenite. Although we have not re-examined these sections, we suggest that the higher sandstone is Muth C of Kumar et al. (1977), is likely to be Upper Devonian, and rests unconformably or disconformably on the Silurian. The Yong Limestone apparently represents an algal-coral build-up in an intertidal setting.

NEPAL

CENTRAL NEPAL (DOLPO, DHAULAGIRI, ANNAPURNA REGIONS) — A more or less complete Lower–Middle Paleozoic sequence is assumed to be present in the Dolpo–Dhaulagiri region of western Nepal (Fuchs, 1967, 1975a, 1975b, 1977; Fuchs et al., 1988; Fig. 1, locality 9). However, there are no compelling paleontologic data for horizons older than the Givetian. Essentially the same sequence is found farther east on the left flank of the Kali Gandaki River (Bordet et al., 1967, 1971; Hagen, 1968). The latter sequence provides unequivocal evidence for the Silurian in the "Dark Band Formation." The "Dark Band" on the northern flank of the Nilgiri massif has upper middle and lower upper Llandovery graptolites (Bodenhausen et al., 1964; Strachan et al., 1964).

A section of the "Dark Band Formation" from inferred Silurian into unequivocal Lower Devonian (Pragian) has been reported by Colchen (1971, 1975, *in* Bordet et al., 1975; Colchen et al., 1987) from the upper Shokang Nala on the eastern flank of the Kali Gandaki. A generally similar Ordovician–Silurian succession continues east through the High Himalaya from Annapurna to Manaslu and Ganesh Himal, where it lacks paleontologic control because of metamorphism (Colchen, 1975; Colchen et al., 1987).

CHANDRAGIRI–PHULCHAUKI SYNCLINORIUM — The Ordovician–Silurian is well displayed in the Chandragiri–Phulchauki synclinorium (Fig. 1, locality 8) on the south flank of the Kathmandu Basin (Medlicott, 1875; Auden, 1935; Bordet et al., 1959, 1960; Arita et al., 1973; Stöcklin et al., 1977; Stöcklin, 1980; Talent et al., 1988, 1991). Corals and stromatoporoids from the limestones of the Chandragiri Range, Chobar, and Godavari are too recrystallized and sheared for specific identification. Excellent outcrops along the Godavari–Phulchauki road (Fig. 7) display a very gradual upward change in lithology. The succession includes the Chandragiri Limestone at the foot of the Phulchauki Range, overlying calcareous and ferruginous siltstone (Chitlang Formation), and then the massive Phulchauki Limestone that forms the highest part of the Phulchauki Range (Fig. 5, column 8). Llandovery, possibly early Llandovery, conodonts from the Chitlang Formation at locality 15 of Talent et al. (1988) on the Godavari–Phulchauki road is about 25 m below their localities 14, 17, and 22. Locality 22 has a Llandovery–Wenlock trilobite fauna described by N. Pillet (*in* Bordet et al., 1960). The Phulchauki Limestone consists of perhaps 200 m of generally massive, often dolomitic, crinoidal limestone. It is assumed to be Silurian, but no biostratigraphically compelling fossils have been found so far.

SOUTHERN XIZANG (TIBET) REGION

Silurian litho- and biostratigraphy of the Xizang region has been reviewed (Anonymous, 1977; Lin and Qiu, 1983; Yang, 1985; Lin, 1989). In this report, we are concerned only with the region south of the Yalung Tsangpo ophiolite belt at the northern margin of the Indian Plate (Fig. 5, column 9).

The Central Himalayan Silurian stratigraphy extends from central Nepal into southernmost Xizang, where a generally more offshore depositional setting is preserved in several areas (Wang, 1987; Burchfield et al., 1992). The Shiqipo Formation near Jiacun village in Nyalam County, close to Nepal (Mu et al., 1973), consists of 90 m of black graptolitic shale and limestone with a basal sandstone. It has graptolites indicative of a generalized Llandovery age. The overlying Pulu Group, reported to be fault-truncated, consists of 46 m of arenaceous limestone and shale with graptolite and nautiloid faunas (Mu and Ni, 1975, 1985; Yin et al., 1983). Corals (Deng, 1982) and conodonts (Qiu, 1985, 1987) have also been reported from this unit. It has a lower Wenlock fauna with the graptolites *Monoclimacis vomerinus* cf. *subgracilis*, *Pristiograptus dubius*, and *P. dubius latus*, and the nautiloid *Michelinoceras jucundum*.

BHUTAN

The Tangchu Group of central Bhutan (Tangri and Pande, 1995; =Tangchu Series of Gansser, 1964; Tongchu Series of Nautiyal et al., 1964; and Tangchu Formation of Singh, 1973; Jangpangi, 1978; Gansser, 1983; Chaturvedi et al., 1983) was thought to be Silurian or Devonian (H. Flügel *in* Gansser, 1964). The correlation was based on specifically indeterminable stromatoporoids and corals. Subsequently, badly sheared, generically indeterminable brachiopods were discovered, and a Middle to Late Devonian (Givetian–Frasnian) age was suggested (Termier and Gansser, 1974). This fauna has been re-evaluated (Talent et al., 1988, p. 36), and determined to be composed of largely unidentifiable specimens. The exceptions are a fenestrate bryozoan (*Fenestella?* sp.) and two spiriferidines, one an indeterminate reticulariacean and the other of uncertain familial affinity. The fauna is Paleozoic, probably Late Paleozoic, and almost certainly post-Silurian. This accords broadly with the Carboniferous–Permian age suggested earlier by some workers (Singh, 1973; Jangpangi, 1978). However, the re-evaluation does not preclude a Devonian age for fossils collected by Gansser (1983) and identified by H. Flügel (*in* Gansser, 1983) as Late Devonian. The general consensus is that the age of the Wachila Formation, the lowest unit of the Tangchu Group, is likely Late Devonian–Early Carboniferous (Tangri and Pande, 1995).

The Tangchu Group rests unconformably on the "Quartzite Formation." On the basis of poorly preserved brachiopods, trilobites, and bryozoans from members G and H of the Maneting Formation of Chaturvedi et al. (1983), the "Quartzite Formation" was thought to be Ordovician. The trilobites are now regarded as Middle–late Late Cambrian kaolishaniids and/or pagodiids (Tangri and Pande, 1995). There is a hiatus in Bhutan, or at least no identified record, that corresponds to the Ordovician, Silurian, and most of the Devonian.

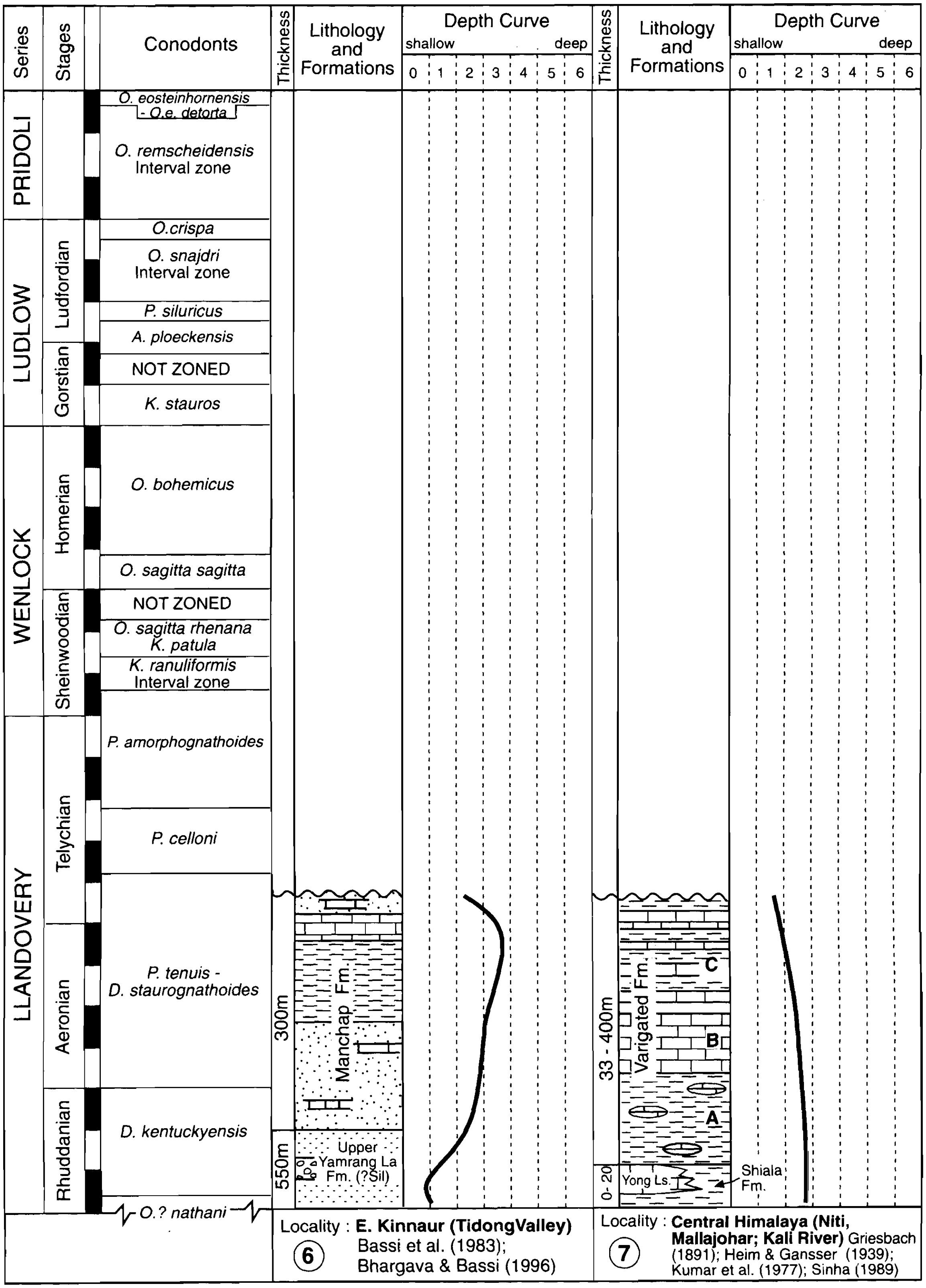

FIGURE 5 — Stratigraphic columns for India (continued), central Nepal, southernmost Xizang, and northern Shan State, Myanmar (Burma).

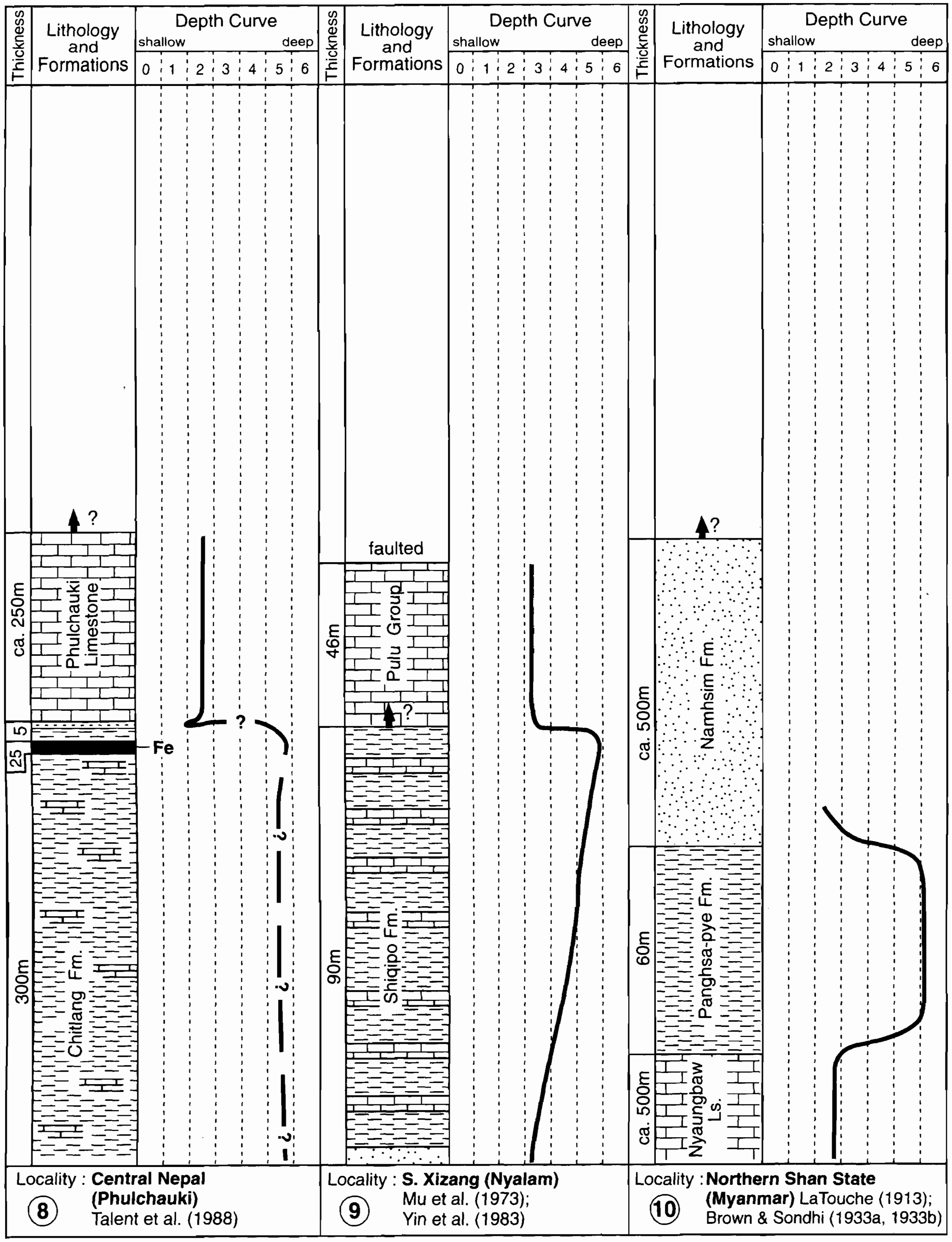

FIGURE 5 *continued.*

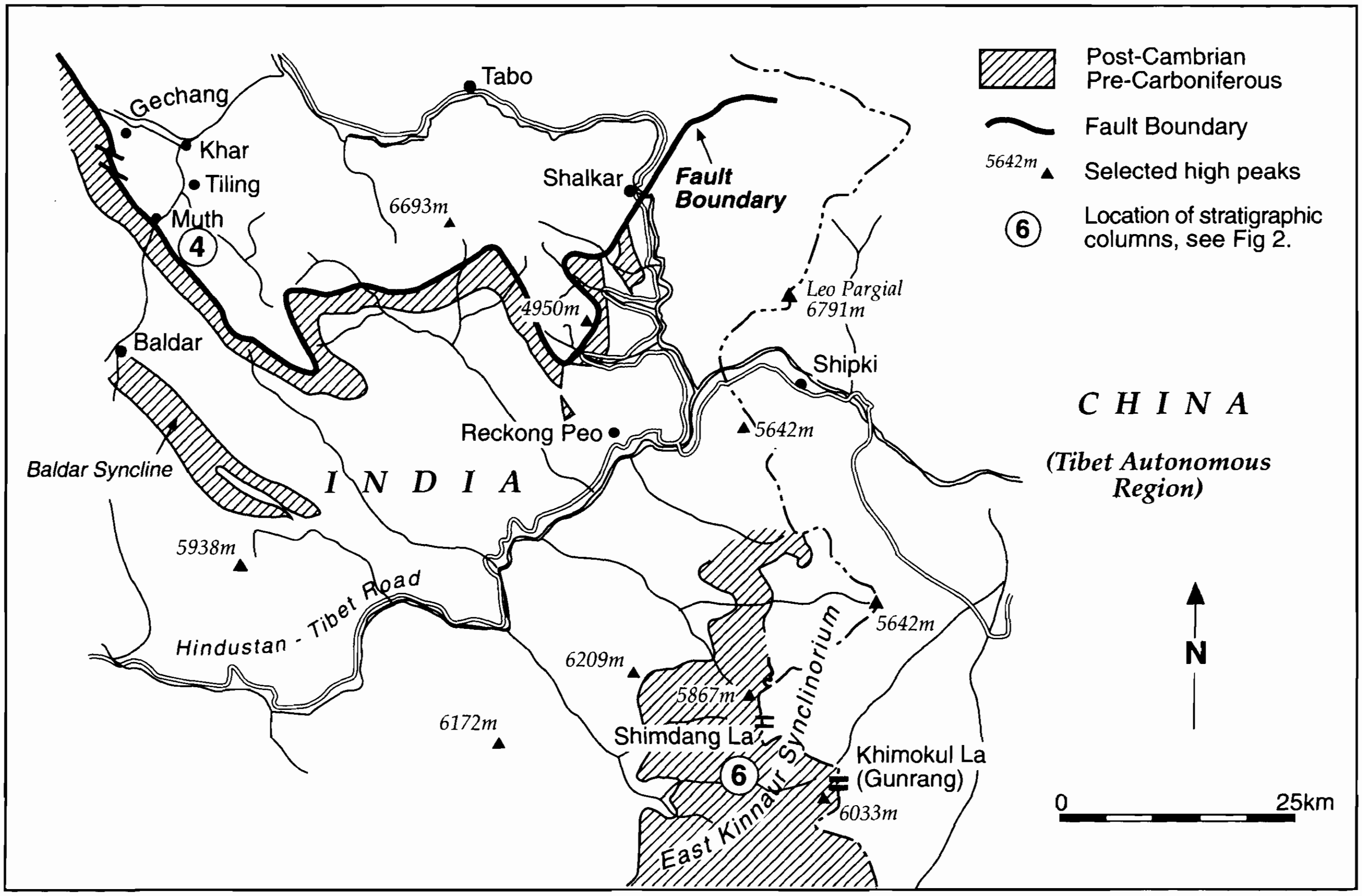

FIGURE 6 — Distribution of post-Cambrian–pre-Carboniferous rocks in Spiti and Kinnaur, India (after Talent et al., 1988).

Myanmar (Burma)

The biostratigraphy of the Myanmar Silurian has been neglected since reconnaissance geologic mapping early in the twentieth century (La Touche, 1913; Brown and Sondhi, 1933a, 1933b). Poorly preserved fossils, generally limited exposures, and deep weathering limit stratigraphic synthesis. Stratigraphic names have tended to proliferate, but the known rocks can be grouped into three stratigraphic "packages." These are the Nyaungbaw Limestone or Linwe Formation (probably extending down in the Ordovician); the Pangsha-pye Formation, a widespread Llandovery graptolitic shale; and the Namshim Formation, a Wenlock and perhaps younger sandstone–marl facies. The unconformably overlying Zebingyi Beds, long thought to be Silurian, have Pragian graptolites and dacryoconarids. This latter unit is now viewed as substantially, if not entirely, Early Devonian.

Approximate correlatives of the Zebingyi Beds occur in northern Thailand (Jaeger et al., 1968, 1969; Jaeger, 1983). Useful reviews of the Silurian of Myanmar are those of Berry and Boucot (1972) and Bender (1983).

Northern Shan State — The Nyaungbaw Limestone (La Touche, 1913, p. 119; Brown and Sondhi, 1933b, p. 219; Pascoe, 1959, p. 616) was long considered to be Late Ordovician (Ashgillian) and an approximate equivalent of the Linwe Formation of southern Shan State. As the Linwe Formation extends into the Llandovery (discussed below), the Nyaungbaw Limestone has also been assumed to span the Ashgillian–Llandovery boundary and, because of the age of the conformably overlying Pangsha-pye Formation, to extend as high as the *Monograptus cyphus* Zone (Fig. 5, column 10).

The approximately 60 m-thick Panghsa-pye Formation takes its name from Panghsa-pye village about 12 km northwest of Hsipaw (Fig. 8). Two members can be discriminated in that area: a lower trilobite-bearing unit with

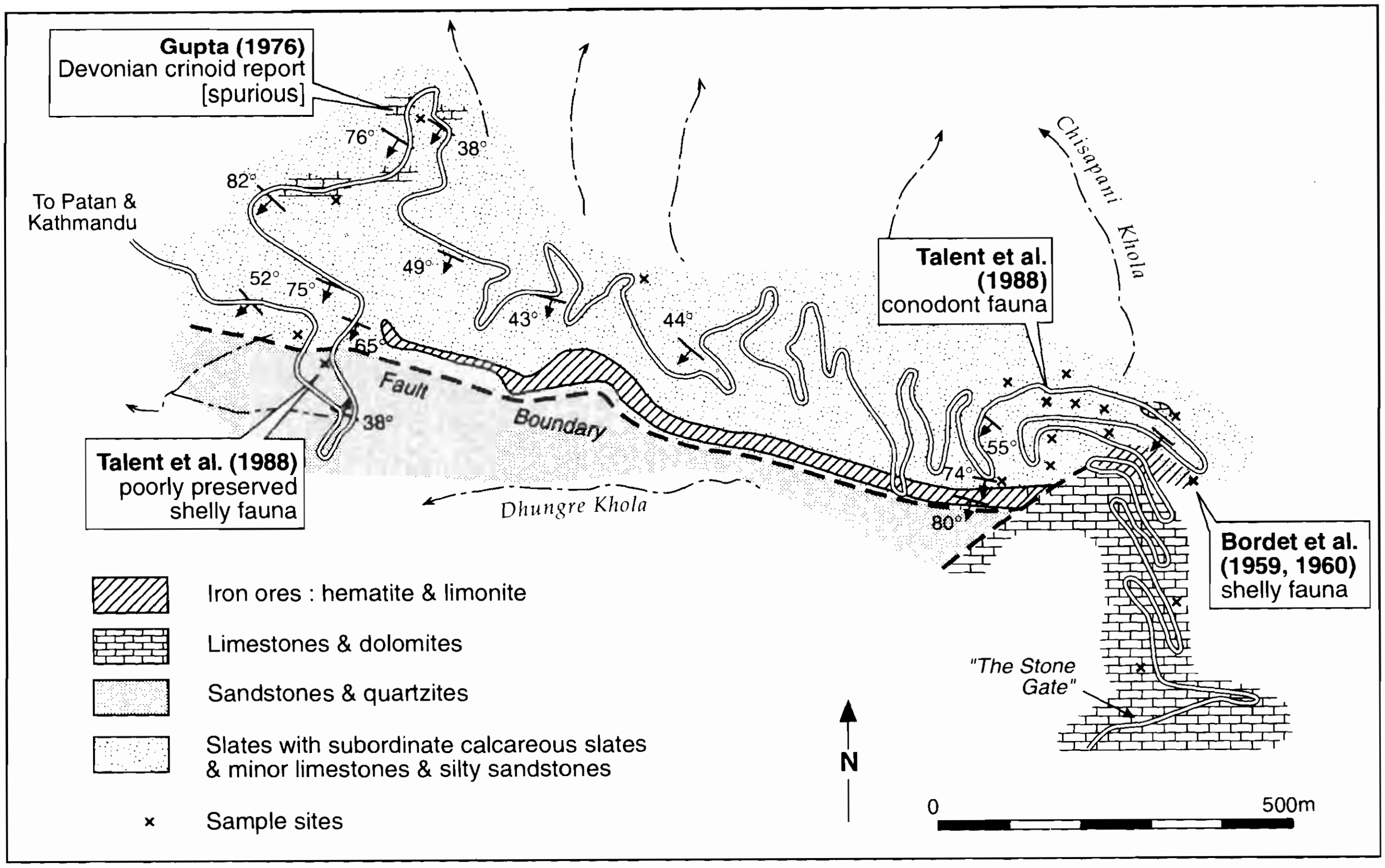

FIGURE 7 — Localities sampled for conodonts and macrofaunas along the Godavari–Phulchauki road, central Nepal (after Talent et al., 1988).

brachiopods, mollusks, and ostracodes, and an upper graptolite-bearing unit. Graptolites from the latter member include *Monograptus tenuis*, *M. gregarius*, *M. cyphus* var. *minor*, *M. concinnus*, *Orthograptus vesiculosus*, *Diplograptus modestus*, *Climacograptus medius*, and *C. rectangularis*. An early Llandovery age, probably the *Monograptus cyphus* Chron, was suggested by G. Elles (*in* La Touche, 1913, p. 126). The same graptolitic interval occurs at several other places in the Nam-tu Valley north of Panghsa-pye (La Touche, 1913, p. 127–129; Berry and Boucot, 1972, p. 21, 22).

The Namhsim Formation (La Touche, 1913, p.130; Pascoe, 1959, p. 646), a widespread unit of up to 500 m of cross-bedded quartzites, overlies the Pangsha-pye Formation. It has poorly preserved late Llandovery–?Ludlow shelly faunas illustrated in a monograph by Reed (1906). Comments on biostratigraphically important taxa from this unit are in Berry and Boucot (1972, p. 17). The Konghsa Marl Member, a sequence reported from several areas in northern Shan State, is assumed to be a facies of the Namhsim Formation (La Touche, 1913); a fauna from this unit west of Kyaukme is regarded as likely upper Ludlow (Mitchell et al., 1977, p. 18). The faunas of the Namhsim Formation need restudy.

SOUTHERN SHAN STATE — The Linwe Formation, with type exposures near the village of Linwe near Ye-ngan (Fig. 8), conformably overlies the Pindaya Group of Myint (1973) and passes, apparently gradationally, into the overlying Wabya Formation. The Linwe Formation has been considered Upper Ordovician (Reed, 1936, p. 58) and mapped as the Upper Ordovician *Orthoceras* Beds by Brown and Sondhi (1933b). However, Myint (1973) reported *Monograptus* from the Linwe Formation, and it is now assumed to be at least partly Silurian and to correlate broadly with the Nyaungbaw Limestone of northern Shan State. The Linwe Formation consists of 150–650 m of red-mottled, phacoidal limestone, siltstone, and fine-grained calcareous sandstone (Myint, 1973; Garson et al., 1976).

The Llandovery graptolites *Monograptus cyphus*, *M.*

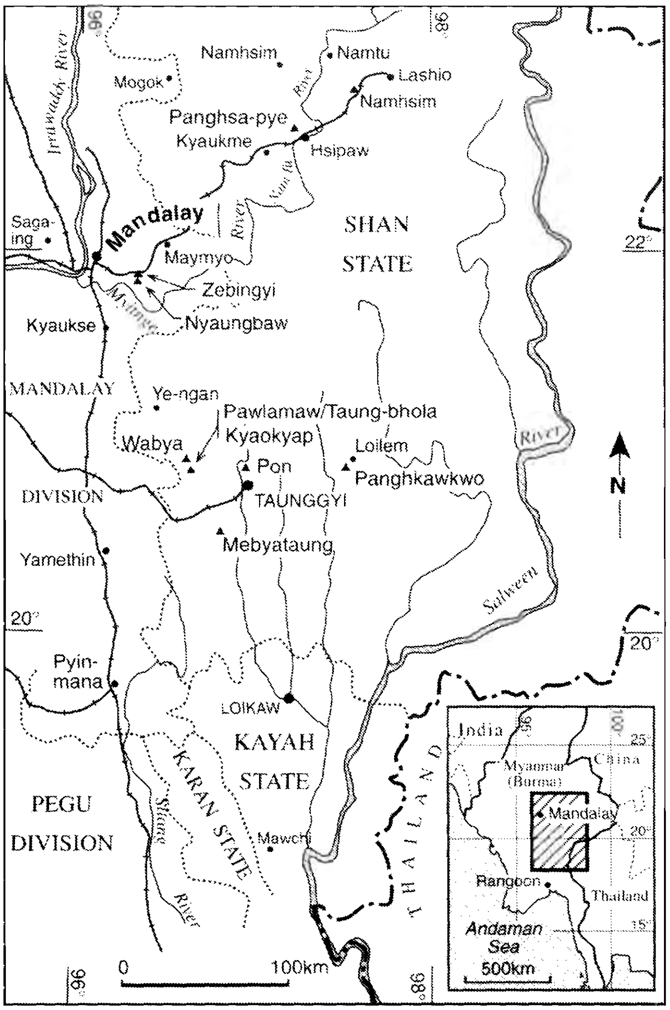

FIGURE 8 — East-central Myanmar (Burma) and localities mentioned in the text.

incommodus, M. sandersoni, Orthograptus vesiculosus, Climacograptus medius, C. rectangularis, and *Glyptograptus tamariscus* var. *incertus* indicate the *Monograptus cyphus* and *M. gregarius* Zones. These graptolites have been reported from the Wabya Formation in the Pindaya Range in southern Shan State (Brown and Sondhi, 1933b, p. 213; Pascoe, 1959, p. 652, 653). A graptolite assemblage near Panghkawkwo and Loilem (Brown and Sondhi, 1933a, p. 144, 145; Reed, 1932, p. 144, 145; Pascoe, 1959, p. 652; Fig. 8) is somewhat younger, and probably referable to the *M. convolutus–M. sedgwicki* interval. Limestones apparently younger than this graptolitic interval are reported near Loilem (Brown and Sondhi, 1933a). Another Llandovery graptolite fauna with *Orthograptus mutabilis, Climacograptus medius, C. innotatus* and *Monograptus fimbriatus* suggests the *M. gregarius* Zone. This fauna occurs on the east slope of Kyawkyap Pagoda Hill, about 15 km north of Heho (Brown and Sondhi, 1933b, p. 216, 217; Pascoe, 1959, p. 653). About 15 km south of Heho at Mebyataung between Pawlamaw and Taungbhola, a limestone–mudstone sequence with three graptolite bands has *Monograptus gemmatus, M. concinnus, M. jaculum,* and *Rhaphidograptus tornquisti* in the lowest horizon. *Monograptus sedgwicki, M. tenuis,* and *Climacograptus scalaris* occur in the middle horizon, and *M. tenuis, M. millepeda, C. scalaris,* and *R. tornquisti* appear in the upper horizon. An interval from the *M. gregarius* Zone to the *M. sedgwicki* Zone may be represented (Brown and Sondhi, 1933b, p. 218, 219; Berry and Boucot, 1972, p. 11). This graptolitic shale and mudstone interval is overlain by limestone and sandy mudstone with no reported faunas. A limestone–shale sequence with shelly fossils east of Pon in the Taunggyi region was thought to be Silurian (Brown and Sondhi, 1933b, p. 226, 227). However, as dacryoconarids also occur, this latter sequence is probably Lower Devonian and broadly correlative with the Zebingyi Beds of northern Shan State.

KAYAH — Well preserved Silurian fossils are reported from limestone and calcareous mudstone (Loikaw Beds of Hobson, 1941, p. 127) in the "Balu Chaung Bed" on a hill near Loikaw, the capital of Kayah State in eastern Myanmar (Fig. 8). Precise data are not available on this fauna and its age (Holland et al., 1956, p. 73). Silurian graptolites have been reported (Ba and Searle, 1961) in the Mawchi "Series" at "Mile 14" on the Kermapyu–Mawchi road.

CONCLUSIONS

Silurian sequences on the Indian subcontinent are less exposed and known than might appear from the litho- and biostratigraphic literature. They are arranged parallel to the northern margin of the Indian sub-continent, and record a paleobathymetry that generally tends to deepen to the north. Continuous sedimentation from the Ordovician to the Llandovery, and perhaps Wenlock, with local carbonate build-ups, is recorded from several regions along the Himalayas, principally from Kashmir, Ladakh, Spiti, Kinnaur, and Kumaon. The biostratigraphic control is limited. Better-documented occurrences with graptolites in southernmost Xizang (Tibet) are part of the same suite of continental margin associations. Paleontologically poorly known Llandovery–Wenlock strata (earlier regarded as Middle and Late Devonian) are documented from the Chandragiri–Phulchauki synclinorium of central Nepal. The only sequence with continuous Pridoli–Lochkovian is in the Nowshera region of northwest Pakistan, where precise biostratigraphic correlation based on conodonts is possible.

The remarkably widespread Muth Quartzite of Kashmir, Spiti, Zanskar, and the High Himalaya of Kinnaur and Kumaon, long thought to be Silurian or Devonian, is a unit that exhibits features indicative of beachs and tidal flats. It is now demonstrated to rest unconformably on Llandovery–?Wenlock sequences formerly included in the "Muth Quartzite."

Most reports and correlations of Silurian fossils from India that post-date the pioneer, nineteenth and early twentieth century monographs of J. W. Salter and F. R. Cowper Reed are discounted. Paleontologically better-dated Silurian sequences have been documented from the smaller terranes adjacent to the Indian Block and separated from it by ophiolitic belts. These include: 1) the Llandovery–Ludlow with brachiopod-trilobite macrofaunas and some conodonts from the Dasht-e-Nawar region of the Helmand Block, southeast Afghanistan; 2) an upper Llandovery graptolite-trilobite occurrence in the Pamir Block in Chitral, Northwest Frontier Province, Pakistan; and 3) the numerous reports of Llandovery graptolites (*Monograptus cyphus* to *M. sedgwicki* Zones) from eastern Myanmar on the Shan-Thai/Sibumasu Block. These areas may have been joined to or have been very close to the Indian Block on the northern Gondwana margin in the Silurian. However, an absence of Silurian paleomagnetic data for these blocks and the generally inadequate paleontologic control for the Indian and adjacent crustal blocks limit reconstruction of facies boundaries through this Asian region.

The reasons for deficiencies in the stratigraphic and biostratigraphic control for this large region are many. The difficulties of geological mapping in the High Himalaya are well known. The Paleozoic macrofaunas are often sheared, and this makes such materials unattractive to all but a very few tenacious workers. As a general rule, organic microfossils have been thermally destroyed in the Paleozoic sequences of the Himalaya and adjoining regions. Still to be seriously tested is the value of conodonts in correlation of major Ordovician–Silurian metamorphosed limestones, dolostones, and marbles that occur through the High Himalaya from Kashmir to Bhutan. It will be no easy task, given the high CAI (color alteration indices) and poor quality of materials anticipated from these often moderately to highly metamorphosed sequences. Even where less-metamorphosed materials are available, as in Spiti, Chitral, and parts of Kashmir and Myanmar, there is still a dearth of high-quality, bed-by-bed biostratigraphic studies, essential for high-precision, inter-regional correlation. There is thus an urgent need for a re-investigation of the Silurian and associated Upper Ordovician and Devonian of the entire Himalayan region of India and the adjacent microcontinental blocks. This type of study is essential for determining the timing of tectonic events, transgression–regression patterns, and changes in paleogeography.

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Part IV: East and North Asian Continents

Silurian Paleogeography of China

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ABSTRACT — Silurian China was composed of five paleoplates and parts of several other insular paleoplates. Complicated interactions between various paleoplates led to intricate sedimentary facies patterns. Collision, subduction, or consumption of paleoplates led to stratigraphic incompleteness. The Silurian of China is dominantly siliciclastic, with or without volcanic input and with local limestone. Llandovery biostromes and bioherms are well developed on the Yangtze platform, Wenlock–Pridoli carbonates are chiefly developed in Tibet and western Yunnan, and Ludlow carbonate mud mounds are known in southern Inner Mongolia. Synecologic analysis of the shelly faunas suggests that most communities were shallow-water (BA 1–3). Deeper water graptolites occur in marginal belts, and radiolarians are present at a few localities in Qinling and western Junggar. Subaerial environments were dominant, and the extent of marine onlap was limited in the Silurian of China. Of the four Llandovery eustatic highstands, three are recognized in South China. Tectonic movements with rapid uplift and subsidence took place in South China during the late Aeronian and early Telychian. A sea-level highstand is recorded in patch reef-bearing areas of South China in the late Ludlow. Many endemic forms in the Llandovery show the isolation of South China from other larger paleoplates. Silurian shelly faunas demonstrate that South China, North China, Tarim, Qaidam, and Indochina, with a *Retziella* fauna, moved north and were located near the equator. In addition, most of western Yunnan (part of Sibumasu) and Tibet, which have a European *Kopaninoceras* nautiloid fauna, were in the southern hemisphere, and separated from the five paleoplates listed above. Finally, the distinctive *Tuvaella* fauna of the Khingan Mountains and northeasternmost Xinjiang (southern Siberian marginal belt) shows that these two areas were in the northern hemisphere and were separate from the other parts of China during the Silurian.

Introduction

Since Yin (1949) published paleogeographic maps for the Early Silurian of South China, a considerable amount of work has been done on the Silurian of China. This has included geologic mapping in almost all regions, detailed measurements of numerous stratigraphic sections, and systematic paleontology of various kinds of fossils in 1950–1980. Based on pre-1980 data, Silurian paleogeographic maps of China were proposed by Lin et al. (1984), Guan et al. (1984), Li and Dai (1985), and Mu et al. (1986). In the 1990s, new data on the Silurian from various parts of China have been obtained. Regional or local paleogeographic maps for different Silurian epochs have been published (Du and Xu, 1990; Zhou et al., 1993, 1996; Liu and Xu, 1994; Ni et al., 1995).

Refined biostratigraphy is one of the most important tasks in paleogeography. Among the maps mentioned above, several feature problems in correlations (e.g., in which the Lojoping and Shamao Formations were assigned to the upper Llandovery and Wenlock, but are now regarded as uppermost Aeronian and lower Telychian, respectively). The Subcommission on Silurian Stratigraphy (SSS) has established an international standard for high-resolution stratigraphic correlation (Holland, 1985; Bassett, 1985; Cocks, 1985). Meanwhile, new studies of graptolites, conodonts, and chitinozoans by Chen X., Wang C.-y., and Geng L.-y., respectively, have revised systematic paleontology. Other projects, such as the Trans-Hemisphere Telychian of South China and the British Isles (Chen and Rong, 1996), improved the syntheses of Silurian correlation, community paleoecology, and paleobiogeography of China.

The Silurian of the South China paleoplate is better known than that of any other Chinese paleoplate because of superior exposure, better access, a longer research history, more detailed and accurate biostratigraphic correlation, and better systematic paleontologic work. Holland (1989) appreciated the Yangtze platform as a gateway to

Chinese geology. Other parts of China, in particular such remote and sparsely populated areas as Xizang (Tibet), Qinghai, Nei Mongol (Inner Mongolia), and Xinjiang, have received less detailed biostratigraphic work than South China. Differences in the detail of Silurian work in various regions in China have resulted in the construction of two paleogeographic maps (Llandovery–Wenlock and Ludlow–Pridoli boundaries) for China. Eight paleogeographic maps have beeen done for South China (i.e., one for the uppermost Ordovician *Normalograptus extraordinarius* Zone; five for the Llandovery *Parakidograptus acuminatus, Demirastries triangulatus, Rastrites convolutus-Stimulograptus sedgwickii, Spirograptus turriculatus*, and *Oktavites spiralis* Zones; and one each for the Wenlock and late Ludlow).

TECTONIC FRAMEWORK

The Ordovician tectonic history of China is summarized in Chen and Rong (1992). Revisions of the Ordovician–Silurian paleoplates of China are given below. Two ranks of tectonic units are used herein: 1) paleoplate [="Domaini" of Wang and Mo (1995)] is the first-rank unit, and includes a continent and surrounding mobile belts; 2) platform, basin (depression), and mobile belt are the second-rank units. "Block" is used for an independent, smaller unit without a precise definition. It is either a separate paleoplate or a tectonic unit within a paleoplate. Various tectonic units of China in the Silurian are illustrated (Fig. 1).

China has a complicated tectonic framework. China in the Silurian consisted of several independent paleoplates, which collided with each other many times through geologic time. Five separate continents (North China, South China, Tarim, Tibet, Qaidam); parts of three continents (Siberia, Kazakhstan, Sibumasu); and several blocks, such as the Bureya–Jamus and Songpan–Ganzi (=Garze) blocks, are recognized in this report. The definition of each of these paleogeographic units of different rank is reviewed below.

BUREYA–JAMUS BLOCK — Chen and Rong (1992) were uncertain about the Bureya–Jamus area and its boundaries (Fig. 1). This block is bounded by the Khingan Mountains, probably with development of early Hercynian convergent zones. It was a land mass in the Silurian and is now covered by Cenozoic rocks. It is still unclear whether this block was related to the Siberian paleoplate or the Northeast Asian paleoplate (Wang and Mo, 1995). It seems more likely that this area was an independent block through its geological history.

SOUTHERN MOBILE BELT OF THE SIBERIAN PALEOPLATE — The extent and definition of the Siberian paleoplate have been discussed by many geologists (e.g., Li et al., 1984; Wang and Mo, 1995). The southern marginal belt, including the Greater and Lesser Khingan Mountains, Mongolia, and Altai, is characterized by a distinctive *Tuvaella* fauna (Su, 1981; Rong and Zhang, 1982), which is absent elsewhere on the Siberian platform. This southern belt has been regarded as an independent tectonic unit because of this brachiopod fauna and the nature of the Ordovician–Silurian stratigraphy (Xue et al., 1980; Su, 1987). A collision between Siberia and North China may have occurred during the Late Devonian to Early Carboniferous, whereas amalgamation of the unit and Siberia was accomplished before the Cambrian.

There is no apparent record of the *Tuvaella* fauna in Kazakhstan (Su, 1981; Rong and Zhang, 1982; Rong et al., 1995). The Barkol area in northeast Xinjiang, with a *Tuvaella gigantea* assemblage (Rong and Zhang, 1982), is regarded as the southern marginal belt of the Siberian paleoplate. It shares a *Tuvaella* Fauna with the Altai in the west and the Khingan Mountains in the east.

KAZAKHSTAN PALEOPLATE WITH JUNGAR, YI'NING, AND BADAINJARIA BLOCKS — With the exception of the area with the *Tuvaella* fauna, northern Xinjiang consists of the Junggar, Turpan–Hami, and Yining Massifs, all of which are assigned to the Kazakhstan paleoplate. The latter definition has been applied by many geologists. However, Scotese and McKerrow (1991) regarded Kazakhstan as an amalgamation of the Kokchetav–North Tianshan island-arc system, the Dzhungaria–Balkhash and Zaisan back-arc basins, the Chingiz–Tarbagatai island-arc system, and parts of the Altai–Sayan shelf (Nikitin et al., 1991). However, a Kazakhstan paleoplate is recognized herein. It is distinguished from Siberia by its Silurian faunas, and it is bounded by the eastern extension of several mobile belts in northern Xinjiang and southwestern Inner Mongolia. The boundary between the Siberian and Kazakhstan paleoplates is clearly represented along the Ertix–Kalameili convergent zone (Chen and Rong, 1992; Wang and Mo, 1995; Ni et al., 1996). Talent et al. (1987) recognized an almost total dissimilarity of the Lower Devonian brachiopod faunas between the southern marginal belt of the Siberian paleoplate and the Kazakhstan paleoplate. Hou and Boucot (1990) proposed the Balkhash–Mongolia–Okhotsk region during the Emsian (Early Devonian), based on numerous endemic genera and the presence of genera shared with the Eastern Americas and Tasman region. Nevertheless, these two paleoplates are known to have collided in the later Carboniferous (Liao and Ruan, 1995).

South of the Ertix–Kalameili suture in the Beitashan area, Badainjaria (Fig. 1), Silurian rocks were thrust southwest onto the eastern edge of the Junggar Massif, which was mainly a land area in the Silurian. The West

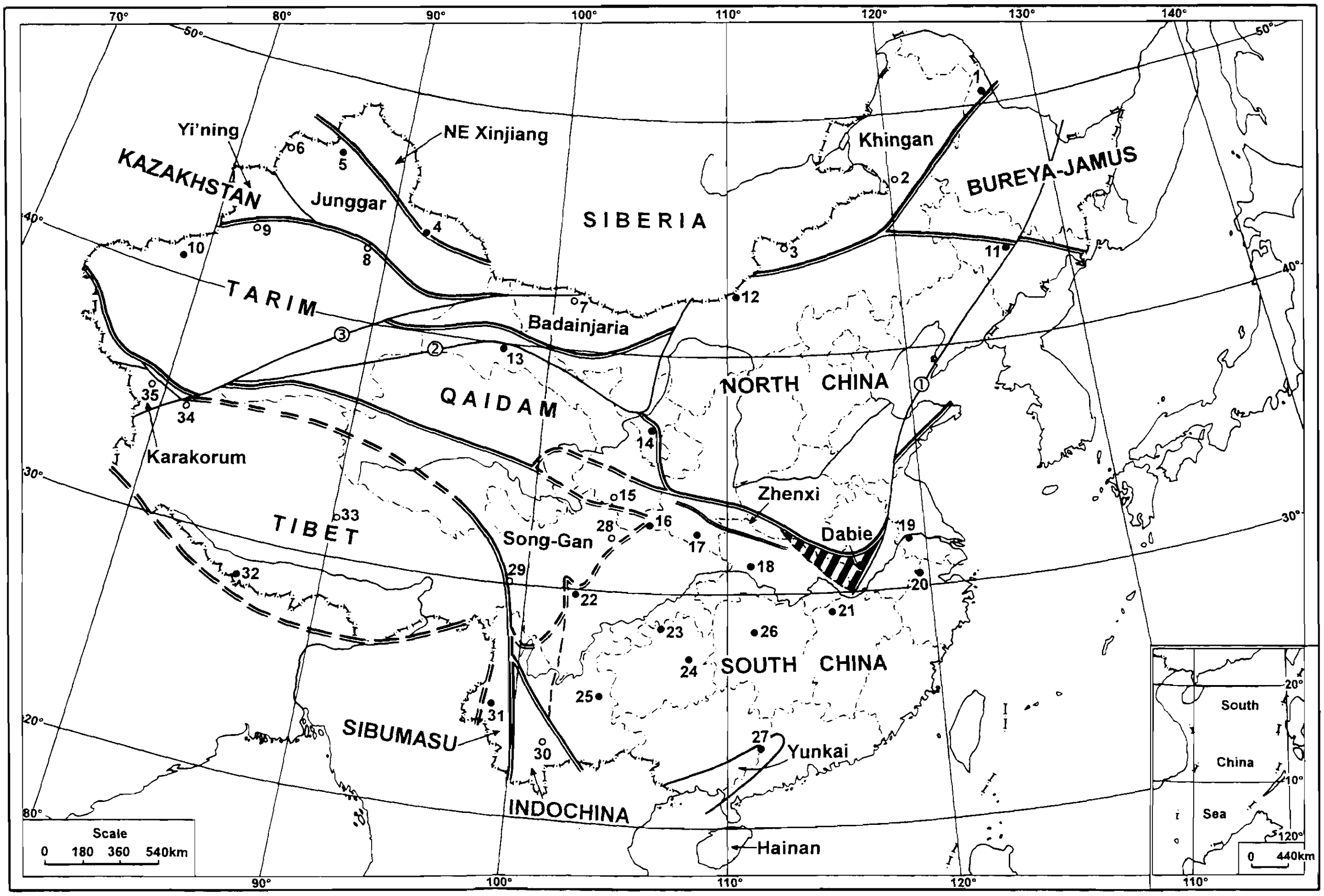

FIGURE 1 — Silurian tectonic units of China. Explanation: 1) Siberian paleoplate with southern mobile belt in northeast Xinjiang and the Greater and Lesser Khingan Mountains; 2) Bureya–Jamus block (Silurian unknown); 3) Kazakhstan paleoplate with its eastern extension as Junggar mobile belt and Junggar, Yi'ning, and Badainjaria blocks; 4) Tarim paleoplate with southern Tianshan mobile belt and Tarim platform; 5) Qaidam paleoplate with Alxa, Qilianshan mobile belt, and Qaidam platform; 6) North China paleoplate (i.e., Sino–Korean paleoplate) with North China platform and northern mobile belt; 7) South China paleoplate (i.e., Yangtze platform, Xianggui Basin, and Cathaysian Oldland) with south Qinling mobile belt, Zhen-Xi block, and Song–Gan [Songpan–Ganzi; =Garze] blocks; 8) Indochina paleoplate with Simao block in eastern west Yunnan; 9) Sibumasu paleoplate in westernmost Yunnan; 10) Tibet paleoplate with Karakorum block (Tibet and Indian paleoplates may have been joined in the Early Paleozoic). 11) Yunkai and Hainan blocks: their origins are debatable. Black dots show sections of Figs. 4–16 and hollow dots show sections referenced: 1. Xigulanhe, Nenjiang, north Heilongjiang; 2. Suhuhe, north Inner Mongolia; 3. Wuburbetobo, central Inner Mongolia; 4. Hongliuxia, Barkol, northeast Xinjiang; 5. Hoboksar, north Xinjiang; 6. Tacheng, northwest Xinjiang; 7. Erjin, west Inner Mongolia; 8. Toksun, central Xinjiang; 9. Baicheng, central Xinjiang; 10. Kalping, west Xinjiang; 11. Zhangjiatun, Yongji, central Jilin; 12. Gashaomiao, Darhan Mumingan Joint Banner, south Inner Mongolia; 13. Yumen, northwest Gansu; 14. Tongxin, south Ningxia; 15. Nanshimengou, Zugqu, southeast Gansu; 16. Ningqiang, southwest Shaanxi; 17. Bajiaokou, Ziyang, south Shaanxi; 18. Fenxiang, Yichang, west Hubei; 19. Jiangning, Nanjing, south Jiangsu; 20. Xiaofeng, Anji, northwest Zhejiang; 21. Wuning–Xiushui, northern Jiangxi; 22. Erlangshan, west Sichuan; 23. Hanjiadian, Tongzi, north Guizhou; 24. Leijiatun, Shiqian, northeast Guizhou; 25. Yuejiadashan–Liaojiaoshan, Qujing, east Yunnan; 26. Anhua, central Hunan; 27. Yu'nan, west Guangdong; 28. Maoxian, west Sichuan; 29. Batang, west Sichuan; 30. Mojiang, eastern west Yunnan; 31. Shidian, westernmost Yunnan; 32. Nyalam, south Xizang; 33. Xainza, north Xizang; 34. Lazhulong, Rutog, northwest Xizang; 35. Tianshuihai, south of Hetian, southwest Xijiang. Strike-slip Faults: ① Tanlu, ② Altun–Longshoushan–Yuanshanzi–Helanshan (Zhou et al., 1995), ③ Qiemo–Xingxingxia.

Junggar mobile belt consists of volcaniclastics with graptolites in a few shale intercalations and shelly fossils in carbonate lenses. It was deformed in the Carboniferous. The well-known late Caledonian Tangbale ophiolite zone resulted from collision of the Yi'ning and Junggar Massifs (Zhu et al., 1987), and was structurally displaced during the Carboniferous.

The Badainjaria block includes Beishan (North Mountain) in the west and the Badainjaria Desert. It is debatable whether the Beishan region belongs to Tarim or Junggar (Xu and Huang, 1990; Ni et al., 1996). According to Zuo et al. (1987, 1990), the Beishan region is composed of two parts separated by an ophiolite zone. The northern part belongs to Kazakhstan and the southern to Tarim.

During the Silurian, the northern part was mostly a land mass that provided considerable siliciclastics to its northern mobile belt. Shelly fossils, mainly corals, prove that the northern mobile belt (Badainjaria) was amalgamated to Kazakhstan in the Devonian (Liao and Ruan, 1995).

TARIM PALEOPLATE — The Tarim paleoplate was previously regarded as a western extension of the Sino–Korean paleoplate (=North China paleoplate herein) through the Paleozoic by many geologists in China (e.g., Li and Wang, 1983) and abroad (e.g., McKerrow and Scotese, 1990). Recent investigations strongly suggest that the bio- and lithofacies of Ordovician–Silurian Tarim are different from those of North China (Zhou and Chen, 1990; Wang and Chen, 1991; Chen and Rong, 1992; Zhou and Dean, 1996). Remarkable differences in geologic history between Tarim and North China are documented (Gao and Wu, 1983; Burret and Stait, 1987; Zhou and Chen 1990; Xu and Huang, 1990; Scotese and McKerrow, 1991; Zhou and Lin, 1995; Liu et al., 1997). Tarim is considered an independent paleoplate during the Early Paleozoic by many geologists (see Zhou and Lin, 1995; Liu et al., 1995; Ni et al., 1996), or even regarded to be part of the same paleoplate as South China (Chen and Norlin, 1995). During the Silurian, Tarim was close to south China (on the bases of paleontological data analyzed herein).

There are many similarities between Silurian shelly faunas, particularly among brachiopods and corals, from south Tianshan in central Xinjiang and southwestern Tianshan on the border of Tadzhikistan and Kirkizistan (Nikiforova and Obut, 1965; Rong et al., 1995). The latter region did not belong to the Kazakhstan paleoplate in the Early Paleozoic (Nikitin et al., 1991), which may indicate that the Karakum region (Talent et al., 1986) was united with the Tarim paleoplate in the Silurian as suggested by Rong et al. (1995). Their present separation was probably caused by very strong crustal compression with the northern movement of the Indian plate during the Himalayan orogeny.

The boundary between the Kazakhstan and Tarim paleoplates is defined either by the boundary between central and south Tianshan (Chen et al., 1995; Gao et al., 1996) or between north and south Tianshan (Nikitin et al., 1991). The Tarim may have been truncated in the east by the Qiemo–Xingxingxia (or Altun) strike-slip fault (Fig. 1) in the Carboniferous (Zheng, 1991). The southern mobile belt of the Tarim platform is not yet well defined. Pan (1996) suggested that there was a proto-Tethys Ocean between northern Kunlun of the Tarim paleoplate and southern Kunlun of the Tibet Paleoplate in the Ordovician–Silurian.

Wang and Mo (1995) suggested that the Yi'ning Massif may have rifted away from and re-accreted to the Tarim paleoplate during the Jinningian orogeny, as suggested by Aksu blue schist, which is 700–900 m.y.a. The Tangbalei and Mishigou volcanic belts have been defined as two separate units by Geng (1992). However, Ordovician melanges within these two belts are overlain by flysch with early Telychian graptolites (Chen X. *in* Zhou and Chen, 1990). Telychian rocks of the Qargaye Formation near Tangbalei belong to the western Junggar mobile belt, and the Telychian rocks of the Mishigou Formation lie in central Tianshan. Ni et al. (1996) suggested that the Yi'ning Massif was part of the Tarim paleoplate before the Wenlock, and became part of the Kazakhstan paleoplate after the Ludlow. They assumed that the North Tianshan Ocean narrowed, while the South Tianshan Ocean originated by gradual extension at the beginning of the Ludlow. As far as we know, the Silurian Tarim paleoplate included two units: 1) the Tarim platform with a northern mobile belt, and 2) the southern Tianshan mobile belt. The Kuluktage–Maggar basin, which was well developed during the Ordovician, disappeared after Late Ordovician uplift following the *Climacograptus (Diplacanthograptus) spiniferus* Chron.

NORTH CHINA (=SINO–KOREAN) PALEOPLATE — Ordovician and Silurian tectonics of the North China paleoplate were similar (Chen and Rong, 1992). The North China platform is bordered on the north by a narrow belt of Cambrian and Ordovician island-arc volcaniclastics with an Early Paleozoic ophiolitic zone (Li, 1987). The rocks are unconformably overlain by later Silurian shallow-water siliciclastics and carbonates in the Baiyunobo area, southern Inner Mongolia (Li et al., 1985). The latter arc is near the Ondur Sum accretional zone in the north (Tang et al., 1993). The western mobile belt of North China was flooded from the Tremadocian to the early Ashgillian, but was uplifted during the middle–late Ashgill, coincident with a eustatic drop. The North China paleoplate is bounded on the south by the northern Qinling belt, where Silurian rocks are largely unknown, with the exception of possible Early Silurian deep-water radiolarian-bearing units near Shangxian, southern Shaanxi (Cui et al., 1995). On its eastern boundary, Silurian rocks (Llandovery–Wenlock Kosa and Woyangri Formations) and fossils (i.e., brachiopods, rugose and tabulate corals, gastropods, and bivalves) were discovered by North Korean geologists in the 1970s (Yang, 1989; An and Ma, 1993). In the Silurian, almost all of North China was a land mass that provided siliciclastic debris to its northern mobile belt (Niu et al., 1993).

QAIDAM PALEOPLATE — "Qaidam" (=Chaidam or Tsaidam in some reports) is the name applied to an Early Paleozoic paleoplate with boundaries well defined by ophiolite zones and distinctive biogeographic evidence. However, Duan et al. (1992) and Zhou and Lin (1995) regarded it as the southern part of the North China (or

Sino–Korean) paleoplate. Three tectonic divisions of the Qaidam paleoplate include the Alxa–Dunhuang Massif in the north, the Qilian mobile belt in the center, and the Qaidam platform in the south. According to Wang and Mo (1995), the Qaidam paleoplate began to split away from the North China paleoplate during the Cambrian, first forming the North Qilian Ocean and then, in the Ordovician, the South Qilian Ocean. Ophiolite and ophiolitic melange zones are well developed in north Qilian. It is probable that the oceanic basins began to narrow in the later Ordovician and closed in the Tongxin area during the later Llandovery, and in the Yumen area in the later Silurian. Zuo and Liu (1987) concluded that the North Qilian Ocean closed in the Middle–Late Devonian, which indicates that the Qaidam and Alxa converged during this interval, earlier than the final convergence of North China with South China or Tarim.

SOUTH CHINA PALEOPLATE WITH SONG-GAN BLOCK — The collision of the North and South China paleoplates during the later Paleozoic and Early Mesozoic has long been recognized by many geologists (e.g., Wang and Mo, 1995). The Qinling–Dabie–Sulu suture zone is the border between these two paleoplates (Nie et al., 1994; Wang and Mo, 1995). Studies of the zone reveal ultra-high pressure (UHP) coesite- and diamond-bearing metamorphic assemblages, which have attracted much attention. These areas represent the largest exposure of crustal rocks that were once buried deeper than 100 km and metamorphosed at pressure >30 kbar (Nie et al., 1994). The continental crust of the South China paleoplate was subducted more than 100 km beneath North China. This is a deeper depth than that reached by the subducting slab of the modern Indian plate under the Tibetan Plateau, as revealed by recent deep seismic reflection (Zhao et al., 1993). The east Qinling fold belt consists of several small units; one of them, the Zhenxi Massif (Yin and Huang, 1995; or Wudang backarc basin of Hao et al., 1996), is accepted as a tectonic unit herein (Fig. 1). This includes the central Qinling belt and part of the South Qinling belt, which form the north border of the South China paleoplate. In the Silurian, the Qinling mobile belt included west Qinling as well. The western end of the Qinling mobile belt during the Early Paleozoic, however, is still undefined due to lack of data.

The Cathaysian Land was recognized as part of the South China paleoplate by Rong and Chen (1987). Wang and Mo (1995) defined it as another tectonic domain within the Shaoxing–Shaoguan convergent zone that marked its west boundary. This interpretation is not accepted herein because definitive evidence of this convergent zone during the Early Paleozoic is not known. Early–Middle Ordovician rocks with a few graptolites from Yong'an, Fujian, southeast China (within the Cathaysian domain of Wang and Mo, 1995) may be regarded as a part of the Zhujiang Basin on the South China paleoplate.

The Lijiang–Ninglang area, north Yunnan, is regarded as the western margin of the South China paleoplate. However, the relationship of Silurian bio- and lithofacies of this area with South China has not yet been defined.

Most parts of the Song–Gan (=Songpan–Ganzi) block are covered by very thick Triassic strata that compose the largest exposure of Triassic rocks on earth. The total volume is 2,200,000 km^3, and this region of the northeastern Tibetan Plateau occupies an area of about 200,000 km^2. Ninety percent of this cover consists mainly of Late Triassic flysch (Nie et al., 1994). Nie et al. (1994) concluded that the majority of these siliciclastics were derived from denudation of the orogenic belt between the North and South China paleoplates after their latest Paleozoic–Triassic collision. The Songpan–Ganzi block probably was part of South China during the Silurian, because it shares the same basement as the Yangtze platform, because it rifted away from South China during the late Early Paleozoic (Yang et al., 1994), and because there are many similarities between the Silurian and Devonian faunas of the Yangtze and Songpan-Ganzi blocks. Thick siliciclastics accumulated in its eastern part during the Silurian, where they define the west mobile belt of the Yangtze platform (Chen and Rong, 1992). Yao and Li (1983) reported small exotic units of Ordovician to Triassic age from Zhuqing near the border between Sichuan and Tibet that indicate a suture between South China and Tibet along the Jinsha River.

YUNKAI AND HAINAN BLOCKS (OR TERRANES) — Yunkai (Qinzhou, Fangcheng, Zhaoqing, and other counties) and Hainan are treated herein as two small blocks that were separate from the South China paleoplate in the Ordovician–Silurian. They may have been located in the ocean between South China and Indochina during most of the Paleozoic.

Chen et al. (1995) considered that the Yunkai block is a nappe characterized by continuous sequence of deep-water, Silurian–Lower Devonian graptolitic strata (Wang X.-f., 1978). Although Silurian shelly fossils in this block are little known, a single species of the Late Silurian brachiopod *Retziella* was found in Chengxi County, west Guangxi (Rong et al., 1995). Early Devonian eospiriferids occur in the Yulin area (Wang and Yang, 1998), but are not known in the South China paleoplate (Wang and Rong, 1986). It is likely that the Yunkai block is probably related to the Indochina paleoplate, but was separate from South China in the Early Paleozoic. It seems to have rifted away from Gondwana a little later than the South China paleoplate, and was finally thrust onto the South China paleoplate during the Indo-Sinian orogeny. Paleomagnetic data on Early Carboniferous limestone in Hepu County in the Yunkai block show a 20.6° S position (Wu et al., 1997). In Shangsi County, Guangxi Province, the Lower Permian is thrust onto the Triassic and Jurassic.

The Hainan block has been suggested to be part of the Indochina paleoplate (Wang, 1989). There are two tectonostratigraphic terranes within the Hainan block, but their boundary is problematical (Metcalfe, 1995). Fossiliferous Cambrian and Ordovician occur mainly in Damao and Yaxian Counties on southernmost Hainan Island. The Early–Middle Cambrian *Xystridura* (trilobite) fauna and phosphate-bearing strata (Damao Formation) are also known in Australia. The early Ashgillian Gan'gou Formation from Yaxian County has low-diversity brachiopods (*Sowerbyella, Leptellina*) and trilobites (*Encrinuroides, Parisoceraurus*) (Ge et al., 1983) that are different from those of Gondwana. The Silurian of southern Hainan Island is composed mainly of slate intercalated with siliceous nodular limestone and with siliceous slate with chitinozoans (*Conochitina* sp. cf. *C. edjelensis*) and acritarchs deposited in deep-water (Wang et al., 1992).

INDOCHINA PALEOPLATE — The Indochina paleoplate is bounded to the northeast by the Song Ma and Ailaoshan suture zones, and to the west by the Changning–Menglian suture zone in west Yunnan, the Nan–Uttaradit suture in Thailand, and the Raub–Bentong suture in Malaysia (Metcalfe, 1995). Western Yunnan is tectonically complicated and is probably composed of four blocks: Simao in the east, Lincang in the center, Baoshan in the west (Fang, 1994), and Tenchong in the far west. The Simao block may belong to the Indochina paleoplate. However, the Early Paleozoic faunas remain poorly known, and their affinities need to be demonstrated. Based on Zhou et al. (1998), Early–Middle Ordovician trilobites of the Indochina paleoplate differ markedly from those of the Sibumasu paleoplate, and suggest separation of the two units during the earlier Ordovician. The Wenlock and Ludlow graptolite-bearing beds from the Mojiang area studied by Zhang and Lenz (1997) are of deep-water origin. The *Retziella* (brachiopod) fauna is known in central Vietnam (Thanh, Boucot, Rong, and Fang, unpublished data), and indicates a close relationship with South China in the Late Silurian.

SIBUMASU PALEOPLATE — The Sibumasu paleoplate (i.e., Than–Thai or Burma–Malaya paleoplate of Scotese and McKerrow, 1991) stretches through most of western Yunnan and the Than States of Burma, through much of Thailand and western Malaysia, and into northwestern Sumatra (Metcalfe, 1988, 1995). Silurian rocks are patchy in this paleoplate, and a variety of facies are represented. They range from graptolitic deposits in the Shan States and on Langkawi Island of northwestern Malaysia to local shelf carbonates (Cocks and Scotese, 1991). The Baoshan block may have formed the northern end of the Sibumasu Paleoplate in Early Paleozoic, and its Silurian rocks are composed chiefly of graptolitic shales and carbonates that yield conodonts and nautiloids with a deeper water affinity (Ni et al., 1982).

The Sanjiang region in northwest Yunnan and southeast Xizang is one of the most geologically complicated areas in the world. Available data on geological history through the Phanerozoic is quite poor for this region. Relationships between this region and the Sibumasu paleoplate or the Tibet paleoplate (see below) are also poorly known. Two possibilities for explaning these two paleoplates have been proposed. They were either united as an independent domain (Metcalfe, 1992; Wang and Mo, 1995), or may have been separate in the later Ordovician and Silurian, which is provisionally accepted herein.

TIBET PALEOPLATE WITH KARAKORUM BLOCK — The Tibet paleoplate may have been joined to the Indian paleoplate in the Ordovician and Silurian (Metcalfe, 1995), but is treated herein as a major unit that includes both the Lhasa and Qiangtang regions. The latter two regions have been regarded as two separate units in the Early Paleozoic (e.g., Scotese and McKerrow, 1990), based on consideration of the Yarlung–Zangbo accretion zone as the suture zone between the Qiangtang and Lhasa regions. The age of the zone, however, is much younger than Paleozoic.

A possible northern mobile belt exists in the Kunlun Mountains. Its northern border possibly coincides with the southern border of the Tarim and Qaidam paleoplates. One of the authors (Chen T.-E.) investigated the Ordovician of southeast Shache County, southwest Xinjiang, and suggested that the Lower Paleozoic of the west Kunlun Mountains may indicate the northern mobile belt of the Tibet platform (see Wen et al., 1996). In Qinghai Province, 70 km south of Nuomuer village, metamorphosed Silurian siliciclastics with volcanics dated at 426 m.y.a. may also indicate a Silurian mobile belt on the north slope of the east Kunlun Mountains in Tibet. At Nachitai village south of Golmud, Ordovician and possible Silurian rocks are unconformably overlain by the Triassic and Jurassic (Ye S.-l., personal commun. 1985). At Xiangrijin, south of Dulan, Ordovician and possible Silurian rocks are unconformably overlain by the Carboniferous (Chen G.-l., pers. comm. 1985). These sequences are located along the boundary between the Tibet and Tarim–Qaidam paleoplates. Most of Tibet belonged to the Tibet platform in the Early Paleozoic. Mu and Chen (1984) reported two lithofacies belts that extend from Tibet to the Kashmir–Nepal region. It should be pointed out that the southern marginal belt of Tibet in the Silurian is poorly known. The Karakorum region adjacent to

Tarim and Pamir on the north and Tibet on the south is geologically complicated, and has been regarded as a block or microplate (e.g., Tongiorgi et al., 1994).

Notes on Correlation

The lower and upper boundaries of the Silurian and the series boundaries agreed on by the SSS in the 1980s are accepted and followed herein. The correlation chart of Silurian rocks in various parts of China is shown in Fig. 2.

Graptolites and conodonts are the two most important groups for age determination and accurate correlation. Chitinozoans have also been used, in particular in rocks where no graptolites or conodonts are known. Brachiopods are commonly abundant in shallow-water Silurian rocks in China, and their assemblages are age-distinctive and -characteristic in several Silurian stages. Trilobites, corals, nautiloids, and other fossil groups are also useful for correlation (Fig. 3). The various stratigraphic columns of China are shown in Figs. 4–16.

Lower and upper boundaries of the Silurian — The *Hirnantia-Dalmanitina* faunal interval and the *Normalograptus extraordinarius* and *Glyptograptus? persculptus* Zones are uppermost Ordovician. Most of the *Hirnantia-Dalmanitina* faunal interval correlates with the *N. extraordinarius* Zone, and some elements persist into the lower *G.? persculptus* Zone. The late *G.? persculptus* Chron was an interval at the beginning of a global sea-level rise with anoxic water conditions and a warmer climate. These conditions caused shallow-water shelly faunas to go extinct before the Silurian. The base of the *Parakidograptus acuminatus* Zone, with *Akidograptus ascensus*, defines the base of the Silurian. Investigations of the taxonomy of key graptolites and brachiopods within a largely continuous sequence across the Ordovician–Silurian boundary in western Hubei have been carried out by Chen X., Rong J.-y., and others since 1996.

The lowest appearance (FAD) of *Monograptus uniformis* defines the base of Devonian. *Icriodus woschmidti* and *Warburgella rugulosa rugulosa* are also of great significance in determining this boundary globally (McLaren, 1977). *Monograptus uniformis* has been recorded in the lowest Devonian at Qinzhou, Guangxi Province, south China (Wang, 1978), where I. *woschmidti* and *W. rugulosa rugulosa* have not been encountered. *Icriodus woschmidti* occurs in the lower Lochkovian in western Qinling (Wang, 1981b), where *M. uniformis* and *W. rugulosa rugulosa* are not found. A question arises regarding the Silurian–Devonian boundary definition if *M. uniformis* and *I. woschmidti* are not found locally in these boundary beds: is it possible to define the boundary in terms of brachiopod faunas? Based on brachiopods in Silurian–Devonian boundary strata of western Qinling, the uppermost Silurian is characterized by the *Retziella-Atrypoidea* fauna, and the lowest Devonian is characterized by the *Protathyris-Lanceomyonia* fauna (Rong et al., 1987; Fig. 3). The older fauna is not associated with the conodont *I. woschmidti*, but rather with *Ozakordina remscheidensis remscheidensis*, which straddles the Silurian–Devonian boundary in western Qinling (Cao et al., 1987). The *Retziella-Atrypoidea* fauna includes the two eponymous genera associated with *Morinorhynchus*, *Proreticularia*, and others. The *Protathyris-Lanceomyonia* fauna has the two eponymous genera along with *Machaeraria*, *Cratorhynchonella*, *Rhynchospirina*, and others. This fauna first occurs ca. 50–60 m below the lowest *I. woscshmidti* in Yanglugou Valley, Zorge County, northern Sichuan; in Xiawunagou Valley, Tewo County, southeastern Gansu; and about 10 m below *I. woscshmidti* in Putonggou Valley, Zorge County (Fig. 3).

Llandovery–Wenlock and Ludlow–Pridoli boundaries — According to the SSS biostratigraphic scheme, the Llandovery–Wenlock boundary (Fig. 4) is the base of the *Cyrtograptus centrifugus-C. murchisoni* Zone (graptolites). We believe that a distinctive, recognizable, and widespread *Cyrtograptus* assemblage has been split into two parts, and this raises a question whether this boundary definition is practical. It is better to treat the *Cyrtograptus* fauna as a unit within the Wenlock rather than partition it. Of course, some *Cyrtograptus* species with a single cladium, such as *C. lapworthi*, may be regarded as progenitors because they are associated with a latest Telychian graptolites. However, the radiation of *Cyrtograptus* started with *C. insectus*, and the *Cyrtograptus* radiation might coincide with the earliest *C. sakmaricus* Chron, based on reports by Ge and Li (1984) and Fu and Song (1986) on the Ziyang section, southeast Shaanxi. It is also difficult to define the Llandovery–Wenlock boundary in sequences without graptolites. For instance, the boundary cannot be precisely defined even if *Pterospathodus amorphognathoides* is found, because the boundary is within the *P. amorphognathoides* Zone. It is not appropriate to define the boundary of a series within a biostratigraphic zone. Therefore, re-consideration of the Llandovery–Wenlock boundary may be needed by the SSS.

The *Ozakordina crispa* Zone is regarded as the top Ludlow conodont zone (Wang, 1980, 1981b). However, this zone straddles the Ludlow–Pridoli boundary (Walliser and Wang, 1989; Kleffner, 1995). How to define this series boundary in a shelly sequence without graptolites or conodonts is a problem. An example is in the shallow-water brachiopod *Retziella* faunal interval of Asia and Australia (Rong et al., 1995). This fauna is Ludlow–Pridoli. A more precise decision cannot be made with brachiopods if pentamerids are absent (Boucot, 1975). In this

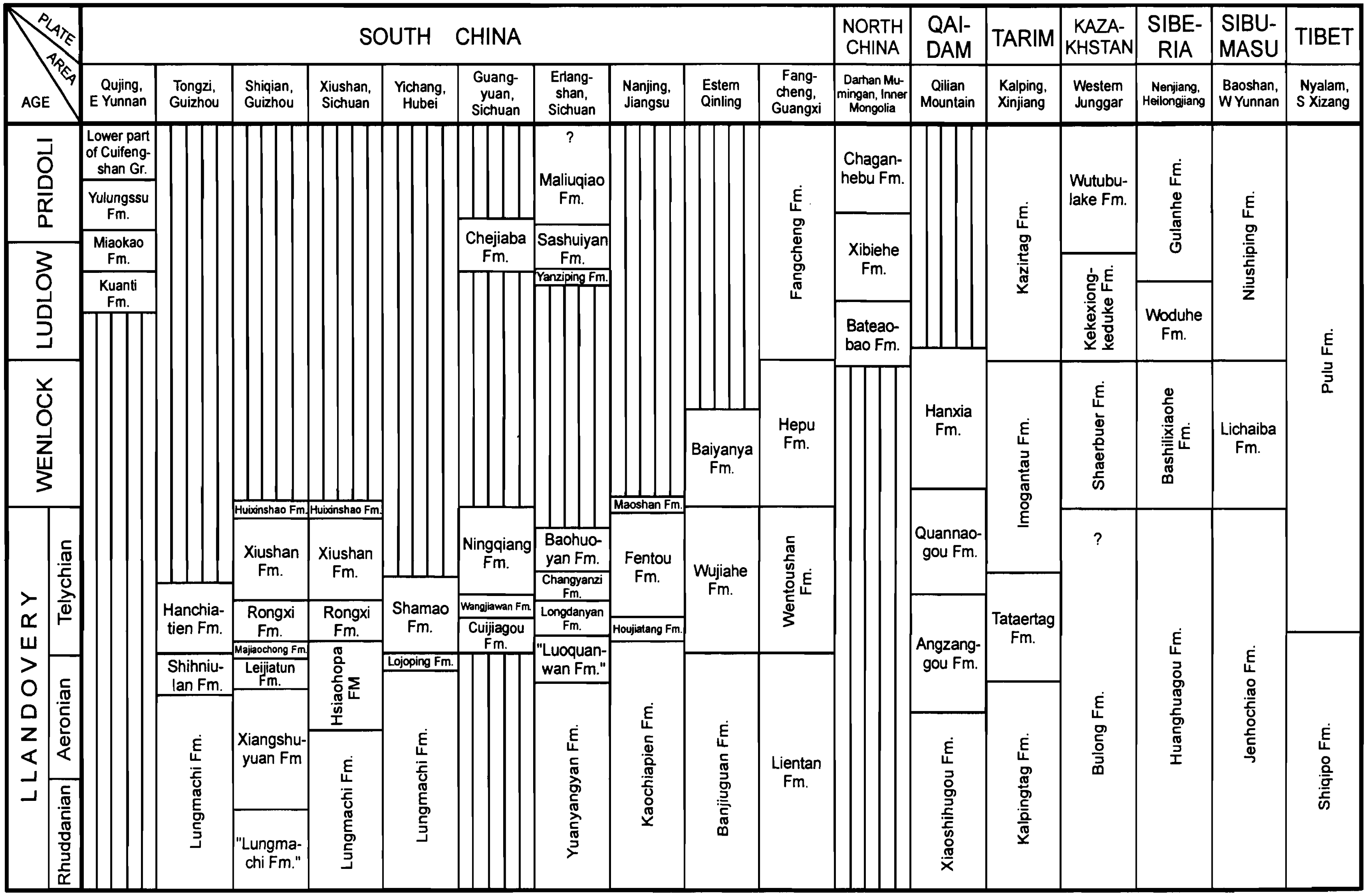

FIGURE 2.–Silurian of various units of China. Lower Cuifengshan Formation in Qujing area, east Yunnan, is upper Pridoli based on thelodont microfossils (Wang N.-Z., 1997).

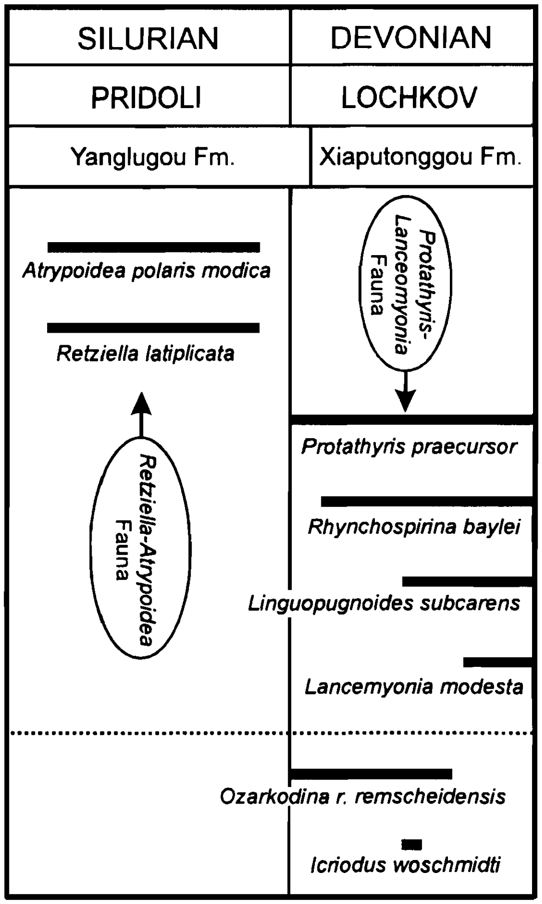

FIGURE 3 — Silurian–Devonian boundary in the west Qinling Mountains based on brachiopods and conodonts.

case, such microfossils as chitinozoans and thelodontids are very useful (Gen et al., 1997; Wang, 1997).

AGE OF THE XIUSHAN FAUNA — The widely distributed, shallow-water Xiushan fauna comes from the middle–upper Xiushan Formation and correlative horizons in many parts of the Yangtze region, including Sichuan, Shaanxi, Guizhou, Hubei, Hunan, Jiangxi, Anhui, and Jiangsu Provinces (Ge et al., 1979; Rong et al., 1990; Chen and Rong, 1996). It is characterized by distinctive brachiopods (e.g., *Atrypoidea lentiformis, Nalivkinia magna, Salopinella minuta, Spinochonetes notata, Xinanospirifer flabellum*), trilobites (e.g., *Coronocephalus, Coronaspis, Senticuculus, Kailia, Parakailia, Rongxiella*), nautiloids (e.g., *Sichuanoceras, Kailiceras, Chuandianoceras, Jialingjiangoceras, Guangyuanoceras, Kionoceras*), and many other fossils (Fig. 4). This fauna was earlier regarded as Wenlock in China. New investigation suggests a middle–late Telychian age (Chen and Rong, 1996). *Monoclimacis griestoniensis?, Oktavites spiralis,* and *Stomatograptus sinensis* have been found in the upper Xiushan Formation of northeastern Guizhou (Fig. 5) and in the upper Ningqiang Formation on the border of Sichuan and Shaanxi (Wang, 1965; Yu et al., 1988; Rong et al., 1990; Chen and Rong, 1996). Conodonts, including *Pterospathodus celloni, P. pennatus, Ozakordina adiutricis, Ambalodus galerus,* and others, are present (Zhou et al. 1981, Wang and Aldridge *in* Chen and Rong, 1996). Associated with the fauna is a low-diversity chitinozoan assemblage characterized by *Angochitina longicollis,* which is upper Telychian but extends into the lower Sheinwoodian of the Wenlock (Paris 1989; Geng et al., 1997). Three higher chitinozoan zones, the *Conochitina visbyensis* (lowest Wenlock), *Lambdachitina tabernaculifera* (latest Wenlock), and *L. crassispina* (middle Ludlow) Zones, are defined in the basal, lower, and upper Xiaoxiyu Formation at Zhangjiajie, Dayong County, northwest Hunan (Geng et al., 1997). It should be pointed out that such elements as *Coronocephalus* sp., *Coronaspis rex* (Grabau), *Salopinella minuta* Rong et al., and *Nalivkinia magna* Yang and Rong of the Xiushan fauna persist from the lower part of the formation at this section (Wang et al., 1988).

AGE OF SHALLOW-MARINE RED BEDS — Shallow-marine red beds are well developed in the Silurian of South China, Qaidam, and Tarim, and they are characteristic of the Silurian across China. They are usually purple or red and locally interbedded yellow–green mudstones, silty mudstones, siltstone, and fine-grained sandstone of a very shallow-water origin (mainly assigned to BA1–2) (Rong et al., 1984; Johnson et al., 1985). Rare macrofossils in associated shallow-marine red beds, such as the brachiopods *Lingula, Nalivkinia,* and *Nucleospira,* are known. Because there is no precise biostratigraphic control, their ages have not been defined. For example, the Rongxi Formation (so-called Lower Red Beds) in the Yangtze region was correlated with the Wenlock (Ge et al., 1979) or middle Telychian (Mu et al., 1982). However, chitinozoans (Geng, 1990; Geng et al., 1997) suggest that the Rongxi Formation, with *Ancyrochitina brevicollis* and *Conochitina daozhenensis,* correlates with the *Spirograptus turriculatus-Monograptus crispus* Zone. Zhou et al. (1981) discovered endemic conodonts (such as *Spathoganathodus parahassi*) in the Rongxi Formation in northeastern Guizhou. Wang C.-y. (*in* Rong et al., 1990) later found *S. parahassi* associated with *Hadrognathodus staurognathoides* in the basal limestone bed of the Wangjiawan Formation in Ningqiang, southwest Shaanxi (Chen et al., 1991; Fig. 6)). This indicates a rough correlation with the Rongxi Formation and an early Telychian age for the shallow-marine red beds.

The Huixingshao Formation (so-called Upper Red Beds) in many places of the Yangtze region is upper Telychian (Rong et al., 1990; Chen and Rong, 1996). Chitinozoans from the lower Xiaoxiyu Formation in an interval correlative with the Huixingshao Formation are lower

AGE	STAGE	GRAPTOLITES	CONODONTS	CHITINOZOA	BRACHIOPODA	
PRIDOLI	Stages not defined yet	*Monograptus? transgrediens* *Monograptus bouceki*	*O. eosteinhornensis*		*Retziella latiplicata-Atrypoidea polaris modica*	*Lingula*
		Monograptus? ultimus	*Ozarkodina crispa*	*Fungichitina kosovensis*	*Retziella minor* *Howellella tingi* *Atrypoidea foxi*	
LUDLOW	Ludfordian	*Saetograptus fritschi linearis*	*Polygnathoides siluricus*	*Angochitina sinica*		*Tuvaella gigantea*
				Grahnichitina philipi		
	Gorstian	*Lobograptus scanicus*	*Ancoradella ploeckensis*	*Lambdochitina crassispina*	*Kirkidium knighti* *Conchidium*	
		Lobograptus progenitor	*Kockella variabilis*			
		Neodiversograptus nilssoni				
WENLOCK	Homerian	*Colonograptus ludensis*	*Ozarkodina bohemicus*	*Lambdochitina tabernaculifera*	*Retziella spiriferoides-Sphaerirhynchioides*	*Tuvaella rackovskii*
		Colonograptus praedeubeli				
		Cyrtograptus lundgreni	*O. sagitta sagitta*		*Hesperorthis Resserella* *Leptaena* cf. *depressa* *Lissatrypa hoboksarensis* *Gotatrypa sarburtensis* *Spirigerina sarburtensis* *Eospirifer radiatus*	
	Shein-woodian		*O. sagitta rhenana*	*Cingulichitina cingulata*		
		Monograptus cf. *riccartonensis*		*Ancyrochitina ansarviensis*		
		C. centrifugus-C. murchisoni				
LLANDOVERY	Telychian	*C. insectus-C.sakmaricus*	*P. amorphognathoides*			
		O. spiralis-S. grandis	*Pterospathodus celloni*	*Angochitina longicollis*	*N. magna-Xinanospirifer flabellum* *Salopinella minuta-Spinochonetes notata*	
		Monoclimacis griestoniensis				
		S. turriculatus-M. crispus	*S. Parahassi-S. guizhouensis*	*Plectochitina brevicollis*	*Nalivkinia elongata-Nucleospira calypta*	
	Aeronian	*Stimulograptus sedgwickii*	Interval A	*Conochitina truncata*	*Paraconchidium shiqianensis*	*Apopentamerus dorsoplanus*
		Demirastrites convolutus	*Spathognathodus obesus*	*Conochitina rossica*	*Zygospiraella duboisi-Spirigerina sinensis*	
		Demirastrites triangulatus		*Conochitina rectangularis*		
	Rhuddanian	*P. (Coronograptus) cyphus*		*Conochitina electa*	*Eospirifer sinensis-Beitaia modica*	*Borealis borealis*
		Cystograptus vesiculosus		*Belonechitina postirobusta*		
		Parakidograptus acuminatus		*Ancyrochitina nodifera*	*Hindella meitanensis*	

FIGURE 4 — Correlation of Silurian graptolite, conodont, and chitinozoan zones with brachiopod, trilobite, nautiloid, and coral assemblages of China.

Wenlock (Geng et al., 1997).

Shallow-marine red beds in the Qujing area, eastern Yunnan, southwest China have upper Ludlow conodonts (*Ozarkodina crispa* in the upper Kuanti Formation; Wang, 1981b) and brachiopods (*Retziella* fauna in the lower Kuanti; Rong and Yang, 1981 Fig. 7).

Shallow-marine red beds of the Hanxia Formation in the northern marginal belt of the Qaidam paleoplate are underlain by the Quannaogou Formation (Fig. 8), which has a coral fauna very similar to that of the upper Telychian Ningqiang Formation on the Shaanxi–Sichuan Province border (Deng Z.-q., personal commun., 1991). We believe that the Hanxia Formation is probably Wenlock, rather than Ludlow–Pridoli, as generally reported (e.g., Lin et al., 1984).

The Tataertag, Imogantau, and Kaziltag Formations have shallow-marine red beds in the Kalping–Bachu area in the western marginal belt of the Tarim paleoplate (Fig. 9). These formations have been considered to be later Silurian–Devonian. However, a conodont assemblage with *Ozakordina* sp. cf. *O. edithae* from the Imogantau Formation (Zhang and Wang, 1995) is upper Llandovery–lower Wenlock, and a chitinozoan assemblage with *Cingulochitina wronai* from the Kazirtag Formation is upper Ludlow–lower Pridoli. A discussion on the age of the associated vertebrates is given below.

AGE OF AGNATHANS AND SINACANTH FIN SPINES — Agnathans (e.g., *Hanyangaspis*) and acanthodians (e.g., fin-spines from *Sinacanthus* and *Neoasiacanthus*) from the Xiushan Formation and correlative units in South China were usually regarded as late Silurian or even Devonian by many Chinese paleontologists (e.g., Pan, 1963). These fossils have also been discovered in the underlying Rongxi Formation in northwest Hunan, South China (Zeng, 1988). Invertebrates from the Xiushan Formation (noted above) indicate that these fish remains are chiefly Telychian. Sinacanth spines, regarded as the oldest known shark-fin spines (Zhu, 1998) are found in the Tataertag and Imogantau Formations of the Kalping–Bachu area, northwestern Tarim, Xinjiang (Wang et al.,

TRILOBITES	NAUTILOIDS		CORALS	
			RUGOSE	TABULATES
Encrinurus tumida *Glossoproetus perconvevus* *Warburgella rugulosa sinensis*	*Yunnanoceras* *Enthyrocycloceras*	*Michelinoceras -Kopaninoceras*	*Denayphyllum-Chavsakia* *Entelophyllia-Pilophyllum*	*Emmonsiella-Squameofavosites* *Laceripora,* *Palaeocorolites,* *Squameolites*
Acanthopyge orientalis	*Heyuncunoceras*		*Micula-Ketophyllum*	*Parastriatopora,* *Syringopora,* *Squameofavosites*
Encrinurus *Sphaerexochus*			*Amplexoides,* *Dinophyllum,* *Fletcheria,* *Gyalophyllum,* *Holmophyllum,* *Pseudamplexus*	*Antherolites,* *Catenipora,* *Halysites,* *Multisolenia*
Coronocephalus-Kailia-Rongxiella	*Sichuanoceras*		*Shenxiphyllum-Idiophyllum*	*Erlangbapora-Carnegiea*
Hypaproetus-Astroproetus (low diversity)				
Songkania-Shiqiania with *Hypaproetus,* *Astroproetus, Ptillaenus*	*Yichangoceras* *Songkanoceras*		*Kodonophyllum-Trypacistiphyllum* *Brachyelasma* *Streptelasma* *Amplexoides*	*Meitanopora-Parafletcheria* *Eoroemerites* *Tetraporella-Troedssonites*
Bumastus, Encrinuroides, *Scutellum* *Leonaspis*	*Mixosiphonoceras* *Kionoceras* *Songkanoceras*		*Brachyelasma* *Grewingkia* *Tongxinophyllum* *Paramplexoides* *Tungussophyllum*	*Proheliolites*

FIGURE 4 *continued.*

1988; Liu, 1995; Wang et al., 1996; Zhu, 1998); these formations were earlier regarded as Devonian. Liu (1995) assigned both formations to the Wenlock, but correlated them with the Fentou Formation in the lower Yangtze region, a unit that correlates with the Telychian Xiushan Formation in the upper Yangtze region (Rong, et al., 1990; Chen and Rong, 1996).

Lithofacies and Biofacies

The many separate Silurian paleoplates or blocks resulted in diversified and intricate sedimentary facies and biofacies across China, particularly in the marginal belts. The lithofacies belts were incompletely preserved or disappeared after fragmentation in many areas near strike-slip faults (Fig. 1). Chen and Rong (1996) recognized five major Telychian litho- and biofacies belts in the Yangtze region. In this report, these belts are also recognized in other areas of China.

Non-Volcanic Siliciclastics

Black graptolitic shale facies – Early Llandovery black graptolite shales were widely deposited in marine basins and on shelves after deglaciation and with global transgression. The shales are widespread on the Yangtze platform (Lungmachi Formation in the upper Yangtze and Kaochiapien or Lishuwo Formations in the lower Yangtze regions; Mu et al., 1986; Zhang, 1989). They appear in the Zhujiang Basin (Lientan Formation; Mu, 1948); the northern Tibet platform (Dewukaxia Formation; Ni, 1986); the northern slope of the upper Yangtze platform (Maliushuwan and Banjiuguan Formations; Fu and Song, 1986); western Yunnan (lower Jenhochiao Formation; Mu, 1959; Ni et al., 1982); and in the northern mobile belt of the North China paleoplate (Taoshan Formation; Liu, 1985). Two types of black graptolitic shale are recognized. The shales in the Yangtze and Tibet regions were deposited on a shallow epicontinental platform, whereas the others are in deeper-water, slope or basinal turbidites.

Leggett (1980) concluded that the black shale probably originated in a restricted basin or in an oxygen-minimum zone. Phytoplankton productivity in modern oceans is concentrated in shelf waters (Deuser, 1975). The oxygen-minimum zone might tend to be better developed during transgression (e.g., Landing et al., 1992). According to Rickards (1964), the black graptolitic shale in British successions contains graptolites to the virtual exclusion of all other fossils, except for planktic or pseudoplanktic organisms. These black shales and mudstones comprise clay- to silt-grade quartz, altered feldspar, mica, iron minerals, carbonates, and clay minerals. A similar mineral composition is also found in Chinese graptolitic black shale. Therefore, there is no significant difference in black shale composition between the shallow- and deep-water types. Reading (1978) defined all black graptolitic shales as pelagic or hemipelagic.

The black graptolite shale is obviously rich in organics. The Fe^{3+}/Fe^{2+} ratio is controlled by the oxidation state, which is ultimately controlled by the amount of organic matter (McBride, 1974). Indeed, all colors in shales are ultimately controlled by the amount of organic matter (Potter et al., 1980). Shales seem divisible into two series, a red–purple–greenish gray series based on Fe^{3+}/Fe^{2+} and a greenish gray-black series based on carbon content (e.g., Landing et al., 1992).

Shelf or platform black graptolite shale — The black, graptolitic, uppermost Ordovician Wufeng Shale (Ashgillian) is overlain by the Lungmachi and Kaochiapien Formations (Fig. 10). This overlying black shale sequence extends from lowest Rhuddanian to middle or upper Aeronian. Laminated organic matter arranged par-

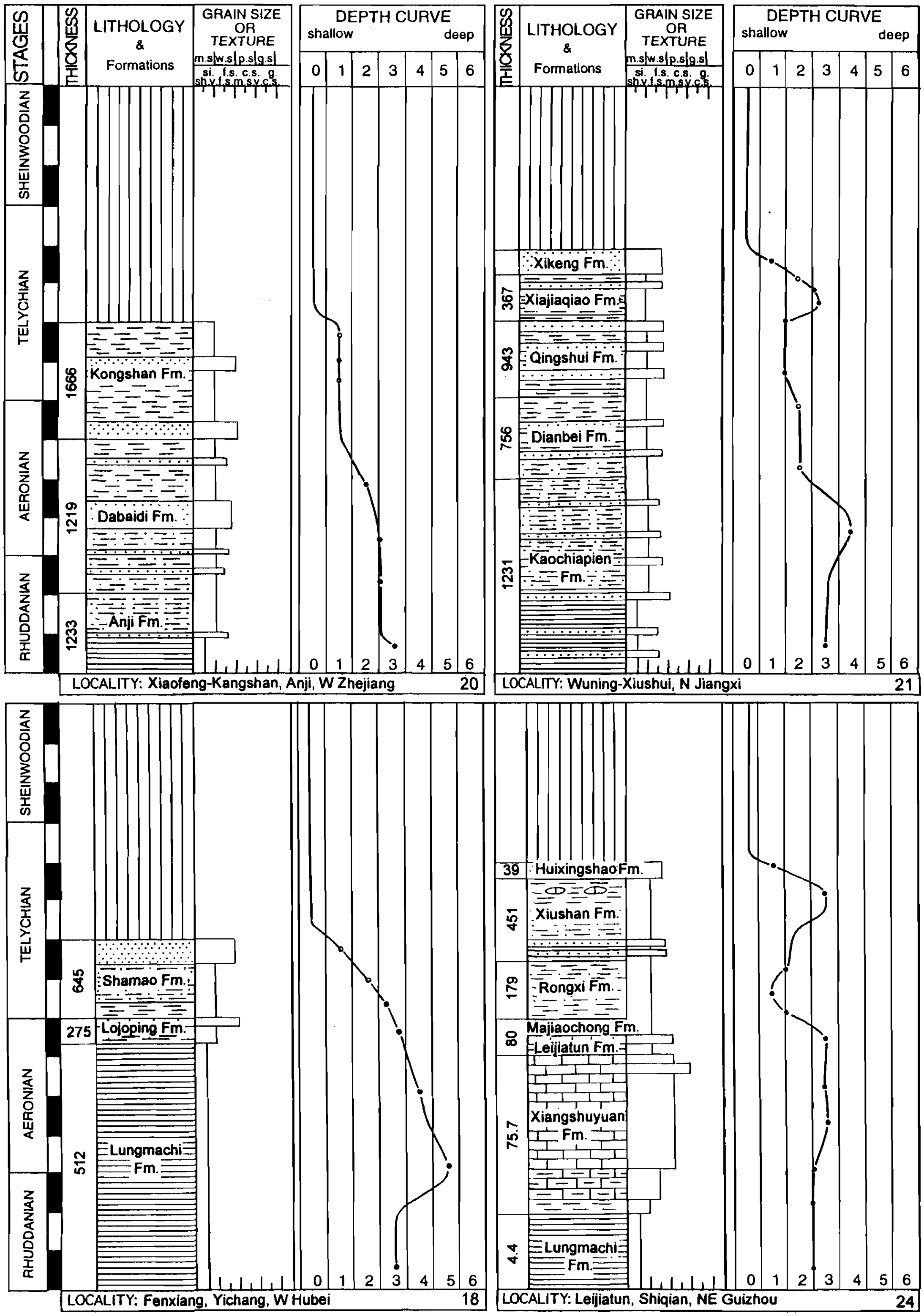

FIGURE 5 — Silurian at Xiaofeng and Kangshan villages, Anji County, west Zhejiang Province; Wuning and Xiushui Counties, north Jiangxi Province; Fengxiang village, Yichang City, west Hubei Province; and Leijiatun village, north of Shiqian County, northeast Guizhou Province. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

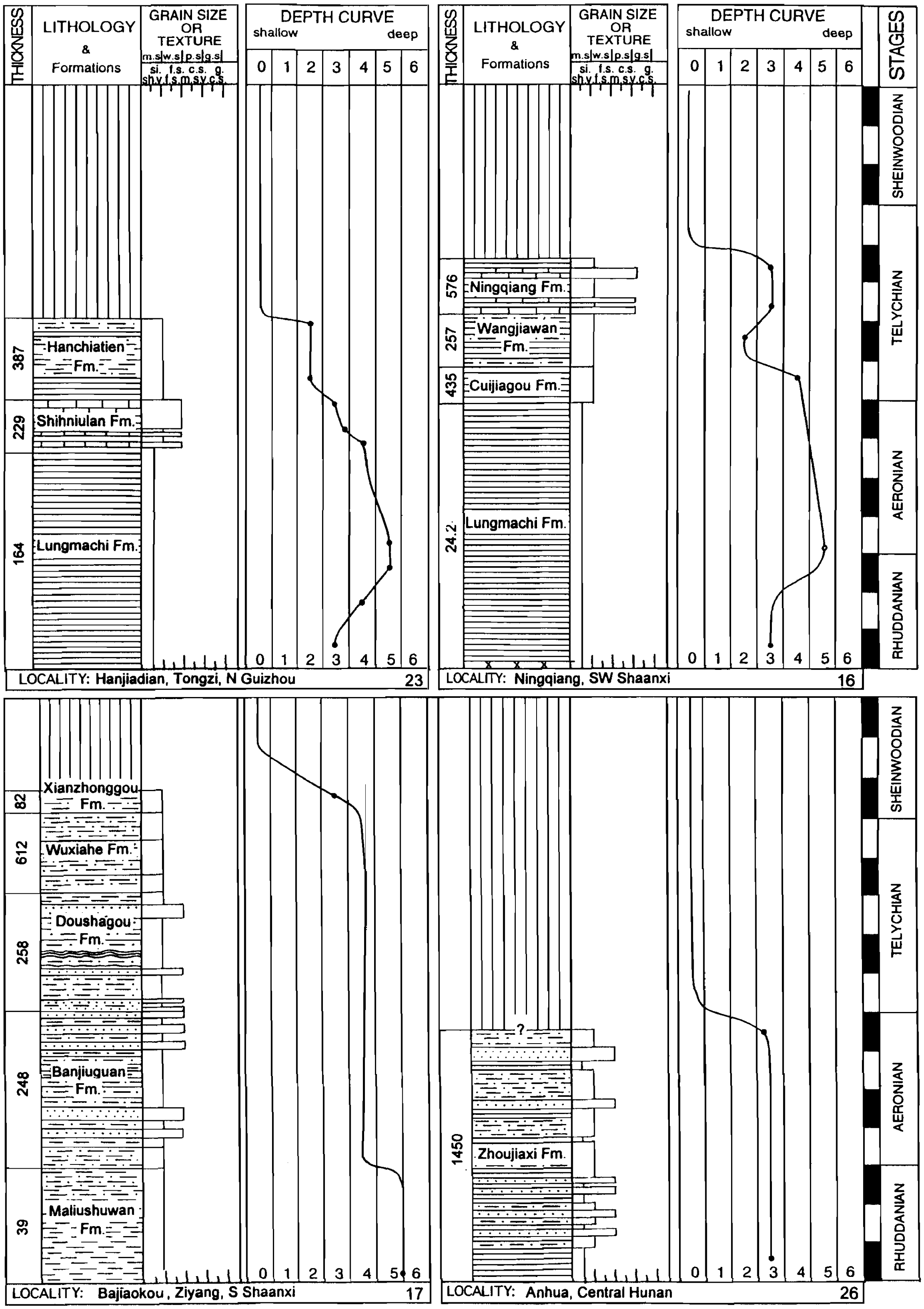

FIGURE 6 — Silurian at Hanjiadian, Tongzi County, north Guizhou Province; Ningqiang County, southwest Shaanxi Province; Bajiaokou Valley, Ziyang County, south Shaanxi Province; and Anhua County, central Hunan Province. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

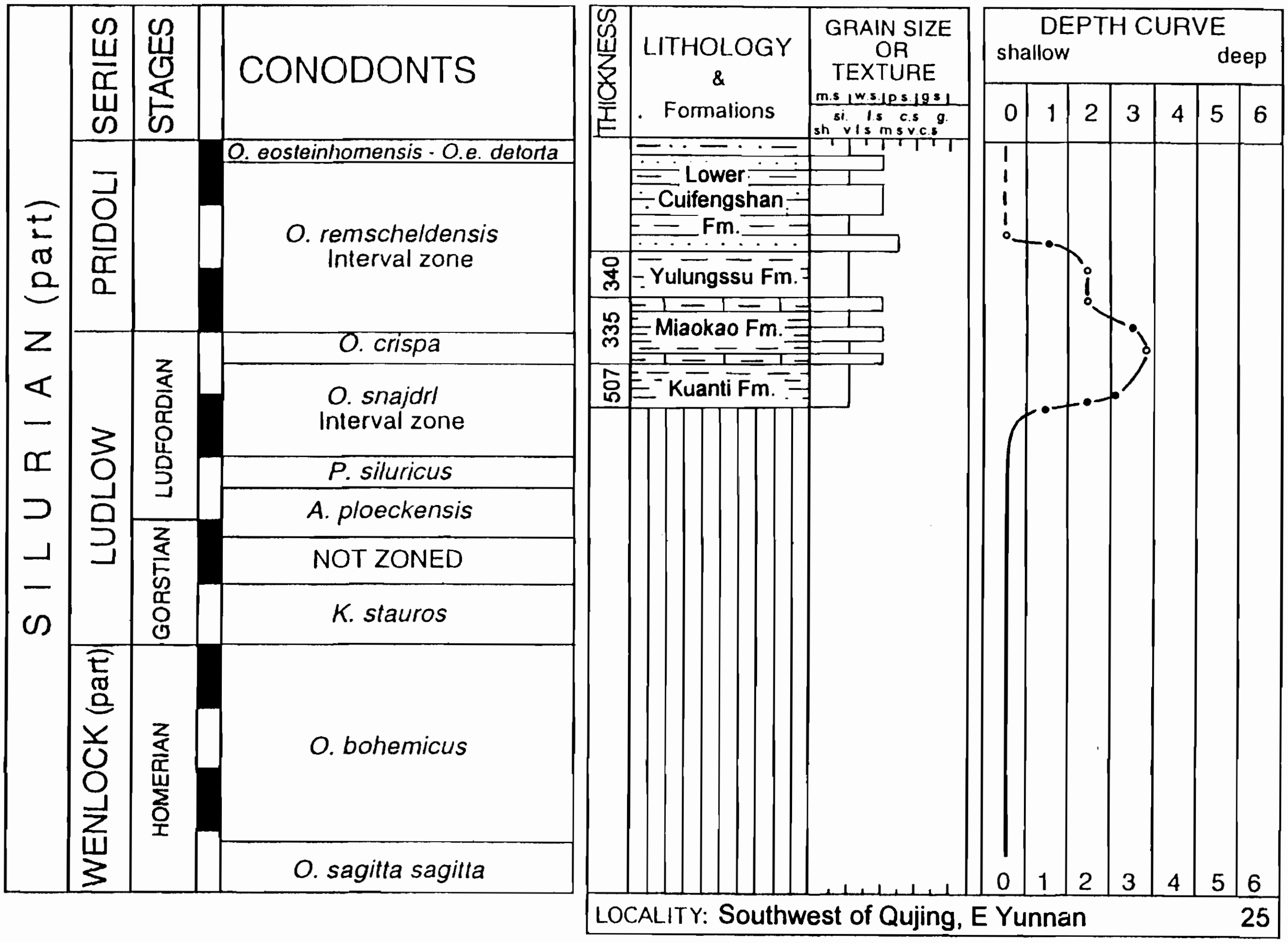

FIGURE 7 — Silurian at Hongmiao, southwest of Qujing County, east Yunnan Province. Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

allel to the bedding planes is rich in the lower parts of these units, and is intercalated with siliceous beds. In some lower Yangtze region localities, organic matter is usually concentrated in the *Cystograptus vesiculosus* and *Pristiograptus (Coronograptus) cyphus-Monoclimacis lunata* Zones. Because of the absence or limited abundance of land plants in this time interval, the organic material must have been derived from algal sources at the base of the food chain. No carbonates are in these black shales. However, a few siliceous carbonate concretions or nodules are present within the Lungmachi Formation (*Demirastrites convolutus* Zone), and isolated graptolites have been recovered from these concretions (Chen, 1986).

The Rhuddanian–Aeronian black shale in the Yangtze region may be related to early Llandovery marine transgressions. However, the restricted Yangtze embayment in the Ashgillian did not completely disappear in the Early Silurian. The Cathaysian (including the Qianzhong Oldland), Chengdu, and Hanzhong Oldlands still existed around the Yangtze platform. The restriction of land area as a result of transgression tends to increase stratification of the water column, with decreased bottom water circulation (Leggett, 1980; Landing et al., 1992). The early Llandovery marine transgression increased primary productivity by algae and other producers and caused an expansion of the oxygen-minimum zone. The expansion or development of this oxygen-minimum zone in this time interval was widespread. It affected the Yangtze and Tibet platforms; the deep-shelf, slope, and basinal facies of the Zhujiang Basin (e.g., central Hunan); the western marginal shelf (e.g., Erlangshan); northern slope of the Yangtze Platform (e.g., Ziyang area); western Yunnan; and northern mobile belt of the North China paleoplate. It placed bottom waters below the reach of

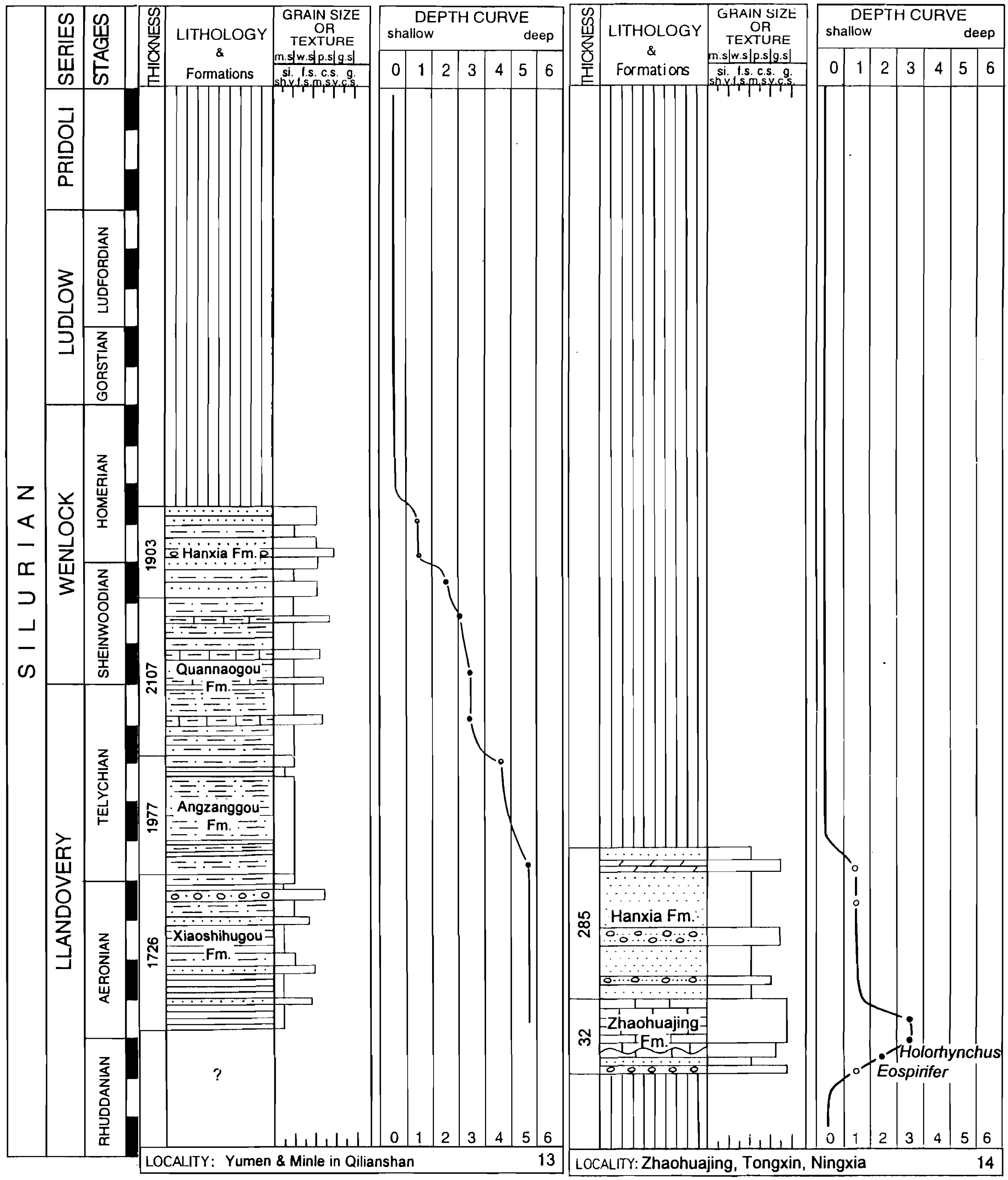

FIGURE 8 — Silurian of Qilianshan Mountain, Yumen and Minle Counties, Gansu Province, and Zhaohuajing, Tongxin, central Ningxia. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

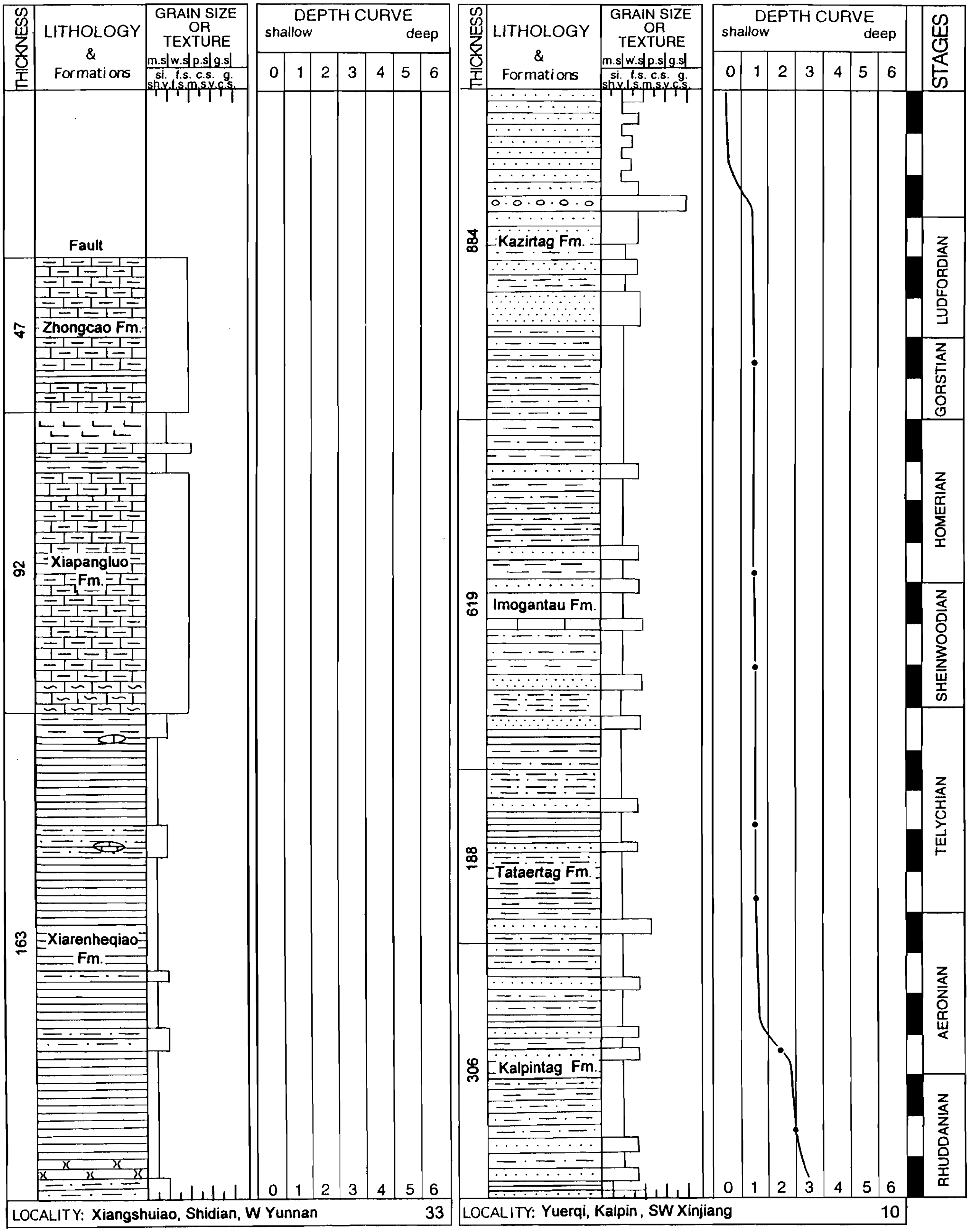

FIGURE 9 — Silurian at Xiangshuiao, Shidian County, west Yunnan Province, and Yuerqi, Kalpin County, southwest Xinjiang Province. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

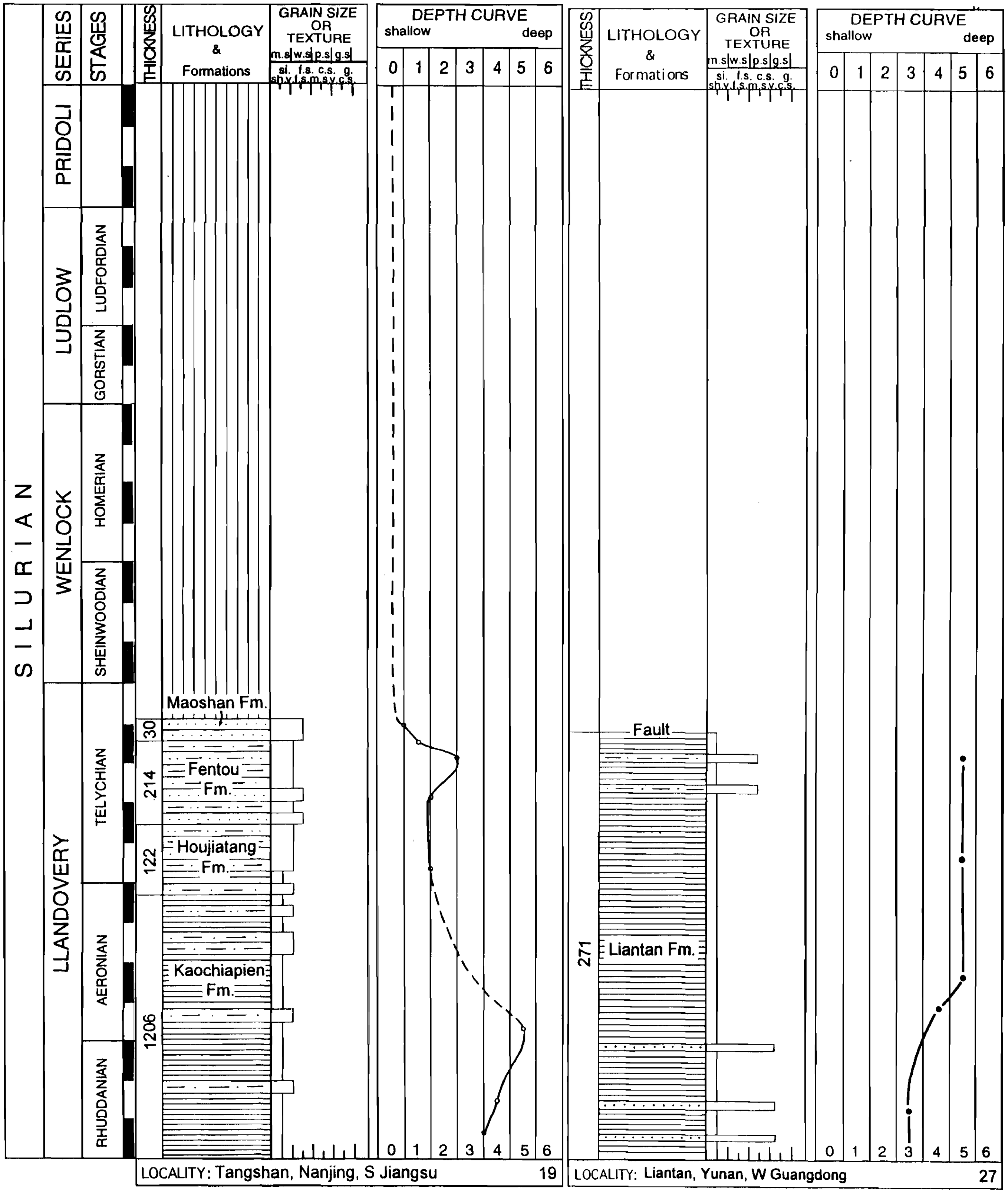

FIGURE 10 — Silurian near Tangshan village, Nanjing City, south Jiangsu Province, and Liantan village, Yunan County, west Guangdong Province. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

wind-driven vertical circulation, and the bottom water became anoxic.

TURBIDITIC BLACK SHALE — Turbidites and black graptolitic shales are well developed in the Ziyang–Lan'gao area on the northern slope of the upper Yangtze Platform (i.e., Banjiuguan Formation, upper Llandovery [Telychian]–Wenlock; Fig. 2). The turbidites prograded northeast. They are best developed in the Bajiaokou section at Ziyang County, southern Shaanxi (Zhao, 1987). The Rhuddanian–Aeronian Maliushuwan Formation features outer-fan turbidite deposits. The Telychian lower Banjiuguan Formation consists chiefly of Bouma CD turbidites, and represents channel deposits on the middle parts of submarine fans. Black graptolite shale usually occurs as the E units at the top of Bouma sequences. The limited occurrences of E-unit black shales might be confined to a zone of upwelling along the west- or northwest-facing margins of the Yangtze platform. The Wenlock upper Banjiuguan Formation suggests the disappearance of turbidites.

SHALLOW GRAY, GREEN, AND YELLOW SILICICLASTICS — This facies was widely distributed in many areas of China in the Silurian. It is characterized by grayish-yellow or grayish-green, very fine-grained (usually shale, mudstone, silty mudstone, argillaceous siltstone, or siltstone) siliciclastics with a small amount of sandstone. The Majiaochong, Hsiaohopa, Hanchiatien, and Xiushan Formations and correlative units on the Yangtze platform are assigned to this facies. They were rapidly accumulated and, in particular, occur in a southern (northeast Guizhou, southeast Sichuan, and northwest Hunan) and a northern (north Sichuan and south Shaanxi) depression. Abundant brachiopods, bivalves, gastropods, trilobites, nautiloids, crinoids, graptolites, chitinozoans, conodonts, fish, machaeridians, and a few other groups occur in this facies. Analysis of community ecology and sedimentology allows environmental interpretations. For example, brachiopods (*Salopinella, Spinochonetes, Nalivkinia, Striispirifer, Xinanospirifer*), trilobites (*Coronocephalus, Rongxiella, Kailia*), nautiloids (*Sichuanoceras*), graptolites (*Stomatograptus sinensis*), conodonts (*Pterospathodus celloni*), and chitinozoans (*Angochitina longicollis*) in the Xiushan Formation suggest a BA 3 facies (Rong et al., 1984; Johnson et al., 1985). A low-diversity BA 2 assemblage in the Majiaochong Formation has rare brachiopods (*Nalivkinia and Nucleospira*), benthic graptolites (*Hunanodendrum*), and a few gastropods and bivalves (Rong et al., 1984; Johnson et al., 1985). Its near-shore distribution supports this conclusion. These representative facies were deposited in epicontinental seas on the Yangtze platform (Fig. 2).

Chen et al. (1988) analyzed the depositional environment of the Telychian Fentou Formation in southern Jiangsu, eastern China, and pointed out that these terrigenous, muddy, siliceous, non-calcareous siliciclastics probably represent estuarine, deltaic, or littoral environments. Muscovite with compressional deformation is common, and suggests rapid accumulation.

SHALLOW-MARINE RED SILICICLASTICS — Widespread Silurian marine red beds were reviewed by Ziegler and McKerrow (1975) in the British Isles, Norway, and the Appalachian Basin. They concluded that Silurian red beds originated in deep-shelf, basinal, and ocean-floor environments, and that red color can not be taken as proof that sediments accumulated in non-marine or shallow-water environments. New investigation of the paleogeographic distribution, lithology, and community associations of Silurian marine red beds, such as the Rongxi and Huixingshao Formations, suggests that they are all of near-shore, shallow-water origin.

Silurian marine red beds are particularly widespread in the Yangtze region. They are composed of purple-red mudstone and siltstone intercalated with gray, yellow, or green mudstone and siltstone, and are chiefly Telychian (Llandovery) and upper Ludlow. Telychian red beds are well exposed close to the Cathaysian Oldland. They include the Rongxi and Huixhingshao Formations of northeast Guizhou, southwest Sichuan, and northwest Hunan in the upper Yangtze region. Other units include the Qingshuihe and Xikeng Formations of north Jiangxi; the Houjiatang and Maoshan Formations of south Jiangsu; and the Tangjiawu Formation of west Zhejiang in the lower Yangtze region. Telychian red beds also occur in the Kaochaitien Formation of the Guiyang area in central Guizhou near the Qianzhong Oldland, in part of the Hanchiatien Formation in the Huayinshan area, and in the Wangjiawan and Ningqiang Formations of the Guangyuan–Ningqiang area near the Chengdu Oldland. The lithology and mineralogy of the red beds in these areas are almost identical, and all have illite, montmorillonite, chlorite, and quartz.

Fossils are rare in the Rongxi Formation. The benthic faunas are low diversity and include brachiopods (*Nalivkinia-Nucleospira* Community; Rong, 1986), trilobites (*Encrinuroides, Astroproetus*), bivalves (*Modiolopsis, Leptodesma*), and benthic graptolites (*Hunanodendrum*). Low-diversity chitinozoan assemblages, with such forms as *Plectochitina brevicollis*, are recorded by Geng (*in* Chen and Rong, 1996). Some ichnofossils are present. Associated with the fossil-bearing horizons or in under- or overlying beds are common shallow-water ripple marks and cross-bedding. Assignment of these red beds to a BA 1–2 habitat was suggested by Rong et al. (1984) and Johnson et al. (1985).

Fossils in the Huixingshao Formation are rare. Only a few specimens of bivalves (*Modiomorpha, Praecardium*), gastropods (*Turbocheilus, Discordichilus*), and pteropods (*pterigotids*) are reported (Ge et al., 1979). Rare trilobites (*Coronocephalus*) and corals (*Tryplasma*) can be present. Distribution of the Huixingshao Formation and its benthos indicate an assignment to BA 1.

The Shengxuanyi Member of the Ningqiang Formation crops out in the Guangyuan–Ningqiang area on the northwest margin of the Yangtze region. This partial correlative of the Huixingshao Formation is composed of red marine beds and yellowish-green mudstone and siltstone intercalated with limestone. These beds yield abundant BA 2–3 faunas, and are definitely deeper than those of the Huixingshao Formation (Chen and Rong, 1996). In the lower Yangtze region, the late Telychian Maoshan and Tangjiawu Formations consist of purple-red siltstone and sandstone interbedded with yellowish-green siliciclastics with few marine fossils. These latter two formations probably record delta or even non-marine deposits (Mu et al., 1986; Chen and Rong, 1996).

In conclusion, the mainly shallow-water Telychian red beds are sedimentologically and paleogeographically different from the European and American deeper shelf, basin, and ocean floor facies discussed by Ziegler and McKerrow (1975). Features associated with their deposition include: 1) a suitably oxidized source area for the supply of a large amount of siliciclastics; 2) stable, probably warm, dry climates with deep weathering that produced fine-grained siliciclastics rich in Fe203; 3) very low organic productivity in areas of red bed accumulation; and 4) a regressive pulse with development of shallow-water, relatively quiet marine environments.

Shallow-marine red beds are also known from the upper Kuanti Formation (upper Ludlow) of Qujing, eastern Yunnan, and the Tataertag, Imogantau, and Karziltag Formations (upper Llandovery–Pridoli) of the Kalping–Bachu area, western Tarim Basin, western Xinjiang. Other examples include the "Hanxia Formation" (Aeronian–Telychian) of Tongxin, southern Ningxia; the Hanxia Formation (Wenlock–Ludlow) of the Yumen area of Gansu Province in the northern marginal belt of the Qaidam paleoplate; and the Ludlow–Pridoli of the Greater and Lesser Khingan Mountains in the southern marginal belt of the Siberian paleoplate. However, sedimentological study of these red beds has not been carried out. Nevertheless, it should be pointed out that no deep-water marine red beds have been encountered in China.

Carbonate Facies

Silurian carbonates are not common in the Chinese paleoplates or blocks. Only recently have some types of carbonates, particularly reef facies, been investigated on the upper Yangtze. There are three types of carbonate facies in China.

Shallow platform carbonates — Typical platform carbonates are known mainly from the later Silurian of the Tibet platform (Fig. 11) and the western Yunnan belt (Ni et al., 1982; Lin et al., 1984; Ni, 1986). Silurian carbonates of Tibet are reviewed by Mu and Chen (1984) and Rao et al. (1988) and are known in three regions: the Himalayan, Gangdise, and Karakorum–Tuanggula regions, from south to north. In the Himalayan region, the Wenlock–Pridoli is limestone. Only low-diversity conodont and nautiloid faunas are known from these later Silurian strata. In the Gangdise region, the Wenlock Zhanongema Formation (32.1–76.5 m) and the Ludlow–Pridoli Mendeeyue Formation (26.5–167.4 m) consist mainly of dolomitic limestone with low-diversity conodonts. In Baingoin in the south Karakorum–Tuanggula region, metamorphosed dolomitic limestone is developed in possible Wenlock or higher strata of the Keermu Group. In the northwest Karakorum–Tuanggula region, possible Telychian–Wenlock limestone of the upper Puercuo Group (300 m) with no fossils are reported (Rao et al., 1988). Nautiloids and conodonts are relatively common, but such benthic forms as brachiopods and corals are rare, and the faunas indicate probable deeper-water conditions.

The Tibet platform consists of three regions (described above) and occupies a huge area of about 10^6 km^2. The actual extent of this platform, however, may have been even larger, because its northern margin was probably consumed by convergence with the Tarim paleoplate. The distribution of central dolomitic and marginal limestone facies on the Tibet platform is similar to that on the North China platform in the Ordovician. This carbonate distribution pattern is similar to that noted in Laurentia in the Silurian (Berry and Boucot, 1970) and Lower Ordovician (Berry, 1972). The thin carbonates of the Tibet platform may indicate slow depositional rates. This is also similar to that of the Ordovician carbonates in North China. Reports of low-diversity shelly faunas from the marginal limestones in Tibet may be due to incomplete investigations and difficult working conditions.

Earlier Silurian (mainly Aeronian–Telychian) shallow platform carbonates also occur in South China (e.g., Ge et al., 1979; Jin et al., 1989; Rong et al., 1990; Chen and Rong, 1996). Aeronian shallow carbonates are well developed in north Guizhou, south Sichuan, and west Hubei. Basic data from the Aeronian carbonates in the upper Yangtze region are described below. Telychian shallow-water carbonates are sporadically developed, and they are usually represented by a few thin-bedded limestones or limestone lenses in many places of the region (Fig. 12). However, relatively thick-bedded limestone is developed in some areas, such as the Ningqiang–Guangyuan area in

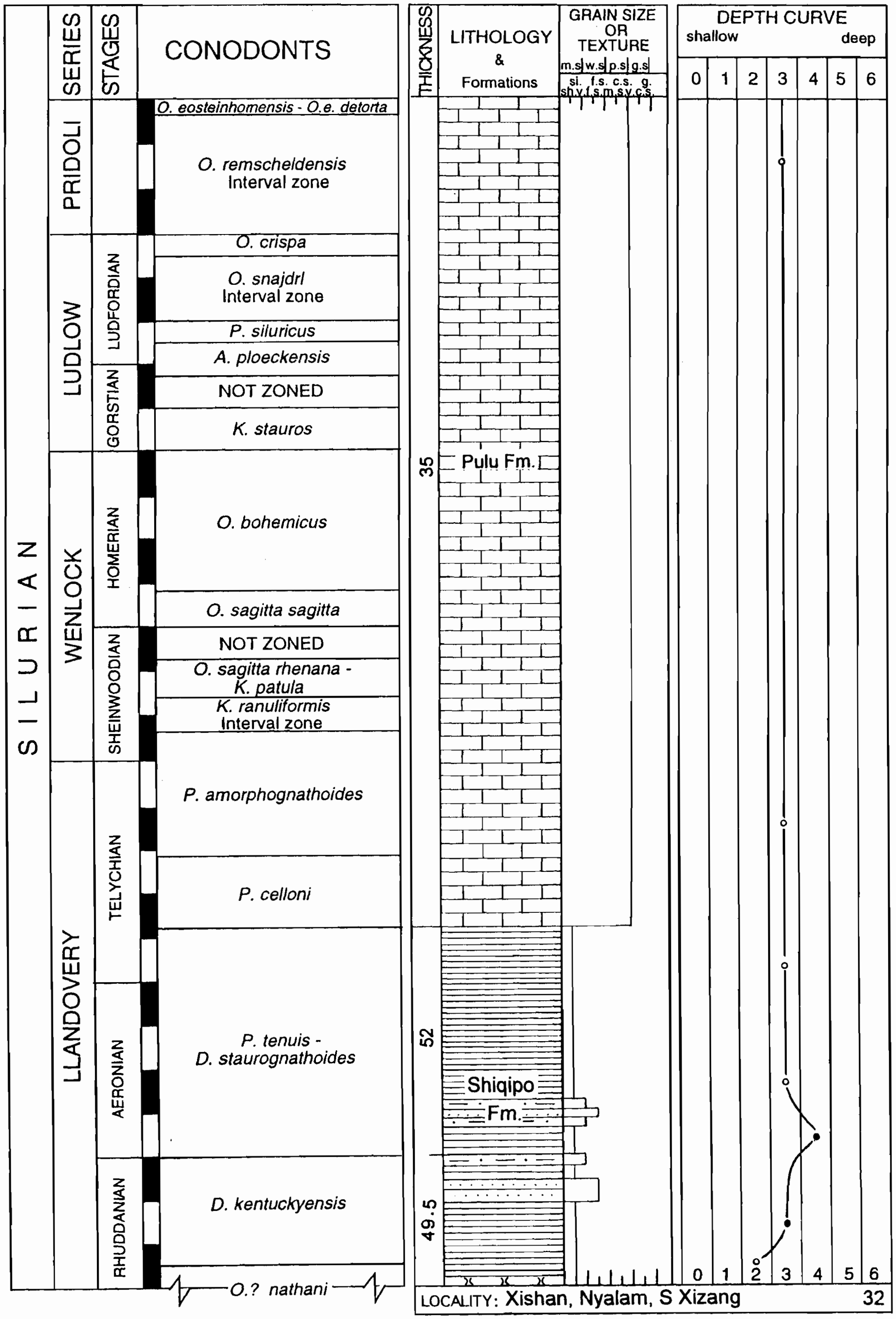

FIGURE 11 — Silurian at Xishan Hill, Nyalam County, south Xizang (Tibet). Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

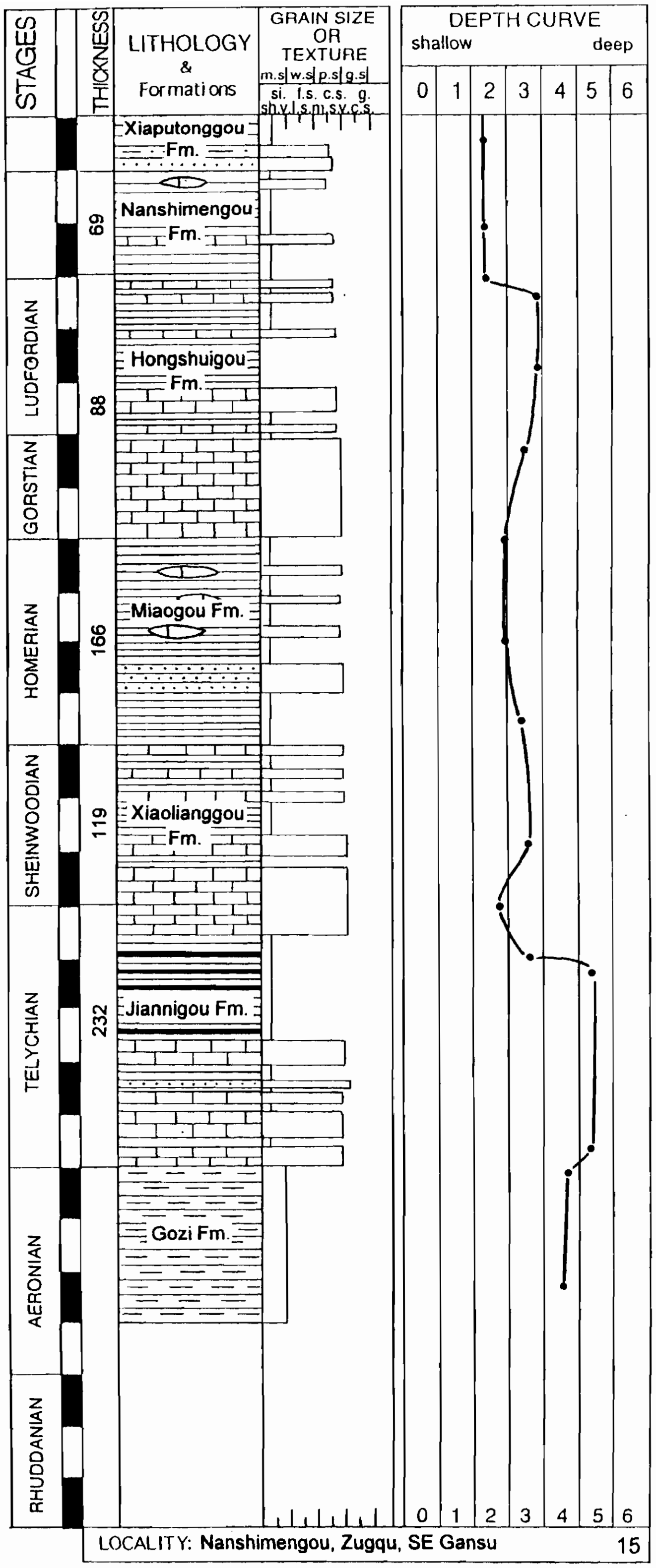

FIGURE 12 — Silurian in Nanshimengou Valley, Zugqu County, southeast Gansu Province. Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

the border region of Shaanxi and Sichuan (Jin et al., 1989; Chen et al., 1991) and the Daguan area of northeast Yunnan (Ye et al., 1983). The Takuan to Daluzhai Formations (upper Aeronian–Telychian) consist of fine-grained siliciclastics and shales irregularly intercalated with limestone and nodular limestone (Ye et al., 1983). A few thin-bedded limestones even occur in the Huixingshao Formation (upper Telychian shallow-marine red beds). This area was not far from the shoreline.

Reef facies — Silurian reefs are rare in China. They are mainly distributed along the northwest edge of the Yangtze platform and the north edge of the North China platform. Bioherms and carbonate mud mounds are recognized.

Telychian bioherms occur in the Shenxuanyi Member of the Ningqiang Formation from Guangyuan to Ningqiang on the northwest Yangtze platform. Their distribution is in the Qinling and Longmenshan Mountains in the north and west mobile belts of the Yangtze platform (Qiu, 1990). The Guangyuan–Ningqiang bioherm belt is in the *Oktavites spiralis–Stomatograptus grandis* Zone, and consists of reef complexes that include reef bases, cores, foreslope talus, and lagoons (Qiu, 1990). The largest known upper Telychian bioherm is exposed at Xuanhe village, Guangyuan, northern Sichuan, and measures 120 m long, 80 m wide, and 30.4 m high, with shale as its enclosing rock. According to Qiu (1990), it can be subdivided into four different facies. The first is a reef-base facies of packstones and grainstones in irregular beds or lenses 2–3 m thick. No intraclasts are present in the packstones; all the grains are skeletal and are dominated by crinoid fragments. The second is a reef-core of irregularly bedded and massive bafflestones. The ascending succession of the reef core includes the reef-base facies, massive bafflestone and coverstone, irregular bedded bafflestone, the reef core, and a reef-top facies with bedded bafflestone. In the bafflestone and coverstone, branching corals (e.g., *Coenites, Aphyllum, Ningqiangophyllum, Halysites*) played an important role in baffling, whereas crustose corals (*Subaveolites, Carnegia, Riphaeolites*) and stromatoporoids (*Labechia*) may have had a role in covering and binding. They may not have constructed a strong, solid, wave-resistant framework; however, a core was present. The third facies is the reef top with bedded bafflestone as a crust on the core of the bioherm. The core and reef-top facies are transitional. The reef-top facies consists chiefly of pelmatozoan-coral-bryozoan gravel. The fourth facies is the reef flank, composed mainly of rud- and floatstone interbedded with purple micrite with oblique laminations. The rudstones are in situ, and the breccia fragments are angular and unsorted. The coral blocks are often overturned. The flank rudstones rapidly change laterally.

Aeronian biostromes and brachiopod coquinas appear in the Xiangshuyuan Formation in the Meitan–Shiqian area of northeast Guizhou and along the Qianzhong Oldland. However, no reefs or bioherms grew in this area due to abundant siliciclastic input from the oldland. Coquina with the abundant virgianid brachiopod *Pseudoconchidium* (=*Paraconchidium*) *shiqianensis* Rong et al. is recorded from the upper Xiangshuyuan Formation (upper Aeronian) in northeast Guizhou (Rong and Yang, 1981; Rong, 1986). Several coquinas composed of the pentamerid *Apopentamerus sichuanensis* (Sheng) are known from the upper Ningqiang Formation (upper Telychian) in north Sichuan (Chen and Rong, 1996). The brachiopod *Pentamerifera oblongiformis* (Nikiforova) forms coquinas in the lower Ludlow of the Emin area of northwest Xinjiang, an area that was part of the Kazakhstan island-arc system (Rong and Zhang, 1988). The acmes of a single species may indicate that a crisis event, probably a storm leading to strong hydrologic segregation of shells, controlled deposition.

Large Silurian carbonate mud mounds are known in a few areas of China. Upper Telychian–Wenlock carbonate mud mounds developed in the Lan'gao area of south Shaanxi in the South Qinling belt. This belt was initially located on the slope in the Ordovician and early–middle Llandovery, but was uplifted in the Telychian. Ludlow mudmounds also occur in the Baterobo Formation at Baterobo in the Daerhan Mumingan Joint Banner of southern Inner Mongolia. Further investigation on these two areas is needed.

RESTRICTED SEA CARBONATES — Ludlow limestones (Sashuiyan Formation) and dolostones (Maliuqiao Formation) with conodonts, brachiopods, and a few corals are recorded in the Erlangshan area on the west edge of the Yangtze platform (Jin et al., 1989; Fig. 13). The Erlangshan area may have been a restricted sea or bay that developed with late Ludlow onlap of the Yangtze platform margin. The rocks and faunas indicate that these carbonates are of very shallow-water origin, and the dolostones may also indicate a restricted environment.

VOLCANICLASTICS

Silurian siliciclastics with a volcaniclastic component are usually well developed in marginal mobile belts in China. These sedimentary rocks are present in six areas.

SOUTHERN MOBILE BELT OF THE SIBERIAN PALEOPLATE — Siliciclastics intercalated with volcanics (e.g., tuffaceous siltstone and sandstone in the Bashilixiaohe and Gulanhe Formations; Fig. 14) occur in the Greater and Lesser Khingan Mountains of northern Inner Mongolia and in Heilongjiang, northeast China (Lin et al., 1984). Argillaceous shale, sandstone, schist, and gneiss with some tuff occur in the Altai, Fuwen, and Qinghe areas of northernmost Xinjiang (Lin et al., 1984; Zhou et Lin, 1995).

EASTERN ISLAND ARCS OF THE KAZAKHSTAN PALEOPLATE — Volcaniclastics occur from eastern to western Junggar, and in the Beishan region. These areas are probably part of the Kazakhstan paleoplate. Tuffaceous silt- and sandstone, tuff, and andesitic detritus are known from the Bulong, Shaerbuer, Kekexiongduke, and Qaergaye Formations in west Junggar (Lin et al., 1984; Fig. 15), and extend into the back-arc Dzhungaria–Balkhash Basins of Kazakhstan (Nikitin et al., 1991). Tuff occurs in the Hongliugou Formation at Kaokesaiergaishan, Mulei, east Xinjiang (Lin et al., 1984; compare Fig. 16). Rhyolite porphyry, breccia lava, andesitic porphyrite, and intermediate and basic tuffaceous lava occur in the Gongpoquan Group (Wenlock–Ludlow) of the Beishan region. Llandovery tuff, tuffaceous slate, sandstone, and lava are found in the Wulanbulage section at Erjin, westernmost Inner Mongolia (Lin et al., 1984).

NORTHERN MOBILE BELTS OF THE TARIM PALEOPLATE — Tuff, andesitic breccia, tuffaceous sandstone, and intermediate and basic lava are present in the Bayinbuluke Formation (Ludlow–Pridoli) in the border area between Jingxian and Xinyuan Counties, northwest Xinjiang. Similar volcanics are found in the Keketiekedaban Group (Wenlock–Ludlow) of Haliketaoshan, Tianshan, central Xinjiang (Lin et al., 1984).

NORTHERN ISLAND ARCS OF THE NORTH CHINA PALEOPLATE — Pebble-bearing tuff, andesitic porphyry, and fine-grained sandstone are known from the Halayan Formation of Baoerhantu, southern Inner Mongolia (Su, 1988). Acidic lava, andesite, and tuffaceous lava occur in the Taoshan Formation (Llandovery; Liu, 1985), and tuff, crystal and vitroclastic tuff, conglomerate, sandstone, and siltstone are known in the Zhangjiatun Formation (Ludlow–Pridoli; Fig. 17) of Jilin Province (Liu and Huang, 1977).

NORTHERN MOBILE BELT OF QAIDAM PALEOPLATE — Andesitic agglomerate, porphyry, basalt, tuffaceous sandstone, siltstone, and andesite are known in the Nannigou Formation, southwest of Gulang village, central Gansu Province (Lin et al., 1984).

NORTHERN MOBILE BELTS OF SOUTH CHINA PALEOPLATE — The Upper Silurian Pailungchiang Formation in west Qinling is composed of slate, sericitic slate, and sandstone intercalated with metamorphosed tuffaceous sandstone (Fu et al., 1983). Tuff, tuffaceous conglomerate, and slate occur in the Llandovery Anzigou Formation of west Qinling (Fu et al., 1983). In the Silurian of east Qinling, volcaniclastics (mainly tuffaceous sandstone) are only present in the basal part of the Telychian Doushangou Formation (ca. 20 m thick) and indicate minor Silurian

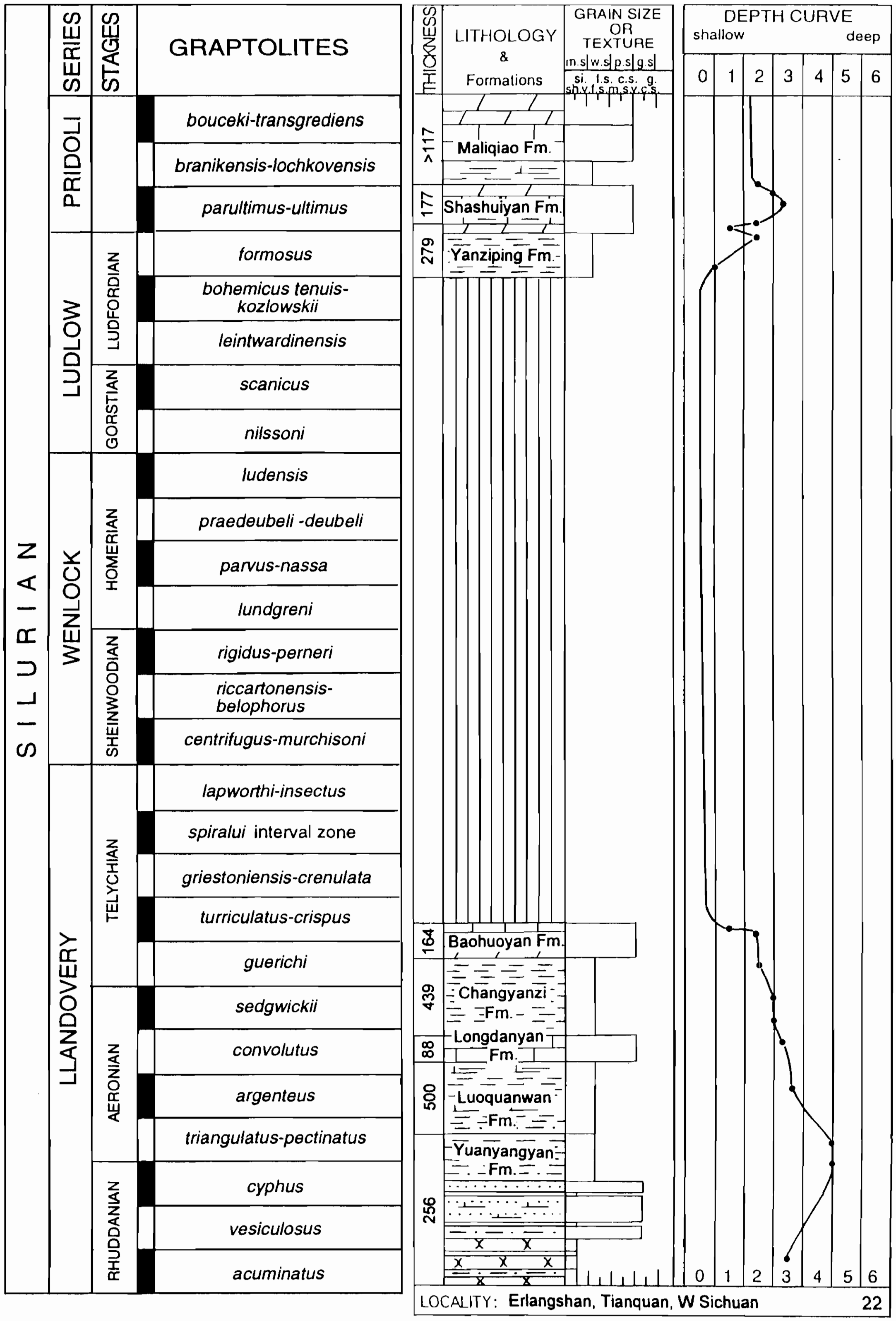

FIGURE 13 — Silurian of Erlangshan Mountain, Tianquan County, west Sichuan Province. Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

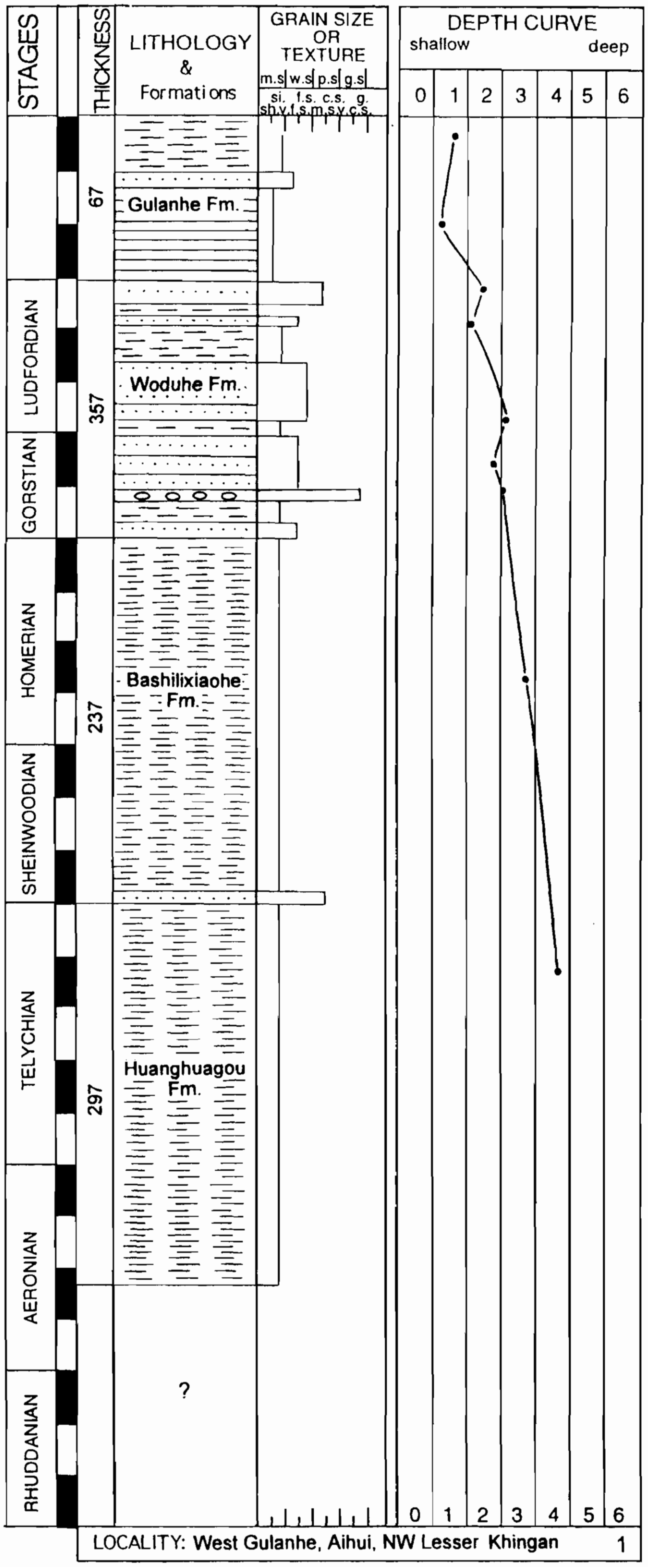

FIGURE 14 — Silurian at west Gulanhe, Aihui County, northwest Lesser Khingan, Heilongjiang Province. Stratigraphic column number noted at bottom right corresponds to locality number marked by black circle in Fig. 1 (the same for Figs. 5–16).

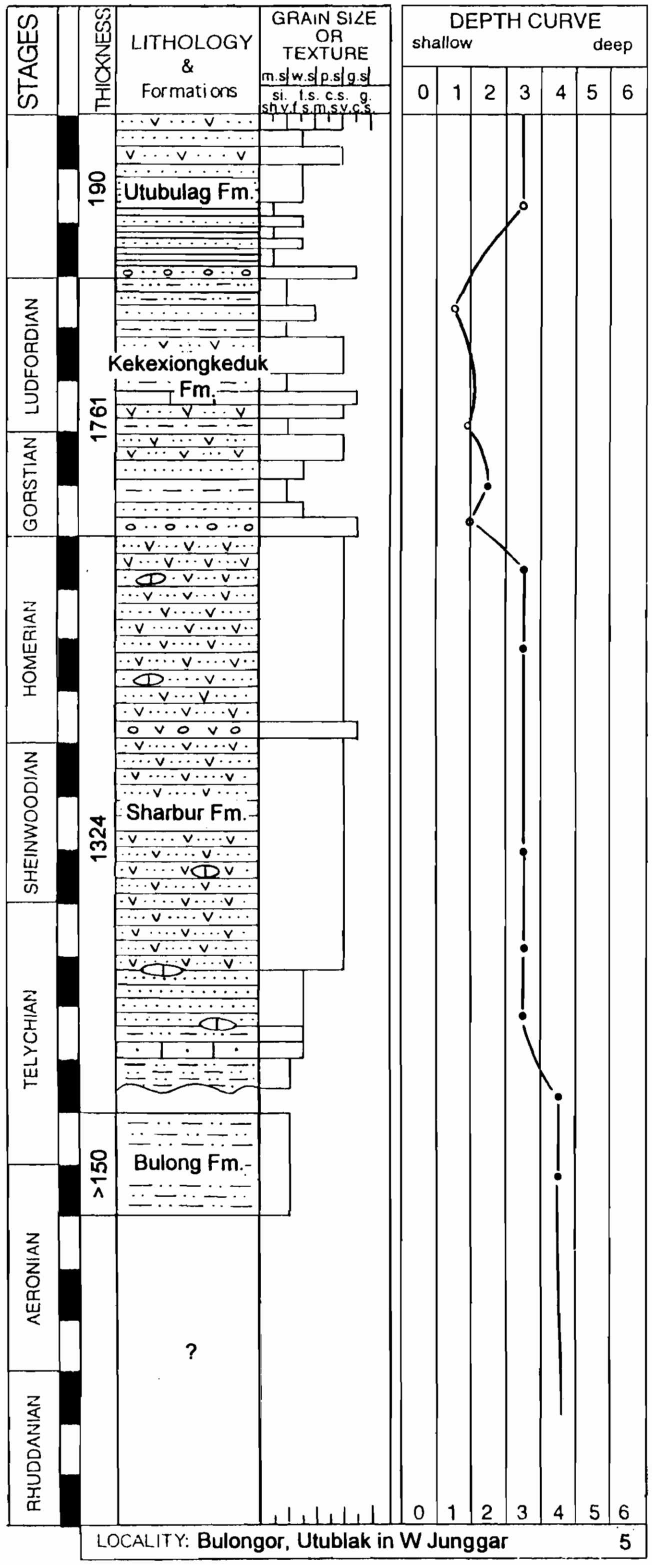

FIGURE 15 — Silurian at Bulongor, Utublak in west Junggar, north Xinjiang Province. Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

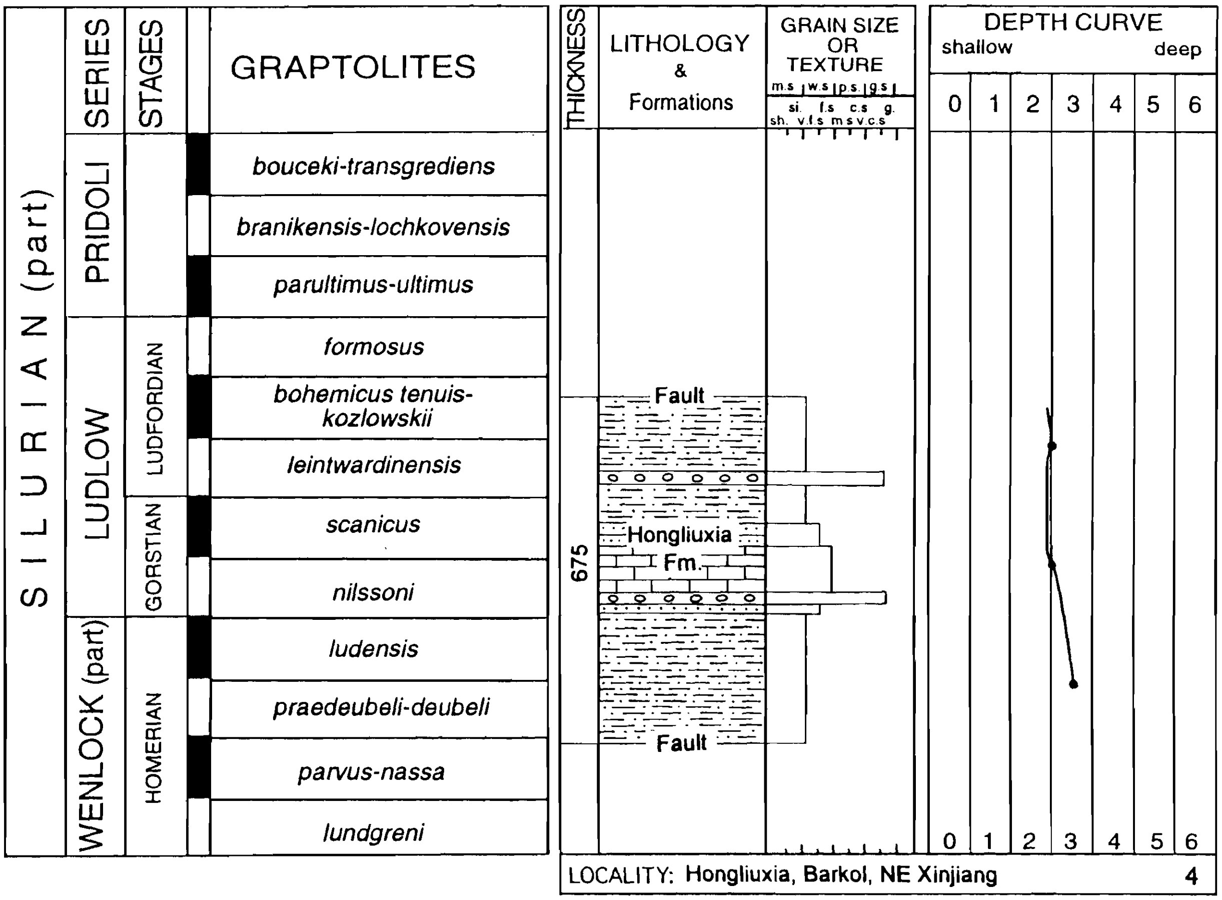

FIGURE 16 — Silurian at Gashaomiao, Darhan Mumingan Joint Banner, southern Inner Mongolia, and Zhangjiatun village, Yongji County, central Jilin Province. Numbers at bottom right correspond to locality numbers marked by black circles in Fig. 1.

volcanism (Fu, 1983).

Silurian volcaniclastics are rarely developed in the Lan'gao area of southeast Shaanxi Province in the south Qinling belt. The tuff in this siliciclastic sequence suggests small-scale Silurian rifting.

Silurian Paleogeography

The Silurian is widespread in China, but remarkable differences exist in outcrop quality between the various regions. In contrast to South China, the rocks in Tibet and Tarim are sporadically or poorly exposed and are poorly studied. Eight paleogeographic maps of China are presented for the following intervals: *Normalograptus extraordinarius* Zone of the uppermost Ordovician (Hirnantian); Llandovery *Parakidograptus acuminatus, Demirastrites triangulatus, Demirastrites convolutus-Stimulograptus sedgwickii, Spirograptus turriculatus-Monograptus crispus*, and *Oktavites spiralis* Zones; the Wenlock, and upper Ludlow–lower Pridoli. Two paleogeographic maps for the Llandovery–Wenlock and Ludlow–Pridoli boundaries are also shown.

As compared with the Ordovician, the epeiric seas were reduced in area, and oldland areas were elevated because the Silurian was one of the most tectonically active periods in China. This tectonism included the Guangxi orogeny in South China. Epeirogenic uplift caused great areal expansion of the oldlands several times in the Silurian.

South China — The latest Ordovician regression was succeeded by global sea-level rise at the beginning of the Silurian. The regression is widespread in South China (Fig. 18). The subsequent transgression flooded many regions of the Yangtze platform (Lungmachi Formation) and expanded the Xianggui Sea (Chouchiachi Formation) in most parts of South China (Fig. 19).

The Rhuddanian was characterized by anoxic condi-

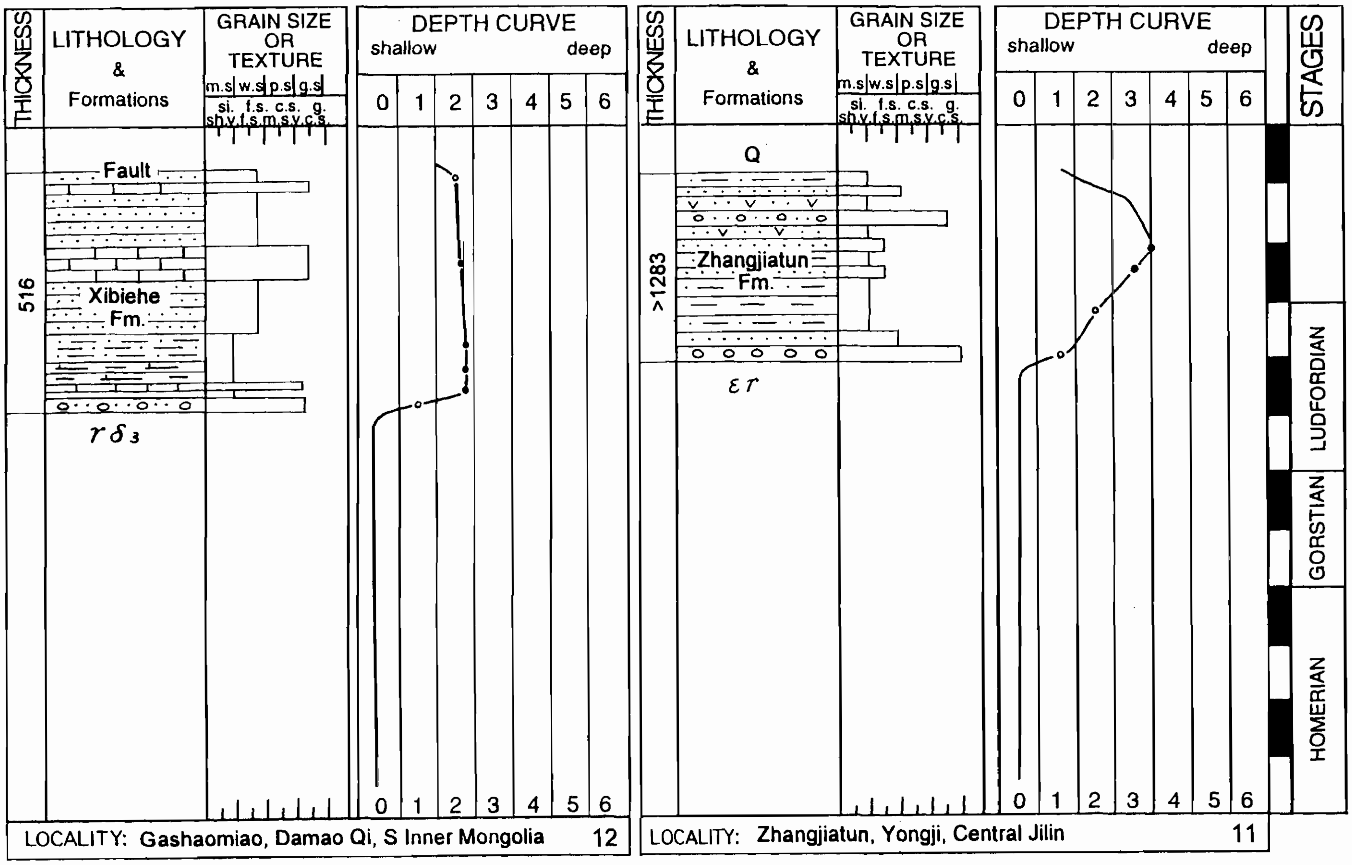

FIGURE 17 — Silurian at Hongliuxia Valley, Barkol County, northeast Xinjiang Province. Number at bottom right corresponds to locality number marked by black circle in Fig. 1.

tions across the Yangtze platform, as recorded in most areas by slowly accumulated black graptolite shale (Chen, 1984; Fig. 20). The anoxic conditions persisted on deeper parts of the platforms and basins until the middle Aeronian in the border area of Guizhou and Sichuan (Lungmachi Formation). Anoxia persisted into the late Aeronian in western Hubei and southern Sichuan (Lungmachi Formation), into the early Telychian in the Nanjiang–Chengkou area, north Sichuan (Nanjiang Formation); and into the Wenlock in the Ziyang area, south Shaanxi (Banjiuguan Formation), upper Yangtze region. The graptolitic sequence on the Yangtze platform was probably not deposited in very deep-water, and it is assigned mainly to BA 3–5 based on the shelly faunas of the overlying strata (Rong et al., 1984; Johnson et al., 1985; Chen, 1990).

The Aeronian of the Yangtze platform is a thick sequence of siliciclastics deposited under high and gradually increasing sedimentation rates (Chen, 1984) that were associated with subsidence and abundant siliciclastic input from the Cathaysian Oldland (Fig. 21). Graptolites, associated locally with low-diversity shelly faunas, are rare. The shallower platforms and basins in the later Rhuddanian and early Aeronian were ventilated, and feature carbonates with diverse benthic organisms in some regions (e.g., northeast Guizhou, southwest China; Tongxin area, Ningxia, north China; and Kalpin–Bachu area, southwest Xinjiang, northwest China) (Chen *in* Zhou and Chen, 1990). These areas became the sources for the recovery of shelly biotas after the latest Ordovician mass extinction (Rong and Harper, 1999).

The Aeronian has the earliest Silurian carbonates in China. These carbonates are common in the Xiangshuyuan and Shihniulan Formations on the south margin of south Sichuan and north Guizhou (Fig. 21). Two types of carbonate facies (Baisha and Yinjiang types) are known.

Baisha-type carbonates include near-shore, shallow-water, biostrome–coquina facies. They are distributed on the northeast margin of the Qianzhong Oldland (e.g., Baisha village, Shiqian County; Wenjiadian village, Sinan County; Balixi and Shijing villages, Fenggang County;

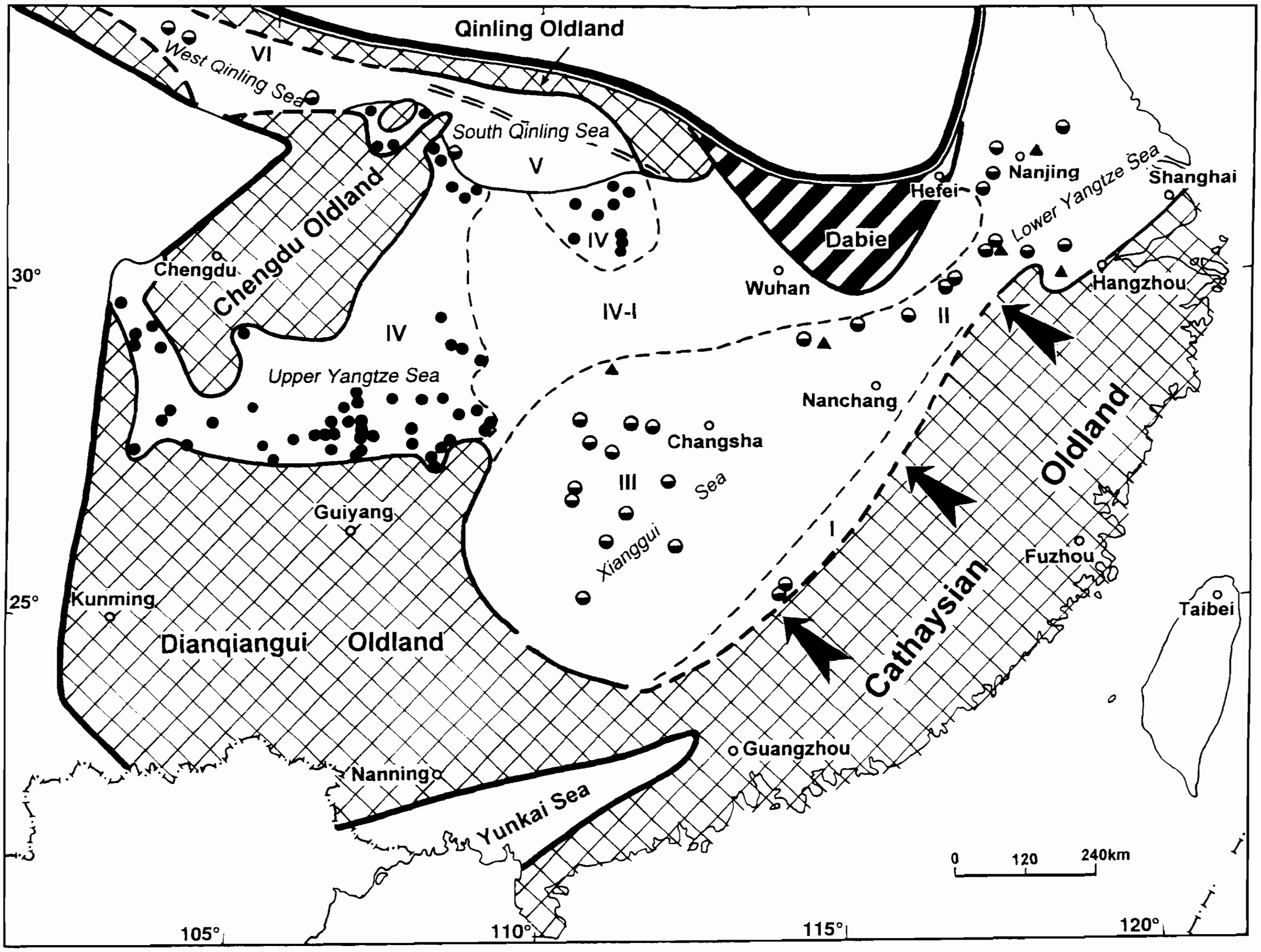

FIGURE 18 — Paleogeography of South China in latest Ordovician (*Normalograptus extraordinarius* Chron). Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas crosshatched. Black circles show shallow water assemblages of *Hirnantia* fauna (BA 2–3) or Hirnantian graptolites. Black triangles show *Paromalomena-Aegiria* Community (deeper water BA 4–5) of *Hirnantia* fauna (see Rong, 1979). Half-black circles show *N. extraordinarius* Zone fossils or rocks. Black arrows show expansion of Cathaysian Oldland in Guangxi orogeny. I. Chongyi foreland basin (molasse), II. lower Yangtze platform (black shale with mudstone), III. Xianggui Basin (thick, black graptolitic shale), IV. upper Yangtze platform (shale, mudstone, siltstone, and argillaceous and dolomitic limestone with *Hirnantia* fauna), IV-I. Uppermost Ordovician unknown, V. South Qinling slope and basin (thick, black graptolitic shale), VI. west Qinling basin (thick, black graptolitic shale).

Xinglongchang village, Meitan County; Tuping village, Zheng'an County; and Wenquan village, Suiyang County) (Ge et al., 1979; this report). This facies is characterized by argillaceous limestone intercalated with thick-bedded coquina limestone. Abundant *Pseudoconchidium shiqianensis,* rare *Virgianella glabera* (Rong and Yang, 1981), and reef-like biostromes with very rich fasciculate and branched rugose corals (*Ceriaster, Stauria, Paleophyllum*), massive tabulates (syringoporids), heliolitids (*Helioplasmolites*), stromatoporoids (*Clathrodictyon, Forolinia*), and algae are found (Ge et al., 1979). These forms indicate BA 2 and, mainly, BA 3 habitats (Rong and Yang, 1981; Rong, 1986; Wang et al., 1987). No reefs have been found in this region (Ge et al., 1979; Mu et al., 1986).

Yinjiang-type carbonates include off shore, deeper-water, argillaceous limestone. This facies occurs in the Yinjiang–Wuchuan area of northeast Guizhou, and was deposited relatively distant from the Qianzhong Oldland (Ge et al., 1979). The brachiopods are characterized by *"Stricklandia" transversa* Grabau, *Merciella striata* Rong et al., and *Lissatrypa magna* (Grabau) (Rong and Yang, 1981). Rugose corals include solitary forms, with common *Rhyzophyllum* (Ge et al., 1979). There are mushroom-shaped and encrustiform corals, rather than fasciculate stauriids

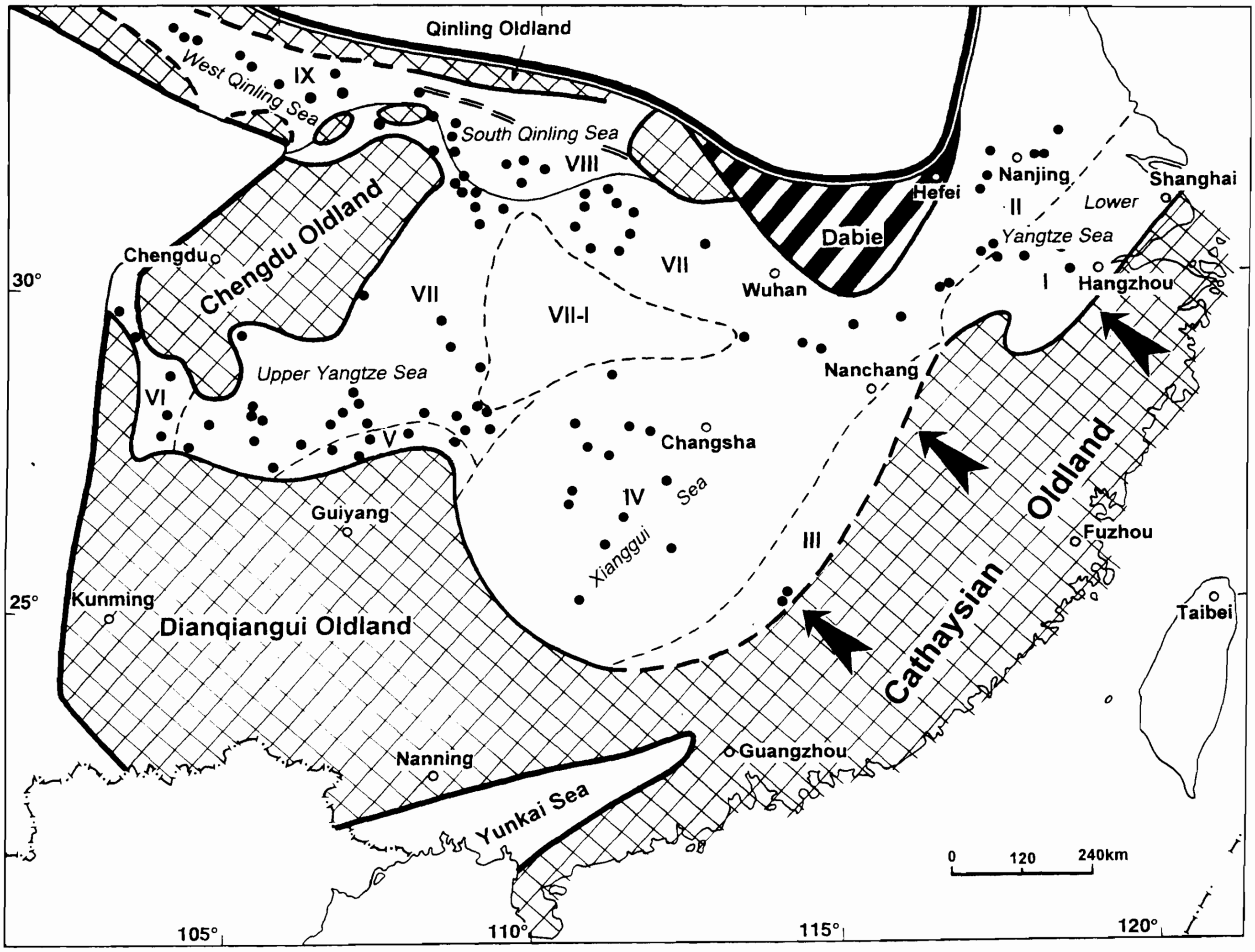

FIGURE 19 — South China in early Rhuddanian (*Parakidograptus acuminatus* Chron), Llandovery. The Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. Black circles show *P. acuminatus* Zone fauna and rock. Black arrows show expansion of Cathaysian Oldland in Guangxian orogeny. I, II, IV, VII–IX are black graptolite shale; III, molasse; V, VI, near-shore mudstones. I. southeast lower Yangtze platform, II. northwest lower Yangtze Platform, III. Chongyi foreland basin, IV. Xianggui Basin, V. northern Guizhou, VI. western Sichuan, VII. upper Yangtze platform, VII-I. Upper Rhuddanian unknown, VIII. south Qinling slope and basin, IX. west Qinling Basin.

and massive syringoporids. Trilobites (*Scutellum, Gaotania*) and rare endemic species of graptolites (*Normalograptus yangtzeensis* Hsu) also occur. An assignment of these fossils to outer BA 3 and inner BA 4 habitats is suggested (Rong, 1986; Wang et al., 1987). Medium- to thick-bedded limestone of the Lojoping Formation (uppermost Aeronian) from the Yichang area, western Hubei, also indicate platform carbonates in the central Yangtze platform. Abundant brachiopods (*Apopentamerus dorsoplanus, "Stricklandia" transversa, Merciella striata, Lissatrypa magna*), trilobites (*Songkania, Kosovopeltis, Aristoharpes*), corals (small, einzoner, solitary *Rhyzophyllum, Densiphylloides, Pycnactis*), and a few stromatoporoids (*Clathrodictyon*) and graptolites (*Monoclimacis arcuata*), indicate outer BA 3 to inner BA 4 habitats (Rong, 1986; Wang et al., 1987).

Isopachs for the Rhuddanian and Aeronian of the upper Yangtze platform and Xianggui Basin are shown in Fig. 22. It can be ascertained that there were two centers of subsidence: the Shimen–Sangzhi area of northwest Hunan and the Anhua–Xupu area of central Hunan, which are separated by the submarine Jiangnan uplift along the Kaili–Tongren–Jishou–Jinshi belt.

Rocky shores mark paleoshorelines. Only one record of a Paleozoic rocky shoreline in China has been documented; it is a latest Aeronian karst shoreline in South

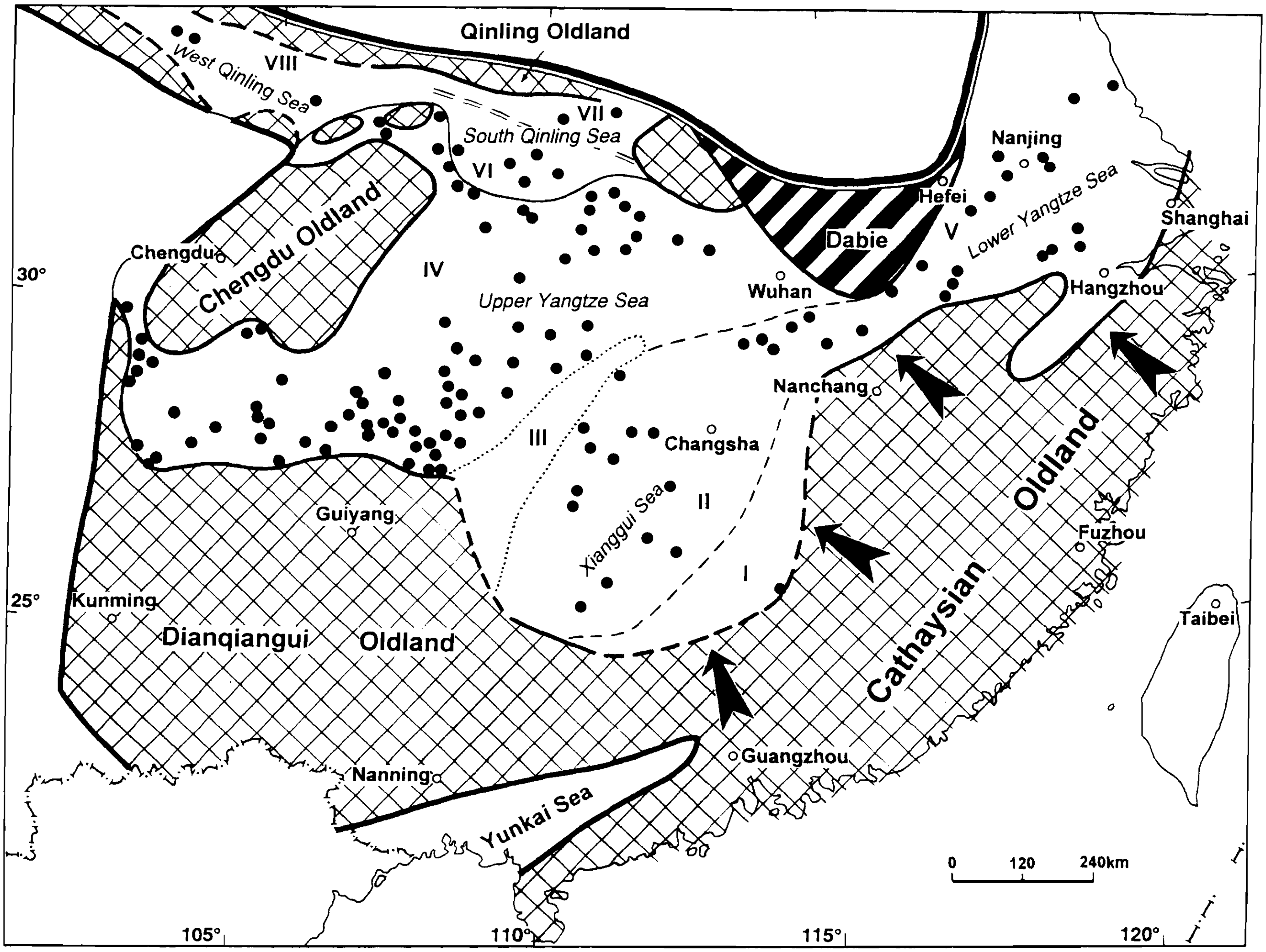

FIGURE 20 — South China in early Aeronian (*Demirastrites triangulatus* Chron), Llandovery. The Lijiang–Ninglang area, north Yunnan, is not included in figure. Land areas cross-hatched. Black circles show *D. triangulatus* Zone faunas and rocks. Black arrows show expansion of Cathaysian Oldland in Guangxian orogeny. I, molasse; II, IV, V–VIII, black graptolite shale; III, lower Aeronian unknown. I. Chongyi foreland basin, II. Xianggui Basin, III. Jiangnan submarine rise, IV. upper Yangtze platform, V. lower Yangtze platform, VI. south Qinling Basin, VII. east Qinling Sea, VIII. west Qinling Basin.

China (Rong and Johnson, 1996) at Huanghuachong, south of Wudang village to the northeast of Guiyang, central Guizhou, upper Yangtze region. In most of the Huanghuachong area, the latest Aeronian–Telychian Kaochaitien Formation (mudstone and siltstone) is underlain disconformably by the Caradocian–early Ashgillian Huanghuachong Formation (limestone). At Yegouchong, the Kaochaitien Formation oversteps 63 m of relief developed on argillaceous limestone of the Llanvirnian Guniutan Formation. These data indicate that a transgressive event took place in the late Aeronian (*Stimulograptus sedgwickii* Chron) in central Guizhou with a local sea-level rise of up to 63 m. The change in sea level may correspond to late Aeronian eustatic rise on several different paleocontinents (Johnson et al., 1991). However, this does not imply a water depth in central Guizhou of 63 m at the peak of Aeronian transgression. Indeed, the opposite was true, as the area was rapidly filled with fine siliciclastics derived from erosion of the upper Arenigian Meitan Formation. The Kaochaitien Formation infills sink holes in the Huanghuachong Formation that are over 5 m deep at Huanghuachong. This example has great geological and geographical significance, and is the first report of a pre-Quaternary rocky shore in China.

The Telychian of the Yangtze platform is characterized by shallow-marine red beds (i.e., lower Telychian

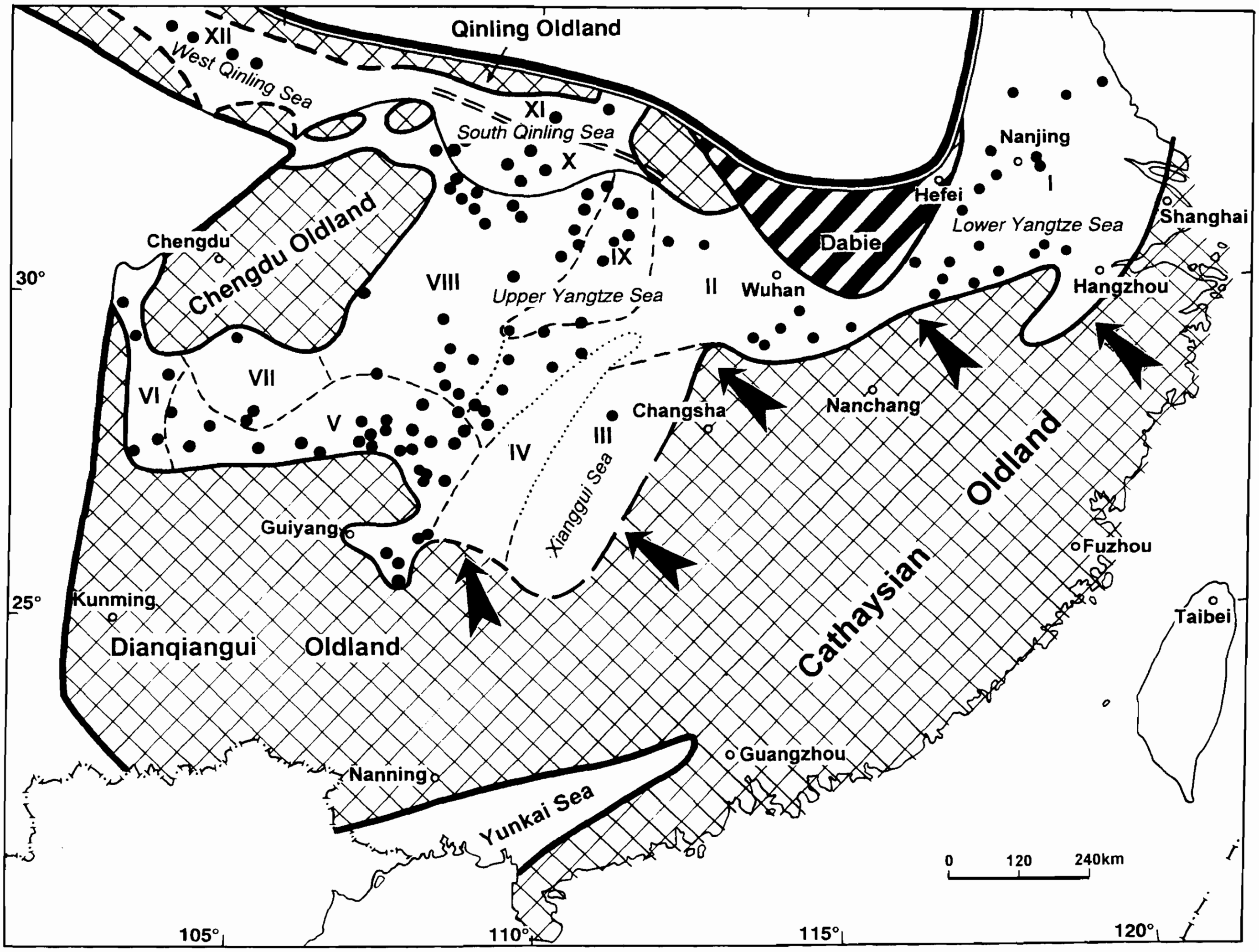

FIGURE 21 — South China in the middle–late Aeronian (*Demirastrites convolutus–Stimulograptus sedgwickii* Chrons), Llandovery. The Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. Black circles show *D. convolutus* or *S. sedgwickii* Zone faunas and rocks. Black arrows show expansion of the Cathaysian Oldland during Guangxian orogeny. I, VII, IX–XII, black graptolite shale; II, VI, VIII, siltstone with sandstone; III, thick, flysch-like rocks; IV, middle–upper Aeronian unknown; V, chiefly limestone. I. lower Yangtze platform, II. central Hubei–southeast Sichuan, III. Xianggui Basin, IV. Jiangnan submarine rise, V. southwestern Sichuan–northern Guizhou, VI. western Sichuan, VII. central Sichuan, VIII. northeast Sichuan and northwest Hubei, IX. western Hubei (with uppermost Aeronian limestone), X. south Qinling Basin, XI. east Qinling Sea, XII. west Qinling Basin.

Rongxi Formation, upper Telychian Huixingshao Formation, and equivalents) that were deposited in depressions near the shoreline of the oldlands (Figs. 23, 24). The Rongxi red beds usually yield a BA 2 *Nucleospira-Nalivkinia* brachiopod community (Rong, 1986; Wang et al., 1987). The Huixingshao red beds have a low-diversity gastropod and bivalve fauna with *Lingula*, which indicates an assignment to BA 1, and rarely to BA 2 (Rong et al., 1984; Johnson et al., 1985).

The isopach map of the Rongxi Formation of the upper Yangtze region is in Fig. 25. Two centers of subsidence are recognized in the Dayong–Longshan area, northwest Hunan, and the Si'nan–Yinjiang area, northeast Guizhou. These centers may have been related to the input of rivers on the Cathaysian Odland (Wang Z.-z., 1996).

A considerable amount of shallow-marine, yellow-gray or green siliciclastics with less carbonate than in the Aeronian (Xiushan Formation, middle Telychian) are sandwiched between the two red beds on the Yangtze platform. The Xiushan fauna (*Spinochonetes-Coronocephalus-Sichuanoceras-Stomatograptus sinensis* Fauna; BA

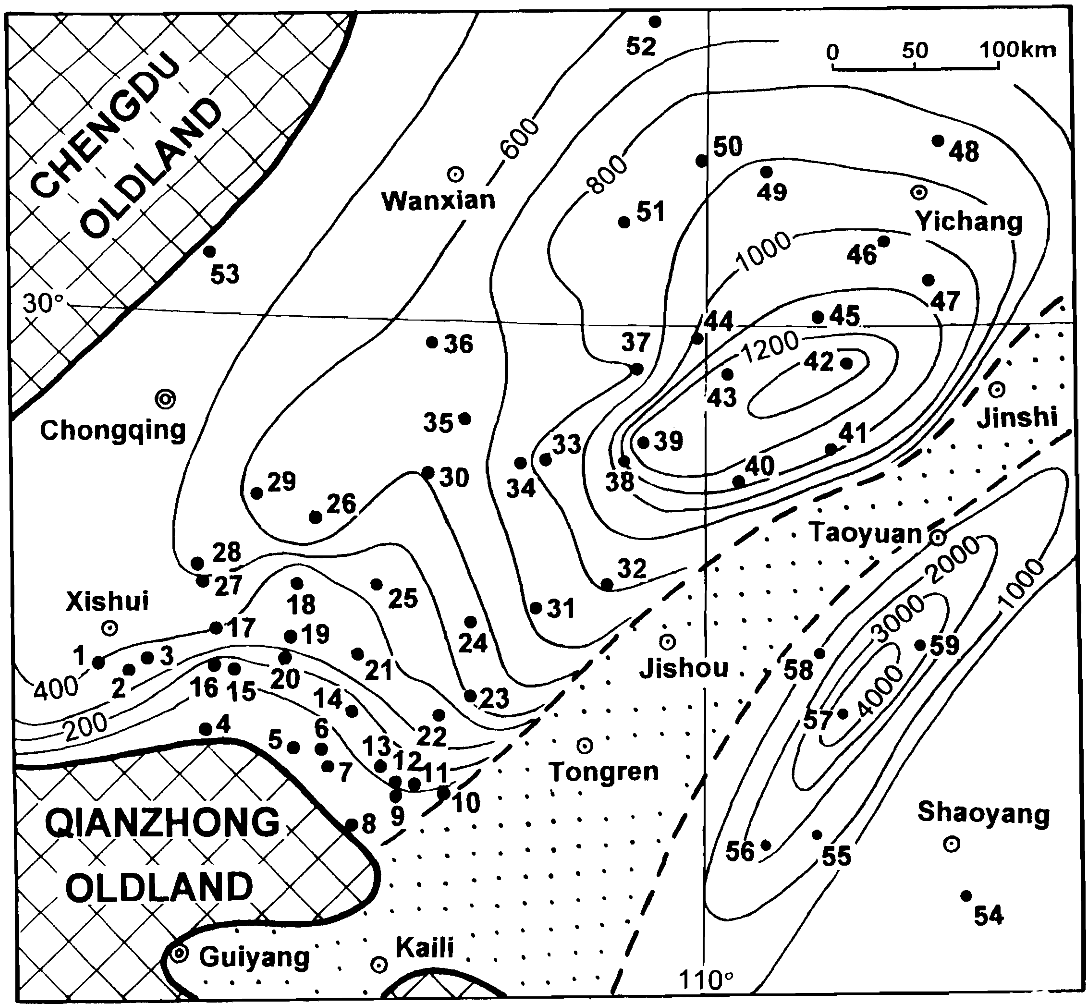

FIGURE 22 — Rhuddanian–Aeronian isopachs for upper Yangtze region, South China. Land area cross-hatched. Black circle: locality named below with thickness in parentheses. Qianzhong Oldland is northern part of Qiangui Oldland. Jiangnan submarine rise (dotted) between upper Yangtze platform and Xianggui Basin. 1. Tiesuoqiao, Gulin (400.3 m); 2. Yangcun, Renhuai (330.7 m); 3. Xiaheba, Tongzi (386.2 m); 4. Donggongsi, Zunyi (7 m); 5. Niuchang, Meitan (>23 m); 6. Xinglongchang, Meitan (63 m); 7. Songyan, Yuqing (26 m); 8. Xiaosai, Yuqing (63 m); 9. Baisha, Shiqian (thickness unknown); 10. Fengxiangping, Shiqian (59 m); 11. Leijiatun, Shiqian (159.3 m); 12. Juntianba, Shiqian (143 m); 13. Wenjiadian, Si'nan (167 m); 14. Dongkala, Fenggang (310.8 m); 15. Xianba, Suiyang (145 m); 16. Sancha, Tongzi (200.7 m); 17. Heishixi, Tongzi (467.1 m); 18. Huangjiawuji, Zheng'an (355 m); 19. Tuping, Zheng'an (345 m); 20. Wenquan, Suiyang (215.7 m); 21. Balixi, Fenggang (328 m); 22. Liangshuijing, Si'nan (319 m); 23. Zhoujiaba, Si'nan (505 m); 24. Qiaojiazhai, Yanhe (495 m); 25. Longjingpo, Wuchuan (492.3 m); 26. Bayu, Daozhen (677.4 m); 27. Hanjiadian, Tongzi (474 m); 28. Guanyinqiao, Xijiang (496.5 m); 29. Sanquan, Nanchuan (737 m); 30. Wanzu, Pengshui (608 m); 31. Rongxi, Xiushan (713 m); 32. Kapeng, Huayuan (873 m); 33. Xiaoxian, Youyang (831 m); 34. Xiaoxitan, Qianjiang (749.3 m); 35. Yushan, Pengshui (>600 m); 36. Qiliao, Shizhu (612.3 m); 37. Gaoluo, Xuan'en (805 m); 38. Bidong, Longshan (762 m); 39. Hongyanxi, Longshan (1,216 m); 40. Wentang, Dayong (1,164 m); 41. Shimaxi, Cili (1,113 m); 42. Hengchang–Gengzishan, Shimen (1,304.7 m); 43. Hongjiayu, Sangzhi (1,293 m); 44. Qiquanyan, Hefeng (1,179.8 m); 45. Longchihe, Shimen (1,029 m); 46. Sandongshui, Changyang (thickness unknown); 47. Zhanxin, Yidu (1,010 m); 48. Dazhongba, Yichang (881 m); 49. Siyangqiao, Badong (942 m); 50. Taohua, Wushan (>882 m); 51. Taiyanghe, Enshi (891 m); 52. Xujiaba, Wuxi (661 m); 53. Huayinshan, Linshui (452 m); 54. Shaoyang (ca. 600 m); 55. Baimashan, Dongkou (2,000 m); 56. east of Dongkou (2,200 m); 57. Jiangdong, Xupu (4,000 m); 58. southwest of Anhua (ca. 1,000 m); 59. Zhoujiaxi, Anhua (>3,000 m).

3) (Rong et al., 1984; Johnson et al., 1985; Wang et al., 1987) is prominent in many areas of the Yangtze Platform. A Llandovery isopach map for the upper Yangtze Platform is shown in Fig. 26. The center of subsidence through the Llandovery was in the Dayong–Taoyuan area, Hunan Province, close to the Cathaysian Oldland (Sun and Gong, 1994).

The late Telychian (late *Monoclimacis griestoniensis* to early *Oktavites spiralis* Chrons) was one of the strongest Silurian transgressive periods on the Yangtze platform (Rong et al., 1984; Johnson et al. 1985). Marine areas were later greatly reduced in South China, because a large-

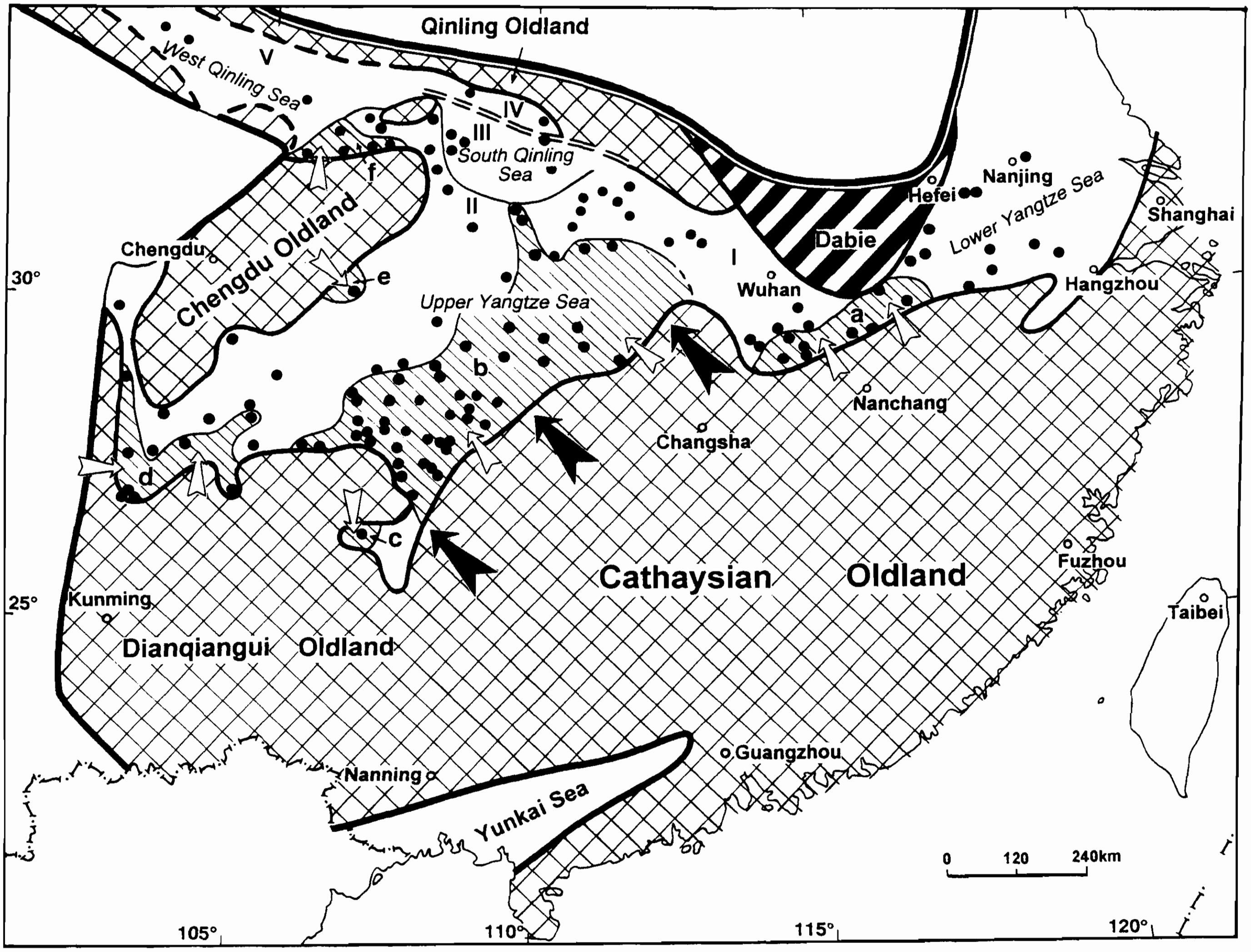

FIGURE 23 — South China in early Telychian (*Spirograptus turriculatus-Monograptus crispus* Chron), Llandovery. The Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. Shallow-marine red beds in diagonal shading. Black circles show localities of Rongxi Formation or equivalent red beds. Black arrows show the direction of Cathaysian Oldland expansion in Guangxian orogeny. Hollow arrows show transport of red sediments. Lower red beds at: a. north Jiangxi; b. northeast Guizhou, southeast Sichuan, and west Hunan; c. central Guizhou; d. northeast Yunnan and southwest Sichuan; e. Huayinshan, central Sichuan; f. Ningqiang–Guangyuan area on the border of Sichuan and Shaanxi Provinces. Areas without lower red beds include: I. Yangtze platform (yellow-green siltstone and mudstone) and black graptolite shale areas in II. Nanjiang–Chengkou, north Sichuan Province; III. south Qinling Basin; IV. east Qinling Basin; and V. west Qinling Basin.

scale regression commenced at the end of the Llandovery and persisted into the Wenlock and early Ludlow. The distribution of late Telychian marine red beds and other lithofacies is shown in Fig. 27.

The central Songpan–Ganzi block was characterized during the Silurian by the emergence of oldlands (Aba, Zorge, Songpan, and adjacent counties), and by marine flooding in the southwest (Batang, Baiyu, Derong, Ganzi, Litang, and Muli Counties). A continuous sequence of Lower Paleozoic carbonates with shallow, warm-water corals, bryozoans, stromatoporoids, and brachiopods is well developed in the region (Yang et al., 1994). A large amount of siliciclastics derived from the Songpan Oldland was deposited in the north in western Qinling, which is known as the Bailongjiang Caledonide rift trough (Yang et al., 1994).

Chen and Mitchell (1996) suggested that the onset of the Guangxi orogeny in South China coincided with the climax of the Taconian orogeny in the *Climacograptus* (*Diplacanthograptus*) *spiniferus Chron* (middle Caradocian). This interpretation is supported by the presence of a clastic wedge in the Nanshichong Formation (about in the *C. spiniferus* Zone) from Taojiang, central Hunan. The Guting Formation is characterized by shallow-water

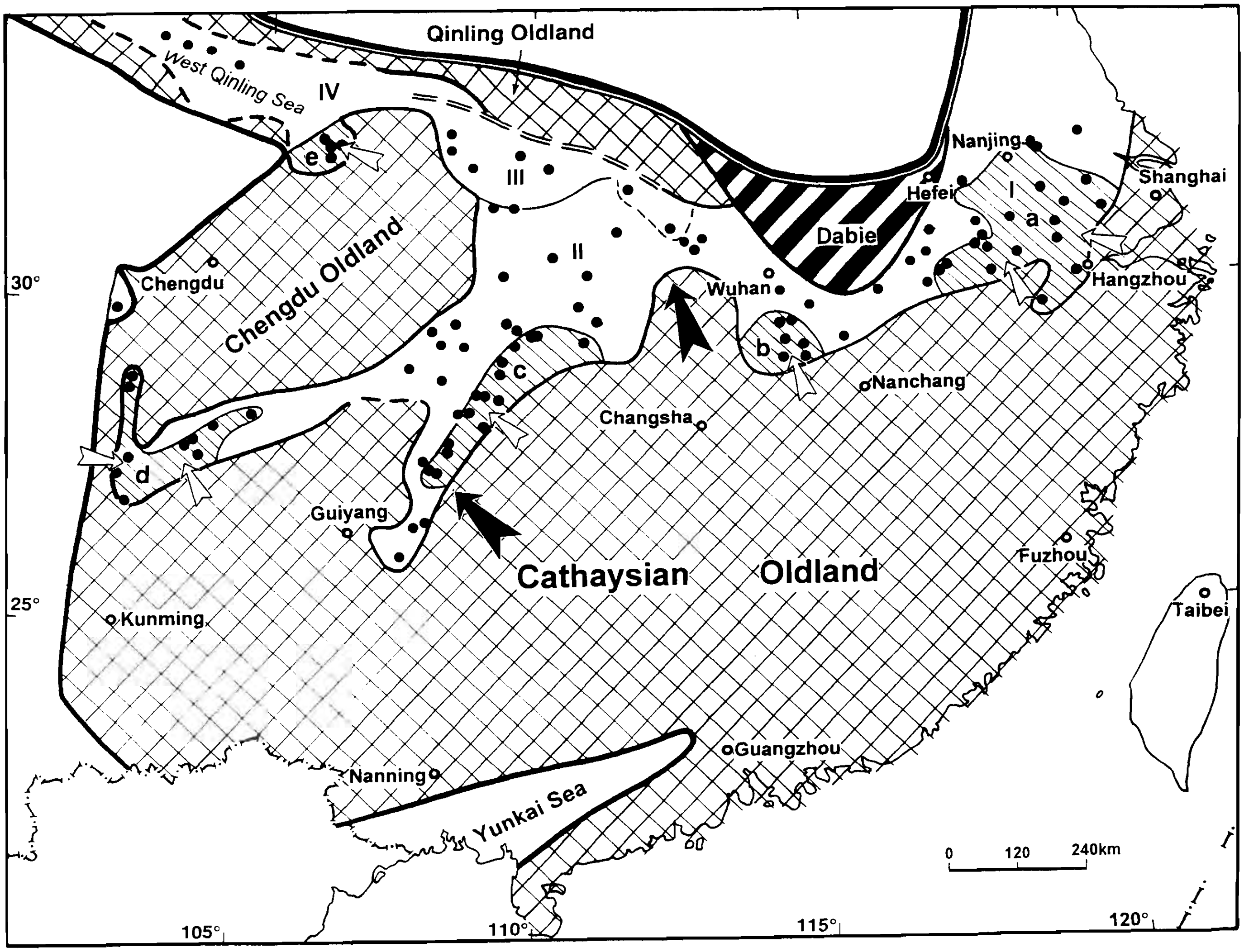

FIGURE 24 — South China in late Telychian (*Oktavites spiralis* Chron), Llandovery. The Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. Hollow arrows show transport of red sediments. Shallow-marine red beds shown by diagonal shading at: a. south Jiangsu, west Zhejiang, and south Anhui; b. north Jiangxi; c. northeast Guizhou, southeast Sichuan, and west Hunan; d. southwest and west Sichuan and northeast Yunnan; e. Ningqiang–Guangyuan area in border area of Sichuan and Shaanxi Provinces. I and II are lower and upper Yangtze platforms, respectively (chiefly yellow-green mudstone with limestone lenses with Xiushan fauna); III. south Qinling Basin with thick flysch-like rock; IV. west Qinling Sea.

limestone within a thick sequence of siliciclastic rocks that yields graptolites in some horizons (middle–upper Caradocian) from Chongyi County, southern Jiangxi Province, and in reef facies in the Xiazhen Formation (middle Ashgillian) from Yushan County, northeast Jiangxi. These units may have accumulated on a submarine rise. The SW–NE trend of this submarine rise may also reflect the Guangxian orogeny in southeast China. Enlargement of the Cathaysian Oldland from the Late Ordovician to the Early Silurian also may have coincided with the development of the Guangxian orogeny. The Cathaysian Oldland expanded northwest to the central Guizhou–central Hunan–south Anhui–east Zhejiang region in the late Aeronian. The lower red marine beds (i.e., Rongxi and Qingshui Formations) of the Yangtze platform were deposited along the northwest side of the Cathaysian Land in the early Telychian. Late Telychian bioherms and the upper marine red beds (Ningqiang Formation) are distributed along the border of the Upper Yangtze platform and the Dabashan uplift. This might also be a consequence of the effects of the Guangxian orogeny. The entire Yangtze platform was uplifted near the end of the Llandovery when the Guangxian orogeny ended (Fig. 28).

During the Wenlock, most of South China was

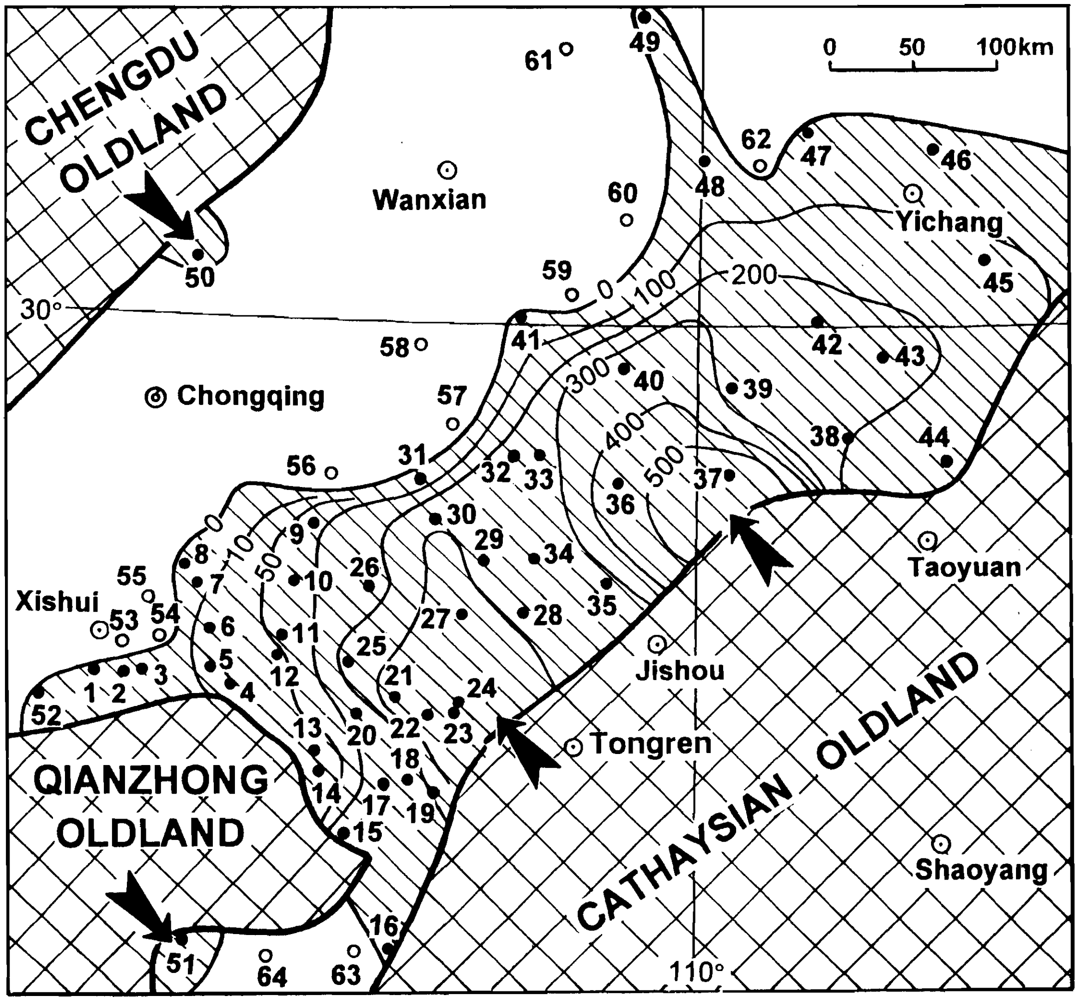

FIGURE 25 — Early Telychian shallow-marine red beds in upper Yangtze region (areas b and c in Fig. 22) with isopachs. Land areas cross-hatched. Red beds in diagonal ruling. Black and hollow circles are sections with and without red beds, respectively. Black arrows show transport of large amounts of fine-grained siliciclastics rich in Fe2O3. 1. Tiesuoqiao, Gulin (9 m of red beds); 2. Mayantan, Gulin (5 m); 3. Yangcun, Renhuai (39.5 m); 4. Xianba, Suiyang (45.1 m); 5. Sancha, Tongzi (41.6 m); 6. Heishixi, Tongzi (86.1 m); 7. Hanjiadian, Tongzi (7.6 m); 8. Guanyinqiao, Xijiang (8 m); 9. Bayu, Daozhen (64.1 m); 10. Huangjiawuji, Zheng'an (95 m); 11. Guankou, Tuping, Zheng'an (53 m); 12. Wenquan, Suiyang (57.3 m); 13. Xinglongchang, Meitan (18 m); 14. Paomozhai, Songyan, Yuqing (20 m); 15. Xiaosai, Suyang, Yuqing (127 m); 16. Shuizhu, Kaili (13.9 m); 17. Juntianba, Shaiqian (156.2 m); 18. Leijiatun, Shiqian (179.2–187.8 m); 19. Fengxiangping, Shiqian (194–207 m); 20. Dongkala, Fenggang (73.1m); 21. Donghuaxi, Si'nan (312.6 m); 22. Liangshuijing, Si'nan (348 m); 23. Chanxi, Yinjiang (347.8m); 24. Zhoujiaba, Si'nan (271 m); 25. Balixi, Fenggang (240 m); 26. Longjingpo, Wuchuan (158.2 m); 27. Qiaojiazhai, Yanhe (381 m); 28. Rongxi, Xiushan (240 m); 29. Dingshi, Youyang (250 m); 30. Hongdu, Yanhe (296 m); 31. Wanzu, Pengshui (15 m); 32. Diheba, Qianjiang (250 m); 33. Xiaoxian, Youyang (250–330 m); 34. Longtan, Youyang (234.8 m); 35. Kapeng, Huayuan (246 m); 36. Bidong, Longshan (467.1 m); 37. Wentang, Dayong (529 m); 38. Shimaxi, Cili (193 m); 39. Hongjiayu, Sangzhi (283 m); 40. Gaoluo, Xuan'en (343 m); 41. south of Lichuan (thickness unknown); 42. Longchihe, Shimen (269 m); 43. Gengzishan, Shimen (283 m); 44. Taifushan, Lingli (139 m); 45. south of Zhijiang City (100 m); 46. Dazhongba, Yichang (1.1 m).

uplifted, and the Qianzhong, Chengdu, and other peninsulas or islands were united with the Cathaysian Oldland (Fig. 29). Residual seas in South China during this interval occurred chiefly in three regions. These include: 1) south Shaanxi (Ziyang, Langao, and Pingli Counties) (Fu, 1983; Ge and Li, 1984) and northwest Hubei (Zhushan County), 2) Jiangsu Province (Geng et al., 1997); and 3) an area west of the northern marginal belt of South China (Fu et al., 1983). The northern marginal belt of the South China paleoplate was uplifted in the later Silurian, and no Ludlow and Pridoli have been found there. The Wenlock paleogeography of the Songpan–Ganzi block is remarkably different from that of other parts of South China because of the development of platform carbonates around the Songpan Oldland in the center of the block.

After the Wenlock, all of South China was probably

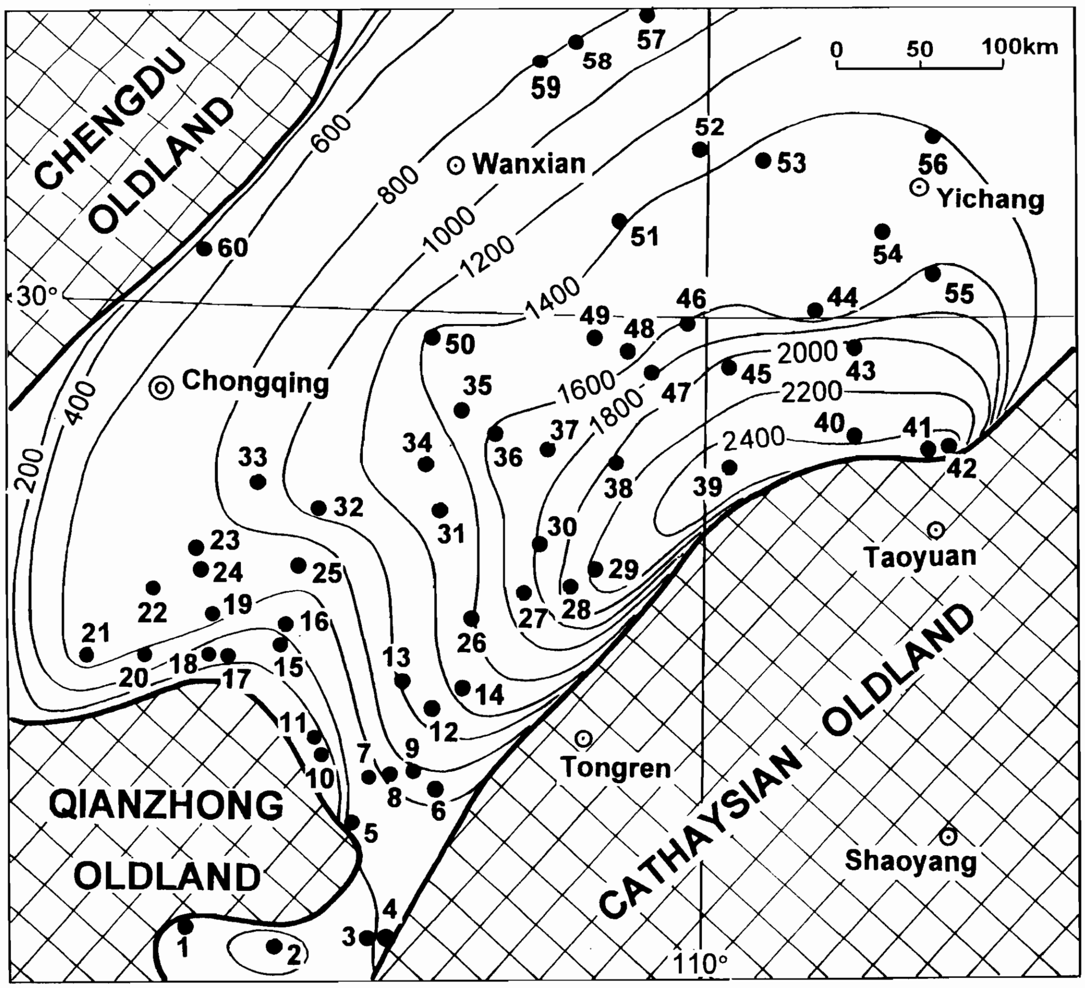

FIGURE 26 — Llandovery isopachs in upper Yangtze region, South China. Land areas cross-hatched. Black circles: 1. Wudang, Guiyang (569.5 m); 2. Dalishu, Guiding (722.5 m); 3. Luomian, Kaili (329 m); 4. Wengxiang, Kaili (562 m); 5. Xiaosai, Yuqing (457 m); 6. Fengxiangping, Shiqian (650.8 m); 7. Benzhuang, Shiqian (115 m); 8. Juntianba, Shiqian (754.5 m); 9. Leijiatun, Shiqian (828.4 m); 10. Songyan, Yuqing (45 m); 11. Xinglongchang, Meitan (>81 m); 12. Liangshuijing, Si'nan (thickness unknown); 13. Donghuaxi, Si'nan (1,006.4 m); 14. Heshui, Yinjiang (1,259 m); 15. Wenquan, Suiyang (343 m); 16. Guankou, Zheng'an (450.5 m); 17. Xianba, Suiyang (226.1 m); 18. Sancha, Tongzi (393.3 m); 19. Heishixi, Tongzi (747.1 m); 20. Yangcun, Renhuai (498.9 m); 21. Tiesuoqiao, Gulin (638.7 m); 22. Wenshui, Xishui (705 m); 23. Guanyinqiao, Xijiang (743 m); 24. Hanjiadian, Tongzi (780.3 m); 25. Huangjiawuji, Zheng'an (675 m); 26. Qiaojiazhai, Yanhe (1,398 m); 27. Rongxi, Xiushan (1,688.9 m); 28. Hujiadong, Xiushan (2,130 m); 29. Kapeng, Huayuan (2,271 m); 30. Longtan, Youyang (1,814.3 m); 31. Hongdu, Yanhe (1,230.7 m); 32. Bayu, Daozhen (1,120.7 m); 33. Sanquan, Nanchuan (unknown thickness); 34. Wanzu, Pengshui (1,563 m); 35. Yushan, Pengshui (1,563 m); 36. Xiaoxitan, Qianjiang (1,640 m); 37. Xiaoxian, Youyang (1,645 m); 38. Bidong, Longshan (1,962 m); 39. Wentang, Dayong (2,498 m); 40. Shimaxi, Cili (1,928 m); 41. Maocaopu, Taoyuan (2,578 m); 42. Taifushan, Lingli (>1,105 m); 43. Hengchang, Shimen (2,100 m); 44. Longchihe, Shimen (1,536 m); 45. Hongjiayu, Sanzhi (2,184 m); 46. Qiyanquan, Hefeng (1,610 m); 47. Lianghekou, Xuan'en (1,818 m); 48. Gaoluo, Xuan'en (1,517 m); 49. Xiaoguan, Xuan'en (1,400 m); 50. Qiliao, Shizhu (1,417.4 m); 51. Taiyanghe, Enshi (1,376.7 m); 52. Taohua, Wushan (>1,128 m); 53. Siyangqiao, Badong (1,437.7 m); 54. Sandongshui, Changyang (1702.1 m); 55. Zhanxin, Changyang (1,680 m); 56. Dazhongba, Yichang (1,441.7 m); 57. Xujiaba, Wuxi (765.1 m); 58. Xianshui, Wuxi (ca. 800 m); 59. Tianba, Wuxi (>700 m); 60. Liziya, Xikou, Huayinshan (505.8 m).

uplifted, with the exception of five small marine bays situated on its western marginal belt. These areas have thick sequences of siliciclastics (mudstone and siltstone) intercalated with nodular limestone, marl, and dolostone. They represent near-shore, shallow-water environments that feature such shelly biotas as the *Retziella* brachiopod fauna (BA 2–3). The few marine embayments include eastern Yunnan and the Erlangshan, Jiangyou, and Guangyuan–Ningqiang areas (Fig. 30). In the latter area, deeper water graptolitic facies with flysch-like siliciclastics are also represented after the Wenlock. The western Qinling region was flooded by seawater from the Ludlow to Pridoli, as indicated by shallow-water biotas (BA 2–3) with brachiopods, tabulate and rugose corals, stromato-

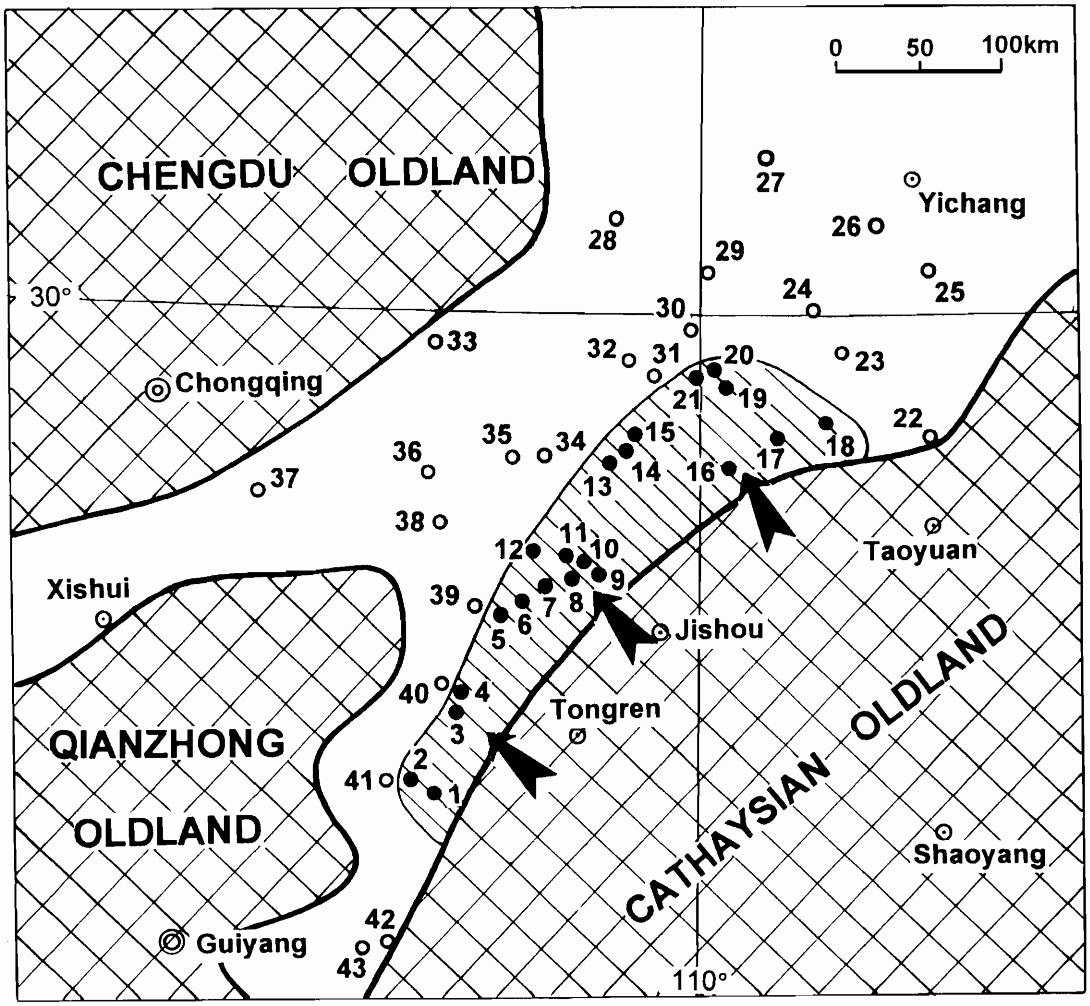

FIGURE 27 — Latest Telychian shallow-marine red beds of upper Yangtze region. Land areas cross-hatched. Areas with red beds shown by diagonal shading. Black and hollow circles mark sections with and without red beds, respectively. Black arrows show transport of large amounts of fine-grained siliciclastics rich in Fe_2O_3. 1, 2. Fengxiangping and Leijiatun, Shiqian; 3, 4. Chanxi and Heshui, Yinjiang; 5. Dongjiaping, Songtao; 6. Rongxi, 7, 8. Miaoquan and Xiaoyakou, Xiushan; 9. Kapeng, Huayuan; 10. Tianjiaping, Baojing; 11. Shidi, Xiushan; 12. Longtan, Youyang; 13–15. Bidong, Xiche, and Hongyanxi, Longshan; 16, 17. Wentang and Zhangjiajie, Dayong; 18. Shimaxi, Cili; 19–21. Xiaoxi, Hongjiayu, and Yunanxi, Sangzhi; 22. Taifushan, Lingli; 23, 24. Hengchang and Longchihe, Shimen; 25. south of Zhicheng; 26. Zhanxin, Changyang; 27. Siyangqiao, Badong; 28, 29. Taiyanghe and Guandiankou, Enshi; 30. Qiyanquan, Hefeng; 31, 32. Lianghekou and Gaoluo, Xuan'en; 33. Qiliaohe, Shizhu; 34. Xiaoxian, Youyang; 35. Dijiaba, Qianjiang; 36. Wanzu, Pengshui; 37. Sanquan, Nanchuan; 38, 39. Hongdu and Qiaojiazhai, Yanhe; 40. Zhoujiaba, Yinjiang; 41. Juntianba, Shiqian; 42, 43. Wengxiang and Luomian, Kaili.

poroids, and trilobites (Rong et al., 1987). A marine embayment may have existed in some parts of Jiangsu Province (Geng et al., 1997). All other parts of the South China paleoplate were continuously uplifted and denudated until the Late Devonian or Early Permian. Most of the Songpan–Ganzi block was still flooded by shallow seawater, and the Songpan Oldland was still situated in the middle of this block during this interval. The Silurian evolution of the South China platforms and basins is shown in Fig. 31.

OTHER REGIONS — Silurian seas flooded the Altai and Khingan regions, and siliciclastics with volcanics were deposited in the southern belt of the Siberian paleoplate. Shelly biotas, such as the *Tuvaella* fauna, inhabited shallow-water, mainly BA 2 to inner BA 3 settings (Su, 1981; Rong and Zhang, 1982). An oldland in the Khingan region and part of the Altai region was the source of a large amount of siliciclastics deposited in the nearby trough. The Llandovery sea was restricted to the Lesser Khingan area, northern Heilongjiang. Shallow-water sandstone and siltstone with tuff (Bashilixiaohe Formation) in the Lesser Khingan Mountains yield a *Tuvaella*

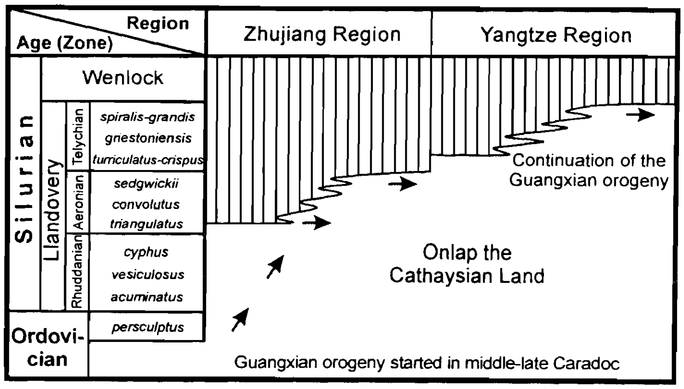

FIGURE 28 — Development of Guangxian orogeny and onlap of the Cathaysian Oldland in South China during later Ordovician–Early Silurian. Vertical ruling shows hiatus.

fauna, which indicates a relationship with Siberia. The shallow Ludlow sea, commonly with a *Tuvaella gigantea* assemblage in near-shore facies, expanded southwest to central Inner Mongolia. During the Pridoli, a regression took place, which is indicated by siliciclastics with a *Lingula* Community (BA 1; Xue et al., 1980).

In contrast to the Ordovician, most of the North China paleoplate was subaerial through the Silurian. Only the Inner Mongolia–Jilin trough had marine deposition in the northern marginal region. Two units can be roughly recognized. In the western part, later Silurian (Ludlow–Pridoli) siliciclastics with carbonates and shallow-water faunas are underlain by Ordovician island-arc deposits. In the eastern part, Llandovery graptolitic shale (Taoshan Formation; Liu, 1985) probably represents continental slope sedimentation, and the Ludlow–Pridoli

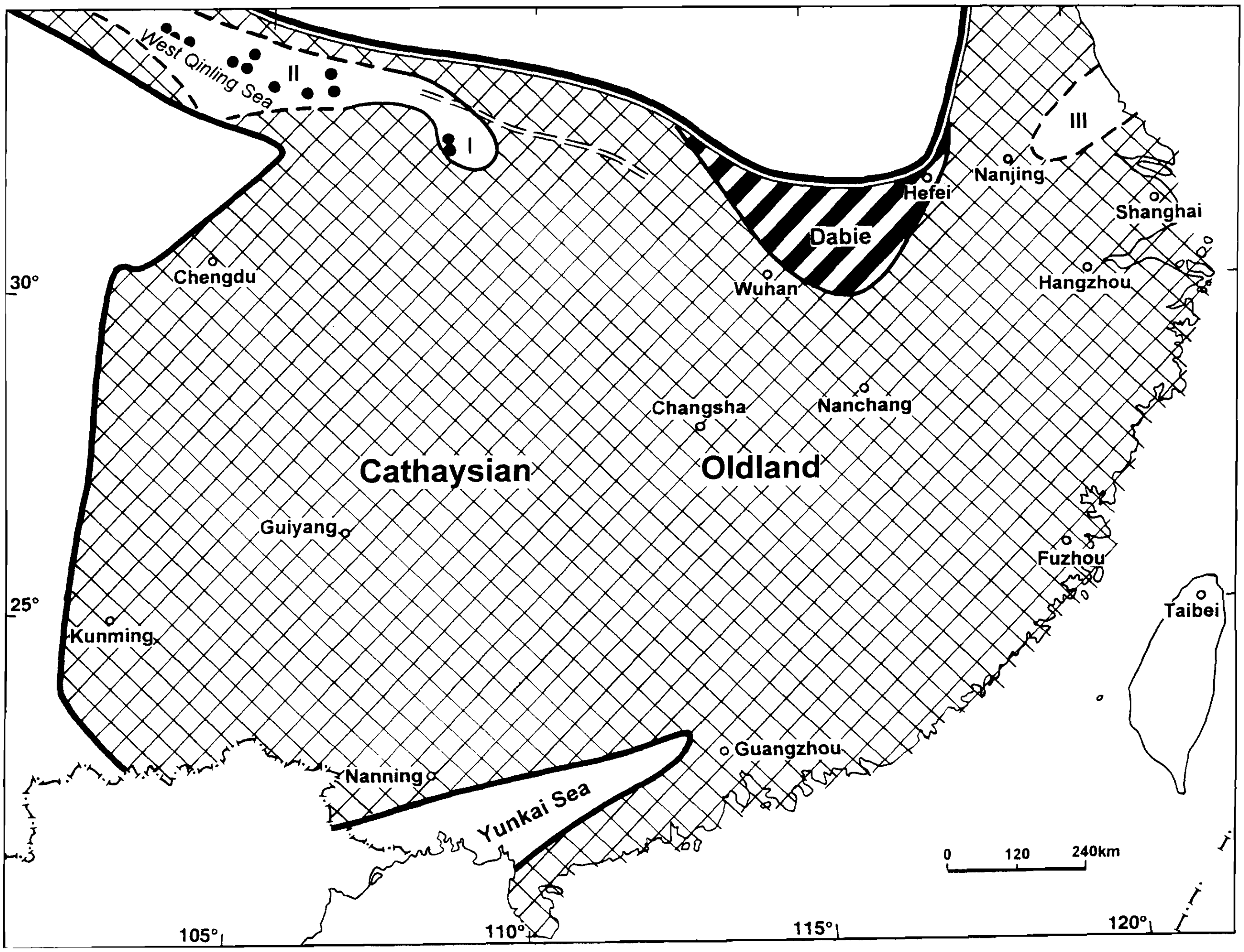

FIGURE 29 — South China in the Wenlock. Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. South China was uplifted during the early Wenlock with exception of I. south Qinling Basin, II. west Qinling Basin, and III. Jiangsu (There was an embayment with near-shore microbiotas (e.g., chitinozoans; Geng et al., 1997).

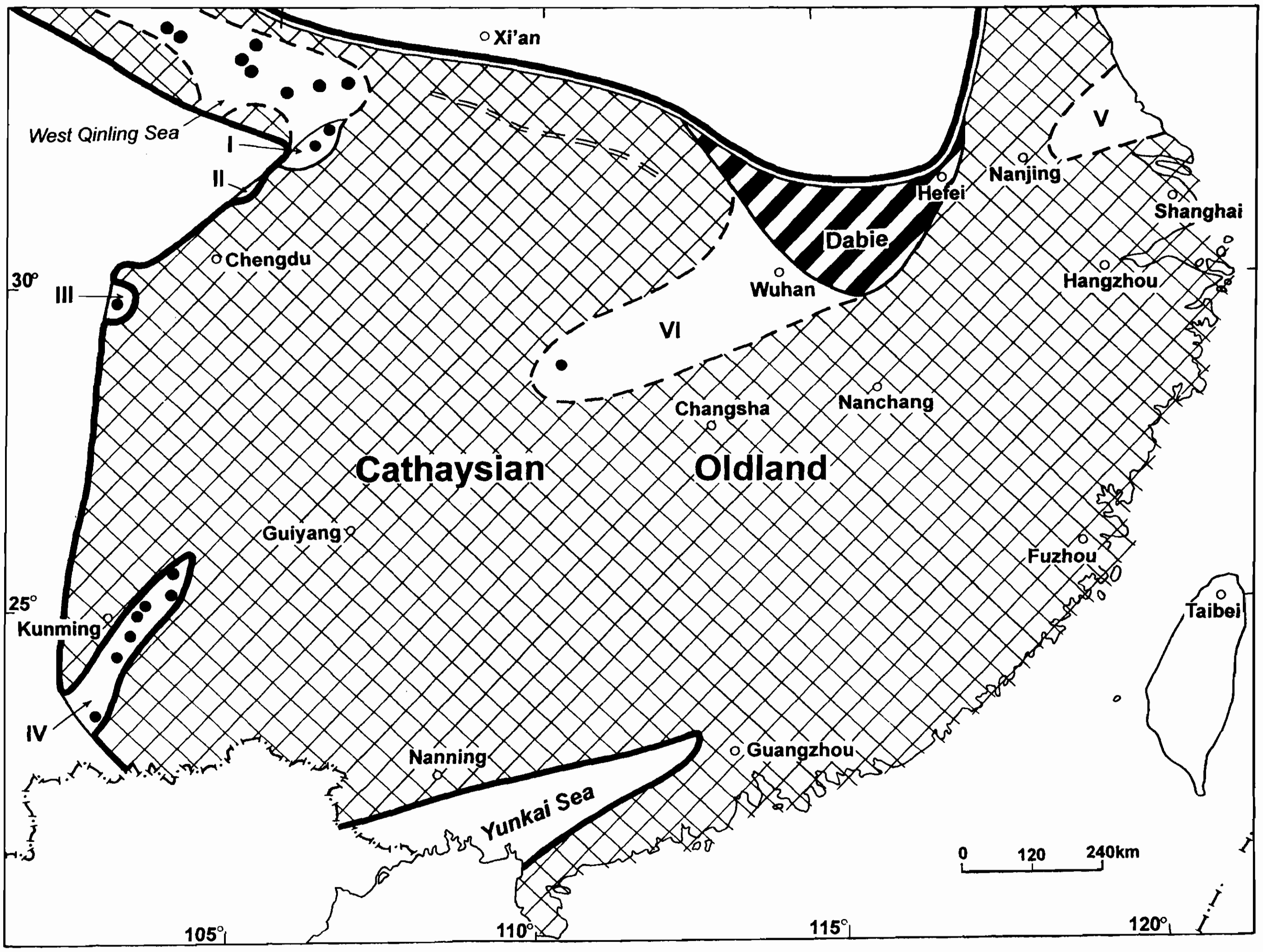

FIGURE 30 — South China in late Ludlow. Lijiang–Ninglang area, north Yunnan, not included in figure. Land areas cross-hatched. Marine embayments: I. Ningqiang–Guangyuan; II. Jiangyou; III. Erlangshan; IV. Qujing; V. Jiangsu (Geng et al., 1997). I–IV with shallow-water *Retziella* fauna. At Chengxi (black dot) on the Yunkai Sea, a near-shore, low-diversity *Retziella* assemblage occurs.

consists of siliciclastics with volcanics (Liu and Huang, 1977) that may represent a back-arc environment. A siliciclastic sequence with abundant intermediate and acid volcanics occurs in southern Mongolia and Jilin, and represents a residual marine enbayment in North China. An unconformity between the Silurian Xibiehe Formation and Ordovician igneous rocks marks the perimeter of a small paleo-island near Bater Obo in central Inner Mongolia near the northern margin of the North China paleoplate. Burial of the island took place with the onset of eustatic rise that peaked in the late Ludlow (Rong et al., 2000; Johnson et al., 2001).

On the Qaidam paleoplate, there were three oldlands (Alxa in the north, central Qilian in the middle, and Qaidam in the south) that were separated by three marine regions (Figs. 32, 33). The Yumen–Tongxin sea in the northern Qaidam area is composed of two parts. The northern part probably represents a foreland with chiefly siliciclastics, and the southern part represents a deeper trough with volcaniclastics. Gray-yellowish and purple-red sandstone and siltstone with shallow-water limestone intercalations (parts of the Quannaogou and Hanxia Formations) appear in the northern Qilian Mountains, and represent a molasse-like facies. In the Tongxin area, the upper Rhuddanian has a shallow-water shelly fauna (e.g., *Eospirifer, Holorhynchus, Tungussophyllum, Paramplexoides*); an Aeronian–Telychian molasse-like facies with red or gray-yellowish siltstone, sandstone, and conglomerate is also prominent. The Tongxin area was uplifted probably in the Wenlock, and the Yumen

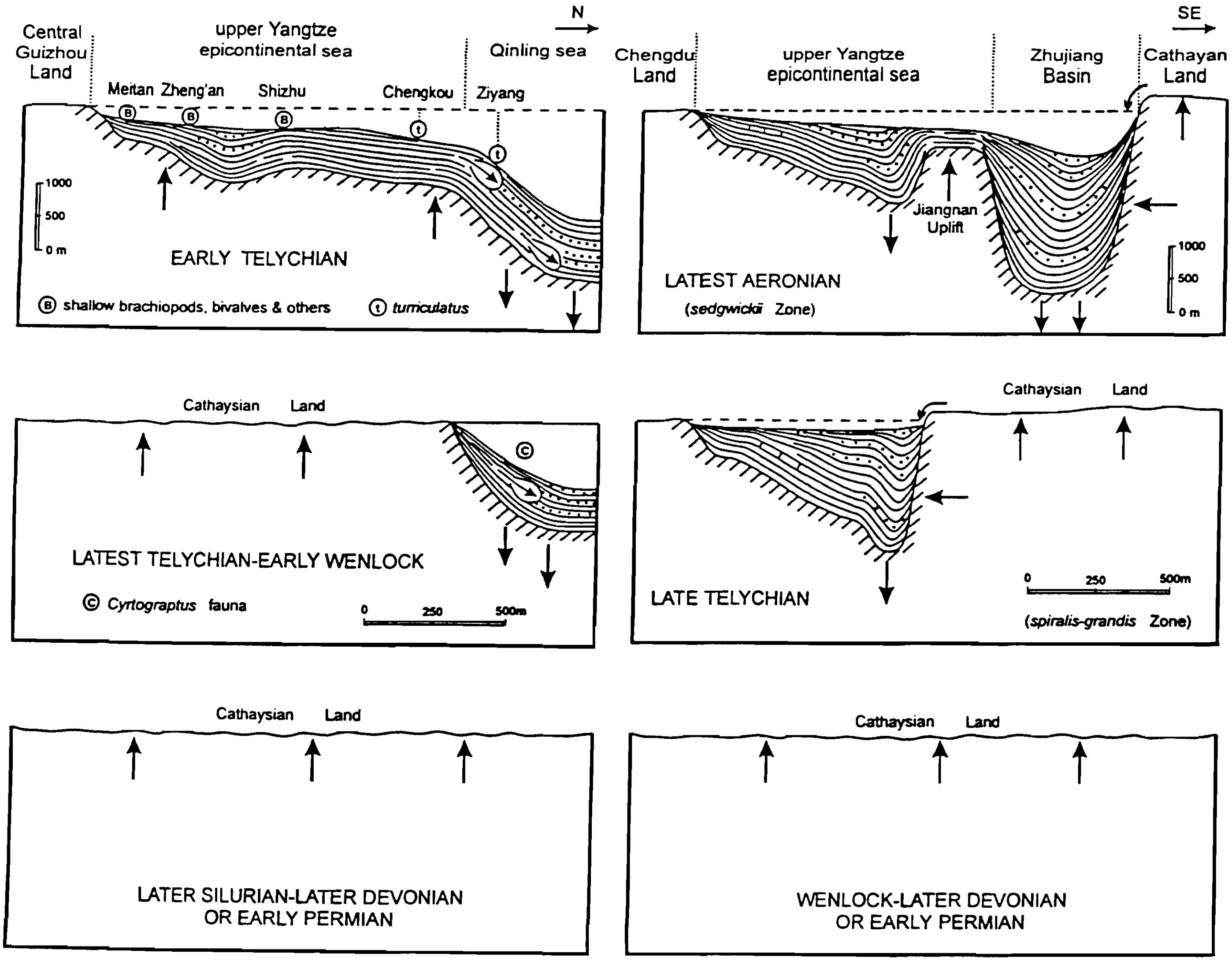

FIGURE 31 — Evolution of submarine topography, facies, and relative subsidence of South China during the Silurian.

area in the later Silurian. Little is known about synecology and sea-level changes for the Silurian in the other areas of the Qaidam Plate. Muddy siliciclastics with abundant volcaniclastics represent an active rift system in north Lüliangshan on the north margin of Qaidam.

On the easternmost Kazakhstan paleoplate, the Silurian occurs mainly in the mobile belts surrounding the Junggar Oldland in Xinjiang. There are two lithofacies belts; one consists of siliciclastics with carbonates, and the other is a volcaniclastic belt along the Altai Mountains. On the west margin of the Junggar Oldland, a belt of volcaniclastics, graptolitic shale, and carbonate lenses occurs northwest of the Ebinur Lake–Klamayi–Hoboksar line (Ni et al., 1995). The Silurian east mobile belt at Junggar may not be preserved, possibly due to Hercynian orogeny (Ni et al., 1995). Two lithofacies belts with siliciclastics, carbonates, and volcaniclastics occur along the north margin of the Yi'ning Oldland. Both the Yi'ning and Junggar Oldlands were the source areas of materials deposited in the surrounding belts. Generally, the geographic framework of northern Xinjiang was stable from the Llandovery to the Pridoli. Wenlock tuff, tuffaceous siltstone and sandstone, andesitic breccia, and intercalated limestone beds and lenses (Shaerbuer Formation) with a shallow-water benthos (rugose and tabulate corals, brachiopods, trilobites, bivalves) occur in the Junggar and Beishan areas (Zhou et al., 1995). These areas belong to the island-arc systems of the Kazakhstan paleo-

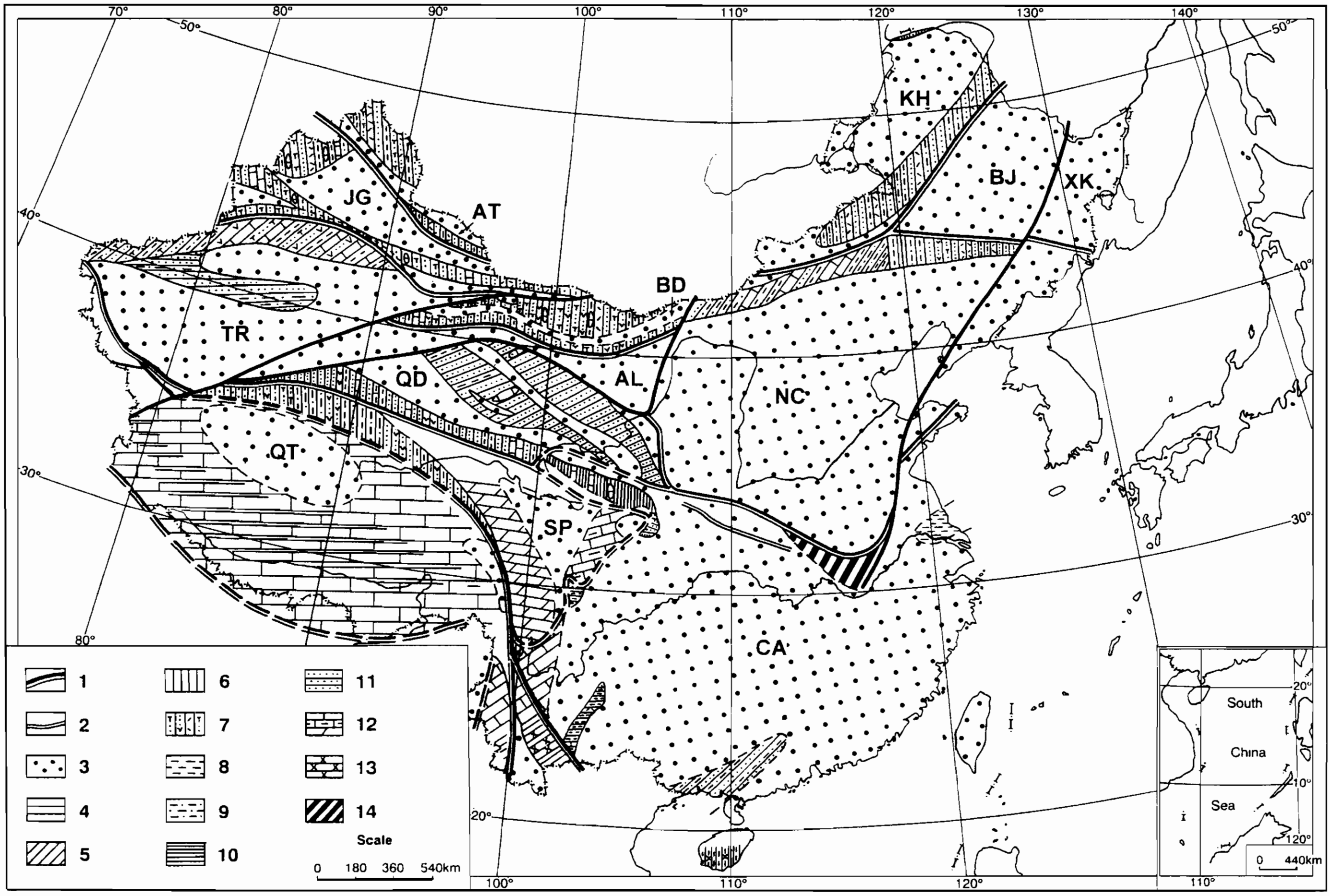

FIGURE 32 — China in the early Wenlock. Probable shoreline at maximum onlap and marine facies shown. Oldlands are dotted: AL, Alaxa–Dunhuang; AT, Altai; BD, Badainjaria; CA, Cathaysia; JG, Junggar; BJ, Bureya–Jamus; KH, Khingan; NC, North China; QD, Qaidam; QT, Qangtang; SP, Songpan; TR, Tarim; XK, Xingkai. Numbers: 1. plate boundary; 2. rift zone; 3. oldland; 4. stable marine platform; 5. transitional marine platform; 6. marine mobile belt; 7. volcaniclastic fragments; 8. mudstone; 9. siltstone; 10. shale; 11. sandstone; 12. limestone and dolostone; 13. siliceous rocks; 14. Dabie block (discussed in text). Extent of marine onlap of Jiangsu Province, east China, based on Geng et al. (1997).

plate.

The Ordovician and Silurian paleogeographies of Tarim are conspicuously different. The Mangar Basin and its west slope existed during the Arenigian to Caradocian (Chen et al., 1995), but disappeared in the Ashgillian with sea-level fall. The Tarim platform became largely subaerial, but a large central bay persisted. Llandovery shallow-water siliciclastics were deposited in this bay. During the Wenlock, the Tarim oldland area expanded northward into the Ahake–Kuluketag belt. Near-shore, very shallow-water, marine red beds (Imogantau Formation, mainly siltstone and sandstone) are well developed in the Kalping–Bachu area, western Tarim. This embayment narrowed during the Ludlow. Littoral and channel deposits filled it, and extended into central Tarim (Geng 1994), a region connected to the South Tianshan deeper-sea trough, or mobile belt. Thick volcaniclastics, siliciclastics, and carbonates were deposited in this belt in the Llandovery–Pridoli (Ni et al., 1996). Information is not available on the Silurian in the southern marginal areas of Tarim. In southern Tianshan, thick detrital, muddy, and organic-rich rocks occur with volcaniclastics in the northern marginal belt of Tarim.

The Silurian paleogeography of Tibet is only roughly known. On the Tibet platform, the Llandovery (Dewukaxia Formation) is mostly composed of graptolitic shale and mudstone, similar to those in western Yunnan (lower Jenhochiao Formation). Most of Tibet was covered by passive margin, shallow- to deeper-water carbonates in the later Silurian. A late Llandovery–early Wenlock reefal buildup was discovered in the neighboring Spiti–Kinnaur area, Himachal Himalaya, India (Bhargava and Bassi, 1986). In the Xainza area, thin Wenlock carbonates are well developed, but fossils are not abundant, with the

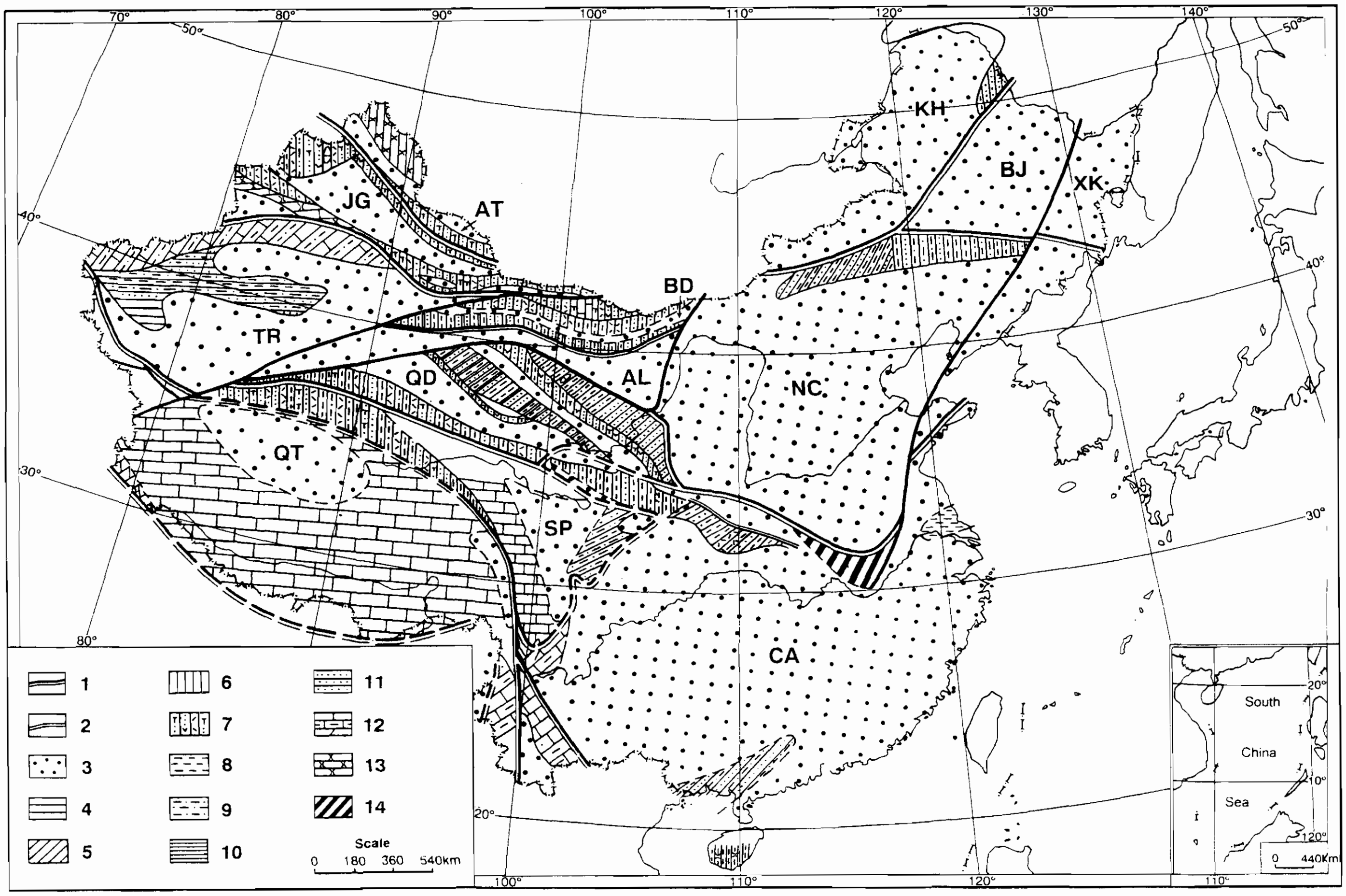

FIGURE 33 — China in the Ludlow. For explanation of symbols, see Fig. 30. Extent of marine onlap of Jiangsu Province, east China, after Geng et al. (1997).

exception of nautiloids and conodonts. Other benthic shelly faunas are not known in this region. In the Lazhulong area at Rotug in northwesternmost Tibet near the Qiangtang Oldland (Wang, 1985), passive-margin, shallow-water conditions with sandy–muddy and carbonate sediments accumulated with common nautiloids and conodonts. A similar Silurian sequence is well developed in the Karakorum region (Wen et al., 1996). Distribution of the Tibetan sea and land during the Ludlow–Pridoli was almost the same as during earlier intervals. Limestones with nautiloids and conodonts are widely distributed on the Tibetan platform (Wang, 1985).

Silurian limestone and marl with siliceous shale occur in western Yunnan (Ni et al., 1982). Nautiloids and conodonts are present in the limestone, and graptolites in the intercalations.

Climate Indications

The Silurian climate gradient was significantly less than those of the Late Ordovician or Early Devonian, and no evidence of widespread continental glaciation is known. However, it is clear that there was a relatively cool southern hemisphere with low-diversity faunas. This is represented by the Malvino-Kaffric Realm, which lacks carbonates but has marine red beds with distinctive fossils (Boucot, 1975, 1991). The Malvino-Kaffric Realm was bounded to the north by the warm-water North Silurian Realm. The Silurian rocks and fossils of these two realms demonstrate a moderate global climate gradient through the period.

All the Silurian paleoplates or blocks of China were in the North Silurian Realm; this is based on biogeographical, lithofacies, and paleomagnetic data. No climatically sensitive sediments, such as bauxites, kaolinites, or evaporites, are recorded in the Silurian of China. In a broad sense, Silurian carbonates are not well developed in China. This is probably due to the fact that Silurian tectonics featured strong uplift and subsidence that brought a large influx of siliciclastic sediments into the Chinese basins. Lower Silurian (Llandovery) carbonate mud mounds and bioherms, nevertheless, are sporadically developed in some regions of China, in particular the upper Yangtze region. Bioherms and carbonate mud mounds occur in the upper Telychian of the Ningqiang–Guangyuan border area and Lan'gao area, respectively, of southeast Shaanxi, and along the northern margin of the upper Yangtze platform of the South China paleoplate. These build-ups feature mainly tabulates and stromatoporoids, sometimes with relatively high diversity at many localities and in certain horizons in the upper Aeronian and upper Telychian on the upper Yangtze Platform. Carbonate mud mounds also occur in the Ludlow of the Baterobo area, Darhan Mumingan Joint Banner, Inner Mongolia, on the northern marginal belt of the North China paleoplate (Li et al., 1985). Such occurrences show that these continents may not have been far from the Silurian equator, and that they had a warm climate. No Silurian calcretes have been documented. However, Early Devonian calcretes are recorded from the Xujiachong Member of the Cuifengshan (Tsuifengshan) Formation in the Qujing area, eastern Yunnan, southwest China (Boucot et al., 1982). They indicate wet–dry seasonality in eastern Yunnan in the Early Devonian (Wang, 1997).

Faunal evidence can help interpret ancient climates. Different graptolite and conodont faunas from the Ordovician are recognized by many paleontologists. These include the Atlantic (cooler and deeper water) and Pacific (warmer and shallower) graptolite faunas (Chen, 1994), as well as the North Atlantic (cooler and deeper) and the Mid-Continent (warmer and shallower) conodont faunas (Wang et al., 1996). However, Silurian graptolites and conodonts show very little endemism, and this is also apparently true for chitinozoans (Laufeld, 1979). All of these data suggest a uniform, warm-water environment.

The origin of marine red clastic beds is contentious. Marine red beds with a shallow-water origin may not be considered as a climatic indicator, in general. However, the continuous sequence of later Llandovery–Pridoli, shallow-marine red beds (Tartaertag, Imogantau, and Kerzirtag Formations) on the northwest margin of the Tarim platform, and those (Hanxia Formation) on the northern Qaidam block, may indicate a low latitude, warm-water environment. A discontinuous sequence of upper Llandovery marine red beds (Rongxi and Huixingshao Formations) along the southeast edge of the upper Yangtze platform near the Cathayasian Oldland in South China may not be a definite climatic indicator. All of these marine red beds were deposited near the oldlands where the Fe_2O_3 developed in the source materials under arid or warm conditions might not have been reduced to FeO when they were deposited in the nearby basins.

Paleogeography of Chinese Paleoplates

The largest Silurian paleocontinent was Gondwana. Cocks and Scotese (1991) included Florida, southern Europe, Turkey, the Arabian Peninsula, Iran, south China, Tibet (Qiangtang and Lhasa blocks), the Shan–Thai (or Sibumasu) block, New Guinea, eastern Australia, and New Zealand in Gondwana. All of these areas have Silurian rocks. They also noted that Lower Ordovician facies and faunas made it probable that South China formed a marginal part of Gondwana (Cocks and Scotese, 1991). Whether or not it formed an actual part of Gondwana in the Silurian is still uncertain. Regarding the Silurian paleogeography of South China and its later geological history, we believe that most of the other paleoplates or blocks may have been related to South China on the basis of biogeographic, sedimentary, and paleomagnetic data.

Paleontology is one of several disciplines used to reconstruct paleogeographical maps. Fossil distributions are important in interpretation of paleogeographic and tectonic problems. Fossils are also independent of tectonic models (Fortey and Cocks, 1986). Furthermore, the fossil record provides a history of biological evolution, which is also useful in reconstructing the histories of various paleoplates. An enormous amount of data is published on Silurian faunas in many parts of China. The best-substantiated view, derived from biogeographical information, demonstrates that three larger units of Chinese paleoplates can be recognized by the presence or absence of the distinctive *Tuvaella* and *Retziella* brachiopod faunas and the *Kopaninoceras* nautiloid fauna.

South China, Qaidam, Tarim, North China, and Indochina Region

South China, Qaidam, and Tarim — A revision of genus-level taxonomy is necessary for paleobiogeographic analysis and reconstruction of South China, Qaidam, and Tarim. Based on the work done by Wu H.-j. and P. Lane (*in* Chen and Rong, 1996), four Aeronian and Telychian trilobite genera of South China (*Luojiashania*

Zhang, 1974; *Chuanqianoproetus* Wu, 1979; *Latiproetus* Lu, 1957; and *Xiushuiproetus* Zhang [*in* Qiu et al., 1983]), as well as *Zhejiangoproetus* Zhang (*in* Qiu et al., 1983), are junior synonyms of *Astroproetus* Begg, 1933. In addition, *Senticuculus* Xia (*in* Zhang, 1974) is a subjective synonym of *Coronocephalus* Grabau, 1924. Two brachiopod genera (*Megaspinochonetes* Yang and Rong, 1982, and *Shiqianella* Xian [*in* Xian and Jiang, 1978]) from the upper Llandovery of South China have been treated as subjective synonyms of *Spinochonetes* Rong et al., 1974. *Paraconchidium* Rong et al., 1974, is a synonym of *Pseudoconchidium* Nikiforova and Sapel'nikov, 1971. Re-evaluation of the geographical distribution of genera is also useful in this type of study. *Amplexoides* Wang and *Ceriaster* Lindström were regarded as endemic coral genera in the Yantzean and Kazakhstan–Sinokorean Provinces (Wang and Chen, 1991). According to He (1994), *Amplexoides* also occurs in the Qinling and Qilian Mountains of Tianshan, Inner Mongolia, Tibet, Australia, Canada, and Greenland. Rare specimens of Ceriaster have been encountered in the British Isles (Deng Z.-q. and C. T. Scrutton in Chen and Rong, 1996). Trilobites in Australia and Japan recorded as *Coronocephalus* are incorrectly identified, and that genus is considered endemic to South China (Wu H.-j. and P. Lane *in* Chen and Rong, 1996). In summary, the most important characters in paleogeography are 1) the site of origination of many groups, 2) presence of endemic genera with a certain dominance, and 3) the unique features of geologic history (discussed below).

There are many faunal groups that originated in South China. The earliest representative of the Foliomenidae (brachiopods) is *Foliomena jielingensis* (Zeng) from the Miaopo Formation (*Nemagraptus gracilis* Zone, lower Caradocian) of western Hubei. This species is associated with other deeper-water brachiopods, and the association is the earliest known record of the *Foliomena* fauna in the world (Rong and Zhan, 1995). A second taxon with earliest representation in China is *Eospirifer praecursor* Rong et al. This earliest known eospiriferine occurs in the Xiazhen and Changwu Formations (middle Ashgillian) of northeast Jiangxi and southwest Zhejiang (Rong et al., 1994). The oldest Retziellidae are recorded mainly from the Ludlow–Pridoli of South China, North China, and Tarim (Rong et al., 1994). Their earliest known genus is *Metathyrisina*, which occurs in the Leijiatun Formation (upper Aeronian) of northeastern Guizhou. The earliest species of *Retziella* occurs in the upper Wenlock in west Qinling on the South China Plate (Fu, 1982). As a fourth example, the earliest known species of *Atrypoidea*, *A. lentiformis* (Wang), occurs in the Xiushan Formation of Changning, southwestern Sichuan (Wang et al., 1980). The genus is commonly recorded from Ludlow–Pridoli rocks in many areas of the world.

Some of the earliest representatives of a number of rugose coral groups are known from the Late Ordovician and early–middle Llandovery of the upper Yangtze region, south China (He, 1994; He and Chen, 1998). These include the calostylids (*Calostylis*, *Yohophyllum*), streptelasmatids (*Tunguselasma*), pycnactids (*Briantelasma*), amplexids (*Synamplexides*, *Amplexoides*), stauriids (*Eostauria*, *Ceriaster*, *Stauria*), and cystiphyllids (*Rhizophyllum*, *Maikottia*).

According to Chen T.-e. (personal commun., 1996), the earliest known Eriditidae, Jovellaniidae, and Nothoceratidae (nautiloids), respectively, include *Malgaoceras* from the Xiangshuyuan Formation (Aeronian), *Mixosiphonoceras* from the Leijiatun Formation (late Aeronian), and *Yichangoceras* from the Xiangshuyuan Formation (Aeronian) of northeastern Guizhou. Similarly, *Virgoceras* is known to occur from the Wenlock and the higher Silurian of southern Xizang, Europe, and North America, but occurs earlier in the Ningqiang Formation (upper Telychian) of northern Sichuan.

Pinnatiramosus is the earliest known vascular plant, and its single-retusoid trilete spores occur in yellow-green argillaceous siltstone of the Hanchiatien Formation at Dongkala, east of Fenggang, northern Guizhou, southwest China (Geng et al., 1997). This formation was regarded as later Silurian in the 1960s, but is now regarded as early Telychian on the basis of brachiopods (*Nalivkinia* sp. cf. *N. elongata* [Wang], *Nucleospira pulchra* Rong et al.), chitinozoans (*Eisenackitina daozhenensis* Geng), and regional stratigraphic correlation (Rong et al., 1990; Cai et al., 1996). These root-like branches were assumed to have penetrated downward from the Permian into the Silurian. The plant bed is 2.2m below the Permian Tongkuangchi Formation, however, and no fossil plants, including rootlets, have been found in the intervening interval (Cai et al., 1996). This locality with the earliest vascular plants was close to the paleo-shoreline. Preservation (usually parallel or slightly oblique to bedding) seems to indicate that they may have been buried in situ or transported only a short distance by near-shore currents. Expansion of the Cathaysian and Qianzhong Oldlands, suitable climatic conditions, and optimum marine environments near shore (mainly BA1–upper BA2) may have promoted the origination of the earliest known vascular plants in South China.

Although Llandovery faunas of South China have many cosmopolitan forms, they include 15–40% endemic genera. These endemics include: 1) brachiopods (e.g., *Athyrisinoides* [=*Kritorhynchia*], *Atrypinopsis*, *Beitaia*, *Dabashanospira*, *Pleurodium*, *Plicidium*, *Quangyuania*, *Spinochonetes* [=*Megaspinochonetes* and *Shiqianella*], *Xinanospirifer*, *Yidurella*) (Rong and Yang, 1981; Rong J.-y. and M. G. Bassett in Chen and Rong, 1996); 2) rugose corals (e.g., *Fengganophyllum, Gyalophylloides, Nanshanophyllum, Neostreptelasma, Paraceriaster, Paramplexoides, Parastauria, Shenxiphyllum, Yangziphyllum*) and tabulates (*Erlangbapora, Lankaolites, Meitanopora, Mesosoleniella, Neofletcheriella, Ningqiangolites, Pachystelliporella, Qianbeilites, Shanxipora, Somphoporella*) (Ge and Yu, 1974; Deng Z.-q. and Scrutton *in* Chen and Rong, 1996; He and Chen, 1998); 3) trilobites (*Coronaspis, Coronocephalus, Kailia, Meitanillaenus, Parakailia, Rongxiella, Shiqiania, Songkania*) (Wu H.-j. and P. Lane *in* Chen and Rong, 1996); 4) nautiloids (*Chuandianoceras, Guangyuanoceras, Jialingjiangoceras, Kailiceras, Neosichuanoceras, Paramixosiphonoceras, Protophragmoceras, Sichuanoceras, Songkanoceras, Yichangoceras*) (Chen et al., 1981; Chen T.-e. and C. H. Holland *in* Chen and Rong, 1996); 5) crinoids (*Spirocrinus, Dazhucrinus*) (Mu and Wu, 1974); 6) vertebrates (acanthodians: *Sinacanthus, Neosinacanthus;* galeaspids: *Eugaleaspis, Hanyangaspis, Sinogaleaspis, Wangolepis, Xiushuiaspis*) (Blieck and Janvier, 1991; Liu, 1995); 7) benthic graptolites (*Hunanodendrum*) (Mu et al., 1974); and 8) conodonts (e.g., *Ozarkodina guizhouensis, Oulodus shiqianensis, Ctenognathoides? qianensis*) (Zhou et al., 1981).

Llandovery brachiopods of South China are characterized by the absence of some widely distributed genera. For example, *Eocoelia and Stricklandia* are common in Europe and North America, but neither genus has yet been recorded from South China. Instead of *Eocoelia,* the atrypoid genus *Nalivkinia* is abundant in South China, where it is usually associated with *Nucleospira* in mostly BA 2–inner BA 3 facies (Rong, 1986). *"Stricklandia" transversa* Grabau, 1925, has well-developed interareas, a wide hinge-line, a shallow ventral sulcus and dorsal fold, and two sets of crossing ornament; these are features that are remarkably different from *Stricklandia* s.s. Species such as *"S." transversa* are under study because they are referable to a new genus related to *Kulumbella* (M. Bassett and Rong J.-y., unpublished data). *Pentamerus oblongus,* one of the most common Llandovery brachiopods elsewhere, is rare in South China, with only a few specimens known from late Aeronian limestones in northeast Guizhou and west Hubei (Rong and Yang, 1981). Furthermore, some Llandovery rugose corals are endemic to South China, and some Llandovery rugose genera (*Amplexoides, Stauria,* and *Rhizophyllum*) that originated in South China did not arrive in Europe and North America until the late Llandovery or Wenlock (He, 1994). These data indicate that barriers existed between South China and other continents in most of the Llandovery.

The *Retziella* fauna is known from the upper Wenlock–Pridoli of South China, North China, central Asia (the border area of Tadzhikistan and Kirkizistan), Indochina (including northern and central Vietnam), east Australia, New Zealand(?), Afghanistan(?), and Pamir(?). This fauna belongs to the Sino–Australian Province of later Silurian brachiopods (Rong et al., 1994, 1995). Work by Rong J.-y. and Zhang Z.-x. (unpublished data) suggests that *Retziella* also occurs in the Ludlow–Pridoli of Heiyingshan in Baicheng County within the northern Tarim marginal belt.

TARIM — Rong and Zhan (1996) suggested that early eospiriferines in the Rhuddanian and lower Aeronian may indicate a closer relationship of Tarim and South China, and may be used to analyze brachiopod biogeography. These fossils have been recorded only from South China, Kazakhstan, and east Australia. *Eospirifer sinensis* Rong et al., *Isorthis qianbeiensis* (Rong et al.), and *Beitaia* sp. cf. *B. modica* Rong et al., as identified by Rong (unpublished data), are known from the middle Kalpingtag Formation in the Kalping–Bachu area of the northwest Tarim marginal belt. An association of these taxa is common in the upper Rhuddanian–lower Aeronian of northeast Guizhou. Furthermore, later Llandovery *Sinacanthus* and *Hanyangaspis,* formerly considered endemic vertebrates in South China, have been found in the overlying Tataertag and Imogantau Formations of the Kalping–Bachu area. These occurrences strongly indicate proximity of Tarim and South China in the Early Silurian. It has been demonstrated that the Tarim region was close to South China, North China, Australia, and Kazakhstan based on Caradocian trilobite faunas with *Lisogorites, Ampyxiella,* and *Taklamakania* (Zhou et al., 1995). This has also been confirmed by the study of Ordovician litho- and biofacies (Zhou and Chen, 1990). Therefore, we conclude that the *Retziella* (brachiopod) fauna, early eospiriferines, *Sinacanthus* fish fossils, extensive development of shallow marine red beds, and other evidence from the preceding Ordovician all show that Tarim was located near the South China plate.

QAIDAM — Qaidam is suggested herein to have been near South China. Some Silurian genera, are common only to South China and Qaidam. *Nanshanophyllum* Yu (rugose coral) is known only from the Telychian and Wenlock of South China and the Qilian Mountains of Qaidam. *Hunanodendrum* Mu et al. (graptolite) is a distinctive, latest Aeronian–early Telychian genus in South China, and is found in coeval horizons in the Minle area of Gansu in the Qaidam. *Eospirifer* and *Striispirifer* are found in the Rhuddanian–lower Aeronian Zhaohuajing Formation of Tongxin, central Ningxia (Fu, 1982), which belonged to the Qaidam plate. The Zhaohuajing Formation has the following corals: 1) *Paramplexoides,* thought to be unique to South China; 2) *Dinophyllum* and *Grewingkia,* known from Europe and South China; 3) *Tungussophyllum* and *Profieviella,* previously known only from Siberia (He X.-y. in Zhu et al., 1987; He, 1994); and 4) the earliest known species of *Tunguselasma,* which is only known in the lower–middle Llandovery of South China and central Ningxia (He and Chen, 1998). These data suggest a close biogeographical relationship of South China and Qaidam and show a degree of Silurian faunal exchange with Siberia.

In the Devonian, northeastern Qaidam may be assigned to the South China biogeographic province. This is suggested by ostracodes (sinoleperditiini), fish (galeaspids, *Remigolepis, Sinolepis,* sarcopterygii), and plants (*Leptophloeum rhombicum, Sublepidodendron mirabile, Eolepidodendron wusihenesi*) in Qaidam that were previously known only from South China (Wang et al., 1995).

NORTH CHINA — The Ashgillian, Silurian, and Devonian are not yet known on the North China Platform. The later Silurian is well developed in the Baterobo area of Darhan Mumingan Joint Banner, southern Inner Mongolia (Li et al., 1985), and in Yongji, central Jilin (Liu and Huang, 1977). Both of these areas are in the northern marginal belt of North China. A *Retziella* fauna with *Retziella, Atrypoidea, Gashaomiaoia,* and many other brachiopods is known in these regions. The most interesting discovery is that a probable *Retziella* fauna occurs south of Pingyong, North Korea (Yang, 1989; An and Ma, 1993). The dominant element of this probable Late Silurian brachiopod fauna was identified as *Protathyrisina*, a junior subjective synonym of *Retziella* (Rong et al., 1994). Moreover, *Xinanospirifer flabellum* (Rong et al.), *Nalivkinia elongata* (Wang), and other brachiopods common in the upper Llandovery of South China are arecorded in North Korea. The latter region, therefore, may be considered to be part of the North China paleoplate and near South China during the Silurian.

INDOCHINA — During the Early Ordovician, Indochina may have been at a higher latitude than that shown by Scotese and McKerrow (1991) and Metcalfe (1995). This proposal is based on the study of trilobites by Zhou et al. (1998), who suggested that Indochina was closer to the south-central Europe block than to South China. However, Rong et al. (1995) concluded that brachiopods of the near-shore, shallow-water *Retziella* fauna suggest the Indochina paleoplate was near South China in the Late Silurian. As the Yunkai block has a Ludlow–Pridoli brachiopod assemblage with a single species, *Retziella* sp., it may have been located between South China and Indochina.

TIBET AND WEST YUNNAN (NORTHERNMOST SIBUMASU PALEOPLATE)

The *Kopaninoceras-Michelinoceras* nautiloid fauna, with *Kopaninoceras dorsatum* (Barrande), *K. jucundum* (Barrande), *K. capax* (Barrande), *Oonoceras plebeium* (Barrande), *Virgoceras palemon* (Barrande), and many others, occurs in western Yunnan and Tibet. *Tuvaella* and *Retziella* faunas are found in these two regions. The Wenlock and Ludlow nautiloids of western Yunnan and Tibet do not seem to have been transported after death (Chen T.-e., personal commun., 1996). The later Silurian nautiloids of Tibet and western Yunnan have a closer affinity with Sardinia and Bohemia (Serpagli and Gnoli, 1977). They are remarkably different from those of South China, which has the endemic *Heyuncunoceras* and *Yunnanoceras* nautiloid faunas (Chen et al., 1981).

Later Homerian graptolite faunas, with *Plectograptus* (*Sokolovograptus*) *textor* Boucek and Munch, *Colonograptus deubeli* (Jaeger), *C. praedeubeli* (Jaeger), *C. ludensis* (Murchison), *C. schedidoneus* Lenz, and other species, occur in the lower Zhongcao Formation in the Shidian area, western Yunnan (Ni, 1997). This formation is underlain by dark grey, thin to medium-thick bedded limestone with the nautiloid *Kopaninoceras juncundum* (Barrande), and is overlain by light to yellowish-gray marl with the conodont *Kockolella variabilis* Walliser (Ni et al., 1982). It should be pointed out that no late Wenlock graptolites have been found on the South China plate, with the exception of those in Yulin, Guangxi (Ni, 1997).

Silurian conodonts in Tibet contain some distinctive species of the European fauna, such as *Kockella variabilis* Walliser, *Polygnathoides siluricus* Branson and Mehl, and *Ancoradella ploeckensis* Walliser. The first species also occurs in the upper Ludlow of western Yunnan, western and central Europe, Malaysia, and Siberia (Qiu, 1988).

Brachiopods, such as *Dnestrina* sp. cf. *D. gutta* Nikiforova and Modzalevskaya, *Lissatrypa leprosa* Kozlowski, and *Plectodonta mimica* Barrande, are common in the uppermost Silurian and lowest Devonian of Podolia, Bohemia, and elsewhere in Europe. These species are also found in the Shidian–Baoshan area, western Yunnan (Jahnke and Shi, 1989), where the *Retziella, Coronocephalus,* and *Sinacanthus* fauna and many of the endemic genera of South China are unknown. However, the *Dayia* brachiopod fauna, common locally in Europe and westernmost Turkey, has not been recorded in Tibet, western Yunnan, Thailand, and Burma. It is still uncertain whether this absence was caused by its isolation or, more likely, reflects local predominance of a graptolitic facies.

Although shallow-water benthic assemblages in Tibet and west Yunnan are not well known, a number of paleogeographical observations can be made: 1) Latest Silurian–earliest Devonian scyphocrinoids (e.g., *Camarocrinus asiaticus* Reed, mainly lower Lochkovian) are well known in the Shidian–Baoshan area, western Yunnan. They are also known from Europe, North Africa, and North America (Jahnke and Shi, 1989), as well as northern Shan State in Burma and peninsular Malaysia (Witzke et al., 1979). None of these bizarre crinoids have been found in South China or Australia (Fang, 1994). 2) Silurian brachiopods and corals from Kuala Lumpur and the Kinta Valley of peninsular Malaysia are closely related to those in Uralian faunas (Boucot et al., 1966; Berry and Boucot, 1972) and Europe (Igo, 1984), respectively. 3) An Eifelian (Middle Devonian) brachiopod fauna known from the Rhenish–Bohemian region appears in the Shidian–Baoshan area, western Yunnan (Hou, 1988) and northern Shan State (Anderson et al., 1969). This brachiopod fauna is evidently different from coeval brachiopod faunas of South China and Australia. 4) Middle Devonian rugose corals from the Shidian–Baoshan area are closely related to those of western Europe (Wang et al., 1989). 5) A typical Hercynian trilobite assemblage of probable Emsian (Early Devonian) age from Satun, southwest Thailand, indicates a close relation to Turkey, Bohemia, Germany, and Morocco; these areas compose the the Rhenish–Bohemian region of the Old World Realm in the Early Devonian (Boucot, 1975; 1991).

Tibet and western Yunnan are located between South China and southern Europe (Fig. 34). The Tibet and Sibumasu paleoplates have been regarded as a single continent (that included Kreios) in the Early Paleozoic and Devonian (Metcalfe, 1992, 1995). This hypothesis can not yet be precluded. Tibet, Turkey, Iran, Afghanistan, and Burma–Malaya (=Sibumasu) formed a Cimmerian continent that rifted away from the northern margin of Gondwana in the later Paleozoic (Ni et al., 1990; Sengor et al., 1993). However, Burrett and Stait (1985) and Fang (1994) proposed that Sibumasu have rifted away from Gondwana approximately during the Middle Ordovician. Tibet was united with India and located on the northern margin of Gondwana during the Pre-Cambrian and Early Paleozoic. It may have rifted away from Gondwana, probably in the later Paleozoic or Triassic. The convergence of Sibumasu with Indochina, and the latter with south China were suggested to be Early Carboniferous and Late Triassic, respectively (Metcalfe, 1988). Indochina may have been joined to Sibumasu along the Uttaradi–Nan–Bentong–Raub suture (Metcalfe, 1992; Fig. 34).

Southern Margin of Siberian Paleoplate

The Greater and Lesser Khingan Mountains of Heilongjiang and northern Inner Mongolia in extreme northeast China have a distinctive *Tuvaella* (brachiopod) fauna (Su, 1981). This region has been considered part of the southern marginal belt of the Siberian paleoplate of the Mongolo-Okhotsk province. This well documented, shallow-marine, benthic fauna is remarkably different from the *Retziella* fauna, which occurred south of the northern marginal belt of North China (Rong et al., 1995). The *Tuvaella gigantea* assemblage is known in the Ludlow of the Barkol area of northeasternmost Xinjiang, which may have been attached to the southern marginal belt of Siberia in the Silurian (Rong and Zhang, 1982).

Easternmost Kazakhstan Paleoplate

The *Tuvaella* and *Retziella* faunas have not been documented in the Junggar region, northwest Xinjiang; this is a distinctive biogeographic feature of this region. *Cooksonella sphaerica* Senkevich, 1978 (=*Junggaria spinosa* Dou and Sun, 1983), earlier recorded only from Kazakhstan, is known from the middle Wutubulake Formation (Pridoli) at Mongkelu, 18 km east of Wutubulake, Hoboksar County, western Junggar (Cai et al., 1993). Western Junggar is the eastern extension of the Kazakhstan paleocontinent (Li et al., 1984). This interpretation has further been confirmed by the occurrence of the brachiopod *Pentamerifera oblongiformis* (Nikiforova), which was earlier recorded in the lower Ludlow of the Pre-Balkash district, Kazakhstan; the Nurata and Alay Mountains of Central Asia; and the Ivdel area in the eastern Urals (Rong and Zhang, 1988). The position of Kazakhstan relative to other paleocontinents is still uncertain, and Kazakhstan developed by an amalgamation of exotic terranes and island-arcs in the Early Paleozoic. Nevertheless, Cocks and Scotese (1991) noted that the dominance of limestones and numerous corals at many localities shows that the continent must have been close to the Silurian paleoequator.

Possible Paleogeographic Position of China

Except for parts of the Siberian and Kazakhstan paleoplates, all of the other units that now compose China may have had an origin as part of Gondwana. They seem to have separated from Gondwana in the Ordovician and Silurian, or later (Wang et al., 1985; Cocks and Fortey, 1986; Chen and Rong, 1992; Metcalfe, 1992, 1995; Wang and Mo, 1995). The time of rifting of these paleoplates from Gondwana remains contentious.

Latest Proterozoic glaciogene deposits are widely distributed in Tarim and North and South China, and indicate that these areas were adjacent during this interval. Metcalfe (1992) postulated that paleomagnetic data show that South North China, and possibly Tarim,

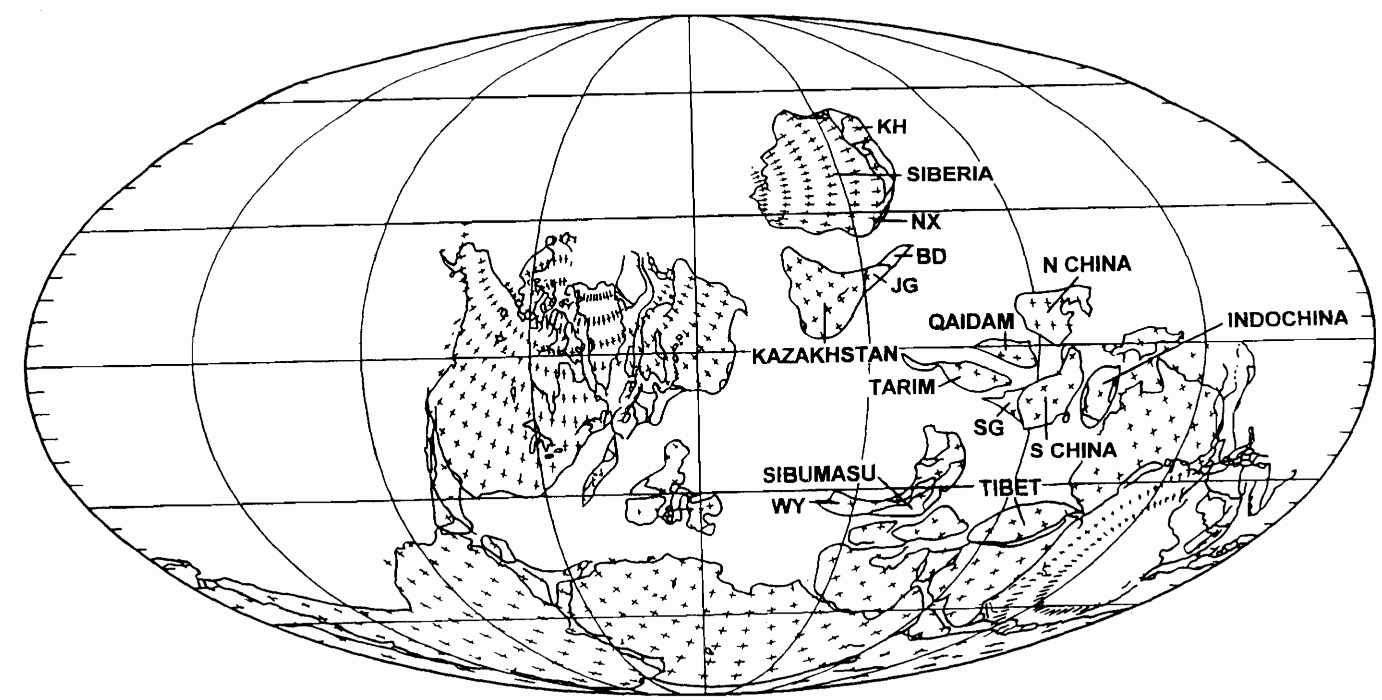

FIGURE 34 — Global Silurian paleogeography modified from Scotese and McKerrow (1990) and Boucot et al. (1995). KH, Khingan; NX, northeasternmost Xinjiang; JG, Junggar; BD, Badainjaria; SG, Songpan–Ganzi; WY, westernmost Yunnan. Indochina included in eastern west Yunnan.

Indochina, and east Malaya, rifted from Gondwana in the Silurian. Thus, South China was effectively isolated from other continents during the Devonian, and this led to the high endemism of its Devonian faunas.

The time of rifting of South China from Gondwana may be earlier than suggested by Metcalfe (1992). Indeed, there are great differences between Late Ordovician and Early Silurian shelly faunas and lithofacies of South China and Gondwana. In addition to the high endemism of Llandovery faunas in South China (Rong et al., 1995), there are several other lines of well-documented evidence. Early–middle Ashgillian faunas, particularly brachiopods (e.g., *Tcherskidium*, *Altaethyrella*, *Ovalospira*, and some endemic genera; Rong et al., 1994; Zhan and Rong, 1995) and tabulates (*Agetolites* fauna; Lin and Webby, 1989), link east China with the Urals, Kolyma, Alaska (*Tcherskidium*), central Asia, east Qinling (Zhenxi), Qaidam (*Altaethyrella*, *Ovalospira*), and Australia (corals; Webby, 1992). Ashgillian conodonts of southeast Qinling show that it probably rifted away to the north from South China. These conodonts represent a warm-water, Laurentian, Mid-Continent-type assemblage (Mei, 1995) and show conodont faunal exchange between east Qinling and Laurentia. A Llanvirnian, warm-water *Aporthophyla* brachiopod fauna, which indicates the Toquima–Table Head province (Neuman and Harper, 1992) of tropical paleocontinents, has been found in central Guizhou, South China (Rong et al., in preparation). This area was geographically isolated from most of the upper Yangtze region by the Qianzhong Oldland during the Middle Ordovician.

The biogeographic relationships of the earlier Ordovician in China are defined by a number of features: 1) Early Ordovician, cool- or deeper-water benthic faunas (chiefly brachiopods) of central–southern Europe and northern Africa have not been found in South and North China, Qaidam, Tarim, Tibet, or west Yunnan (Indochina and Sibumasu paleoplates). The Arenigian brachiopods of South China are characterized by *Sinorthis*, *Yangtzeella*, *Martellia*, *Metorthis*, and others that indicate a high degree of endemism and strong faunal differences between South China and central–southern Europe and northern Africa (Xu and Liu, 1984). 2) Tremadocian–Arenigian chitinozoans and acritarchs from South China and Karakorum, Pakistan, are similar to those of North Africa and southern Europe in the North Gondwana realm, or Peri-Gondwana province (Paris, 1989; Tongiorgi et al., 1994; Playford et al., 1995). This suggests that cold currents led to biotic exchange with Gondwana (Li, 1987; Wang and Chen, 1994). However, the acritarchs of South China also show some differences with those of the Peri-Gondwana province (Playford et al., 1995). Chen et al.

(1995) further pointed out that the chitinozoans of South China are also similar to those of the Baltic, a region characterized by low-latitude, warmer-water elements after the earliest Arenigian. 3) South China has Ordovician conodont assemblages of the Baltic type, which are of cold- or cool-water origin (Wang et al., 1996).

It should be noted again that the Guangxian orogeny started in the Caradocian and ended in the latest Llandovery (Chen and Mitchell, 1996). This indicates a collision between South China and an unidentified eastern block. This suggests that South China moved north after it rifted from Gondwana. Based on the data noted above, the rifting of South China from Gondwana was after the early Arenigian and before the early Caradocian. During this interval, South China was near Australia and far from southern Europe, with biotic exchanges taking place by oceanic currents. This interpretation is supported by paleomagnetic data from the Silurian of South China. Determinations of 6° S and 8° S have been made for southeast Hubei and south Anhui, respectively, in the Llandovery (Meng, 1993; Fang et al., 1992). Qujing, eastern Yunnan, in southwest China was located at the equator in the Ludlow–Pridoli (Liu C. *in* Fang et al., 1985).

North China had warm-water conodont assemblages during the Middle–Late Ordovician, and is distinct from South China with its cool-water North Atlantic conodonts (Wang et al., 1996). North China may have rifted away from Gondwana before South China. The Tremadocian–Arenigian is suggested to be the time of rifting of North China from Gondwana. The Silurian shelly faunas of North and South China are similar, and the former is herein located to the north, but still near South China based on biogeographic, tectonic, and sedimentological data.

Qaidam and Tarim are suggested (Wang and Mo, 1995) to have been close from the latest Proterozoic. They also suggested that the north Qilian–north Qinling Caledonian oceanic basin was subducted under the North China paleoplate after the Ordovician. The *Retziella* fauna (chiefly Ludlow–Pridoli) occurs on the north Tarim margin and indicates a close relationship with North China, South China, and Indochina. Paleomagnetic data suggest the Tarim paleoplate was in low south latitudes during the late Early Paleozoic (see Liu et al., 1997), may have collided with the Yili block of Kazakhstan in the Devonian, and subsequently collided with the North China paleoplate in the later Paleozoic (Zhou and Dean, 1996).

Qaidam, North and South China, and Tarim (probably including Tadzhikistan and Kirkizistan) seem to have been adjacent (Zhou and Chen, 1990). A deep ocean did not separate them from Australia in the Cambrian and Ordovician. The exchange of shallow benthic organisms between these paleocontinents was controlled by oceanic currents. All of these paleocontinents have the Late Silurian *Retziella* fauna (Rong et al., 1994; 1995), which is also recorded from Indochina. Indochina was not far from South China during the Late Silurian, and was possibly joined in the Late Paleozoic to South China along the Song Ma–Song Da suture zone in northern Vietnam (Metcalfe, 1992, 1995).

Cambrian and earlier Ordovician biotas of Sibumasu have strong Australian affinities. Southern Tibet has the Llanvirnian *Aporthophyla* brachiopod fauna (Liu, 1976; Neuman and Harper, 1992) and corals (e.g., *Yohophyllum, Calostylis, Ningnanophyllum*), which are similar to those of South China (He X.-y. *in* Yang et al., 1990). Ordovician conodonts of south and north Tibet are similar to those of South China, Tarim, and Baltica, and belong to the North Atlantic province (Qiu, 1988). However, there was a remarkable later change of faunal provincialism in Sibumasu in the Late Ordovician and Early Silurian (Fang, 1994). This indicates that barriers to faunal exchange between Sibumasu and South China must have developed. In the Late Ordovician, Sibumasu was located outboard of Australia and Tibet and between India and Indochina (see Scotese and McKerrow, 1991; Metcalfe, 1992). Two observations have been made on shelly faunas: 1) The *Retziella* fauna has not been found in Tibet, Sibumasu, or Afghanistan. Similarly, these regions do not seem to have the European, Late Silurian *Dayia* fauna (Rong et al., 1995). 2) The *Kopaninoceras* fauna occurs in southern Europe, Tibet, and Sibumasu, but not South China. Faunal exchange of some swimming animals (e.g., nautiloids) occurred between Sibumasu, Tibet, and southern Europe during the Late Silurian. Thus, we propose that Sibumasu was located between southern Europe and South China, whereas Tibet was still joined to India.

In the Karakorum region, there is no good record of Silurian brachiopods. A Late Silurian *Kopaninoceras-Michelinoceras* fauna is reported from the Purco Formation (Wen et al., 1996). This nautiloid fauna is also known from north Tibet and west Yunnan. It indicates a Late Silurian proximity of Karakorum to Tibet–west Yunnan. However, in the Early Ordovician, the Karakorum block was located in a latitude intermediate between the Mediterranean region and South China. This interpretation is based on its Peri-Gondwana province palynomorphs and *Sinorthis* fauna brachiopods (Tongiorgi et al., 1994; Xu and Su, 1998).

A number of paleogeographic conclusions are possible: 1) North China, Tarim, Qaidam, and South China successively rifted away from Gondwana through the Ordovician and moved north in the Silurian. These continents may have been tropical. Silurian North China was in the northern tropics near northwest Australia. South

China was south-tropical and near North China. Tarim and Qaidam were adjacent to South China, but possibly somewhat further north. Indochina was near South China, and Yunkai was between these paleocontinents. 2) Sibumasu lay between South China and southern Europe. Tibet, with Karakorum, may have been part of the Indian paleoplate in the Silurian. 3) The *Tuvaella* fauna occurred on insular continents in more northern latitudes and indicates distance from all the other Chinese terranes (Fig. 34).

It is fairly easy to reconstruct the relative positions of various paleoplates, and to know when two paleoplates collided, based on geological evidence. However, it is more difficult to define the precise location of each continent; one of the reasons is that interpreting the width of the oceans between continents and terranes is difficult. We only know if Silurian oceans were wide enough to restrict biotic exchange and define a number of faunal provinces on paleocontinents.

CONCLUSIONS

Silurian China was not a consolidated paleogeographic unit. Its tectonic history was more complicated that that of Laurentia, Siberia, Australia, or other major units. Modern China is composed of five paleoplates (South China, North China, Qaidam, Tarim, Tibet); parts of four other paleoplates (west Yunnan–Sibumasu and Indochina, Jungar and Badaninjaria–Kazakhstan, and Khingan Mountains and northeasternmost Xinjiang–Siberian paleoplates); and many smaller blocks (e.g., Bureya–Jamus, Songpan–Ganzi, Yunkai, and Hainan). These plates or blocks collided with each other beginning in the Late Silurian. However, most collisions were in the Late Paleozoic (e.g., the west and east Junggar regions of the Kazakhstan paleoplate with the Siberian paleoplate) or Late Paleozoic–Early Mesozoic (e.g., the South China paleoplate with the North China paleoplate).

The complicated tectonic relationships of various paleoplates commonly led to intricate sedimentary facies patterns. Collision, subduction, or consumption of two paleoplates led to incompleteness of the sedimentary record in many areas, particularly along the border of two paleoplates. A sedimentary trough usually developed parallel to the shoreline of the uplifted region. This depocenter can be considered a foreland basin. Its thick siliciclastics commonly had a shallow-water biota that lived in a region with a high rate of sediment accumulation.

As opposed to the Ordovician, the Silurian of China is dominantly siliciclastic with or without volcanic input and local bedded limestones. This is related to the deposition of widespread graptolitic shales on the platform, successive collisions of continents and terranes, rifting and uplift, and strong weathering and degradation. Landmasses such as the Cathaysian and the Tarim Oldlands expanded remarkably during the Silurian. The Silurian of the upper Yangtze region contains a large volume of siliciclastics, with limited bedded limestone in the Llandovery and Late Silurian. Silurian platform carbonates are well developed, chiefly in Tibet and western Yunnan (part of Sibumasu). Aeronian biostromes and late Telychian bioherms are prominent in northeast Guizhou and on the northwest margin of the Yangtze platform. Ludlow(?) carbonate mud mounds are known in southern Inner Mongolia.

Analysis of the community paleoecology of shelly faunas, particularly brachiopods, suggests that almost all the Silurian of China represents a shallow-water regime (BA 1–3). Deeper-water shelly faunas, such as brachiopods of the Late Ordovician *Foliomena* fauna or Early–Middle Devonian *Costanoplia* fauna, have not been documented in the Silurian of China. It is likely that BA 4–5 shelly biotas will be found in the marginal belts. Deeper-water graptolites occur in marginal basins (e.g., Qinling, Tianshan, western Yunnan), and a few localities with deep-water radiolarians are known from Qinling and western Junggar. Graptolites in the lower Lungmachi Formation (Rhuddanian–lower Aeronian) appeared in the epicontinental sea on the Yangtze platform and are presumed to represent a BA 4–5, dysaerobic-water regime.

Paleogeographic changes took place across Silurian China, particularly in the Wenlock–Pridoli. Subaerian environments were dominant in the latter interval. Most of the Yangtze platform was covered by shallow seawater during the Llandovery, was uplifted in Wenlock, and a few marginal areas were flooded again in the Ludlow–Pridoli. Most parts of North China, Tarim, Qaidam, the south margin of the Siberian paleoplate, the Bureya–Jamus block with the Yanbian and Xingkai terranes, and some other regions were separate, subaerial continents in the Silurian. The Tibet paleoplate is an exception, because it was flooded through the entire Silurian, as were possibly the Qangtang and Himalayan Oldlands.

The study of Silurian sea-level changes in China has focused on the Yangtze platform. Of the four global high stands during the Llandovery (Johnson et al., 1985, 1997), three have been recognized in South China: 1) near the *Pristiograptus* (*Coronograptus*) *cyphus–Demirastrites triangulatus* zonal boundary; 2) in the basal *Spirograptus turriculatus-Monograptus crispus* Zone; and 3) in the upper *Monoclimacis griestoniensis*–lower *Oktavites spiralis* Zones. Highstands 1 and 3 can be observed on all parts of the platform, and 2 is evident only in the Nanjiang–Chengkou area of northern Sichuan. Active tectonic movements with rapid uplift and subsidence took place in South China during the *Stimulograptus sedgwickii* and *Spirograptus turriculatus-Monograptus crispus* Chrons. The largest regression in South China commenced at the end of the Llandovery and persisted into the Wenlock and early Ludlow. The local presence of *Ozakordina crispa* in South China shows that a highstand in sea-level took place in the late Ludlow.

New investigation indicates many endemic taxa and the earliest-known representatives of clades in South China during the Late Ordovician–Early Silurian, especially during the Telychian. They include trilobites, nautiloids, brachiopods, graptolites, crinoids, vertebrates, and other forms (Chen and Rong, 1996). This shows the isolation of South China and nearby paleocontinents from other continents (e.g., Siberia, Laurentia, Avalonia, and Gondwanan southern Europe). This isolation was probably caused by the restriction of South China platform seas during the Guangxian orogeny (Chen et al., 1997), and is shown by the biogeographic distinctiveness of South China in the Llandovery (Rong et al., 1995).

Paleobiogeographic analysis of Silurian shelly faunas allow some conclusions: 1) South China, North China, Tarim, and Qaidam, all areas with the *Retziella* fauna, moved north and were equatorial. 2) Western Yunnan (Sibumasu) and Tibet have a European-like *Kopaninoceras* nautiloid fauna. These areas were isolated from the other four paleoplates named above. 3) The southern marginal belt of the Siberian Paleoplate (Greater and Lesser Khingan Mountains and extreme northeastern Xinjiang) have a distinctive *Tuvaella* fauna. These areas were geographically separate from the other parts of China in the Silurian.

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Silurian Paleogeography Along the Southwest Margin of the Siberian Continent: Altai-Sayan Folded Area

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ABSTRACT — This report summarizes data on the Silurian shelf on the southwestern margin of the Siberian paleocontinent. The Silurian is widespread in this area of the Altai–Sayan folded area. The Altai–Sayan is divided into four subareas: Salair, Altai, Western Sayan, and Tuva. Most representative of the Altai–Sayan, the Silurian of the Altai region is reviewed in detail. In this territory, all of the standard Silurian divisions are recognizable, and the stratigraphic schemes for the other regions are correlated into the Altai. Silurian paleogeography is reconstructed on a late Telychian map that shows the position of the shelf and its inner and outer zones. Locations of carbonate platforms and back-reef basins are summarized, with correlation charts for reference districts and cross-sections. The species-level faunal similarity between the Altaian and Tuvan late Llandovery brachiopod associations is very low. Sea-level fluctuations and patterns for depth curves are analyzed. Climatic indicators support the Silurian position of the Siberian paleocontinent close to, or within, the tropical realm. The Siberian paleocontinent was separate from the Tuva area (microcontinent) during the Silurian.

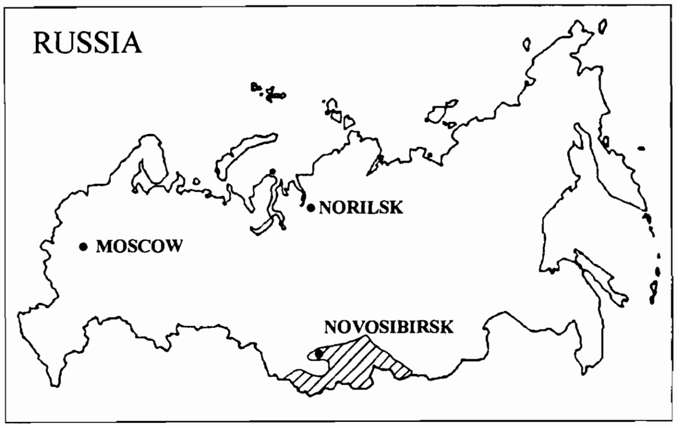


FIGURE 1 — Index map of Russia showing the location of the Altai–Sayan folded area (oblique lines) southeast of Novosibirsk.

Introduction

The Altai-Sayan folded area (ASFA) embraces a large part of southern Russia (Fig. 1). It is located in the geographic center of Asia, and a special monument marks this point in the capital city of the Tuva (Tyva) Republic. Rocks exposed here are Precambrian–Permian, but the Middle Carboniferous–Permian is non-marine. These sedimentary rocks are regarded as the best studied components of the central Asiatic fold belt.

The geology of the ASFA displays a complicated combination of tectonic blocks (Fig. 2), which are usually separated by shear zones or major faults. These blocks were earlier considered to be down-folded zones (troughs or synclinoria) and uplifts (massifs or anticlinoria) that developed within a geosyncline (Kuznetsov, 1954; Nikiforova and Obut, 1965). These local structures were united in a single geosynclinal megastructure called the Altai–Sayan folded area on the margin of the old Siberian craton (Fig. 3). The ASFA and its continuation to the northwest and southeast, constitute the central Asian fold belt that formed in the Paleoasian ocean (Zonenshain et al., 1976).

Much attention was paid to this great fold belt under IGCP project 283: "Geodynamic evolution of the Paleoasian Ocean" (Dobretsov et al., 1994). Many models have been proposed for the relationships through space and time and accretion history of the volcanic island arcs and microcontinents located between the Siberian, North Chinese, and East European platforms. Paleogeographic reconstructions suggest that Vendian–Early Paleozoic volcanic arcs and microcontinents of the ASFA were accreted to the Siberian platform at the beginning of the Early Ordovician (Yolkin et al., 1994). It has been shown that the southwestern shelf of the Siberian paleocontinent (in terms of modern coordinates) occupied a stable position within the western part of the ASFA (Salair, Altai). During the Silurian, these regions were a passive continental margin.

Silurian deposits have a patch-like distribution throughout the Salair, Gorny Altai, western Sayan, and Tuva regions (Fig. 4). In this report, the main focus is on the southern grouping of the Altai, western Sayan, and Tuva regions. Within this large area, Silurian rocks are relatively well exposed and often feature complete sections. They are characterized by well known biogeographically distinct areas. One is the Old World realm, and the other is represented by the so-called Tuvan faunas. Until recently, these paleobiogeographic differences allowed only provisional correlations of the Altaian and Tuvan Silurian (Kul'kov et al., 1985; Vladimirskaya et al., 1986). A direct connection of the Altaian and Tuvan seas (Naumenko, 1970; Kul'kov, 1974) is unlikely. A solution to this problem was proposed by the use of Silurian paleomagnetic and geological data to separate the Tuvan microcontinent from the ASFA as a part of the Siberian paleocontinent (Zonenshain et al., 1990, p. 292, Fig. 125).

The purpose of this report is to document the Silurian of the Altai–Sayan segment of the southwest shelf of the Siberian paleocontinent. This shelf has been earlier described in a general way (Yolkin et al., 1994). The cyclic character of Silurian sedimentation has been illustrated in the Altai (Yolkin and Zheltonogova, 1974; Yolkin et al., 1997a) and related to transgressive–regressive (T–R) cycles and trilobite evolutionary stages (Yolkin, 1983, 1998). Another objective is to show the errors that have led to illustrations of a rotated Siberian paleocontinent in the Silurian, as commonly seen in the recent literature. This is supported by evidence for barrier reefs that developed during the late Llandovery and late Wenlock on the outer shelf and by the presence of warm-water faunal associations. A topic of special interest involves the paleogeography of the Tuva area, a region assumed on the basis of faunal affinity to lie in the southern hemisphere during the Silurian. The bases for our conclusions include improved correlations of individual sections, composite sections for reference districts that are now related to eustatic history, cross-sections across outer- and inner-shelf zones, a paleogeographic map that shows rapid development of a late Llandovery barrier reef, and the

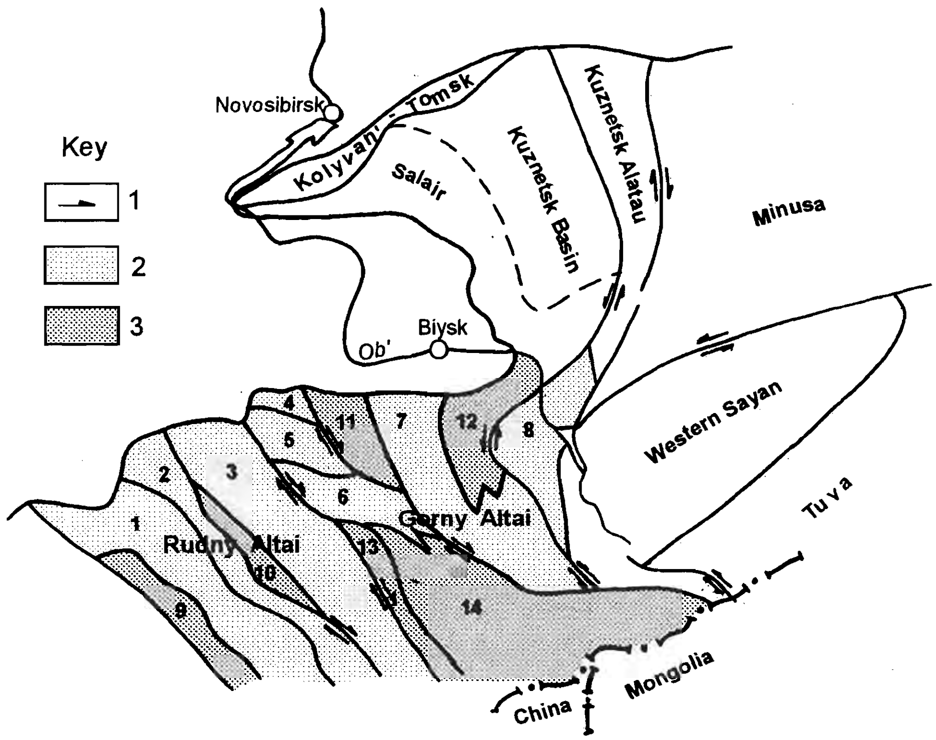


FIGURE 2 — Tectonic structure of the Altai–Sayan area showing separate blocks in the Rudny and Gorny Altai regions (from Yolkin et al., 1994). Key: 1, shear zones and blocks; 2, folded zones; 3, uplifts. Tectonic blocks and uplifted areas: 1, western Kalba; 2, Kalba–Narym; 3, Rudny Altai; 4, Kur'ya; 5, Charysh–Inya; 6, Korgon; 7, Anui–Chuya; 8, Uimen'–Lebed'; 9, Chara;10, Irtysh; 11, Talitsa; 12, Biya-Katun'; 13, Kholzun; 14, southern Altai.

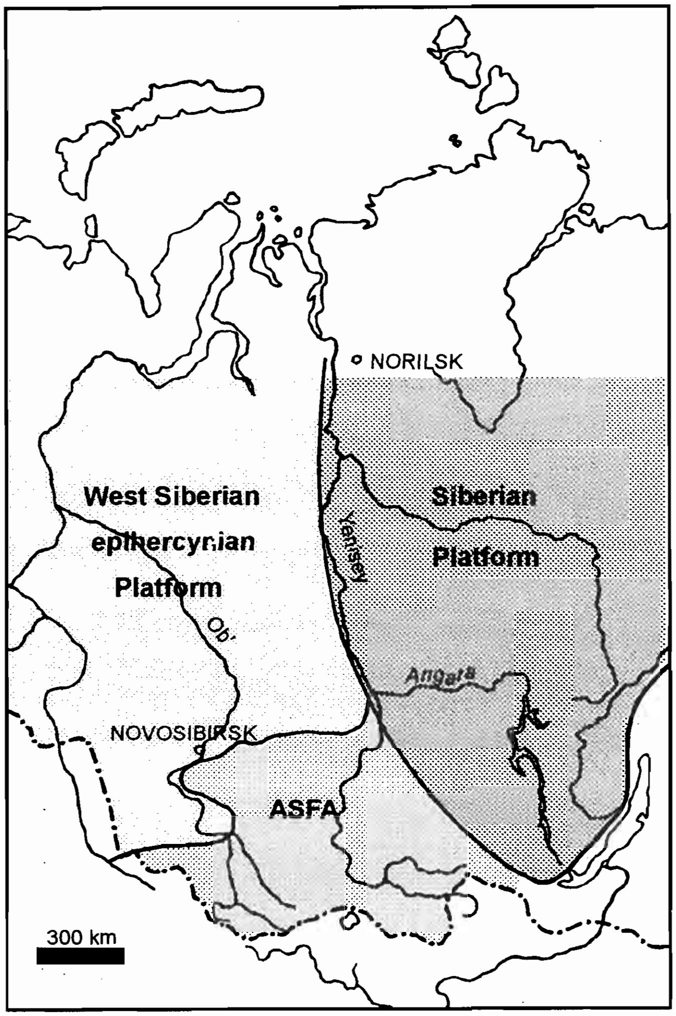


FIGURE 3 — Principal geological structures of Siberia: Siberian platform, Altai–Sayan folded area (ASFA), west Siberian epihercynian platform.

interpretation of flooding surfaces as recording eustatic and sedimentary events.

Geological Setting

Except for Tuva, this part of the Siberian paleocontinent was passive in the Silurian. The main tectonic events happened before or after this time interval. The Siberian craton had developed by expansion to the west through collisions and accretions during the Baikalian (Pre-Vendian) and Salairian or Early Caledonian (Late Cambrian) orogenies (Berzin and Dobretsov, 1994). The latter created collisional–accretional complexes of Vendian–Cambrian age between the upper reaches of the Yenisey and Ob' Rivers. At the beginning of the Early Ordovician, the southwestern shelf of the Siberian paleocontinent shifted to the west approximately from the longitude of the mouth of the Angara River to Novosibirsk. This territory now is named the Altai–Sayan folded area (Fig. 3). Changes in geodynamic regime from a volcanic island arc to a continental margin occurred during the Tremadocian (Yolkin et al., 1994). In the Arenigian–Early Devonian, this margin of the Siberian paleoontinent was characterized by passive margins. The belt is subdivided into two facies belts (inner and outer) characterized by moderately thick deposits that do not contain volcanics. Individual facies are easily traceable along these belts, but clearly change across the shelf. The essential feature of sedimentation from the latest Ordovician–Middle Devonian was a cyclic alternation of siliciclastics and reef limestones (Yolkin and Zheltonogova, 1974; Yolkin et al., 1997a). Large-scale buildups (barrier reefs) periodically formed on the outer shelf (Yolkin et al., 1994). The late Llandovery and late Wenlock were major reef-building intervals. These barrier reefs are found in a 600 km-long, 10–50 km-wide belt that is strongly curved to the east and includes the Salair and northern Altai areas.

Strata deposited at the beginning of the Silurian occur in Altaian sections. A sharp transition occurs from the topmost Ordovician reef limestones to Silurian black shales rich in graptolites (Sennikov, 1976, 1996; Yolkin et al., 1978, 1988). There are exposures where this boundary is represented by fine-grained siliciclastics. The oldest Silurian barrier reef was described from the upper Llandovery outer shelf of the northern Altai (Sennikov et al., 1988; Gladkikh et al., 1989; Yolkin et al., 1994). It is recognizable even in the southwestern Altai (Sennikov, 1987; Sennikov et al., 1990, 1992) and western Sayan (Naumenko, 1970), and extends more than 1,500 km. The late Wenlock carbonate platform was wider and extended across part of the latest Llandovery inner shelf (Yolkin et al., 1994, fig 3B). Within the eastern part of this area, however, it is poorly known. A single section in the central Altai shows the final late Ludlow–Pridoli stage of Silurian deposition (Yolkin and Zheltonogova, 1974). It consists of red, non-marine siliciclastic strata with a limestone package in its middle part. The Lower Devonian rests unconformably on the Silurian, possibly without significant biostratigraphic break, but also transgressed more ancient deposits (Yolkin, 1968). Thus, the Altai Silurian is a sequence initiated by the global early Llandovery transgression, and is bracketed by another global regression.

In contrast to other regions of the ASFA, the Tuva region has its own Silurian geological history. Tuva was characterized in the Silurian by extremely shallow-water sedimentation, by red siliciclastics after the Wenlock, and

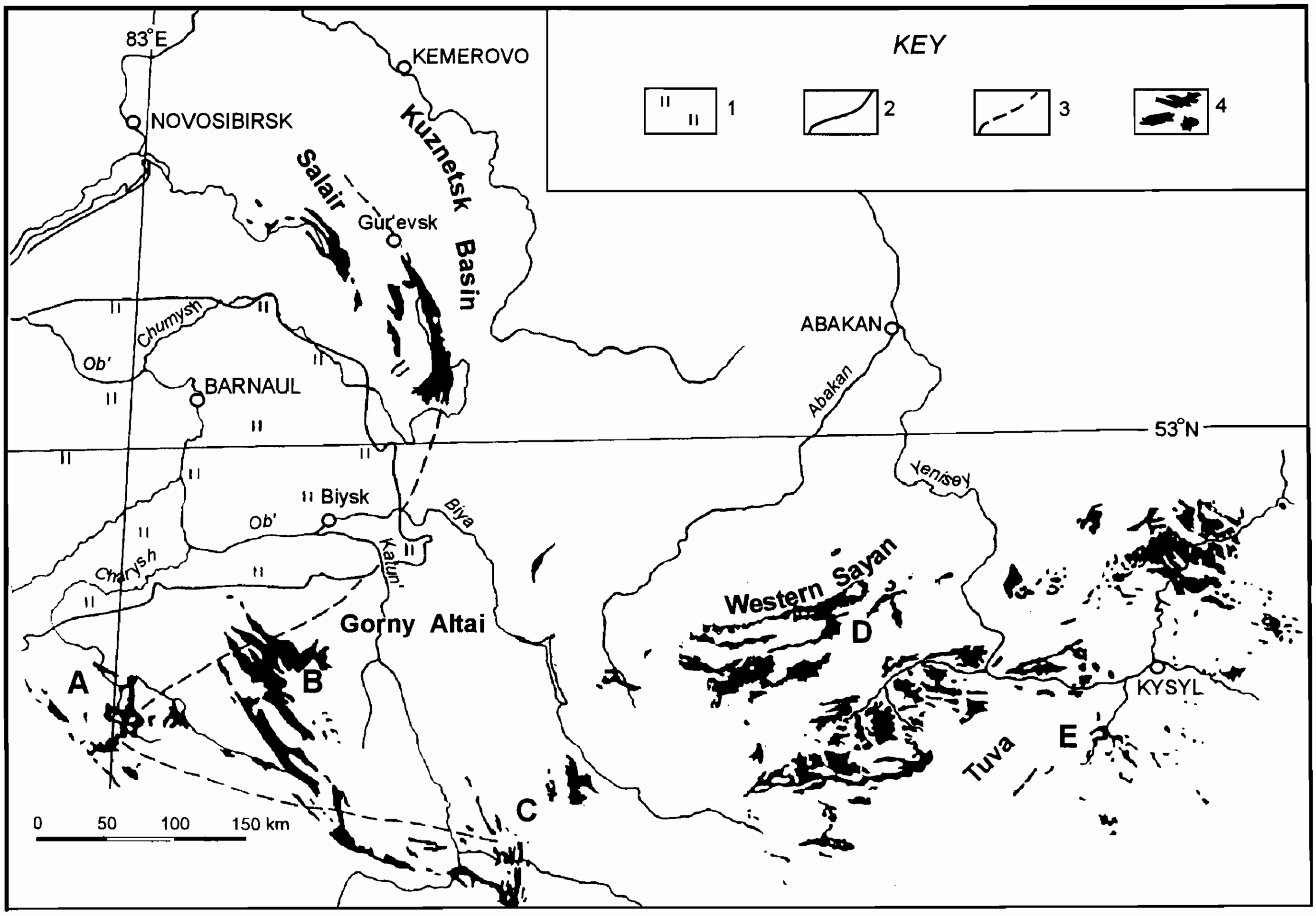


FIGURE 4 — Distribution of Silurian outcrops in the Altai–Sayan folded area with locations of territories noted in text: A, northwest Altai; B, north-central Altai; C, southeastern Altai; D, central part of the Western Sayan; E, Elegest River. Key: 1, Cenozoic; 2, western limit of ASFA; 3, location of inner–outer shelf boundary; 4, Silurian outcrop.

by high faunal endemism. These features say nothing about the relationships between the neighboring Tuva and Altai basins. Nevertheless, they were considered parts of a single sea (Naumenko, 1970; Kul'kov, 1974). According to paleomagnetic and geological data, however, the Tuva area was a separate microcontinent that joined with the Siberian paleocontinent during the Silurian–Devonian boundary interval (Zonenshein et al., 1990; Berzin et al., 1994). This view is supported by faunal evidence (Yolkin and Sennikov, 1998). Moreover, a rotation of the Siberian continent during the Middle Paleozoic is erroneous. The actual paleogeographic position of the ASFA segment of the Siberian continent should be in the tropics, somewhere between 5° and 20° N. The Tuva microcontinent has elements of the Silurian Gondwanan fauna (Tchernysheva, 1937; Khalfin (ed.), 1961), and was located ca. 20° S (Fig. 5).

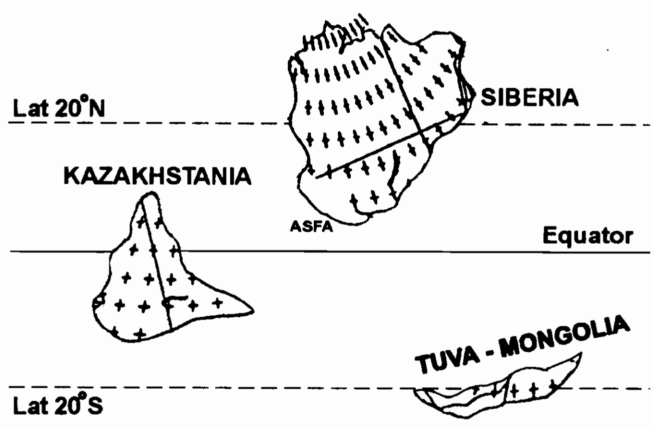


FIGURE 5 — Location of Siberian and Kazakhstan paleocontinents and Tuva–Mongolian microcontinent in late Llandovery. Continents after Scotese and McKerrow (1990).

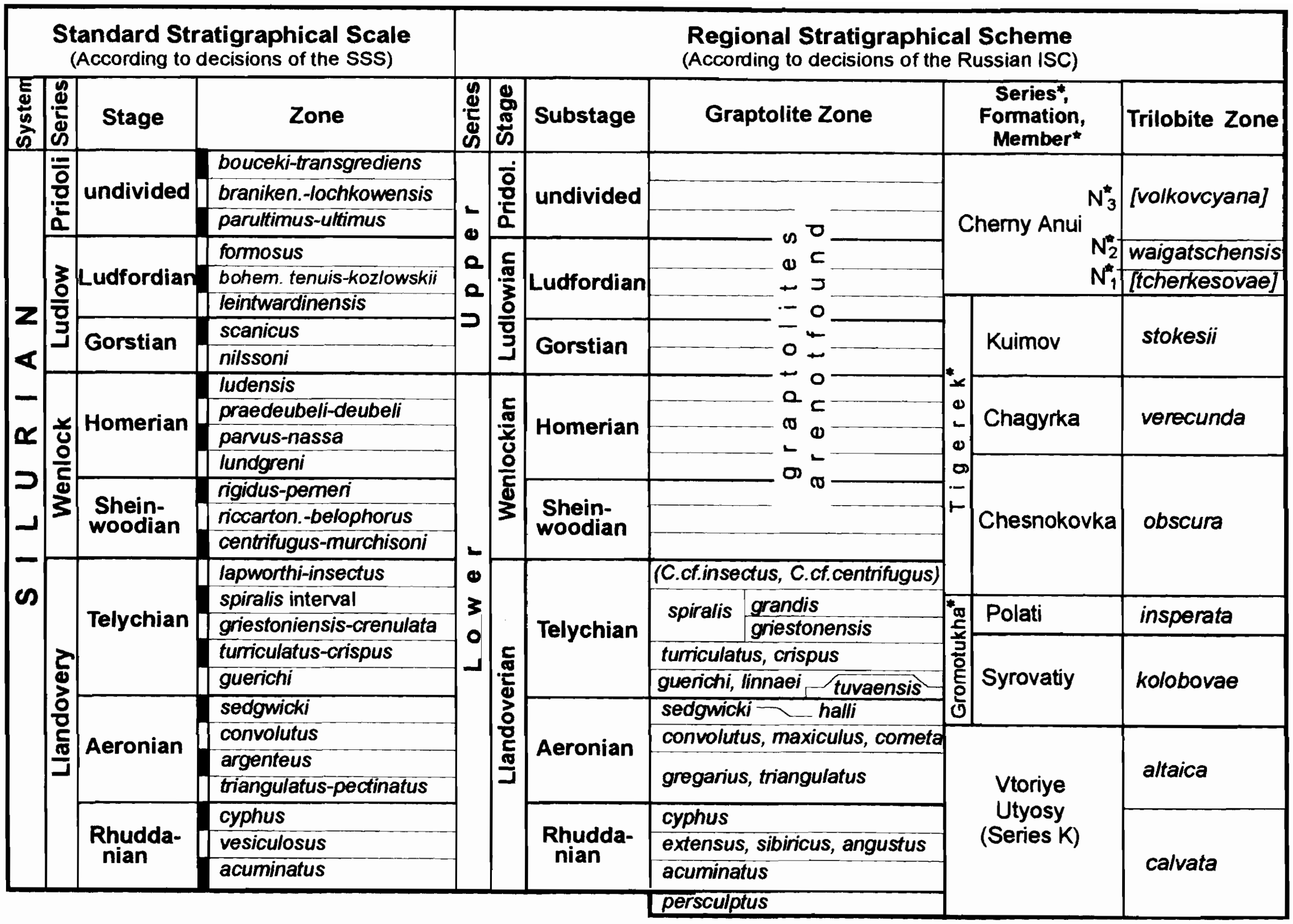


FIGURE 6 — Silurian of Gorny Altai.

STRATIGRAPHY

In the four regions of the ASFA, the Silurian is subdivided on the basis of litho- and biostratigraphic features. Faunas from the Salairian, Altaian, and western Sayanian Silurian are similar. They belong to the widely distributed Old World realm fauna. In contrast, Tuvan Silurian fossils are highly endemic. Among these regions, the Altaian sections are the best studied. They are quite complete, and include diverse pelagic (graptolites, conodonts) and benthic (brachiopods, trilobites, corals, stromatoporoids, etc.) fossils. For these reasons, the Altaian succession is regarded as the reference for the entire region.

ALTAI — The currently defined Silurian stratigraphy for this region is in Fig. 6. Development of the understanding of this succession was rapid (see Yolkin and Zheltonogova, 1974, table 1). The Silurian of the northwest Altai was first described by Nikonov (1931). The second stage in the development of a Silurian stratigraphic synthesis was the proposal of two formations after regional geologic mapping in the late 1940s and 1950s (Gintsinger, 1959; Resheniya, 1959). These are the Chagyrka Formation (Ludlow, massive limestones) and the Pod-Chagyrka or Chineta Formation (Llandovery–Wenlock, siliciclastics and carbonates). It should be noted that massive limestones, which accumulated during the latest Llandovery and late Wenlock reef-building episodes, were previously referred to the Chagyrka Formation and interpreted as Ludlow. The red-colored Cherny Anui Formation was regarded as Devonian.

Since the early 1960s, paleontological investigations

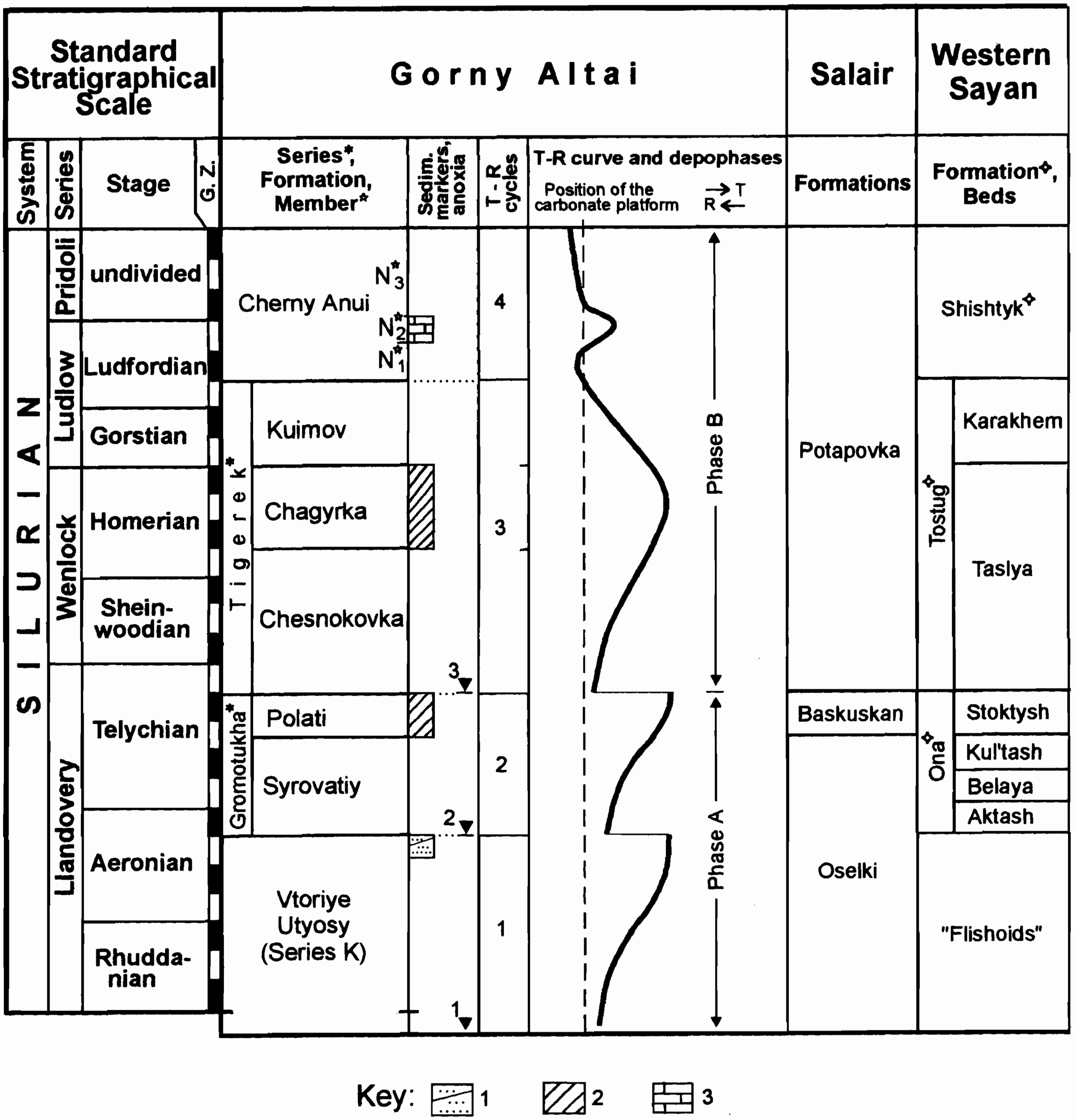


FIGURE 7 — Silurian of Gorny Altai, Salair, and western Sayan. Key: 1, cross-bedded siliciclastics; 2, reef limestone; 3, bedded limestone within red beds. Black triangles correspond to anoxic events: 1, Chineta (*Glyptograptus persculptus* Zone) event; 2, Syrovatiy (*Monograptus sedgwickii* Zone) event; 3, Chesnokovka (*Cyrtograptus* Zone) event. T-R curve and depo-phases after Yolkin (1998).

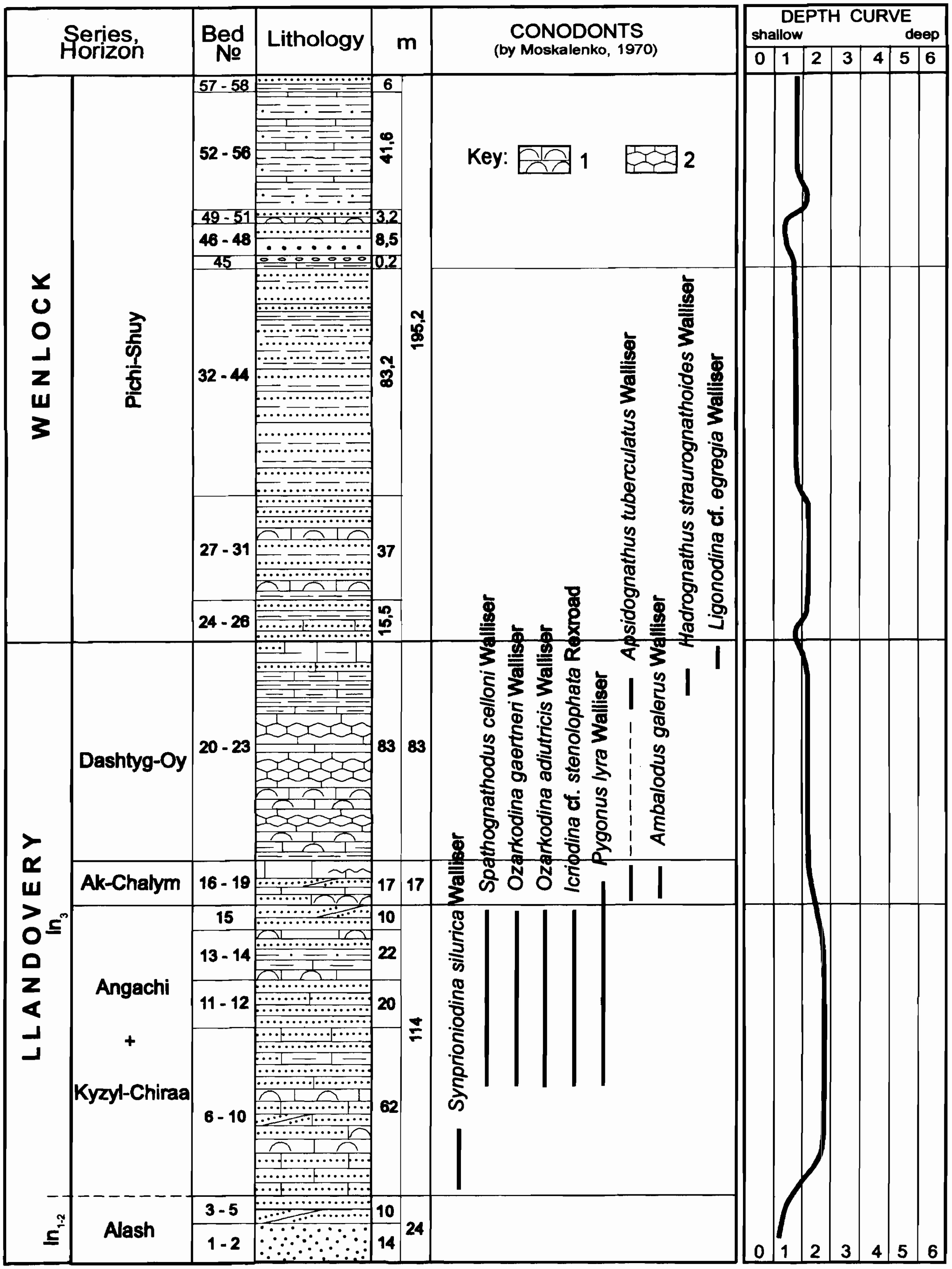


FIGURE 8 — Columnar section, biostratigraphy, relative sea-levels of Elegest section (Tuva). Key: 1, shelly limestone; 2, nodular limestone. Other explanations in Fig. 13.

Groups of horizons	Horizon	Graptolites	Conodonts	Evolutionary changes: Brachiopods	Stromatoporoids	Heliolitids	Vertebrates	Sedimentology	Kul'kov et al. 1985	Vladimirskaya et al. 1986	This paper: Tuva	Altai	
Baital	Taugan-Teli			*Tuvaella gigantea*			III	Sedimentary event level ▼	S_2	pr	w	Chesnokovka to Cherny Anui formations	▼ Chesnokovka-E.
	Pichi-Shuy				IV	IV				ld-pr			
Elegest	Dashtyg-Oy			*Tuvaella rackovskii*		III	II		w	ld	In_3	Polati Formation	▼ Syrovatiy-Event
	Ak-Chalym	*M. priodon* *R. angustidens*	*amorph.*-Z.		III				In_3	w			
	Angachi	*S. exiguus* *St. tuvaensis*	*celloni*-Z.							In_3		Syrovatiy Formation	
	Kyzyl-Chiraa				II	II			In_2				
Alavelyk	Alash			*Dicoeromyonia asiatica*				...	In_1	In_{1-2}	In_{1-2}	Vtoriye Utyosy	
	Khondelen				I	I	I		O_3	O_3	O_3	O_3	

FIGURE 9 — Silurian biostratigraphy and correlations of the Tuva area into the Altai region. Evolutionary events after Vladimirskaya et al. (1986). Grey strips are barren intervals. Two boundary lines mean sharpest changes. Conodont zones from Chadan (Vorozhbitov, 1992) and Elegest (Moskalenko, 1970) reference sections. Graptolites from the Chadan section (Kul'kov and Obut, 1973; Sennikov, 1996).

were undertaken to clarify the age of individual sections and stratigraphic subdivisions (Khalfin, 1961; Kul'kov, 1967; Ivanovsky and Kul'kov, 1974; Yolkin and Zheltonogova, 1974; Sennikov, 1976, 1996; Yolkin, 1983; Kul'kov and Severgina, 1989). The Chagyrka Formation was reassigned to the Wenlock based on fossils from its stratotype section (Kul'kov, 1967). Two new formations were named, the Kuimov and Cherny Anui Formations, in the lower and upper Ludlow, respectively (Kul'kov, 1967). Four additional formations (Chesnokovka, Polati, Syrovatiy, and Vtoriye Utyosy) were later proposed (Yolkin and Zheltonogova, 1974; Sennikov, 1976). They are based on bed-by-bed descriptions of many sections that demonstrate a cyclic pattern of deposition. Re-evaluation of previous correlations of these sections was accomplished with key fossil groups (graptolites, brachiopods, trilobites, rugose and tabulate corals). Symmetrical and asymmetrical cycles, regarded as sedimentary series (Yolkin and Zheltonogova, 1974, figs. 4, 12), were also recognized. It so happened that all formational boundaries were originally defined in a way that coincided with cycle boundaries (Yolkin and Zheltonogova, 1974, fig. 17) and are allostratigraphic units. They demonstrate lateral changes in facies within individual units that do not require a change in name (e.g., Brett et al., 1998). This approach permits Silurian subdivisions to be traced

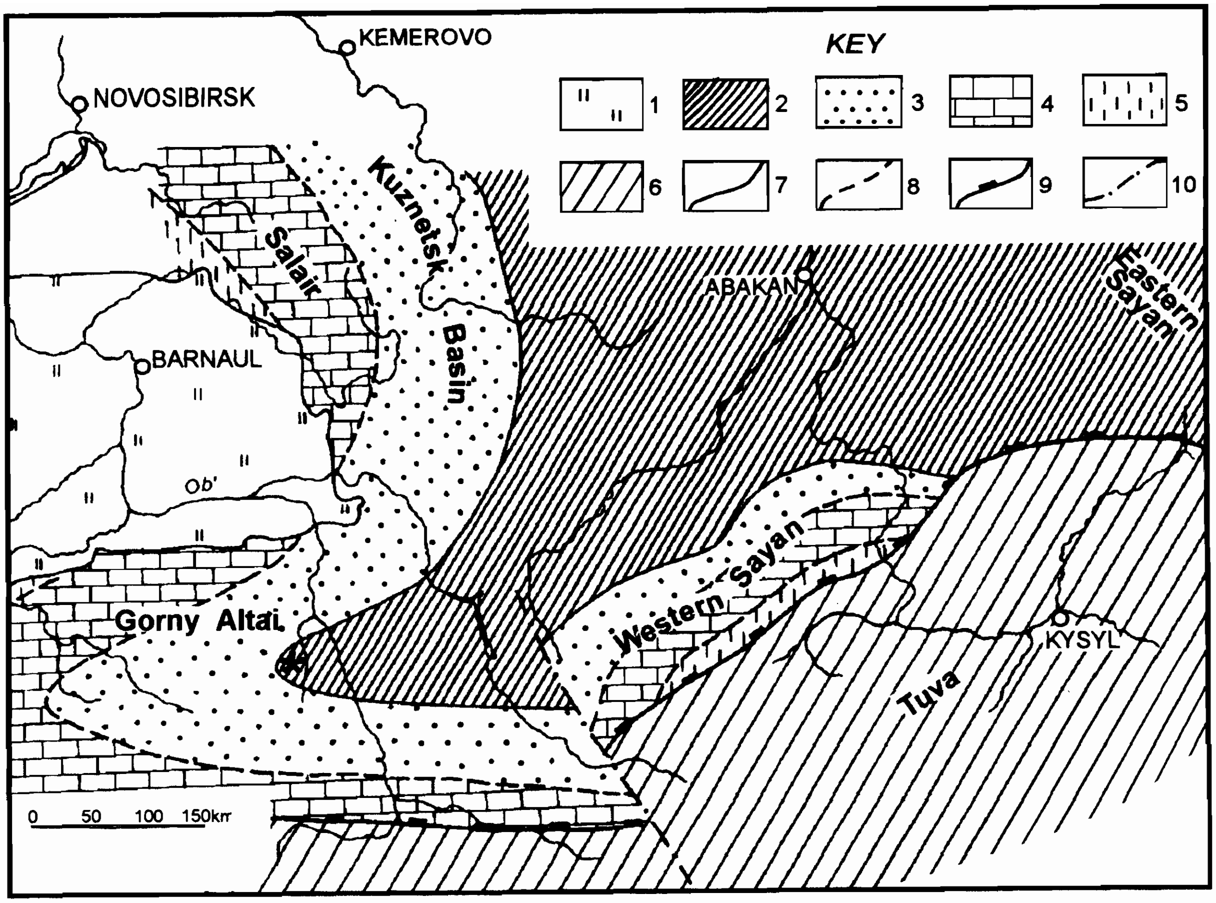


FIGURE 10 — Late Llandovery (Telychian) paleogeography of Altai–Sayan folded area. Key: 1, Cenozoic; 2, land; 3, inner shelf; 4, outer shelf (carbonate platform); 5, slope; 6, Tuvan Silurian with *Tuvaella* fauna; 7, land–sea boundary; 8, boundaries of shelf zones; 9, north limit of deposits with *Tuvaella* fauna; 10, fault. Asterisk shows angular unconformity of Devonian on Cambrian (Yolkin, 1968).

throughout the entire Altai region and allows biotic–sedimentary events to be used for inter-regional correlations (Yolkin et al., 1997a; Yolkin, 1998).

SALAIR AND WESTERN SAYAN — The Silurian of these regions is poorly known and poorly exposed. It often occurs in tectonically complicated areas or in nearly inaccessible mountainous territory. Nevertheless, available data have been synthesized into proposed stratigraphic schemes (Fig. 7). Although still provisional, some of the intervals are recognized as equivalents of Altaian units.

Three Silurian formations in the Salair region that are represented in composite sections have diverse Llandovery–Ludlow faunas (Kharin, 1960, 1968; Kharin et al., 1965; Ivanovsky and Kul'kov, 1974). Much more work needs to be done on the known sections to clarify their correlation and faunas. There is only one known interval in the Salair region (Baskuskan Formation) that corresponds precisely to the Polati Formation in the Altai (Fig. 7).

The Silurian of western Sayan (Fig. 7) is known in greater detail than that in the Salair region. This is the result of extensive work carried out by the late A. I. Naumenko (1964a, 1964b, 1969, 1970) in the 1960s. He studied many sections, some of which are metamorphosed and exposed in difficult mountainous terrain. His large collections include brachiopods, corals, stromatoporoids, bryozoans, and other fossil groups. He identified only the most widely distributed of these fossils, which are tabulate corals and brachiopods. Unfortunately, only a few of these taxa are described and illustrated, and this material requires re-examination. Some his identifications are useful, particularly those of *Pentamerus oblongus* (Sowerby) from the Aktash and Belaya Beds, *Tuvaella rackovskii* Tchernysheva from the Stoktysh Beds, and *T. gigantea* Tchernysheva from the Karakhem Beds. These guide fossils, together with sharp lithologic changes at the base of the Tostug Formation (=Chesnokovka/*Cyrtograptus* Event),

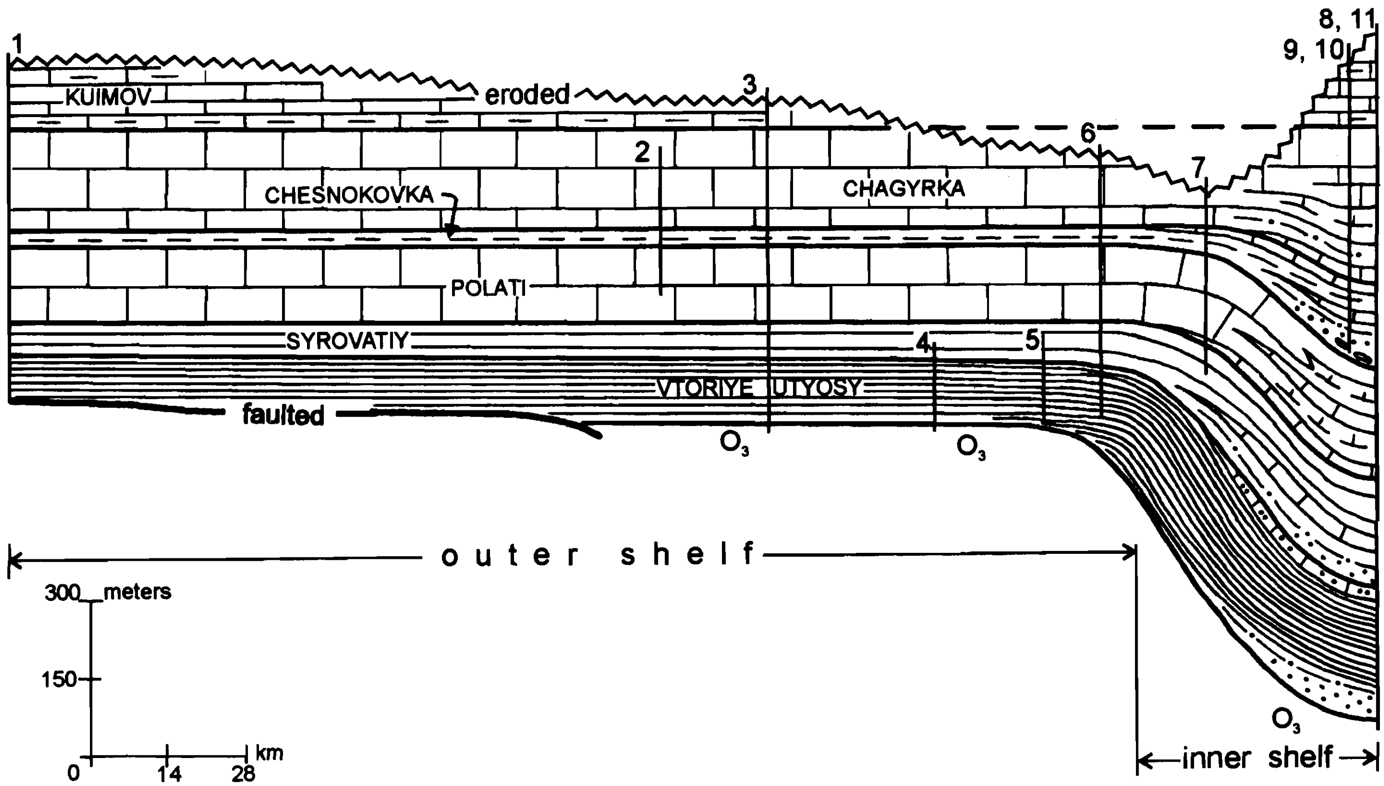


FIGURE 11 — Cross-section of Silurian in northwest Altai between Krasnoshchekovo village and Tigerek River. Vertical lines are measured sections (see Fig. 12). Other explanations in Fig. 13.

permit correlation of the Ona Formation in the western Sayan with the Syrovaty and Polati Formations of the Gorny Altai (Fig. 7).

TUVA — The Silurian of the Tuva Depression is well studied (e.g., Nikiforova and Obut (eds.), 1965; Vladimirskaya and Chekhovich, 1969; Vladimirskaya, 1978a, 1978b; Kul'kov et al., 1985; Vladimirskaya et al., 1986). It includes a highly endemic, shallow-water fauna characterized by many specimens of a few local species and genera. The most prominent genus is *Tuvaella.* This name is applied to an entire fauna that is widely distributed in Mongolia and the Tuva region, but also present in the Altai, Russian Far East, and North China (e.g., Nikiforova and Obut, 1965; Kul'kov, 1967; Kul'kov and Kozlov, 1978; Rong and Zhang, 1982). This large region is the Mongolo–Tuvan province (Vladimirskaya and Chekhovich, 1969; Vladimirskaya, 1973) or Mongolo–Okhotsk region (Boucot, 1975). The potential of the *Tuvaella* fauna for subdivision and correlation are limited to this biogeographic region. Biostratigraphic divisions established here may be correlated into other regions only by pelagic fossils (graptolites, conodonts). Two of the key Tuvan Silurian sections at Chadan and Elegest (Fig. 7) have these pelagic fossils (Moskalenko, 1970; Kul'kov and Obut, 1973; Sennikov, 1979, 1996; Vladimirskaya and Krivobodrova, 1985; Kul'kov et al., 1985). These fossils and eustatic event horizons allow correlation of a widely traceable, upper Llandovery interval (Fig. 8). However, there are still problems with correlation of the Pichi–Shuy and Taugan–Teli intervals. It should be pointed out that the Elegest section (Fig. 7) shows monotonous shallow-water deposits and a surprisingly smooth sea-level curve.

POSITION OF STANDARD BOUNDARIES — Modern knowledge of ASFA pelagic and benthic faunas permits recognition of most Silurian chronostratigraphic boundaries. Graptolites permit recognition of all of the standard Llandovery divisions in the Altaian sections (Sennikov, 1976, 1996). The boundaries of the Telychian Stage and the base of the Silurian are located just above the basal of the regional stratigraphic units (Fig. 6). In this respect, the earlier two- or three-fold or two-fold divisions of the Llandovery (Cocks et al., 1970) are more applicable to the Altaian Silurian.

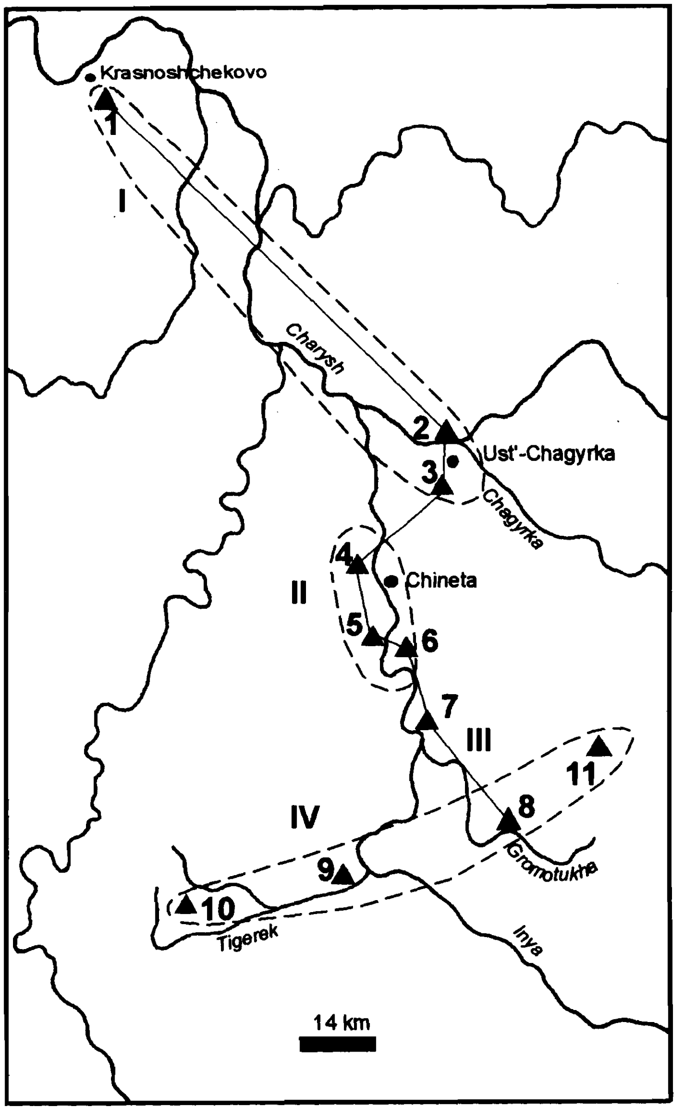


FIGURE 12 — Map of northwest Altai (see Fig. 4A) shows location of four reference districts: I, Chagyrka–Krasnoshchekovo; II, Chineta–Rossypnaya; III, Syrovatiy; IV, Tigerek–Gromotukha–Generalka. Triangles with numbers show locations of selected sections: 1, Krasnoshchekovo village area; 2, right bank of Charysh River; 3, left bank of Chagyrka River; 4, Burovlyanka Brook; 5, Vtoriye Utyosy [Second Rocks] on left bank of Inya River; 6, Rossypnaya Mount; 7, Syrovatiy Ravine; 8, right bank of Gromotukha River; 9, Strawberry Gully; 10, Chesnokovka Brook; 11, Generalka village area.

In contrast to the Llandovery, the Wenlock–Ludlow in the Altai–Sayan (except in Tuva) is characterized by benthic associations with wide distribution. They include stromatoporoids, corals, brachiopods, and trilobites (Khalfin, 1961; Naumenko, 1964a, 1970; Kul'kov, 1967; Ivanovsky and Kul'kov, 1974; Yolkin and Zheltonogova, 1974). These latter reports allow definition of the base of the Ludlow Series at the base of the Kuimov Formation (Fig. 6). This level probably corresponds to the standard definition of the Ludlow base at the base of the *Neodiversograptus nilssoni* Zone (Holland and Bassett, 1989). The base of the Kuimov Formation corresponds, as in many regions of the world (Johnson, 1996), to a transgressive pulse and to a significant change in faunas (Yolkin and Zheltonogova, 1974) that coincide with the early Gorstian, *Monograptus ludensis* bio-event (Kaljo et al., 1996).

There are no paleontological data available that help define the base of the Pridoli Series in the ASFA sequences. The Pridoli base is usually correlated with the lower boundary of the Cherny Anui Formation (Ivanovsky and Kul'kov, 1974; Yolkin and Zheltonogova, 1974; Yolkin, 1983) and its equivalents in other regions of Russia (e.g., Tsyganko and Chermnykh, 1983). However, indirect evidence may be used to shift the Pridoli base upward into the Cherny Anui Formation (Yolkin and Zheltonogova, 1974). It should be noted that the stratotype of this formation comprises a symmetrical sedimentary cycle (Yolkin and Zheltonogova, 1974, fig. 16). This cycle is considered to reflect eustasy, and its base may correspond to the transgressive, early Ludfordian bioevent of the Ludlow (Johnson et al., 1991; Kaljo et al., 1996). If so, only the upper, regressive stage of the Cherny Anui Formation cycle could be assigned to the Pridoli (Yolkin, 1998, fig. 5).

The lower Homerian and Ludfordian boundaries may be defined only provisionally in the Altai. Very likely, the base of the Homerian is located slightly below the Chagyrka Formation, and coincides with the start of the stillstand stage of T-R cycle 3 (Fig. 7) and the origination of new morphological forms of dechenellid trilobites (Yolkin and Zheltonogova, 1974). This biotic–depositional event may be coeval with other developments near the base of the *Monograptus duebeli–Gothograptus nassa* Zone (Kaljo et al., 1996). The lower Ludfordian boundary is placed provisionally within the Kuimov Formation, and at the level of a clearly expressed T-R event (Fig. 7).

Regional flooding surfaces located near some of the standard Silurian boundaries are traceable from the Altai into other regions of the ASFA. The most recognizable levels are the base and top of the upper Llandovery or Syrovatiy–Cyrtograptus sedgwickii and Chesnokovka–*Cyrtograptus* events (Yolkin, 1998; Fig. 7).

Paleogeography and Depositional Environments

The first attempt to outline the trend of the Silurian shelf on the southwestern margin of the Siberian paleocontinent was done in the Salair and northern Altai regions (Yolkin et al., 1994). A set of maps for time slices from the

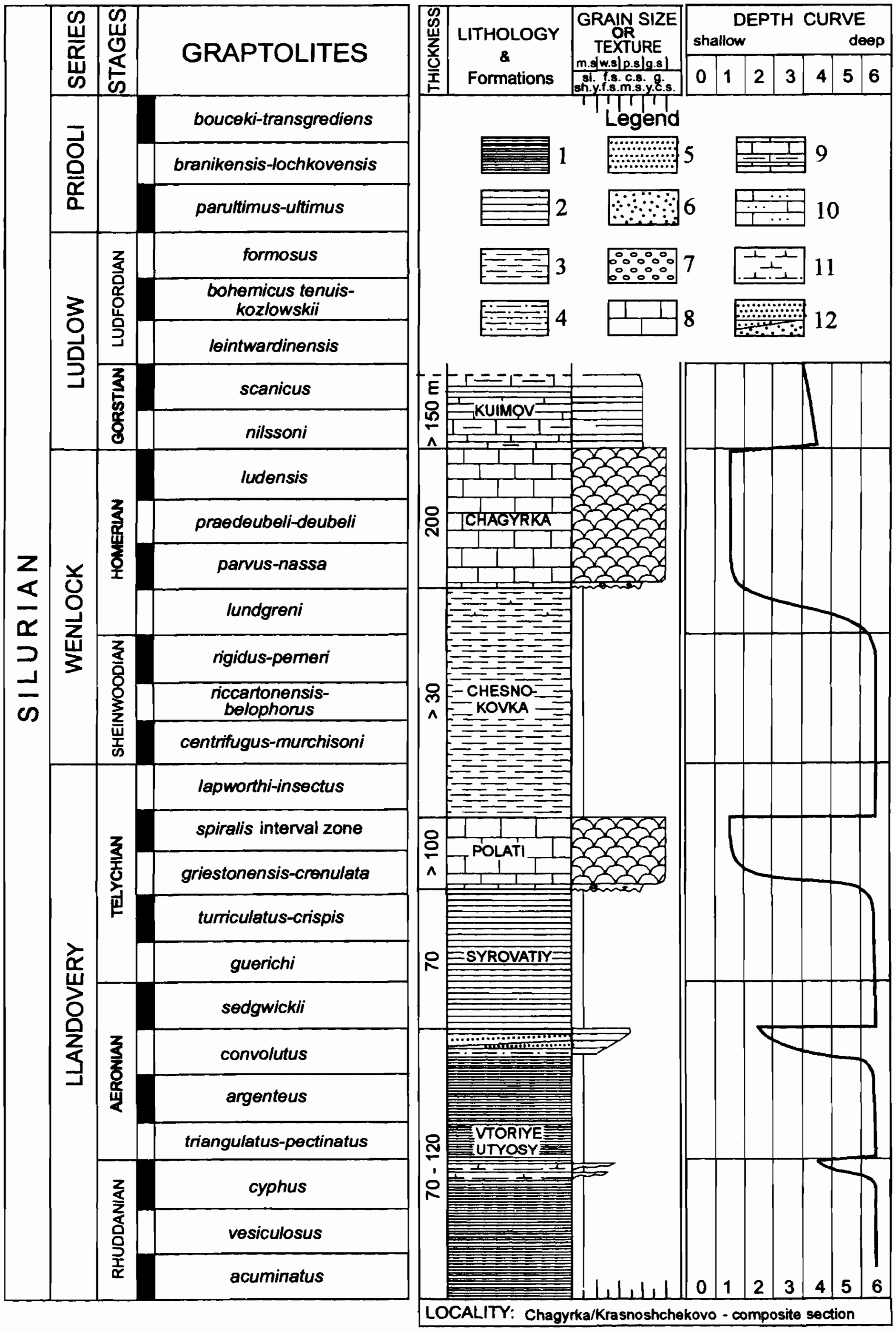


FIGURE 13 — Correlation chart for Chagyrka–Krasnoshchekovo reference district (see Fig. 12). Sea-level curve calibrated by benthic assemblages and sedimentary features (after Brett et al., 1993). Legend: 1, black or dark shale; 2, green or grey shale; 3, mudstone; 4, siltstone; 5, sandstone; 6, coarse-grained sandstone; 7, conglomerate; 8, massive limestone; 9, argillaceous limestone; 10, sandy limestone; 11, calcareous shale; 12, cross-bedded sandstone.

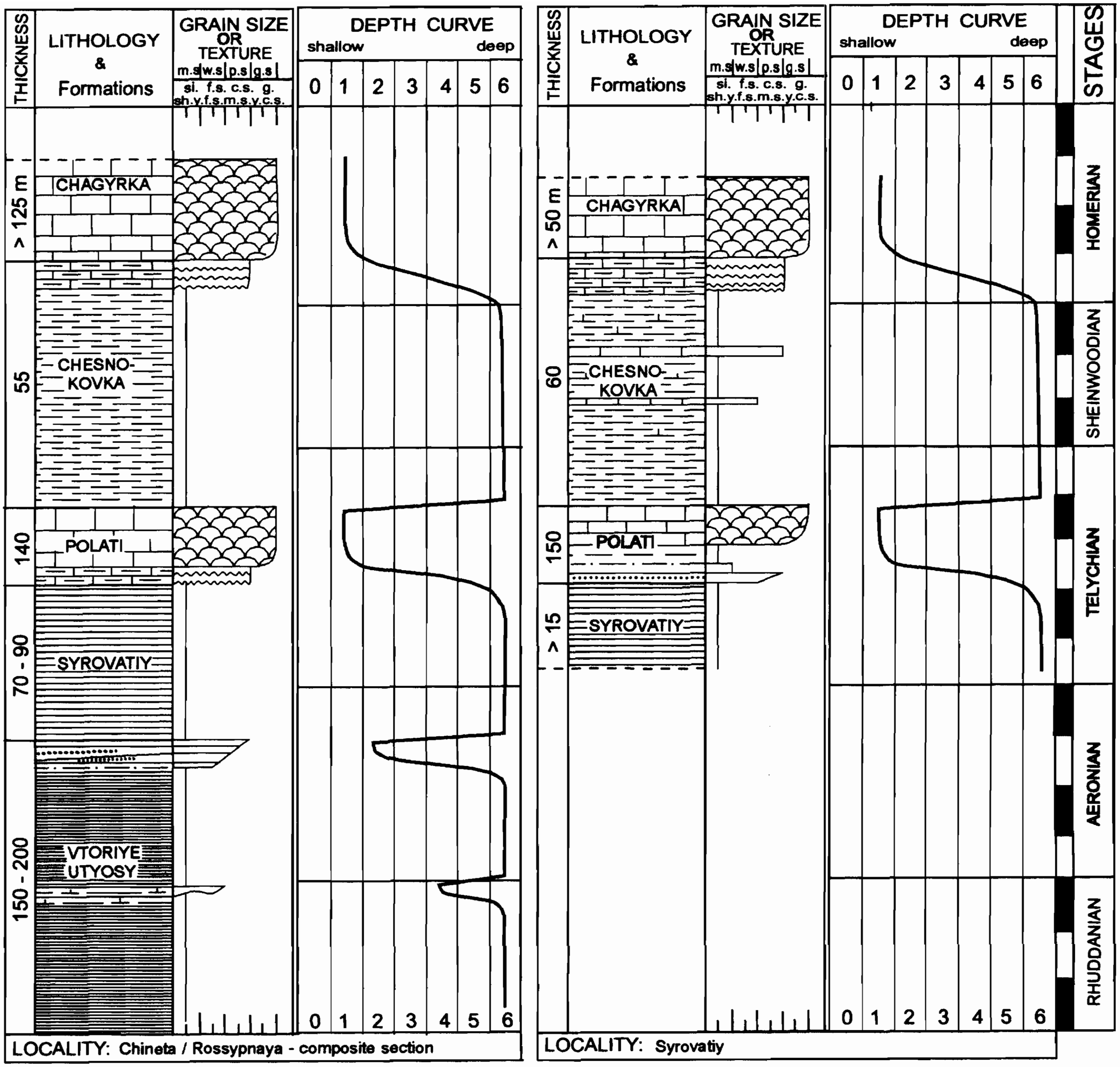


FIGURE 14 — Correlation chart for Chineta–Rossypnaya and Syrovatiy reference district (see Fig. 12). Explanation of symbols in Fig. 13.

Tremadoc to the Middle Devonian prepared at that time focused mainly on reef-building episodes. A general picture was shown of the developing continental margin that could be interpreted in terms of geodynamics.

In this report, we used all available data on the Silurian in the ASFA. The most appropriate interval for a paleogeographic reconstruction is the late Telychian. In terms of local stratigraphy, the appropriate unit is the Polati Formation in the Altai and its equivalents in other regions (Fig. 9). All of these units are represented mainly by reef limestones that are easily identified even in small exposures, and provide a good feature for paleogeographic analysis. In addition, lateral and vertical facies changes are shown for the Silurian on the correlation

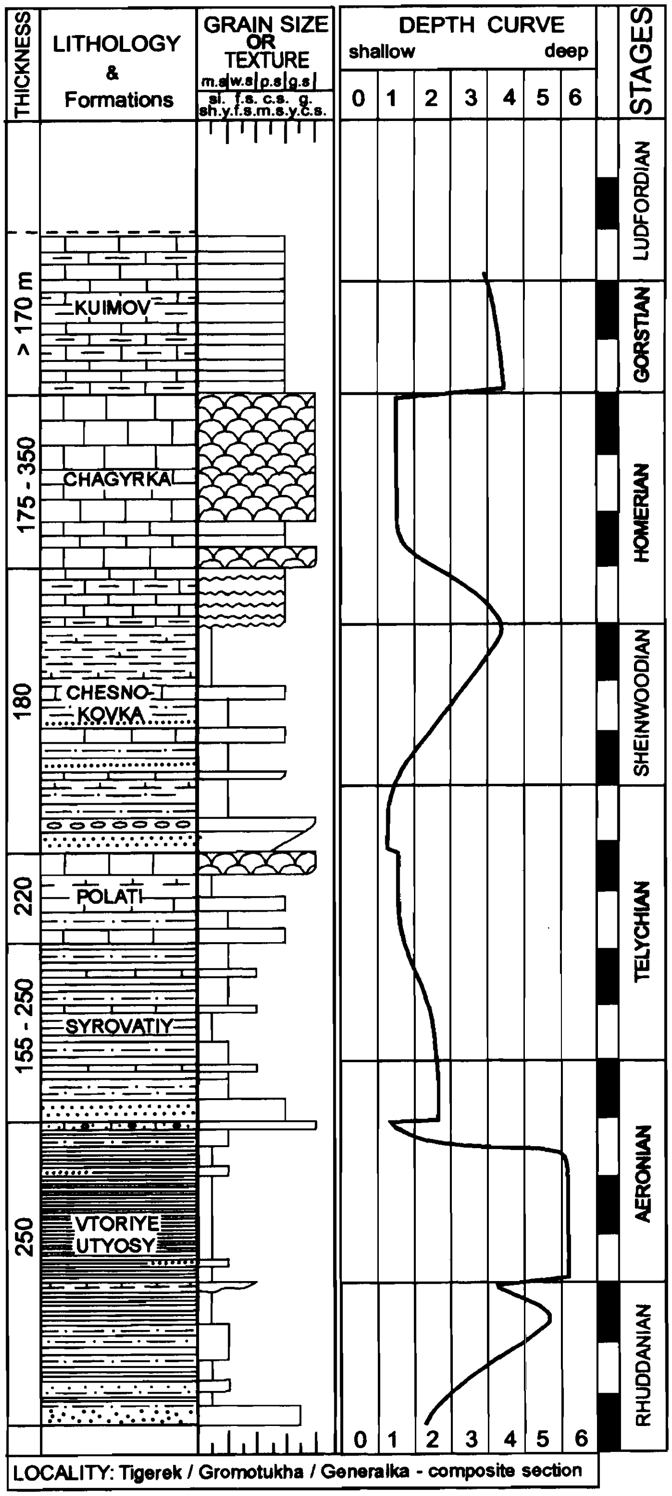


FIGURE 15 — Correlation chart for Tigerek–Gromotukha–Generalka reference district (see Fig. 12). Explanations of symbols in Fig. 13.

charts and cross-sections. These data also apply to considerations of biofacies and sea-level fluctuations.

The upper Llandovery (Telychian) is correlated by graptolites or benthic fossils (Fig. 4). These units help delineate the principal facies boundaries that separated land and sea and the inner and outer shelf zones (Fig. 10). The slope zone is tentatively shown in this figure. Inner and outer shelf facies zones are seen in the Silurian correlations in three territories of the Gorny Altai (Fig. 4A, B, and C). The lithologic columns in each chart are composite sections created by correlating marker horizons between local sections. These sections are detailed in a number of reports (Kul'kov, 1967; Ivanovsky and Kul'kov, 1974; Yolkin and Zheltonogova, 1974; Sennikov, 1976, 1996; Yolkin et al., 1978, 1988; Sennikov et al., 1984; Gladkikh et al., 1989).

Cross-section of the northwestern Altai — Lateral facies changes for the entire Silurian in this region are shown in the cross-sections (Fig. 11). The axis for this section follows the Charysh and Inya Rivers (Fig. 12). Each of the four districts in this territory is represented by a correlation chart (Figs. 13–15) to demonstrate vertical facies changes and depth curves. Well-defined outer- and inner-shelf zones are shown. The last section is in a tectonized foreland trough filled with siliciclastics and carbonates of variable thicknesses. The outer shelf has two well-developed (upper Llandovery and upper Wenlock) carbonate platforms. They are algal-reef limestones with locally rich benthic faunas. The basal parts of the outer-shelf sequences include mainly fine siliciclastics with diverse graptolites. Sea-level curves for the outer-shelf sections consistently show four shallowing episodes that successively become more severe (Figs. 13, 14). Inner-shelf sections display the same shoaling episodes (Fig. 15).

Cross-section of north-central Altai — This cross-section (Fig. 16) cuts through a part of the ASFA with a complete Silurian succession. It is located about 100 km further northeast along the strike of the shelf from the northwest Altai cross-section described above (Fig. 4). It shows the same lithofacies. The southeast end of this cross-section is located near the shoreline (Figs. 10, 17) and shows coarser-grained siliciclastics (Fig. 19). Neither the late Llandovery nor late Wenlock carbonate platform deposits record any significant facies changes. The late Wenlock platform slightly expanded toward the foreland trough. Sea-level curves also show the same shallowing intervals (Figs. 18, 19).

Cross-section of the southeast Altai — This cross-section (Fig. 20) is located on the other side of the "Altaian Silurian peninsula" (Figs. 4, 10). It is comprised mainly by Llandovery deposits. In contrast to the above cross-sections described above, the facies of the southeast

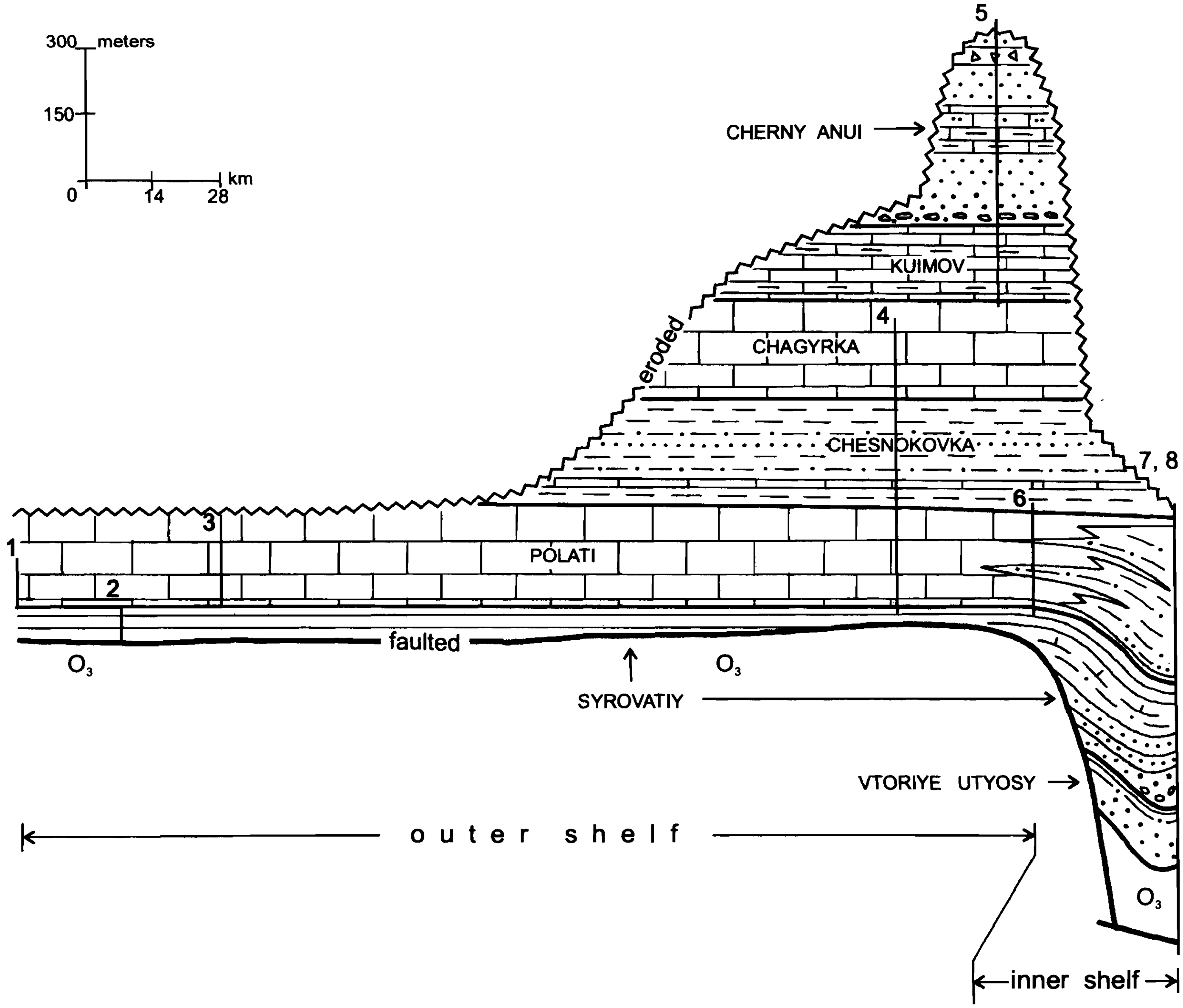
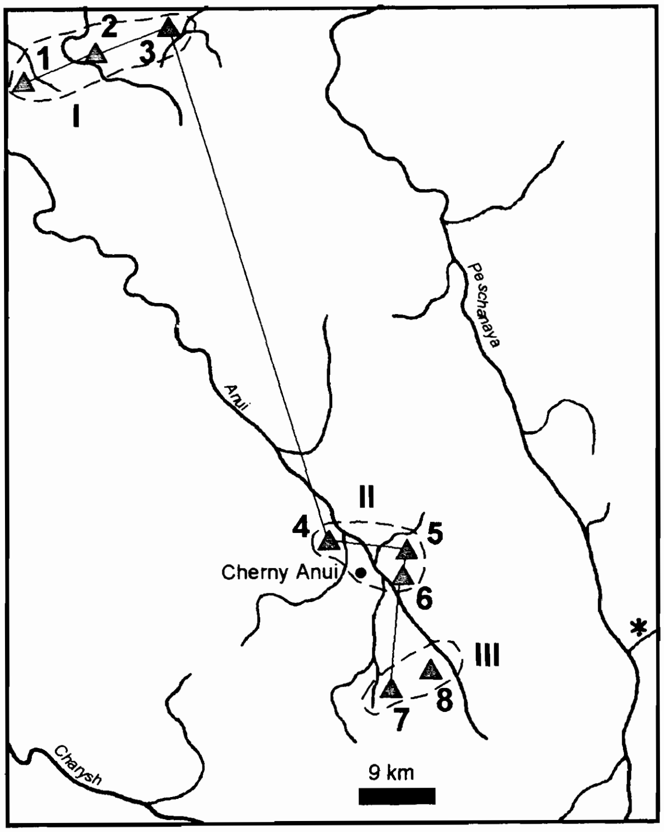


FIGURE 16 — Cross-section of Silurian in northwest Altai along Anui River. Vertical lines show locations of studied sections (see Fig. 17). Explanation of symbols in Fig. 13.

Altai cross-section are quite different (Fig. 21). The inner–outer shelf zone boundary is defined only provisionally, because of gradual proximal–distal facies changes. However, the sea-level curves (Figs. 22, 23) exhibit the same shallowing intervals as the carbonate platform sections on the another side of the "Altaian Silurian peninsula" (Figs. 13, 14).

Thus, the Altaian part of the late Llandovery ASFA shelf records some differences on the north and south sides of a "peninsula." On the north side, the shelf is well differentiated into two facies belts. The inner shelf could be considered here as a back-reef basin. The carbonate platform, or outer-shelf zone, is easily traced into the Salair region, where it forms a Silurian barrier reef. This major buildup also continues around the "peninsula" into western Sayan (Naumenko, 1970). As in the southeastern Altai, the western Sayan shows lateral transitions from near-shore trough siliciclastics to reef limestones on



the carbonate platform and into distal, flysh-like, fine-grained rocks. Another type of sedimentation is recorded in neighboring Tuva. It is characterized by continuous, extremely shallow-water deposits (Fig. 8). The sea-level curve for this section shows only very weak shallowing and deepening events. These features support the view that Silurian deposition in the Tuva and ASFA seas was very different.

Communities and Biogeographic Faunal Affinity

An attempt to recognize brachiopod communities in the Altai–Sayan was made by Kul'kov (1974). He identified a *Tuvaella* Community and a *Pentamerus–Pentameroides* Community and proposed a new guide fossil (*Aegirina*) for the *Clorinda* Community. The potential for a more

FIGURE 17 (left) — Map of north-central Altai (see Fig. 4B) with locations of three reference districts: I, Solov'ikha–Kamyshenka; II, Cherny Anui; III, Dietken–Beliy Anui. Triangles with numbers locate individual sections: 1, Ganin Brook; 2, Solov'ikha River; 3, Kamyshenka River; 4, Karakol River; 5, Cherga River; 6, Turata village; 7, Dietken Brook; 8, Beliy Anui village. Asterisk marks Cambrian–Devonian unconformity (Yolkin, 1968).

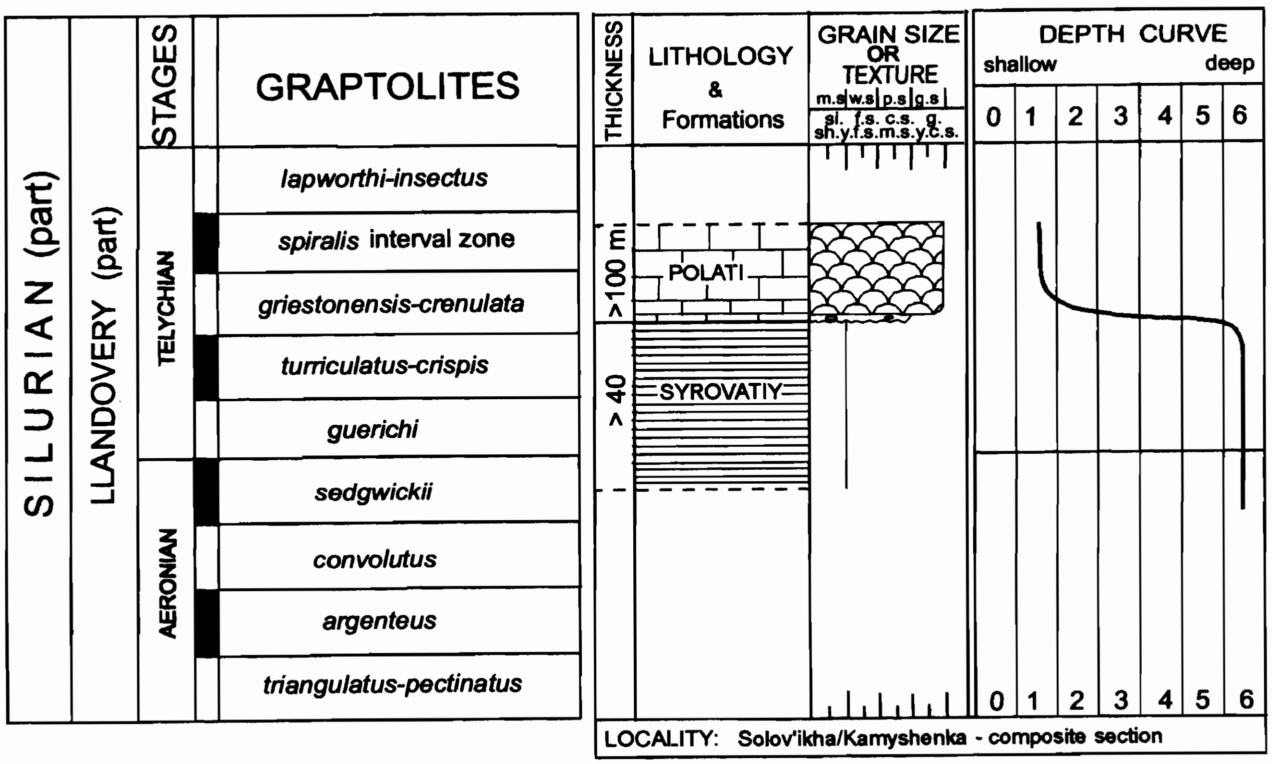


FIGURE 18 — Correlation chart for Solov'ikha–Kamyshenka reference district (see Fig. 17). Explanations of lithologic symbols in Fig. 13.

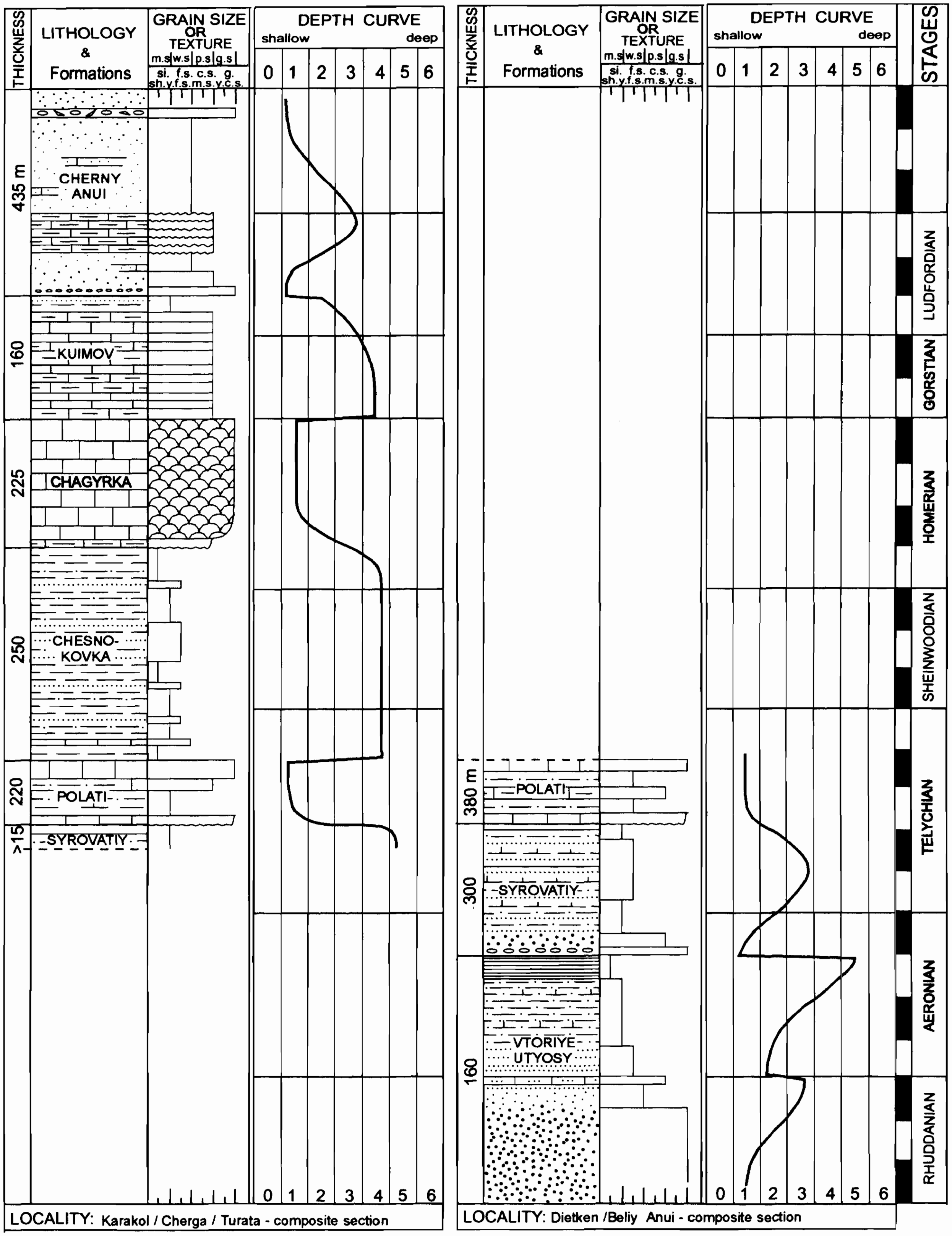


FIGURE 19 — Correlation charts for the Cherny Anui and Dietken–Beliy Anui districts (see Fig. 17). Explanations of lithologic symbols in Fig. 13.

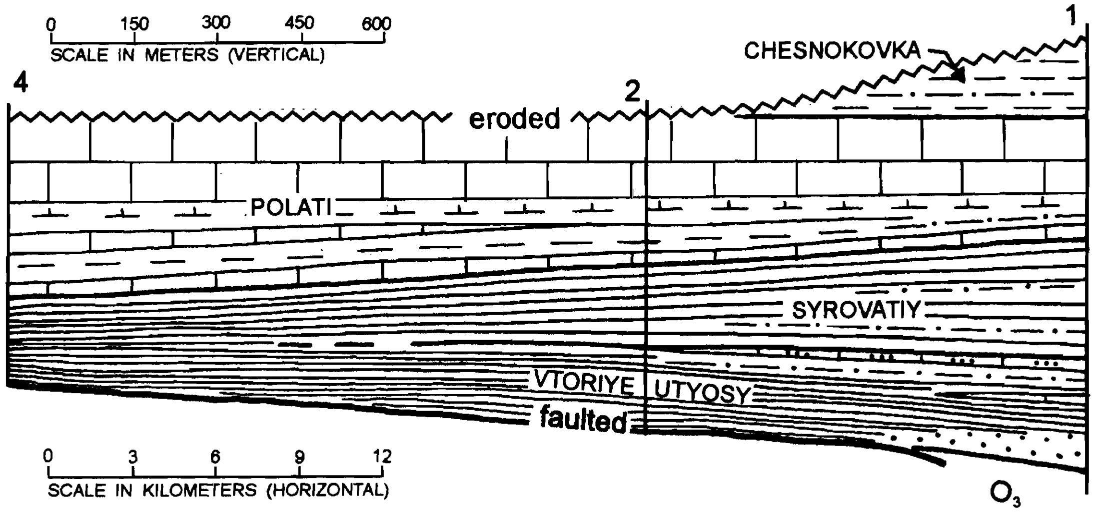


FIGURE 20 — Cross-section of Llandovery in southeast Altai. Vertical lines show locations of sections (see Fig. 21). Explanation of symbols in Fig.13.

sophisticated paleocommunity analysis in the Altai–Sayan is limited. As shown by the synthesis of late Llandovery paleogeography, only two very different lithofacies occur: deep-water, fine-grained siliciclastics with graptolites, and a shallow-water facies. Both of these lithofacies abruptly alternate within cyclic successions.

Biogeographic affinity is a problem only for the *Tuvaella* fauna. This problem has been examined on the basis of affinity indices for brachiopod genera by Rong et al. (1995). These authors compared Llandovery brachiopods from South China, Tuva, Altai, Kazakhstan, and the Siberian platform. They found such a high degree of affinity between the Altai and Tuva faunas that they combined them in a comparison with brachiopods elsewhere in east Asia. We have listed late Llandovery brachiopod species from the Altai, Tuva, and Siberian platform (Fig. 24) and calculated affinity indices using the same formulas. The results show a very limited affinity and suggest significant barriers (climatic or oceanic) between the Altai and Tuva regions in the Llandovery.

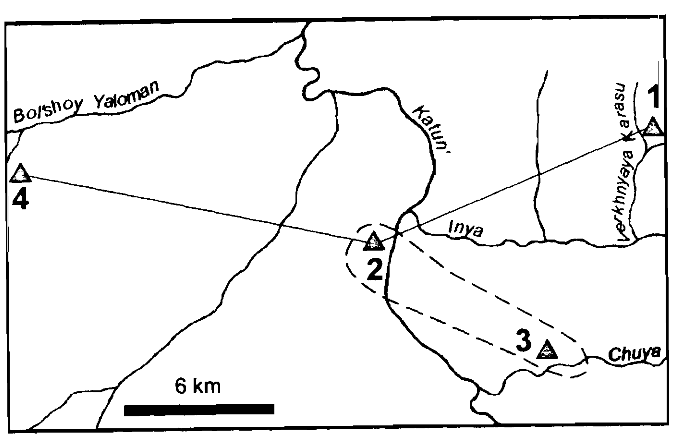


FIGURE 21 — Map of southeast Altai (see Fig. 4C) with locations of reference sections: 1, Karasu; 2, Katun'; 3, Chuya; 4, Yaloman.

Sea-Level Fluctuation and Depth Curves

Sea-level fluctuations in the Middle Paleozoic of the Gorny Altai and Salair regions were recently detailed (Yolkin et al., 1997 a, b; Yolkin, 1998). Each correlation chart in this report is accompanied by depth curves. It is useful to examine depth curves for adjacent outer- and inner-shelf zones and compare them with curves from other regions (Fig. 25). All three Altaian depth curves are generally similar. The left column (seaward) demonstrates cyclic sediment accumulation and a progressive shallowing trend in the basin. The next two curves reflect near-shore sediment accumulation conditions. In this case, the right column shows Silurian deposition from the start of transgression to final regression. The depth curve for the Tuva Silurian section (also Fig. 8) is absolutely different. This could mean that the Tuva region comprises a separate terrane with a distinctly different epeirogenic

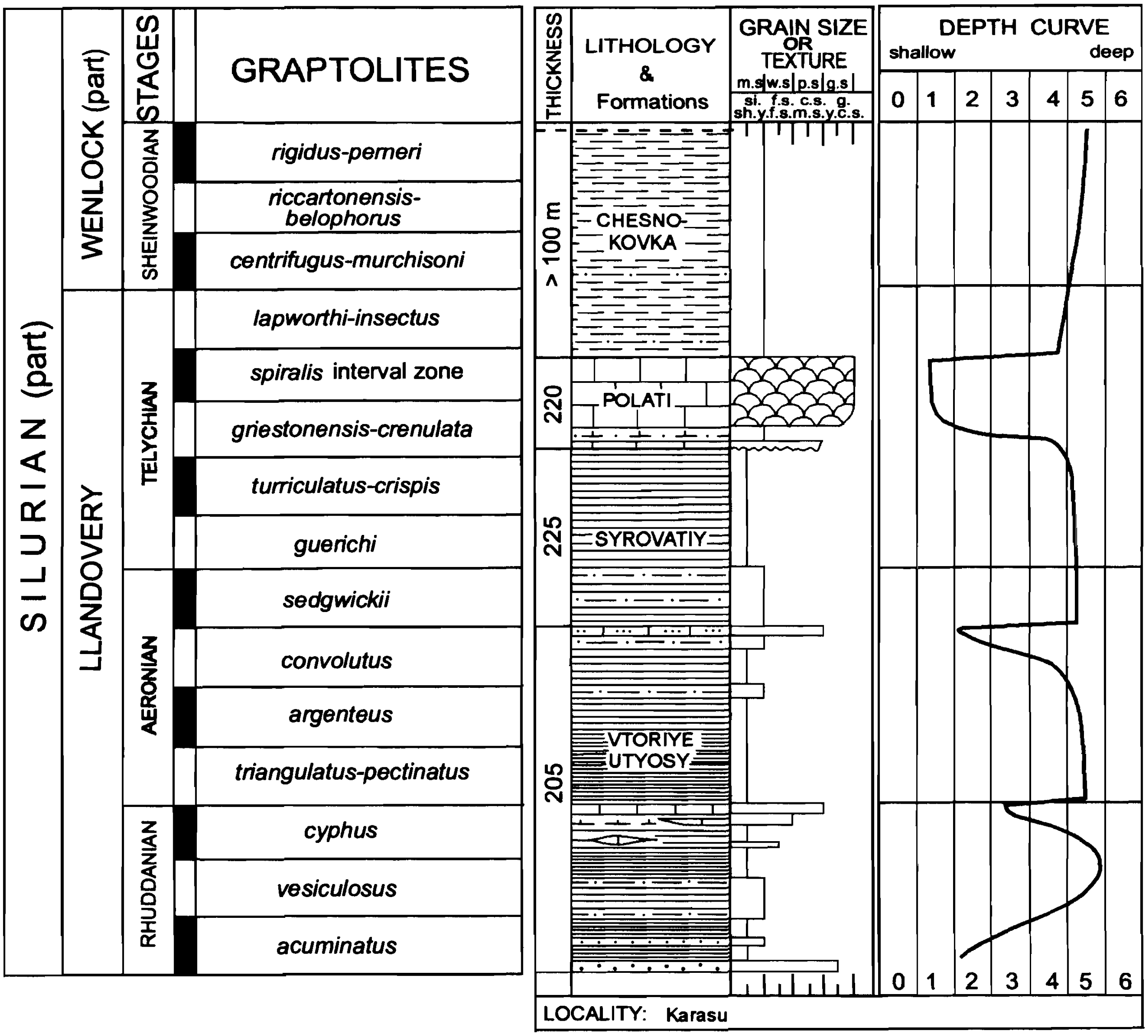


FIGURE 22 — Correlation chart for the Verkhnyaya Karasu River (see Fig. 21). Explanations of lithologic symbols in Fig. 13.

history from that of the Altaian margin of the Siberian paleocontinent.

Climatic Indicators and Application to Silurian Geography

The paleogeographic reconstructions by Scotese and McKerrow (1990) are widely applied. However, their location of the Siberian paleocontinent during the Silurian is greatly in error. This ancient craton should be located nearly in its present orientation, but was shifted by them along its ASFA margin to the tropics (Fig. 5). The main evidence for this interpretation includes: 1) wide distribution in the ASFA of Late Ordovician–Late Devonian barrier reefs and warm-water faunal associations, as confirmed by Pedder and Oliver (1990); and 2) great differences between late Llandovery brachiopod associations from the Altai and the northwest Siberian platform (Fig. 24). These regions are interpreted as warm- and cool-water areas, respectively.

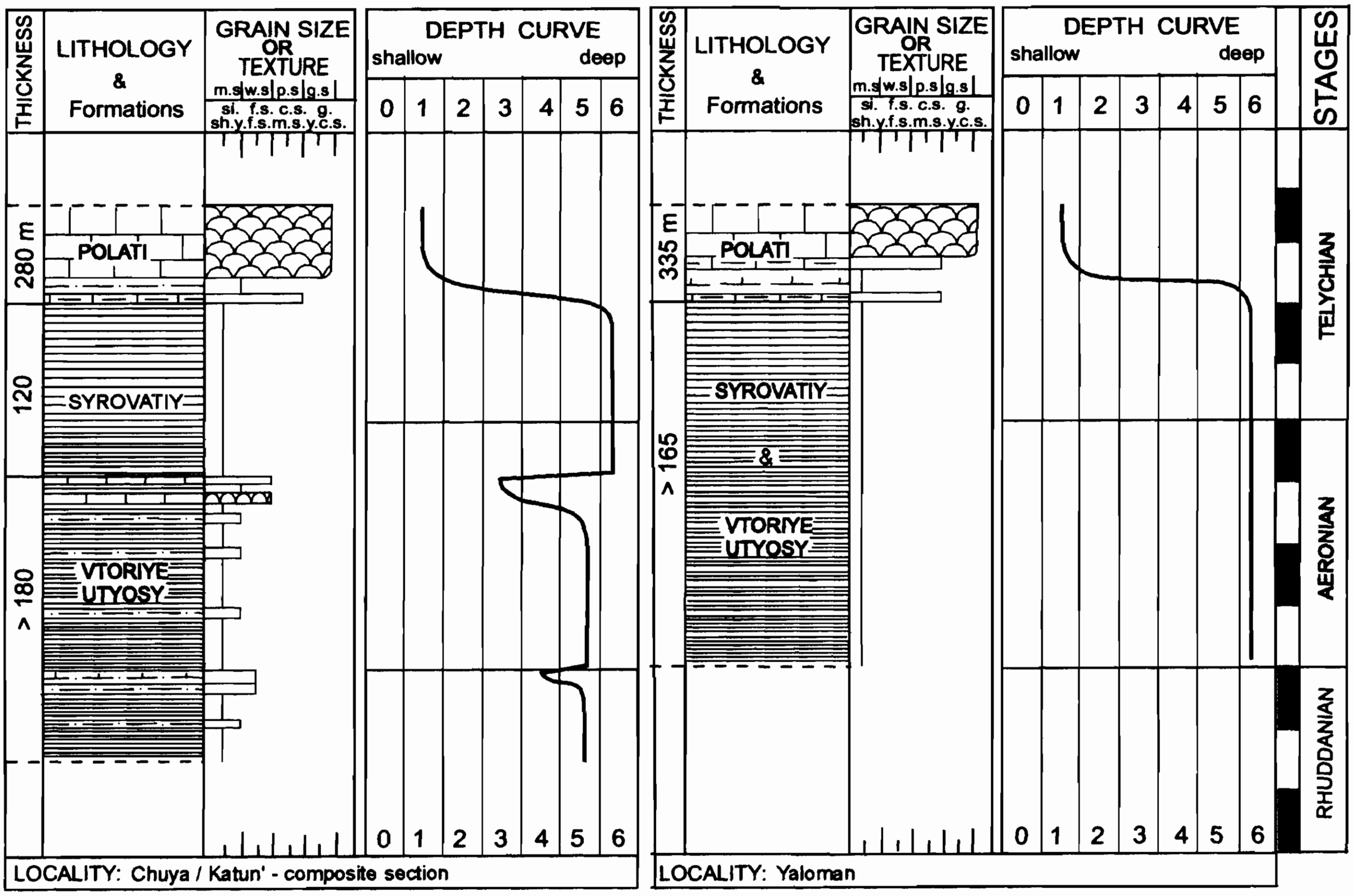


FIGURE 23 — Correlation charts for the Chuya–Katun' and Yaloman reference sections (see Fig. 21). Explanations of lithologic symbols in Fig. 13.

Conclusions

Data from four regions in the Altai–Sayan folded area (Altai, Salair, western Sayan and Tuva) allow reconstruction of the southwestern shelf of the Siberian paleocontinent and constrain its geographic position during the Silurian. These data involve stratigraphy, cyclicity, sedimentary and biotic events, faunal affinity, and paleogeography. The principal conclusions are as follows: First, the Silurian shelf of the Siberian paleocontinent lay on the western part of the ASFA territory and was a passive continental margin. This margin may be traced from the Salair to the Altai and further into western Sayan. Second, the Tuva area was separate during the Silurian from the rest of the ASFA. Third, large barrier reefs in the ASFA and warm-water faunas locate this part of the Siberian paleocontinent in the tropical realm.

Acknowledgments

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Taxa	Altai	Siberian Platform	Tuva
1	2	3	4
1. *Aerigia norilskensis* Lop.		+	
2. *Alispira gracilis* Nikif.		+	
3. *Alispira ? rotundata* Nikif. et T.Modz.		+	
4. *Alispira tenuicostata* Nikif.		+	
5. *Amphistrophia* cf. *funiculata* (McCoy)			+
6. *Amphistrophia striata* (Hall)			+
7. *Amphistrophia tchernychevi* Kulk.			+
8. *Anabaria rara* (Nikif.)		+	
9. *Arctomeristina tchadanica* (Kulk.)			+
10. *Atrypa* ex. gr. *orbiculatus* (Sow.)	+		
11. *Atrypa hedei* Struve			+
12. *Atrypa reticularis* (Linnaeus)			+
13. *Atrypa subquadrata* Rybk.			+
14. *Atrypina dichotoma* Kulk.	+		
15. *Atrypopsis absimilis* Rybk.			+
16. *Atrypopsis chondelensis* Rybk.			+
17. *Atrypopsis legrinus* Kulk.	+		+
18. *Brachyprion (Protomegastrophia) basseti* Kulk.			+
19. *Brachyprion (P.) semiglobosa*(Dav.) *pygmaea* Kulk.			+
20. *Brachyprion arenacea* (Dav.)	+		
21. *Brevilamnulella undatiformis* Rozman		+	
22. *Camarium protadjunctus* Kulk.	+		
23. *Carinatina praearimaspus* Nikif.	+		
24. *Clorinda* (?) *minor* Kulk.	+		
25. *Clorinda substantiva* Kulk.	+		
26. *Clorinda undata* (Sow.)	+	+	
27. *Coolinia gorbiyatchense* Lop.		+	
28. *Coolinia gracilis* (Andr.)		+	
29. *Cordatomyonia disjuncta* Vlad.			+
30. *Cryptatrypa praecordata* Kulk.	+		
31. *Cryptothyrella lacrima* (Nikif.)		+	
32. *Cryptothyrella tchadanica* Kulk.			+
33. *Dalejina* ex gr. *hybrida* (Sow.)		+	
34. *Dalejina tchernychevi* Vlad.			+
35. *Dolerorthis* aff. *interplicata* (Foerste)	+		
36. *Dolerorthis karasugensis* Vlad.			+
37. *Dolerorthis ? llandoveriensis* Lop.		+	
38. *Elegesta nikiforovae* Vlad.			+
39. *Eocoelia hemisphaerica* (Sow.)		+	
40. *Eohowellella yadrenkinae* Lop.		+	
41. *Eoplectodonta* cf. *penkillensis* (Reed)	+		+
42. *Eoplectodonta duvali* (Davidson)			+
43. *Eoplectodonta ? pumila* Lop.		+	
44. *Eospirifer decorus* Kulk.	+		
45. *Eospirifer parvus* Kulk.	+		
46. *Eospirifer radiatus* (Sow.)			+
47. *Eospirifer tuvaensis* Tchern.			+
48. *Eospirigerina* (?) *groenlandica* (Poulsen)			+
49. *Eridorthis ? siluriensis* Lop.		+	
50. *Gacella originata* Kulk.			+
51. *Glassia minuta* Rybk.			+
52. *Hedeina araargensis* Vlad.			+
53. *Hesperorthis rubeli* Lop.		+	
54. *Holtedahlina* (?) *parva* Bancroft	+		
55. *Howellella* cf. *splendens* (Thomas)	+		
56. *Howellella dashtygoica* Vlad.			+
57. *Howellella elevataeformis* Lop.		+	
58. *Howellella tapsaensis* (Tchern.)			+
59. *Idiospira khetaensis* (Nikif.)		+	
60. *Idiospira kuntikahina* Lop.		+	
61. *Isorthis angaciensis* Vlad.			+
62. *Isorthis neocrassa* (Nikif.)		+	
63. *Isorthis tannuolis* Vlad.			+
64. *Janius exsul* (Barrande)			+
65. *Kulumbella kulumbensis* Nikif.		+	
66. *Leangella scissa* (Dav.)	+		
67. *Lenatoechia elegans* (Nikif.)		+	
68. *Lenatoechia ramosa* (Nikif.)		+	
69. *Leptaena depressa* (Sow.)			+
70. *Leptaena kyziltchiraensis* Kulk.			+
71. *Leptaena parvissima* Kulk.	+		
72. *Leptaena rhomboidalis minuta* Lop.		+	
73. *Leptostrophia andreevae* Lop.		+	
74. *Leptostrophia (?) compressa* (Sow.)			+
75. *Leptostrophia filosa* (Sow.)			+
76. *Leptostrophia petrakovi* Lop.		+	
77. *Leptostrophia talikitensis* Lop.		+	
78. *Mclearnites prosperus* Kulk.			+
79. *Meifodia recta* (Nikif.)		+	
80. *Mendacella tungussensis* Nikif.		+	
81. *Merista fabulosa* Kulk.	+		
82. *Merista protadjuncta* Kulk.	+		
83. *Meristina obtusa* (Sow.)			+
84. *Nalivkinia grunewaldtiaeformis* (Peetz)	+		+
85. *Nalivkinia tuvinica* Vlad.			+
86. *Parastrophinela* aff. *altaica* Kulk.	+		
87. *Parastrophinela altaica* Kulk.	+		
88. *Pentameroides exactus* Kulk.	+		
89. *Pentamerus kamyschenskiensis* Kulk.	+		
90. *Pentamerus oblongus* (Sow.)	+	+	
91. *Pentlandina parva* (Bancroft)	+		
92. *Pentlandina subcostatula* (Lop.)		+	
93. *Pholidostrophia (Eoph.) sefinensis* (Williams)			+
94. *Pholidostrophia (Mesoph.) salopiensis* Cocks			+
95. *Pinquispirifer kadensis* Ivanova			+
96. *Plectatrypa imbricata* (Sow.)			+
97. *Plectatrypa lamellosa* (Lindstrom)			+
98. *Protatrypa alia* (Nikif.)		+	
99. *Protatrypa olga* Kulk.	+		
100. *Protatrypa septentrionalis* (Nikif.)		+	
101. *Rafinequina ? inaequicostata* Lop.		+	
102. *Rostricellula* (?) *lewisi* (Dawson)			+
103. *Rostricellula* (?) *nalivkini* (Tchern.)			+
104. *Rseudocamarotoechia ubsuensis* (Tchern.)			+
105. *Schizonema (?) kyzilchiraensis* Vlad.			+
106. *Septatrypa antiquata* Nikif.		+	
107. *Septatrypa lantenoisi* (Termier)	+		
108. *Septatrypa letnyaensis* Lop.		+	
109. *Septatrypa magna* Nikif.		+	
110. *Septatrypa pentagonalis* Nikif.		+	
111. *Sericoidea postrestricta* Kulk.			+
112. *Spirigerina grata* Kulk.	+		
113. *Spirigerina groenlandica* (Poulsen)	+		
114. *Spondylostrophia lata* Kulk.	+		
115. *Spondylostrophia sibirica* Kulk.	+		
116. *Stegerhynchus extendilatus* Lop.		+	
117. *Steger. decempeicatus duplex* Nikif. et T.Modz.		+	
118. *Stegerhynchus pseudonuculus* Nikif.		+	
119. *Stegerhynchus* aff. *praecursor* Foerste	+		
120. *Stegerhynchus tungussensis* Lop.		+	
121. *Stegerhynchella* (?) *angaciensis* (Tchern.)	+		+
122. *Stricklandia lens* (Sow.)		+	
123. *Stricklandia salteri* (Billings)		+	
124. *Stropheodonta aurita* Kulk.	+		
125. *Stropheodonta polaris* Andr.		+	
126. *Strophomena kulumbensis* Lop.		+	
127. *Strophomena pectenoides* Andr.		+	
128. *Strophomena sibirica* Andr.		+	
129. *Strophomena ? striatissima* (Poulsen)		+	
130. *Strophonella euglypha* (Dalman)			+
131. *Strophonella ? kulumbeana* Lop.		+	
132. *Tannuspirifer pedaschenkoi* (Tchern.)			+
133. *Tuvaella gigantea* Tchern.			+
134. *Tuvaella rackovski* Tchern.	+		+
135. *Tuvaerhynchus khalfini* Kulk.			+
136. *Zygospiraella duboisi* (Verneuil)		+	
137. *Zygospiraella planoconvexa* (Hall)		+	
Total	**37**	**53**	**53**

FIGURE 24 — Late Llandovery brachiopods from the Altai (after Kul'kov, 1967; Ivanovsky and Kul'kov, 1974; Kul'kov and Severgina, 1989), Tuva (after Kul'kov et al., 1985), and Siberian platform (after T.V. Lopushinskaya, unpub. data).

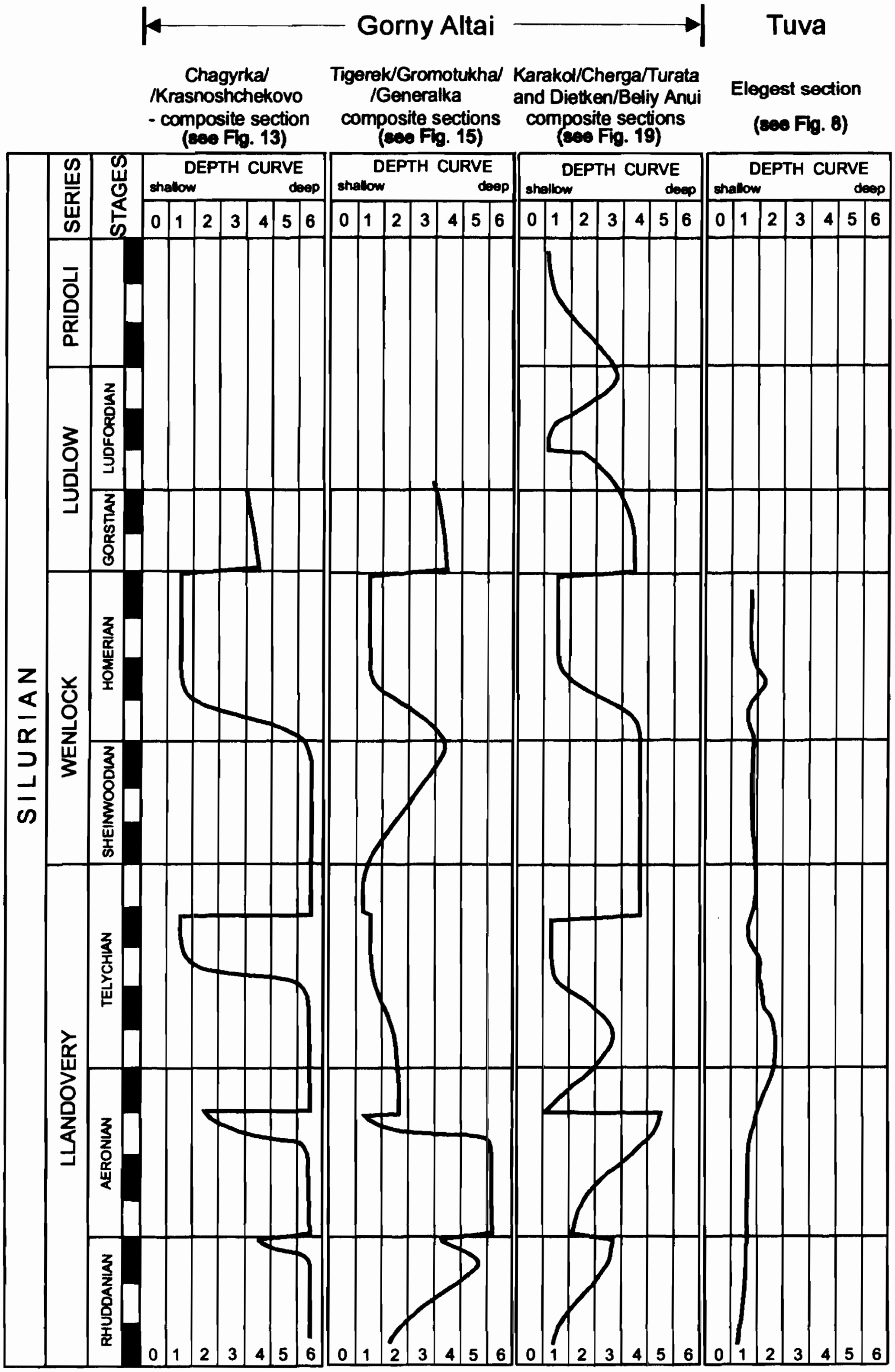

FIGURE 25 — Silurian depth curves for Altai and Tuva.

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Kazakhstan in the Silurian

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ABSTRACT — Siliciclastic and volcaniclastic Silurian rocks are widespread in central and eastern Kazakhstan. Tectono-stratigraphic zones (TSZ) are characterized by different compositions and thicknesses of sedimentary rocks, as well as by the presence or absence of volcanics formed in volcanic arcs and deposited in deep-water basins. The regional stratigraphic standard includes the Lower Silurian Alpeis, Donenzhal, and Bogut Stages and the Upper Silurian Akkan and Tokrau Stages. Brachiopods and tabulate and rugose corals are used for regional biostratigraphy, and correlation with the Silurian Global Standard is mainly based on graptolites. The lower three stages are richly fossiliferous, while the Akkan and Tokrau Stages are mostly barren, red siliciclastics and pyroclastics, with marine, siliciclastic, graptolite-bearing deposits of limited distribution. The Late Ordovician tectonic evolution of Kazakhstan suggests large eastern and western terranes that collided by the beginning of the Silurian. This resulted in a shallow basin with thick monotonous siliciclastic deposits across the unified plate. A distinctive transgressive event is middle Aeronian. The Agadyr oceanic basin with cherty and volcanic rocks separated the southern segments of the collided microplates and persisted from the Ordovician. The Dzhungaro–Balkhash and Irtysh–Zysan volcanic arc basins were separated by the Chingiz–Tarbagataj volcanic-arc system, which persisted until the end of the Wenlock. A volcanic belt in the Mointy–South Dzhungaria TSZ lasted from the Ordovician until the end of the Silurian and had maximum volcanism in the Wenlock. Benthic faunal affinities and paleomagnetic data suggest that Kazakhstan was located between the Baltic and South China Plates in equatorial paleolatitudes (6–12°or –19° N). These paleogeographic data are consistent with known spatial relationships of the TSZ, especially if a subsequent 180° rotation of the Kazakhstan Plate is considered. Faunal affinities with South China are strongest during the Ashgillian and Llandovery, but rapidly decrease by the end of the Silurian. Faunal affinities with the Siberian Plate are weak.

Introduction

Kazakhstan lies in the center of the Central Asian fold belt, and is surrounded by the Uralian, South Tien Shan, and Irtysh–Zajsan orogenic belts (Fig. 1A). The Silurian is widespread in central and eastern Kazakhstan, but absent in the west (Fig. 1B). It comprises thick, monotonous, predominantly siliciclastic and volcaniclastic sequences without significant metamorphism. The Lower Silurian is more lithologically complex, fossiliferous, and widely distributed than the Upper Silurian. The latter is composed mostly of unfossiliferous red siliciclastics and pyroclastics, but marine siliciclastics have limited geographic distribution (Nikitin and Bandaletov, 1986; Beliaev et al., 1989).

The Ordovician–Silurian boundary in Kazakhstan lies at the base of the *Akidograptus ascensus–Parakidograptus acuminatus* Zone and coincides with distinctive changes in tectonic evolution. The Upper Ordovician includes carbonates, volcanics, and volcano-sedimentary rocks in nappes and slump structures (olistostromes) (Nikitin, 1972; Nikitin et al., 1991). In contrast, Silurian rocks are moderately folded and faulted but not typically part of nappes, although they are included in nappes in the Agadyr and western Balkhash areas.

Several Silurian tectonostratigraphic zones (TSZ) with generally monotonous lithofacies are recognized. These include the Selety–Chu-Ili, Mointy–South Dzhungaria, Chingiz–Tarbagataj, Predchingiz–North Karaganda, Dzhungaro–Balkhash, and Agadyr TSZs (Bandaletov, 1969, Nikitin, 1991; Fig. 1B). These zones had contrasting tectonic settings, as indicated by the composition and thickness of the sedimentary rocks and presence or absence of volcanics. They include a variety of different shelf environments with predominantly siliciclastic successions. Volcanic and cherty-basaltic rocks characteristic of certain TSZs were deposited in volcanic-arc and deep-water basinal environments, respectively.

In this report, attention is focused on the Silurian stratigraphy, litho- and biofacies, and paleogeography of

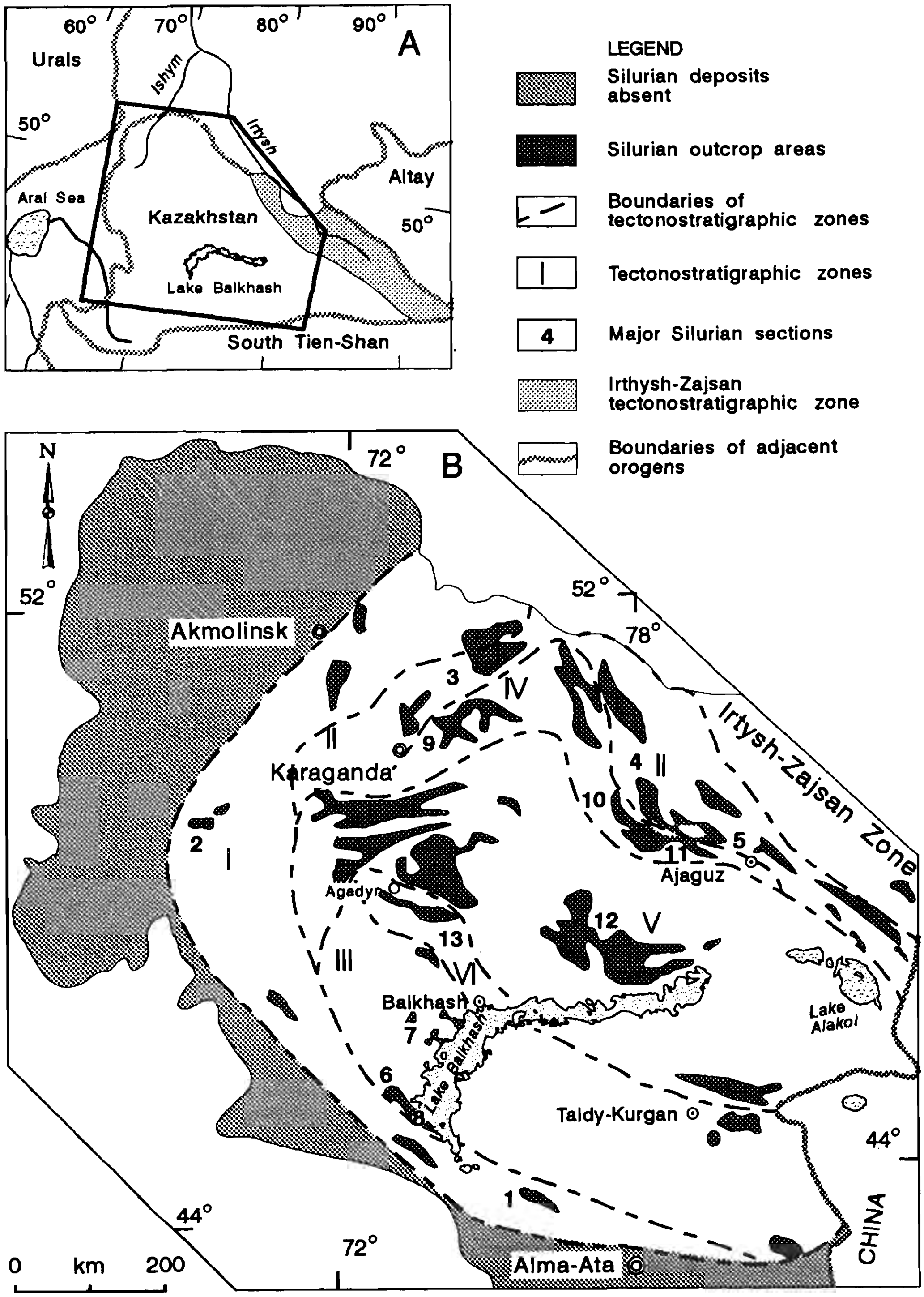

FIGURE 1 — A. Schematic map of Kazakhstan showing boundaries of orogens; B. Enlarged map of Kazakhstan showing distribution of Silurian, tectonostratigraphic zones, and location of key sections. Tectonostratigraphic zones in Roman numbers: I, Selety–Chu-Ili; II, Chingiz–Tarbagataj; III, Mointy–South Djungaria; IV, Prtedchingiz–South Karaganda; V, Djungaro–Balkhash; VI, Agadyr. Arabic numbers mark key sections: 1, Chu-Ili Mountains; 2, Sarysu–Teniz; 3, Karaaigyr Mountains; 4, Chingiz Range; 5, Ajaguz River; 6, Sholpan Valley; 7, Navaly Valley; 8, Mynaral village, 9, Akzhar–Zhartas; 10, Otyzbes; 11, Algabaz River; 12, Kok–Baital Mountains; 13, Agadyr.

Kazakhstan. An overview of the tectonic evolution of the Kazkhstan Plate during the Silurian is attempted.

Regional Stratigraphy

Historical background — Mapping and stratigraphic investigations during the 1950s (Borissiak, 1955, 1955a; Keller, 1958a, 1958b; Keller et al., 1958; Kovalevsky, 1959; Borissiak et al., 1961) led to the first correlation chart of the Kazakhstan Silurian. Three regional subdivisions, the Alpeis and Zhumak Stages of the Llandovery and the Akkan Stage of the Ludlow, were proposed on the basis of benthic faunas. These faunas were considered highly endemic and not appropriate for precise correlations with the British Silurian Series.

By the late 1950s, few graptolites were known in the region, and systematic sampling was followed by taxonomic and biostratigraphic studies only in the middle 1960s and early 1970s. Improved graptolite data (Obut and Sobolevskaya, 1965; Mikhajlova, 1971) were of utmost importance for correlation of the Silurian lithostratigraphy of central Kazakhstan. This interval also saw extensive study of benthic groups from many regions (Rukavishnikova, 1961; Borissiak, 1964, 1965; Kovalevsky, 1965; Senkevich, 1968; Stukalina, 1971). The biostratigraphic data were summarized by Bandaletov (1969) in his monograph on the Kazakhstan Silurian; his overview of the tectonic and paleogeographic setting followed Kassin's (1947) ideas. A revised stratigraphy was the basis for the second correlation chart of the Kazakhstan Silurian (Schlygin, 1976). This included an improved definition of the regional stages and a correlation with the British Silurian subdivisions.

Since the middle 1970s, regional studies concentrated on the definition of the Silurian's boundaries and on revision of upper Llandovery and Wenlock stratigraphy in Kazakhstan (Sapelnikov and Rukavishnikova, 1975; Apollonov et al., 1980, 1988; Sultanbekova, 1986; Stukalina, 1986; Paletz, 1987, 1990; Koren' and Popov, 1991; Nikitin, 1991). Studies of Upper Silurian graptolites and benthic faunas (Menner, 1975; Mikhaylova, 1976; Bandaletov, 1979; Olenicheva, 1983; Koren', 1983, 1989; Ushatinskaya, 1983; Nikitin and Bandaletov, 1986) led to precise boundary definitions for the Akkan Stage and established the Tokrau Stage as the uppermost Silurian of Kazakhstan.

The Alpeis, Donenzhal, and Bogut Stages comprise the Lower Silurian. The Upper Silurian is divided into the Akkan and Tokrau Stages (Fig. 2).

Lower Silurian — The lower boundary of the Silurian is the base of the *Akidograptus ascensus–Parakidograptus acuminatus* Zone. This horizon has been located in strata overlying the *Persculptograptus persculptus* Zone, where graptolites occur with the *Dalmanitina–Hirnantia* fauna in several sections in the Selety–Chu-Ili TSZ (Fig. 3). The boundary is marked by the appearance of *Akidograptus ascensus* Davies. The systemic boundary is defined in shallow-shelf regions (Fig. 4, Chingiz Range) by the lowest appearance of the brachiopods *Eospirifer cinghizicus* Borissiak, *Stricklandia lens mullochensis* Reed, and such tabulate corals as *Mesofavosites fleximurinus* Sokolov and *Halysites nitidus* (Lambe).

Lower Silurian sedimentary and volcanic sequences are widely distributed in central Kazakhstan and assigned to the Alpeis, Donenzhal and Bogut Regional Stages (Fig. 2). They are characterized by benthic faunas, predominantly brachiopods and tabulates with rugose corals, crinoids, and rare trilobites. Correlation of the regional stages with the Silurian standard units is based mainly on graptolites (Fig. 2). These fossils are numerous in Alpeis, lower Donenzhal, and Tokrau siliciclastics, but occur only at a few marine levels in the higher Silurian. The uppermost Donenzhal Stage is tentatively correlated with the *Cyrtograptus centrifugus–Cyrtograptus murchisoni* Zone of the lowest Wenlock. No key graptolite taxa occur at this level, with the exception of the long-ranging *Monograptus priodon* Bronn, which occurs in the Telychian and lower Sheinwoodian.

An interval between the Donenzhal and Akkan Stages that is equivalent to most of the Wenlock is the Bogut Regional Stage. This is a poorly understood interval in central Kazakhstan with widespread, thick, terrestrial sedimentary rocks that are almost unfossiliferous and have extensive pyroclastics.

In the Donenzhal and Bogut Stages, non-marine, unfossiliferous, red siliciclastics and volcanic rocks appear along the periphery of the Predchingiz–North Karaganda and Dzhungaria–Balkhash TSZ (Fig. 1, IV and V). The inner part of both zones is characterized by persistent siliciclastic sedimentation (Kurkovskaya, 1985; Jakubchuk, 1989).

Upper Silurian — Upper Silurian marine carbonates and siliciclastics with benthic faunas and graptolites are restricted to the Mointy–South Dzhungaria and Dzhungaro–Balkhash TSZ (Fig. 1). Unfossiliferous, red, molasse-type siliciclastics that overlie the Lower Silurian in the Selety–Chu-Ili TSZ are tentatively assigned to the Upper Silurian.

The Upper Silurian in central Kazakhstan is subdivided into the Akkan and Tokrau Regional Stages. The former corresponds to most of the Ludlow Series; the latter is equivalent to the upper Ludfordian (*Monograptus formosus* Zone) and Pridoli Series.

Fossiliferous upper Ludlow and Pridoli are known in the Dzhungaria–Balkhash TSZ and sporadically in the

System	Series	Stage	Graptolite zone	Regional biostratigraphy			
				Stage	Graptolites	Brachiopods	Tabulate corals
Silurian	Pridoli		*bouceki - transgrediens*	Tokrau	*microdon* *perneri*	*Stegerhynchella - Tastaria* assemblage	*Axuolites borissiakae* assemblage
			branikensis - lochkovensis		*lochkovensis - beatus*		
			parultimus - ultimus				
	Ludlow	Ludfordian	*formosus*		*formosus/ bessobaensis*		
			bohemicus tenuis - kozlovskii	Akkan	*bohemicus tenuis-kozlovskii*		*Favosites stepanovi* assemblage
			leintwardinensis			*Kirkidium knighti vogulicum*	
		Gorstian	*scanicus*			*Brooksina striata*	
			nilssoni		*nilssoni/ colonus*	*Pentamerus oblongiformis*	
	Wenlock	Homerian	*ludensis*	Bogout	*ludensis - nassa*		
			praedeoubeli - deubeli				
			parvus - nassa				
			lundgreni		*flemingii*		
		Shein-woodian	*rigidus - perneri*				
			riccartonensis - belophorus		*riccartonensis*		
			centrifugus - murchisoni	Donenzhal			*Sapporipora tarbagataica* Beds
	Llandovery	Telychian	*lapworthi - insectus*		*spiralis - griestonensis*	*Nalivkinia* assemblage	*Propora obrutschewi* Beds
			spiralis interval zone				
			griestonensis - crenulata				
			turriculatus - crispus		*crispus - turriculatus*	*Pentamerus longiseptatus* Beds	
			guerichi				
		Aeronian	*sedgwickii*		*sedgwickii - convolutus*		
			convolutus	Alpeis		*Eospirifer cinghizicus* Beds	*Mesofavosites fleximurinus* Beds
			argenteus		*gregarius*		
			triangulatus - pectinatus				
		Rhuddanian	*cyphus*		*cyphus*	*Holorhynchus cinghizicus* Beds	
			vesiculosus		*vesiculosus*		
			acuminatus		*ascensus - acuminatus*		

FIGURE 2 — Regional subdivisions of Kazakhstan Silurian and correlation with the standard units of the system.

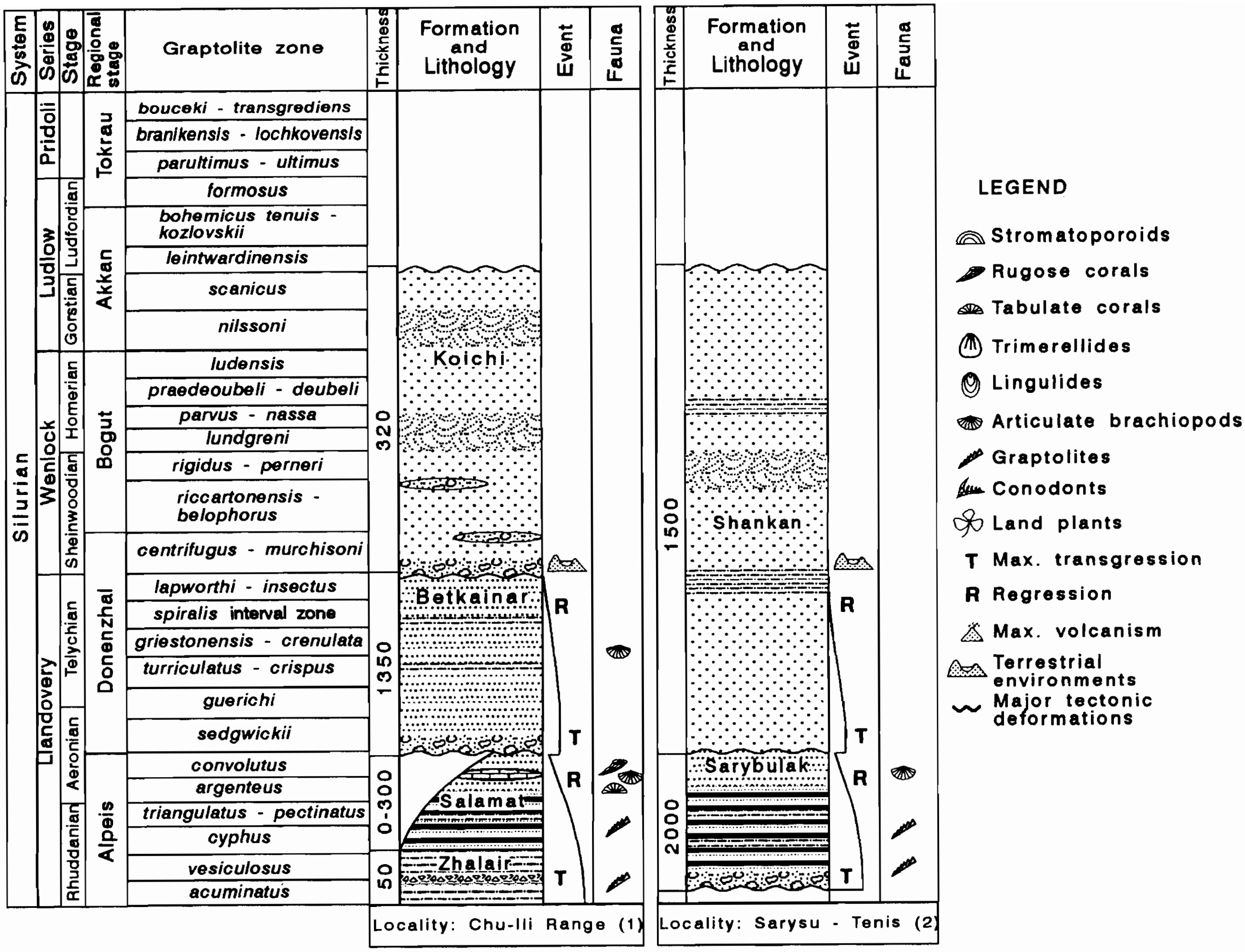

FIGURE 3 — Correlation of Silurian sections in the Selety–Chu-Ili tectonostratigraphic zones. The same legend is used for the faunas, land plants, and events in Figs. 4–7. Locality numbers correspond those of Fig. 1. Lithologic symbols are those of Compton (1962).

Mointy–South Dzhungaria TSZ (Figs. 1, 5). They comprise the Tokrau Stage, the uppermost regional unit in the Kazakhstan Silurian (Bandaletov, 1979; Nikitin and Bandaletov, 1986).

The base of the Devonian in Kazakhstan cannot be recognized by graptolites because of distinct facies changes close to the Tokrau–Ajnasu Stage boundary. It is conditionally defined 29–60 m above uppermost Tokrau rocks with the graptolites *Monograptus prognatus* Koren' and *M. microdon aksajensis* Koren', and at the base of sandstones and siltstones with a rich benthic assemblage. This benthic fauna is similar to that from the Ajnasu Regional Stage (Lower Devonian) in its type area in the southern part of the Karaganda region. The lowest type Ajnasa has numerous Lower Devonian brachiopods such as *Sieberella*, *Iridistrophia*, and *Eodevonaria* (Menner, 1975). This lowest Devonian brachiopod assemblage has numerous coarse-ribbed gypidulids, rather than the smooth and weakly ribbed Ludlow and Pridoli forms (Nikitin and Bandaletov, 1986). The associated crinoids include abundant scyphocrinitids, and indicate proximity to the systemic boundary (Stukalina *in* Menner, 1975).

The Selety–Chu-Ili TSZ (Fig. 1) is the only area in central Kazakhstan with a continuous record of marine-shelf deposition. In the middle Ashgillian (*Climacograptus supernus* Zone), a southern region in the Selety–Chu-Ili TSZ (Fig. 3, locality 1) illustrates black graptolitic shale deposiiton (Chokpar Formation). The Chokpar was suc-

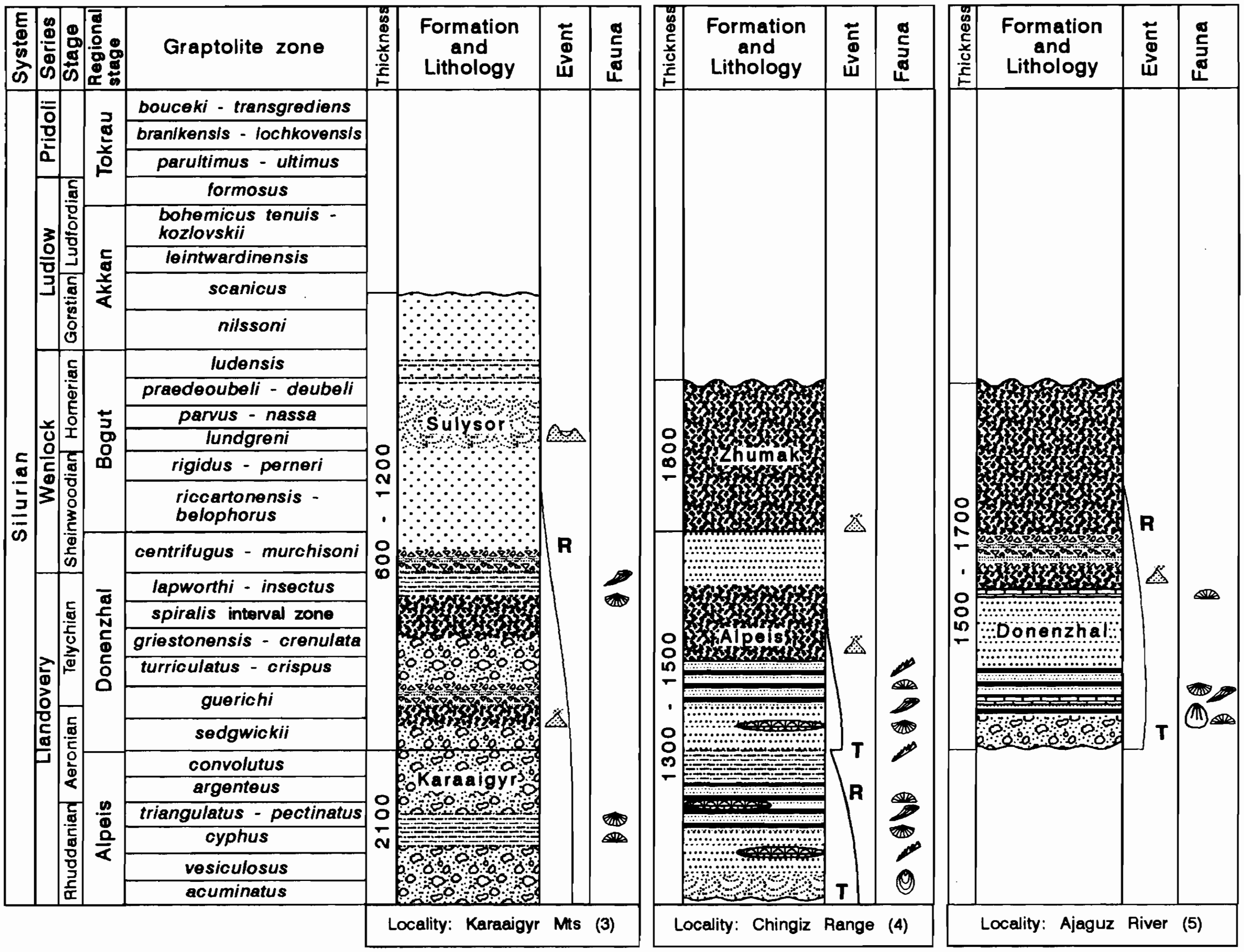

FIGURE 4 — Correlation of Silurian sections in Chingiz–Tarbagataj TSZ.

ceeded in the Hirnantian (*Normalograptus? extraordinarius* and *Persculptograptus persculptus* Zones) and Rhuddanian by graptolite-bearing rhythmites and locally turbidite sandstones up to 15–60 m thick (Zhalair Formation, *Parakidograptus acuminatus*–lower *Coronograptus cyphus* Zones) (Apollonov et al., 1980; Nikitin, 1991). Thin beds of acid tuff suggest nearby volcanism.

The upper *Coronograptus cyphus–Demirastrites convolutus* Zones (i.e., Salamat and upper Sarybulak Formations, Fig. 3) display depositional features characteristic of shelf and slope environments. The deposits are mostly graptolite-bearing sand and silt turbidites that are 300–2,000 m thick. These are replaced upward by lithologically varied siliciclastics with local conglomerates and lenses of limestone with benthic assemblages dominated by brachiopods with rugose and tabulate corals (Palets et al. *in* Abdulin, 1980). The Alpeis Stage in the Selety–Chu-Ili TSZ was terminated by a major sea-level fall and followed by a transgression.

The overlying shallow-marine deposits of the Donenzhal Stage (i.e., Betkajnar and Shankan Formations, Fig. 3) rest unconformably on various Silurian deposits (Bandaletov, 1957; Nikitin, 1991). These coarse- to fine-grained siliciclastics with lenses of shell hash limestone and polymict conglomerates are 300–1350 m thick. A low-diversity brachiopod assemblage with *Pentamerus* sp., *Tschatkalia nikiforovae* Rukavishnikova, *Whitfieldella* sp., and *Howellella* sp. occurs in the lowest Betkajnar Formation (Nikitin, 1991). There is a record of progressive shallowing in the basin, and by the end of the Llandovery

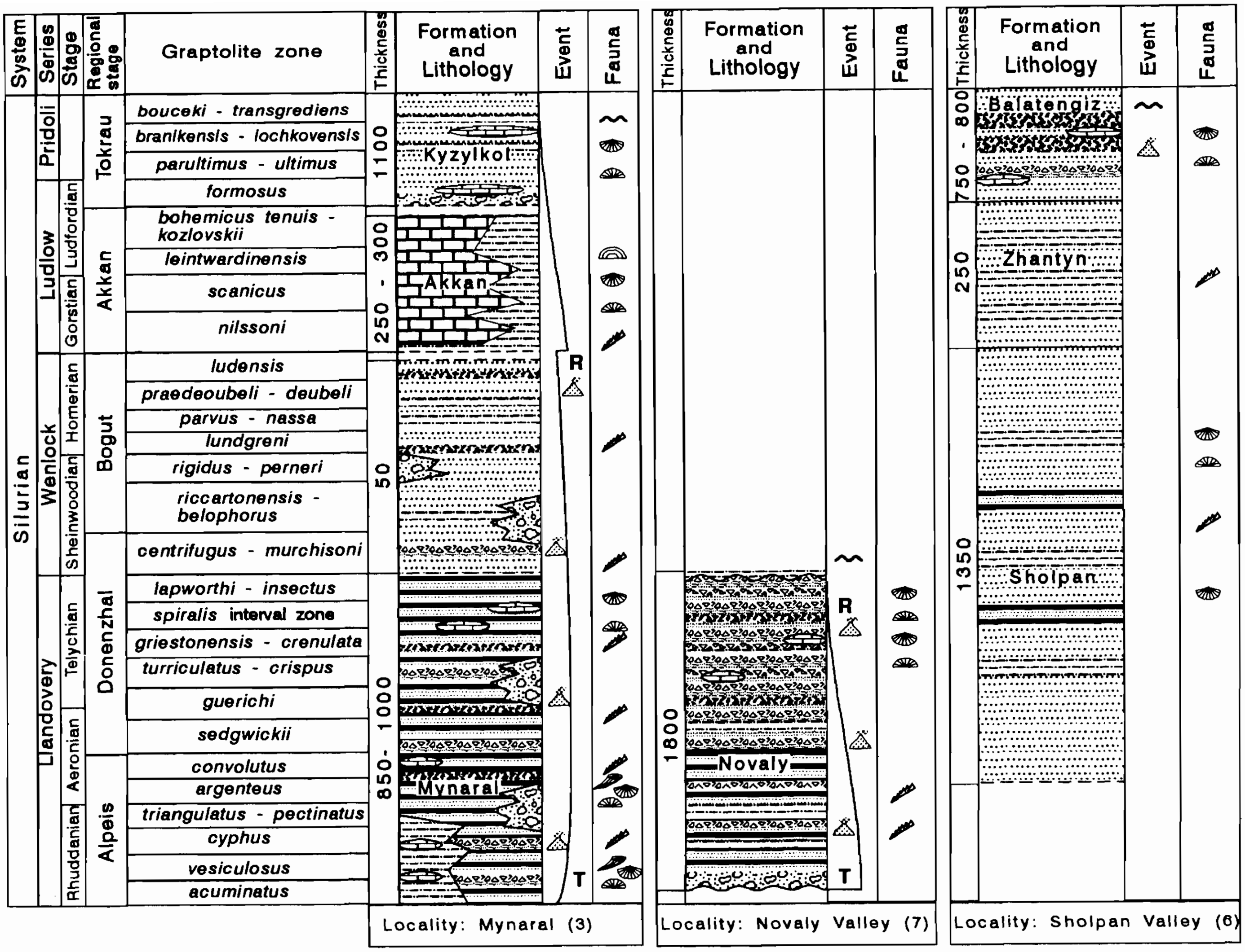

FIGURE 5 — Correlation of Silurian sections in Mointy–South Dzhungaria TSZ.

shallow-marine sediments were replaced over the area by red, cross-bedded sandstones, siltstones, and polymict conglomerates deposited in subaerial environments (i.e., Koichi and upper Shankan Formations, Fig. 3). They include fluvial and lake deposits with occasional rain imprints (Popov, field observation). Their Late Silurian age is based primarily on their position beneath fossiliferous Lower Devonian (Paletz et al. *in* Abdulin, 1980; Nikitin, 1991).

The Chingiz Range in the central part of the Chingiz–Tarbagataj TSZ (Fig. 1) has an uppermost Ordovician succession with interbedded volcaniclastics, pyroclastics, and carbonates that comprise the Akdombak Formation (Nikitin, 1972, 1991). This formation demonstrates the progressive shallowing of the basin. The Hirnantian age of the upper Akdombak Formation is confirmed by occurrence of the trilobite *Dalmanitina mucronata* (Brongniart). This trilobite appears above the highest occurrence of a *Holorhynchus giganteus* brachiopod assemblage and graptolites of the *Climacograptus supernus* Zone (Apollonov et al., 1980). Red and variegated siliciclastics with shallow-marine assemblages (lingulates, bivalves, gastropods) and polymict conglomerate occur near the Ordovician–Silurian boundary. Their correlation with the Hirnantian or lowest Rhuddanian is provisional because of the absence of age-diagnostic benthic fossils and graptolites (Bandaletov, 1969; Nikitin, 1972; L. E. Popov, unpublished data).

The Alpeis Formation (Alpeis and Donenzhal Stages) consists of up to 1.5 km of monotonous green and variegated siliciclastics with occasional conglomerates and limestone lenses (Fig. 4). Benthic assemblages in the lower Alpeis Formation suggest a transgressive trend interrupted by a regressive episode under the base of the Donenzhal Stage (*Pentamerus longiseptatus* Beds). Tidally influenced clastic deposits with lingulates at the base of the Alpeis Formation are replaced at its top by coarse- to fine-grained siliciclastics with coquina lenses. These lenses have a low-diversity brachiopod assemblage with *Holorhynchis cinghizicus* (Benthic Assemblage Zone 2–3) and a medium-diversity *Eospirifer cinghizicus* assemblage (Benthic Assemblage Zone 3–4). Associated tabulate corals (*Paleofavosites alveolaris* [Goldfuss], *P. maximus* [Tchernyshev], *P. groenlandicus tarbagataicus* Bondarenko, *P. poulseni* Teichert, *Mesophavosites fleximurinus* Sokolov, *Catenipora gotlandica* Yabe, *Halysites nitidus* [Lambe]) are diagnostic of the *M. fleximurinus* Beds (Koren' and Popov, 1991). In the same sections, a *Coronograptus gregarius* Zone assemblage occurs near the top of the *E. cinghizicus* Beds.

Unfossiliferous red and variegated siliciclastics, similar to those of the Hirnantian and possibly deposited within the shore-face zone, re-appear near the base of the Donenzhal Stage. The overlying *Pentamerus longiseptatis* Beds have the brachiopods *Stricklandia lens progressa* Williams, *P. longiseptatus* M. Borissiak, and *P. oblongus* (Sowerby), and graptolites of the *Spirograptus turriculatus-Globosograptus crispus* Zone.

The *Propora obrutschewi* Beds and *Sapporipora tarbagataica* Beds, with distinctive tabulates, brachiopods, and crinoids, appear in Donenzhal Stage carbonates and siliciclastics of the southern Chingiz–Tarbagataj TSZ (Figs 2–4, locality 5). The *Propora obrutschewi* Beds have *Antherolites septosus* Sokolov, *Multisolenia tortuosa* Fritz, *Catenipora panga* Klaamann, *Halysites interstinctus* (Linnaeus), and *P. obrutschewi* Kovalevsky. The *Sapporipora tarbagataica* Beds are characterized by *Favosites kennihoensis* Ozaki, *Halysites junior* Klaamann, and *S. tarbagataica* Barskaja. In the Tarbagataj Range, tabulate corals from the *Propora obrutschewi* Beds are associated with such Telychian graptolites as *Stimulograptus halli* (Barrande), *Monograptus rickardsi* Hutt, *Spirograptus turriculatus* (Barrande), and *Oktavites* sp. cf. *O. spiralis* (Geinitz).

The Alpeis Stage in the northern Chingiz–Tarbagataj TSZ (i.e., Karaaigyr Formation, Fig. 4, locality 3) is interpreted as a progradational siliciclastic sequence that is more than 2,000 m thick. It includes a lower turbidite fan complex that is replaced upward by shallow-marine calcareous sandstone and siltstone with the brachiopod *Eospirifer shidertensis* Borissjak.

Beginning in the late Llandovery–early Wenlock (Donenzhal Stage), volcanic and pyroclastic rocks become widespread in the Chingiz–Tarbagataj fold belt (Sulysor, Zhumak, and Donenzhal Formations; Fig. 4). Red lavas, lava breccias, dacite, andesite, and andesite–basalt tuffs are about 600–1,700 m thick. These rocks rest conformably on underlying Silurian marine deposits or unconformably on the Lower Palaeozoic. The volcanics are calc-alkaline. The siliciclastics include red to variegated sandstones and conglomerates with siltstone and limestone lenses with a benthic fauna (Velikovskaya et al., 1980; Degtjarev et al., 1994). Volcanic and siliciclastic rocks often replace each other laterally. The composition and thicknesses of the volcanics are variable. Paleontological data show that the start of volcanism was diachronous in the Donenzhal.

In the northern Chingiz–Tarbagataj TSZ, the red, predominantly non-marine siliciclastics presumably belong to the Bogut Stage and form the top of the Silurian sequence (e.g., upper Sulysor Formation. Fig. 4, locality 3). They consist of up to 1,000 m of red and variegated conglomerates, sandstones, and siltstones with acid tuffs and volcaniclastic intercalations. These volcaniclastics are above the fossiliferous Donenzhal deposits. A sparse benthic fauna with graptolites suggests a correlation with the Wenlock (Degtjarev and Rjazantzev, 1993).

The Lower Silurian (Alpeis, Donenzhal, and Bogut Stages) of the Mointy–South Dzhungaria TSZ has 1,000–1,500 m of volcanic and sedimentary rocks (e.g., Novaly Formation and Mynaral Group, Fig. 6, localities 8, 9). Depositional environments indicative of the inner and outer shelf, as well as turbidites and mass flow deposits, are known. The shelf-to-basin transition is imperfectly known due to strong tectonic deformation and absence of sedimentological study.

Silurian rocks usually unconformably overlie the Lower Palaeozoic (Abdulin, 1986; Paletz, 1990). The volcanics include flows, tuff, and volcaniclastics of rhyolite, dacite, and andesite composition. The volcaniclastic sandstones and siltstones occasionally yield graptolites of the *Coronograptus gregarius* to *Oktavites spiralis–Monoclimacis griestonensis* Zones. Beds, lenses, and olistoliths of shell hash limestone bear a rich benthic fauna (brachiopods, tabulate and rugose corals) at some levels.

The lower Rhuddanian is recognized with certainty only in the central Mointy–South Dzhungaria TSZ (Fig. 5, locality 6). This interval has a characteristic brachiopod assemblage with *Stricklandia lens mullochensis* Reed (Modzalevskaya and Popov, 1995).

During the Wenlock–Pridoli, continuous marine sedimentation persisted only in the central and southern Mointy–South Dzhungaria TSZ. Fossils in the Bogut Stage are very sparse. In some localities in the central part of the zone, such Homerian and Sheinwoodian graptolites as *Monograptus priodon* Bronn, *M. flemingii* Salter, and *Monoclimacis vomerina* Nicholson are known, and the upper Sholpan Formation (Fig. 5, locality 6) has the corals

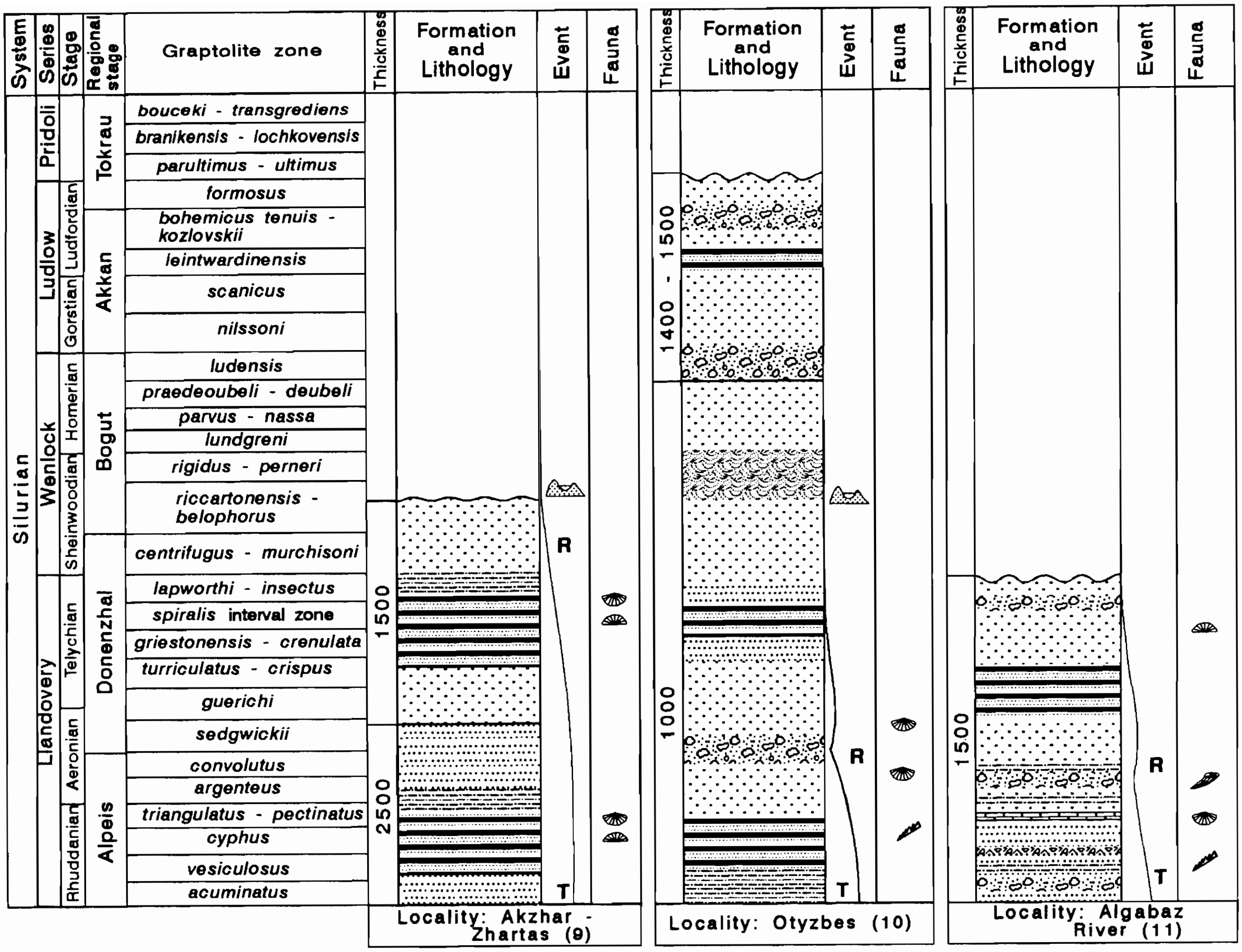

FIGURE 6 — Correlation of the Silurian sections in Predchingiz–North Karaganda TSZ.

Heliolites balkhashensis Kovalevsky and *Halysites* sp. and the brachiopods *Dicoelosia osloensis* Wright and *Eoplectodonta* cf. *duvalii* (Davidson), among others.

A narrow, predominantly carbonate belt is present in the central and southern part of the Mointy–South Dzhungaria TSZ and correlates with the Ludlow (Akkan Stage). The carbonate unit is up to 300 m thick (Akkan Limestone) and contains stromatoporoid-algal buildups. The Bogut may represent a barrier reef system developed along the accretionary wedge of an active volcanic arc (Keller, 1958a; Keller et al., 1958; Bandaletov, 1969). The Akkan Limestone may be traced southward in the subsurface, and crops out in outhern Dzhungaria near the Chinese border.

The lowermost Akkan Limestone is a thin siliciclastic with intercalations of limestone that contain such graptolites as *Gothograptus nassa* Holm, *Neodiversograptus* sp. cf. *N. nilssoni* (Lapworth), *Colonograptus colonus* (Barrande), *Plectograptus* sp. cf. *P. macilentus* (Tornquist), and *Saetograptus* sp. Several diagnostic brachiopod assemblages occur in the Akkan Limestone. They form the basis for three successive local biostratigraphic units: the *Pentamerus oblongiformis*, *Brooksina striata*, and *"Conchidium" knighti vogulicum* Zones (Figs. 2, 5). Each zone is characterized by pentamerid species diagnostic of the Elton, Bringewood, and probably Leintwardine Formations, respectively, of the Ludlow Series in Great Britain (Lawson and White, 1989). *Favosites stepanovi* Kovalevsky, *F. effusus* Klaamann, *Halysites opimus* Kovalevsky, and other tabulate corals are characteristic of the lower Akkan

Limestone (Kovalevsky, 1959, 1965). A rich crinoid fauna and trilobites such as *Staurocephalus murchisoni* Barrande and *Calymene blumenbachii asiatica* Weber occur as well (Nikitin, 1991). The uppermost Akkan Limestone in the type area has a fault contact with overlying red barren siltstones that are tentatively assigned to the Ludlow Series (Fig. 5, locality 8).

The red, green, and variegated unfossiliferous siliciclastic rocks of the Zhantyn Formation in the central part of the Mointy–South Dzhungaria TSZ are provisionally assigned to the Akkan Stage (Fig. 5, locality 6). This assignment is based on their stratigraphic position above the upper Sholpan Formation (Wenlock) and below the Lower–Middle Devonian. An occurrence of such Gorstian graptolites as *Pseudomonoclimacis dalejensis* Boucek with *Pristiograptus dubius* Suess was reported from an isolated locality that is provisionally referred to the Zhantyn Formation. By the end of the Silurian, marine sedimentation within the Mointy–South Dzhungaria TSZ continued only in south Dzungaria.

Variegated siliciclastics, 1.0–1.5 km in thickness, form the Lower Silurian (Alpeis and Donenzhal Stages) of the Predchingiz–North Karaganda TSZ (Fig. 6). They are transitional into the green siliciclastics of the Dzhungaro–Balkhash TSZ. The sections consist of sandstones and siltstones with minor conglomerates and lenses of nodular limestones, the latter with rich benthic faunas. Rare graptolites occur sporadically at some levels, and a sequence that includes the *Akidograptus ascensus–Parakidograptus acuminatus*, *Cystograptus vesiculosus*, *Coronograptus cyphus*, and *Coronograptus gregarius* Zones can be recognized. Pebbles of intermediate volcanics, compositionally similar to those known in the Chingiz–Tarbagataj TSZ, occur in these clastic sections. The southeastern sections (Fig. 6, locality 11) also contain the *Holorhynchus cinghizicus* brachiopod assemblage, in association with corals characteristic of the *Mesofavosites fleximurinus* Beds of the Alpeis Stage (Rhuddanian–Aeronian). To the west of this zone, brachiopod assemblages with *Pentamerus longiseptatus* and *Nalivkinia* and characteristic of the Donenzhal Stage (Aeronian–lower Sheinwoodian) are reported (Bandaletov, 1969; Nikitin, 1991).

The Upper Silurian in the Predchingiz–North Karaganda TSZ (Fig. 6, locality 10) is known from unfossilliferous red and variegated polymict cross-bedded sandstones and siltstones with lenses of conglomerates. It is up to 1.5 km thick, and presumably formed under terrestrial conditions (Bandaletov, 1969).

Predominantly fine-grained sandstones and siltstones, up to 1.5 km thick with graptolites and local occurrences of benthic fauna, are characteristic of the Alpeis, Donenzhal, and Bogut Stages in the Dzhungaria–Balkhash TSZ (Fig.7, locality 12). Sequences of two somewhat different Upper Silurian (Akkan and Tokrau Stage) facies may be recognized in the Dzhungaria–Balkhash TSZ. In the outer part of this zone south of the city of Karaganda (Nura–Ajnasu gegion), the Isen' Formation (Akkan and Tokrau Stages) includes up to 1.8 km of variegated sandstones, siltstones, and lenses of shell hash and argillaceous limestone with tabulates and brachiopods (Nikitin, 1991). This unit presumably formed under inner-shelf depositonal environments.

In the inner Dzhungaro–Balkhash TSZ (Fig. 7, locality 12), Akkan and lower Tokrau Stage deposits include a progradational sequence of siliciclastic and volcaniclastic turbidites with several units of tuffaceous sandstone, pebbly sandstone, and acid tuff. At several levels, siltstones yield numerous graptolites and vascular plants. Volcaniclastic and tuffaceous siltstone units dominate the upper part of this sequence; however, lenses and interbeds of shell hash limestone with tabulate corals, trilobites, brachiopods, and occasional crinoids. Coeval strata, about 800 m thick and characterized by benthic faunas in scattered carbonate lenses (upper Isen' Formation), are recognized south of Karaganda City (Nurin–Ajnasu region) in the Dzhungaria–Balkhash TSZ (Velikovskaya et al., 1980; Nikitin, 1991). Uppermost Silurian siliciclastic and volcanic rocks are known in south Dzhungaria to the south of the Balkhash region (Fig. 5, Sholpan Valley, Balatengiz Formation). Carbonate lenses here have Tokrau Stage tabulate corals and brachiopods such as *Favosites similis* Sokolov, *F. pactum* Chekhovich, and *Stegerhynchella angaciensis* Tchernyshev. The Tokrau deposits form a continuous sequence with the underlying Akkan Stage (Ludlow) and the overlying Ajnasu Stage (Lower Devonian). The stages are further subdivided mainly by graptolites, brachiopods, and tabulate corals (Koren', 1983, 1989; Nikitin and Bandaletov, 1986).

In the lower Tokrau Stage in the Kok-Bajtal Mountains (Fig. 7, locality 12), the uppermost Ludfordian graptolites *Monograptus* (*Formosograptus*) *formosus* Boucek and *Monograptus bessobaensis* Koren' are known. The base of the Tokrau Stage is defined at the boundary between beds with *Neocucullograptus kozlowskii* and M *(F.). formosus* zonal assemblages. The lowermost Tokrau has sparse, non-diagnostic pristiograptids and is conditionally correlated with the lowermost Pridoli (the *Neocolonograptus parultimus* to *Neocolonograptus branikensis* Zones). The middle and upper Tokrau is subdivided into successive graptolite assemblages characteristic of the following local zones (Fig. 2): 1) *Neocolonograptus lochkovensis* and *Monograptus beatus;* 2) *Monograptus bouceki* and *Pseudomonoclimacis bandaletovi;* 3) *Monograptus perneri kazakhstanensis;* and 4) *Monograptus microdon aksajensis.* These four units are correlated with the middle–upper Pridoli graptolite zonal sequence in the Barrandium

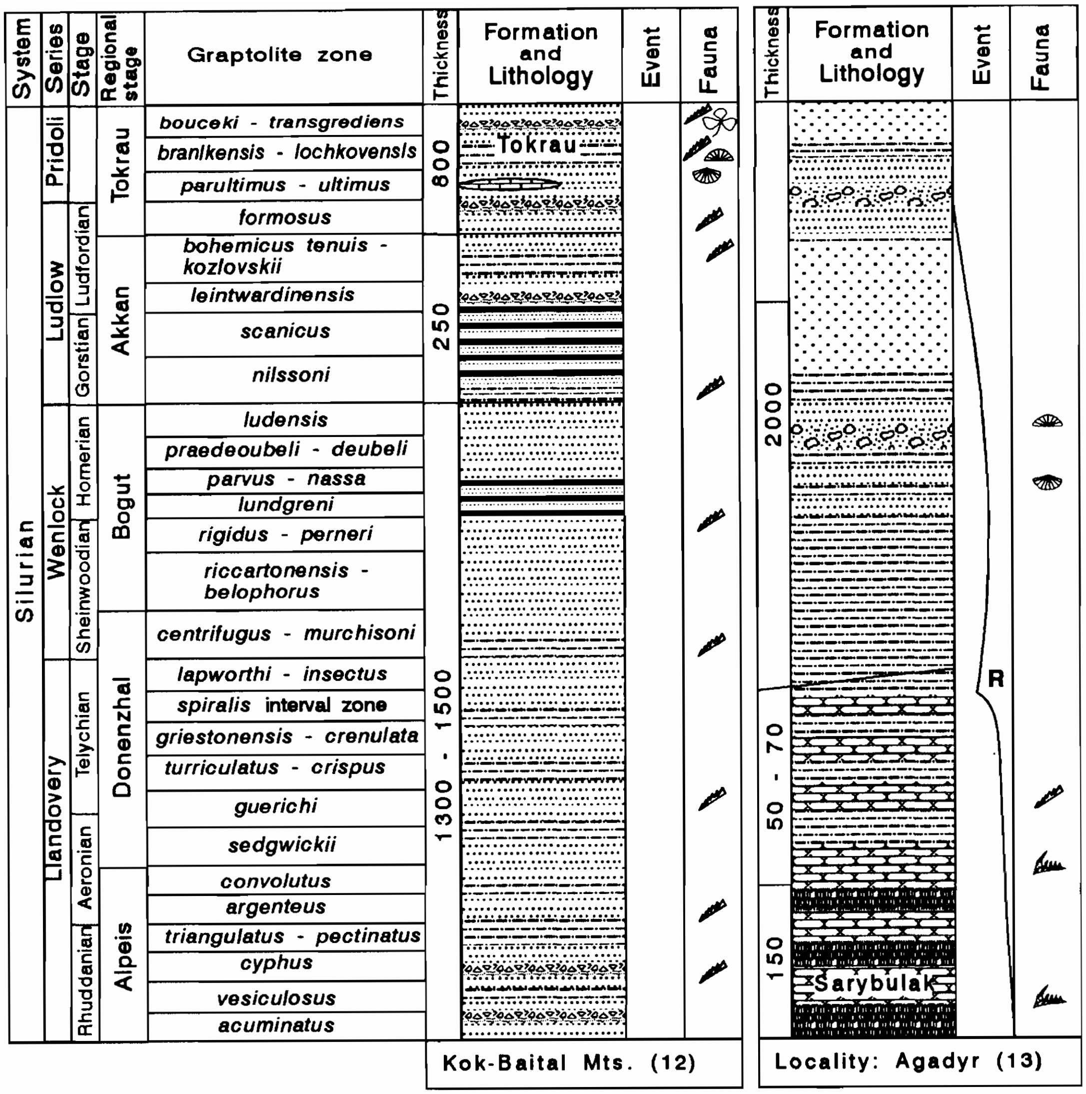

FIGURE 7 — Correlation of the Silurian sections in Dzhungaro–Balkhash TSZ.

(Pribyl, 1983).

The Tokrau Stage has a *Mesodouvillina-Tastaria* brachiopod assemblage, with *Mesodouvillina costatula* (Barrande), *Tastaria asiatica* Borissiak, *Gypidula optata* Barrande, *G. integra* Barrande, *Clorinda pseudolinguifera* Kozlowsky, and *Stegerhynchella angaciensis* Tchernyshev. The first species appears in the Lochkov of the Barrandium and Podolia, as well as the Isphara Regional Stage of Tien Shan (Nikitin and Bandaletov, eds., 1986). *Gypidula optata*, *G. integra*, and *C. pseudolinguifera* are present in the uppermost Silurian (Pridoli) along the eastern slope of the central and northern Urals (Sapelnikov, 1972). In general, the Tokrau brachiopod association shows a predominance of Silurian genera. This, if taken together with the absence of typical Ludlow pentamerids, suggests a Pridoli correlation of the Tokrau Stage.

Tokrau tabulate corals are closer taxonomically and in colonial structure to those of the Lower Devonian Ajnasu Stage than to the preceding Akkan Stage. Typical Devonian genera such as *Squameofavosites*, *Rhiphaeolites*, *Axuolites*, and massive *Favosites* colonies make their first appearance in the Tokrau. Among the tabulate corals, the following species are most characterisic: *Favosites favositiformis* (Holtedahl), *F. pactum* Chekhovich, *Squameofavosites uralensis* Yanet, S. incredibilis Chekhovich, and *Squameolites kirgizicum* (Chernova). They are also well known from the uppermost Silurian (Pridoli) in central Asia and the Urals. Some of them, such as *S. uralensis*

Yanet and *S. gurievskiensis* Mironova, are typical Lower Devonian taxa on the southern margin of the Siberian platform (Kuzbass).

Vascular plants, such as *Cooksonella sphaerica* Senkevich, *Taeniocrada* sp., and *Jugumella burubaensis* Senkevich, are present in the upper Tokrau (Menner, 1975; Nikitin and Bandaletov, 1986). However, they have not been systematically studied.

The Lower Silurian (Alpeis Stage) in the Agadyr TSZ comprises 150–200 m of radiolarian cherts and amygdaloidal and aphyric basalts and tuffs (Fig. 7), with rare graptolites and conodonts. The upper Llandovery–Wenlock (Donenzhal and Bogut Stages) in the Agadyr TSZ includes gray-green and purple tuffaceous siltstones, siliceous tuffs, and argillaceous cherts that are 50–100 m thick. The absence of non-volcanic siliciclastic material is characteristic. These rocks may represent a deep-water sediment apron deposited on cherty oceanic basalts erupted during Alpeis time.

The Upper Silurian is represented by up to 2 km of green, graptolite-bearing siltstone and sandstone with tuffaceous sandstone and tuff that rest conformably on a cherty, basaltic sequence of late Llandovery–Wenlock age. Calc-alkaline volcanics occasionally occur. They can be compared with the rocks formed in volcanic arcs.

PALEOENVIRONMENTS

Paleogeographic reconstructions show two distinct sedimentological and structural phases in the Early Silurian. These phases are early–middle Llandovery and late Llandovery–Wenlock.

Condensed Early Silurian (Llandovery) radiolarian cherts deposited on the oceanic crust in the Agadyr TSZ suggest a relic oceanic basin with abyssal depths. The basin originated as a result of the oblique collision of the east and west Kazakhstanian terranes (Fig. 8A). A connection of the basin to the open ocean was maintained only in the southeast (present coordinates), via northwestern Dzhungaria. To the east and north, the basin was surrounded by land massifs with wide siliciclastic shelves. Shallow-marine environments with peritidal and upper subtidal deposits predominated during the Llandovery in the outer part of the Chingiz–Tarbagataj TSZ. The inner North Karaganda–Predchingiz and Dzhungaria–Balkhash zones had turbiditic and mass flow deposition. These silicilastic sequences thin offshore from 2,000 m to 500 m, and fine-grained sandstones and siltstones with graptolites accumulated adjacent to the the Agadyr TSZ. To the southwest, the relic oceanic basin was bordered by an active volcanic arc (Mynaral–South Dzhungaria TSZ) and, possibly, a trench. Most of the Selety–Chu-Ili TSZ was below storm wave-base at the beginning of the Llandovery, and apparently was a back-arc basin that filled with sediment before the Wenlock.

The latest Aeronian was marked by a distinct paleogeographic change, as indicated by the onset of intense calc-alkalic volcanic activity in the Chingiz–Tarbagataj TSZ. This resulted in the Mointy–South Dzhungaria volcanic belt (Fig. 8B, III) and the Chingiz–Tarbagataj volcanic arc (Fig. 8B, IV) along the margin of the central Kazakhstan basin. An association of tuffs, pyroclastic rocks, carbonates, and shallow-water to subaerial siliciclastics formed along this margin.

At the end of the Aeronian, the size of the central Kazakhstan relic oceanic basin began to decrease, and the shoreline was displaced toward the center of the Predchingiz–North Karaganda TSZ (Fig. 8B, IV). Thick variegated siliciclastics with benthic faunas accumulated on a shallow shelf. Simultaneously in the Dzhungaria–Balkhash TSZ, green, fine-grained flysch with graptolites was deposited, possibly representing progradation of the siliciclastic shelf from the Chingiz–Tarbagataj island-arc basin (Fig. 8B, V). Presumably coeval cherty and basaltic rocks, not more than 50 m thick and without siliciclastic input from outside the basin, accumulated in the deep-water Agadyr basin (Fig. 8B, I).

The Wenlock–Ludlow boundary coincides with further paleogeographic changes. Volcanic activity stopped in regions bordering the Dghungaro–Balkhash zone in the northeast (Fig. 8 C), and marine sedimentation was replaced by thick, subaerial red beds. In the Mointy–South Dzhungaria TSZ, a narrow belt of carbonate deposits can be interpreted as Ludlow barrier reefs on the rim of the fore-arc basin. Marine sedimentation continued in the latter region during the Pridoli. The local Pridoli includes siliciclastic and carbonate deposits with a benthic fauna and intermediate calc-alkalic tuffs.

Another important change featured the infilling of the relic Agadyr deep-water basin with Upper Silurian siliciclastics. In the Chingiz–Tarbagataj TSZ, volcanic activity was interrupted and marine sedimentation was replaced by thick, unfossiliferous, coarse-grained red siliciclastics that suggest subaerial environments. The Dzhungaria–Balkhash TSZ has a shelf-to-basin transition predominantly in siliciclastics. (Fig. 8 C, V). A rapidly prograding clastic shelf resulted in areal reduction of the Late Silurian basin. Shallowing and progradation continued until the Middle Devonian, when subaerial sedimentation prevailed.

SILURIAN DEPOSITIONAL HISTORY OF CENTRAL KAZAKHSTAN — During the Early Silurian, most of central Kazakhstan was a shallow basin with siliciclastic shelf sedimentation. A relic oceanic basin with cherty-basaltic

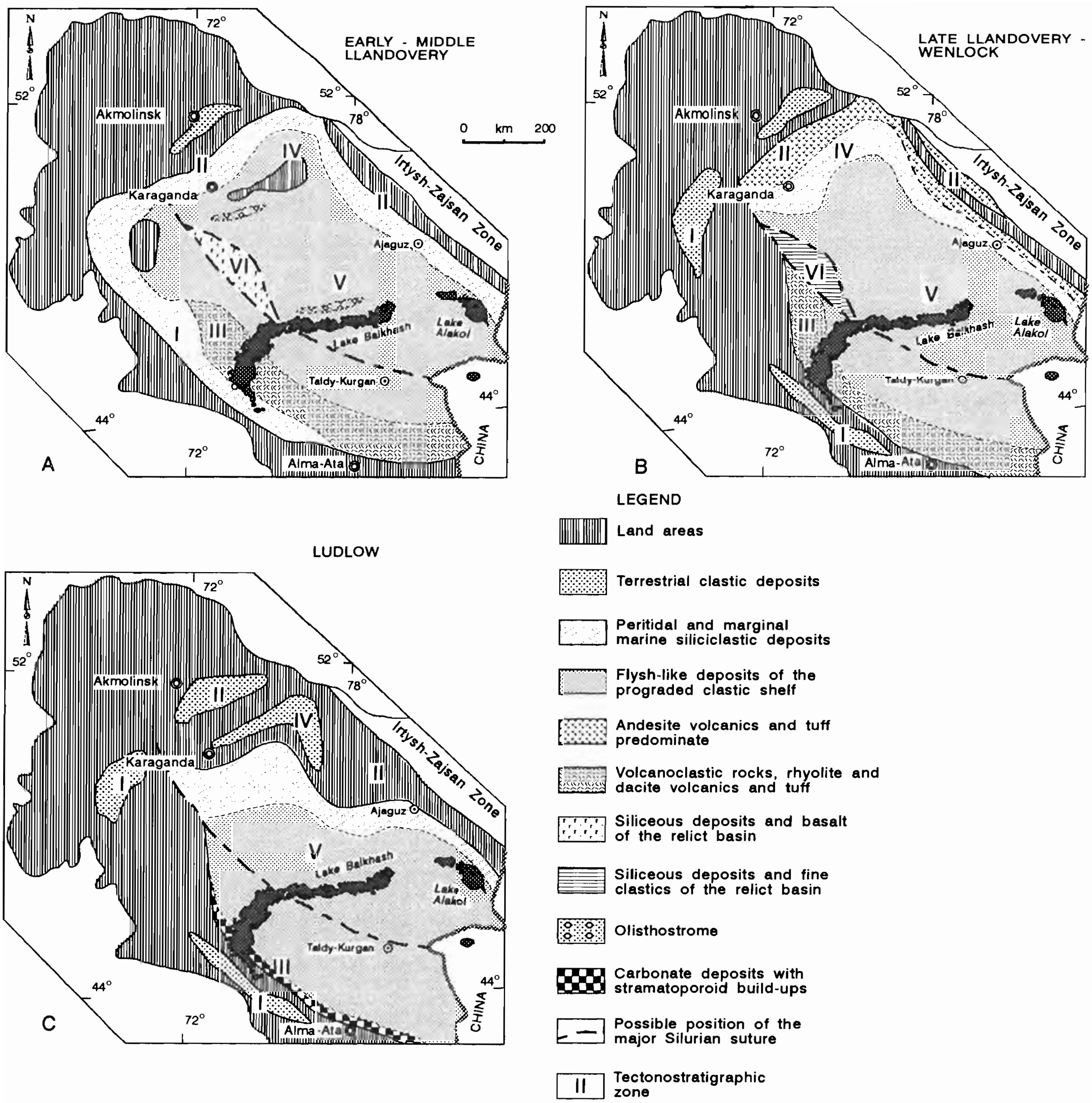

FIGURE 8 — A–C, Generalized paleogeographic reconstruction for Silurian intervals.

sedimentation was inherited from the Ordovician. Middle Aeronian (early Donenzhal) time coincided with a transgressive event that can be traced all over the territory. The lower Donenzhal, namely the *Pentamerus longiseptatus* Beds in shallow-marine areas, shows a distinct transgressive interval (Chu-Ili Mounbtains, Chingiz Range; Figs. 3, 6). The end of the Aeronian featured development of a volcanic arc along the periphery of the central Kazakhstan marginal sea. The early Late Silurian was marked by tectonic changes that lead to a general cessation of volcanic activity and marine sedimentation in the outer Chingiz–Tarbagataj, North Karaganda–Predchingiz, and Selety–Chu-Ili TSZs. Almost no changes occurred in the Dzhungaro–Balkhash basin in the middle part of central Kazakhstan (Fig. 9C). Here, continuous siliciclastic sequences were formed during the Silurian and Early Devonian. Thick (2.5–5.0 km), monotonous, graptolite- and plant-bearing sandstones, siltstones, and shales indicate steady subsidence and a stable water depth in the marginal basin during most of the Silurian, which makes the construction of a local sea-level curve difficult. However, starting with the end of the Aeronian and continuing to the end of the Silurian, a distinct regression, shallowing, and areal reduction of marine sedimentation is recorded.

Faunal Dynamics and Affinities

The published information does not represent the complete taxonomic diversity of the Silurian benthos in Kazakhstan. This makes detailed paleobiogeographic analysis difficult.

Tabulate corals — Late Ordovician tabulate corals, brachiopods, and trilobites display distinct provincialism. The Kazakhstan tabulate coral associations have strong similarities with the Eurasian rhelm faunas from southern Tien Shan and China (Webby, 1992). *Plasmoporella*, *Agetolites*, and *Amsassia* are abundant, but Late Ordovician tetradiids, calapoeciids, lichenariids, and cyrtophyllids, typical of the other regions, are not found in either region. Ashgillian tabulates from Kazakhstan show striking differences from those of the American–Siberian realm. In paleobiogeographic studies, the presence of *Sarcinul* in Kazakhstan and China is of utmost interest. Contrary to the Baltic, the range of *Sarcinul* is restricted to the middle Ashgillian.

At the end of the Ordovician, tabulate corals underwent an abrupt mass extinction recognizable at the generic and higher taxonomic levels. Extinct taxa include the Lichenariida, Sarcinulida, Tetradiida, Protaraeida, Agetolitidae, Fletcheriidae, Plasmoporellidae, Proheliolitidae, and Cyrtophyllidae. Tabulates are unknown in the Durben Regional Stage (Hirnantian) in Kazakhstan. However, the first cosmopolitan Silurian elements, such as *Paleofavosites*, *Mesofavosites*, *Heliolites*, and *Propora*, show a stepwise appearence prior to the extinction event. From their origin, they displayed a wide geographic distribution.

The Alpeis Stage (Rhuddanian–Aeronian) has numerous but taxonomically monotonous coral associations that consist of pandemic genera and widely distributed species. In addition to the above-mentioned genera, *Halysites* appeared for the first time.

During the Donenzhal Age, taxonomic diversification increased, and the first endemics appear in association with numerous pandemic taxa. Some of the endemics are known in only one or two regions outside the eastern part of central Kazakhstan. *Antherolites* (south Tien Shan and Gornyi Altaj), *Somphopora* (China), and *Helioplasmolites* (Wenlock and Ludlow of south Tien Shan, China, and the Urals) make their first appearance. *Propora obrutschewi*, a diagnostic Donenzhal species, was described as *P. conferta* Milne–Edwards & Haime from China (Lindström, 1899). *Sapporipora* is widely distributed in the Urals, Siberia, Tajmyr, northeast Russia, and Korea, but is unknown in south Tien Shan and China.

A majority of Akkan (Ludlow) tabulates are pandemics, although such species as *Squameofavosites tchernyshevi* Chekhovich and *Helioplasmolites nalivkini* Chekhovich have a restricted geographic range. *Squameofavosites tchernyshevi* could be the earliest representative of this genus. These tabulates show the closest affinities with tabulate associations from the Paadla Stage of Estonia and the Dalijan Stage of south Tien Shan. Some genera, such as *Thecia* and *Laceripora*, are very characteristic of the Wenlock and Ludlow on the East European Plate and Urals, but are absent in Kazakhstan.

Among the predominantly favositid associations of the Tokrau Regional Stage (Pridoli), many species are also recorded from the uppermost Silurian of the Urals, Tien Shan, and Altai–Sayan, as well as Canada, China, and Australia. However, there are no species shared with the Upper Silurian of the East European and Siberian Plates, including northeast Russia.

Brachiopods — Knowledge of the Silurian brachiopods of Kazakhstan is incomplete. Only Rhuddanian and Late Silurian brachiopods of the Dzhungaria–Balkhash TSZ are well known from the publications of Nikiforova (1937), Olenicheva (1983, 1990), Modzalevskaya and Popov (1995), and Rukavishnikova (1961). Silurian pentamerids were studied by Sapelnikov and Rukavishnikova (1975). Information on the lingulates, orthids, strophomenids, and atrypids is incomplete, and is dispersed in a number of short publications (Borissiak, 1972; Gorjansky, 1972; Misius and Ushatinskaya, 1977; Rukavishnikova, 1972a, 1972b, 1977; Ushatinskaya, 1977a, 1977b; Ushatinskaya and Alekseeva, 1983). The majority of taxa described by Borissiak (1955a, 1955b) are from the Lower Silurian of the Chingiz Range. As only part of the brachiopod fauna has been monographed and a zonation has never been proposed, only preliminary notes are possible.

The early–middle Ashgillian brachiopods and trilobites of Kazakhstan have many similarities with coeval South China faunas. This is shown by the distribution of diverse trimerellid assemblages (Percival, 1995; Holmer et al., 1996) and the *Pliomerina* trilobite fauna (Jaanusson, 1979). Affinities with Late Ordovician faunas of the Baltic Plate, including the north Urals, are evident, as well as affinities with benthic assemblages of the Hiberno–Salairian type (Jaanusson, 1979; Nikitin et al., 1996) and the *Holorhynchus* brachiopod assemblage (Rozman, 1967; Rong et al., 1989).

A significant part of the Late Ordovician benthos in Kazakhstan, including numerous genera and higher taxa of the rhynchonelliformean (articulate) brachiopods and trimerellids, became extinct in a stepwise fashion before the Hirnantian. In Kazakhstan, a diverse *Hirnantia* brachiopod fauna is recorded only in the Chu-Ili Mountains (Selety–Chu-Ili TSZ), where it is in association with a *Dalmanitina* trilobite assemblage. Both faunas show close affinities with those in Europe and South China (Apollonov et al., 1980).

The earliest Rhuddanian brachiopods in the Mointy–South Dzhungaria TSZ include *Dolerorthis sowerbiana* (Davidson), *Isorthis* (*Protocortezorthis*) *prima* Walmsley and Boucot, *Girardiella* sp., *Leangella scissa* (Davidson), *Eopholidostrophia sefinensis allisae* Hurst, *Eostropheodonta* sp., *Stricklandia lens mullochensis* Reed, *Eospirigerina porkuniana* (Jaanusson), *Zygospiraella duboisi* (Verneuil), *Meifodia tulkulensis* Modzalevskaya and Popov, and *Eospirifer* (Modzalevskaya and Popov, 1995). The occurrence of *S. lens mullochensis*, the earliest member of the *Stricklandia–Costistricklandia* lineage, correlates this brachiopod assemblage with the *Akidograptus ascensus-Parakidograptus acuminatus* Zone to the lower *Coronograptus cyphus* Zone (Baarli, 1987). Most likely, this assemblage replaced the *Hirnantia* fauna in the western Kazakhstan terrane. The relationships of the Late Ordovician and Early Silurian brachiopod faunas in Kazakhstan are somewhat uncertain. The majority of the Early Silurian genera are newcomers, with the exception of *Eostropheodonta* and such Lazarus genera as *Dolerorthis* and *Eospirigerina* that are known prior to Hirnantian time. This assemblage shows close affinities with coeval Baltic and British faunas. However, the association of *Stricklandia lens mullochensis* Reed and *Eospirifer cinghizicus* Borissiak is an important difference, as in the latter regions *Eospirifer* does not appear earlier than the late Llandovery. The succeding late Rhuddanian and Aeronian eospiriferids are widely known in Kazakhstan and South China.

Rhuddanian brachiopods from the Alpeis Formation in the eastern Kazakhstan terrane (the Chingiz–Tarbagataj TSZ; Fig. 4) are dominated by *Eospirifer chinghizicus* Borissiak and *Holorhynchus cinghizicus* Borissiak (Bandaletov, 1969). Although their precise correlation into the Lower Silurian graptolite zones remains somewhat uncertain, they form a distinct, recurring, low-diversity assemblage that is remarkably different from those mentioned above. In particular, *Holorhynchus cinghizicus* is absent in the western Kazakhstan terrane, and *Stricklandia* makes its first appearance with *Pentamerus longiseptatus* Borissiak not earlier than the Aeronian (in the upper Alpeis Formation in the Chingiz Range) in the eastern Kazakhstan terrane (Sapelnikov and Rukavishnikova, 1975). Other components of Rhuddanian faunas in these areas are little known. However, *Zygospiraella duboisi* (Verneuil) is also locally abundant.

The early Rhuddanian association in the Mointy–South Dzhungaria TSZ has close affinities with coeval brachiopods in Britain, Norway, and Estonia (Modzalevskaya and Popov, 1995). However, the co-occurrence of *S. lens mullochensis* Reed and *E. cinghizicus* Borissiak is an important difference from the Baltic and Britain, where *Eospirifer* appears no earlier than the late Llandovery (Cocks et al., 1984). The succeeding late Rhuddanian and Aeronian spiriferids are widely known from many regions of Kazakhstan and South China.

A distinctive character of the Rhuddanian and Aeronian brachiopods of the BA 3 Zone in Kazakhstan is the absence of *Borealis* and *Virgiana*, which are probably replaced by *Holorhynchus cinghizicus* Borissiak. *Borealis* is one of the most distinctive taxa of Lower Silurian BA 3 assemblages in Balto-Scandia (Rubel, 1970), and also is reported from South China (Wang et al., 1987). *Virgiana* occupies a similar position in Lower Silurian brachiopod biofacies of North America and Siberia. The Aeronian trimerellids in the Predchingiz–North Karaganda TSZ reappear from the Rhuddanian (*Dinobolus*) and range into the lower Sheinwoodian (*Trimerella*). They are likely immigrants from South China, rather than direct descendents of local Ordovician lineages.

Aeronian and Telychian brachiopods (the *Pentamerus longiseptatus* Beds) are widespread in central Kazakhstan. Outside the Selety–Chu-Ili and Mointy–South Dzhungaria TSZs, these brachiopods have a wide distribution within the Predchingiz–North Karaganda TSZ. The overlying *P. longiseptatus* Beds in the Chingiz–Tarbagataj TSZ are distinguished by the appearance of numerous pentamerids (e.g., *Clorinda, Antirhynchonella, Stricklandia, Pentamerus*). The brachiopod assemblages also include smooth and ribbed atrypids (*Atrypa, Nalivkinia, Zygospiraella, Atrypopsis*), as well as spiriferids (e.g., *Eospirifer and Striispirifer*). The athyridids are characteristic of the Mointy–South Dzhungaria and Dzhungaro–Balkhash

TSZs, where they are represented by *Tschatkalia*, *Whitfieldella*, and *Nucleospira*. By the early Sheinwoodian (upper Donenzhal, *P. obrutschewi* Beds), the *Nalivkinia* Community was part of BA 2–3 associations (Boucot, 1975; Wang et al., 1987), and flourished in the shallow-shelf biotopes with sandy bottoms. Nalivkiniids (*Pronalivkinia*) made their first appearance in the early Llandovery of the Chingiz–Tarbagataj TSZ and became a characteristic element of the late Aeronian–Telychian brachiopod faunas in central Kazakhstan, the Altai–Sayan, and the Siberian and Yangtze Plates. They are, however, unknown in Britain, Balto-Scandia, and Australia.

These brachiopod distributions and the comparative studies by Rong et al. (1995) of Llandovery brachiopods in several Asian regions suggest a closer proximity of Kazakhstan to the East European and Yangtze Plates than to the Siberian Plate in the Early Silurian. This interpretation is supported by the presence of numerous shared genera, and the absence of the *Tuvaella* and *Virgiana* faunas in Kazakhstan and South China (Rong et al., 1995).

Late Donenzhal time was typified by species brachiopod taxa known from Britain and Balto-Scandia. Among these are *Pentamerus*, *Stricklandia*, *Resserella*, *Dalejina*, *Dolerorthis*, *Eospirifer*, *Leangella*, *Atrypa*, and *Striispirifer*. The number of taxa with a restricted geographic distribution also increased. The Kazakhstan endemics include *Laevispirifer zhamankonensis* Ushatinskaya, *Kjaerina kazakhstanensis* Ushatinskaya, *Macropleura? otarensis* Rukavishnikova, *Stricklandiella aseptata* Sapelnikov and Rukavishnikova, *Marianella costatula* Sapelnikov and Rukavishnikova, and other species. Taxa such as *Atrypopsis asiatica* Menakova and *Isorthis planoconvexa* Kulkov are shared with coeval brachiopod associations from the south Tien Shan and Altai–Sayan (Beliaev et al., 1989; Paletz, 1990).

Brachiopod diversity decreased strongly by the late Wenlock. In the shallow-water biotopes of that time, brachiopod communities were replaced by tabulate coral and stromatoporoid associations (the *Propora obrutschevi* Beds). *Stricklandia*, *Pentamerus*, and *Clorinda* persisted until the early Sheinwoodian, and are reported from the uppermost Donenzhal of the Chingiz–Tarbagataj TSZ. Outside of this region, as in the Mointy–Dzhungaria Zone, brachiopods are extremely rare and are represented by sparse specimens of *Resserella*, *Dicoelosia*, *Dolerorthis*, *Eoplectodonta*, *Leangella*, *Isorthis*, and *Strophonella*.

Ludlow brachiopods, known only in the Akkan Limestone in the inner part of central Kazakhstan, are dominated by such pentamerids as *Anastrophia*, *Capelniella*, *Lissocoelina*, *Brooksina*, *Eokirkidium*, *Rhipidium*, *Vagranell*, and *Wyella*. Atrypids and spiriferids are also abundant, but orthids and rhynchonellids are less common. These assemblages inhabited shallow-shelf environments with algal-stromatoporoid reefs. Outside Kazakhstan, *"Conchidium" knighti vogulicum* (Verneuil) and *C. biloculare asiaticum* Sapelnikov and Rukavishnikova form a separate low-diversity benthic community (BA 3) characteristic of the upper Akkan Limestone. Outside of Kazakhstan, the *"Conchidium"* community formed extensive banks over a vast region from the north Urals to the southern Tien Shan. Some components of this pentamerid fauna are recorded from Gorny Altai, northeast Russia, Novaya Zemlya, Balto-Scandia, Britain, Nevada, and Australia.

The Pridoli (Tokrau) brachiopod fauna is less diverse and has a limited areal distribution in the Dzhungaria–Balkhash TSZ. Some pentamerids (e.g. *Gypidula*, *Clorinda*, *Anastrophia*) cross the Ludlow–Pridoli boundary, but are represented by different species, such as *Gypidula nux* Khodalevich and *G. kokbajtalensis* Olenicheva. Pridoli brachiopod assemblages usually include *Resserella*, *Mesodouvillina*, *Strophonella*, *Leptostrophia*, *Iridistrophia*, *Chonetes*, *Stegerhynchella*, *Tastaria*, *Lissatrypa*, *Eospirifer*, *Delthyris*, and *Howellella*. The Pridoli brachiopod fauna of Kazakhstan is characterized by numerous genera that make their first appearance in the Tokrau Stage. The majority of these newcomers are later diagnostic of Lower Devonian faunas in the Reinish–Bohemian Province of the Old World rhelm (Boucot, 1975). The Retziellinae, characteristic of Upper Silurian–Lower Devonian brachiopod faunas of China, Australia, and central Asia, are apparently absent in the Upper Silurian of Kazakhstan.

The Silurian brachiopod fauna of Kazakhstan has close affinities with coeval faunas of the East European and Yangtze Plates. The affinities with the Yangtze Plate were strongest in the Llandovery, as shown by the early appearance of spiriferids and Silurian trimmerellids in both areas and the distribution of *Nalivkinia* assemblages. Affinities between these regions decreased significantly before the end of the Silurian, and, in particular, the Retziella fauna is absent in the Upper Silurian of Kazakhstan.

GRAPTOLITES — The lower Llandovery *Akidograptus ascensus–Parakidograptus acuminatus* Zone graptolites have distinct affinities with Arctic Canadian and South American faunas (Guerda et al., 1988; Melchin, 1989). They differ from the south Tien Shan and south Urals faunas with diplograptids. The late Rhuddanian–Telychian graptolites feature cosmopolitan species. Wenlock graptolites are sparse in Kazakhstan, and Ludlow graptolites are recorded from few localities, although they are represented by pandemic taxa. The Ludfordian graptolites have European and Tien Shan affinities, and include a few pandemic species also known in Arctic Canada and Australia (Koren', 1983, 1989; Lenz, 1990). Pridoli graptolites from the Balkhash region, south-central Kazakhstan, are very diverse and are represented by numerous new taxa, some of which were recently discovered in northwest China and Australia.

Tectonics

The Late Ordovician tectonic evolution of Kazakhstan suggests two large terranes usually called the Eastern and Western blocks. These terranes have contrasting depositional and structural histories. The Western terrane includes west-central Kazakhstan and north Tien Shan. It is characterized by widely distributed Precambrian sialic massifs, which formed a unified block of continental crust by the end of the Ordovician. The Eastern terrane includes east-central Kazakhstan and western Dzhungaria. Only sporadic exposures of the Precambrian massifs are known. Towards the end of the Ordovician, complex nappes were formed in the Eastern terrane with the collision and accretion of volcanic arcs and closure of back-arc basins. These events were not accompanied by granite intrusions or by the formation of thick continental crust. Although the Late Ordovician Kazakhstan terranes were surrounded to the west and south (in modern coordinates) by the Uralian and Turkestan Oceans, a passive margin can be reconstructed only along the Turkestan Ocean (Kurenkov and Aristov, 1995).

By the beginning of the Silurian, an oblique collision between the Western and Eastern terranes occurred, and a new basin developed across a unified plate with a heterogenous basement. During the Rhuddanian–Aeronian, thick monotonous siliciclastic sediments accumulated.

The Agadyr relic oceanic basin separated the southern segments of the collided microplates. The accumulation of cherty and basaltic rocks continued uninterrupted from the Ordovician. The presence of a dismembered, Early Silurian ophiolite association (oceanic basalts and radiolarian cherts) in the Agadyr TSZ suggests abyssal depths, a deficiency of siliciclastic sediment input in the central part of the basin, and existence of a spreading zone that remained active until the end of the Llandovery. The Agadyr TSZ is separated by large-scale faults from volcaniclastics and carbonates in the Mointy–South Dzhungaria volcanic belt. This suggests the possible northwest displacement of the Agadyr complex by a distance of at least several hundred kilometers. Fragments of comparable sequences with Early Silurian radiolarians occur in the serpentinite melanges in western Dzhungaria (Zhang Chi et al., 1993). The Dzhungaro–Balkhash volcanic-arc basin widened to the southeast, where it was connected to the Irtysh–Zajsan Basin. These two basins were divided by the the Chingiz–Tarbagataj volcanic-arc system, which may represent a remnant Ordovician island arc in the Llandovery. Volcanism was reactivated at the end of Aeronian and continued through the Wenlock.

The Mointy–South Dzhungaria volcanic belt originated before the Silurian. Middle Ordovician island arc volcanism is recorded within a belt extending into the Sarytuma zone, about 100 km to the northeast. During the Ashgillian and possibly Early Silurian, the northeast margin of the Western Kazakhstan terrane was re-organized. The center of volcanic activity and zone of subduction migrated as a result of the accretion of small fragments of continental crust or a relict arc. Fine-grained pyroclastic material in the Hirnantian and Rhuddanian (the Zhalair Formation, Fig. 3) of the Chu-Ili Range suggests that volcanic activity was reduced, but did not completely cease during this interval. The active plate margin to the northeast could have been the source of the pyroclastics. Maximum volcanic activity occurred in the Wenlock, although volcanic and pyroclastic rocks of Pridoli age suggest that the Mointy–South Dzhungaria volcanic belt remained active until the end of the Silurian.

Several palinspastic reconstructions showing the position of the Kazakhstan Plate in the Paleozoic have been attempted (Zonenshein et al., 1987; Scotese and McKerrow, 1990). However, paleomagnetic data from Kazakhstan have been used more recently (Didenko et al., 1994). Paleontological data were not considered previously for two reasons. First, these data are dispersed in numerous papers, published mostly in Russian. Secondly, modern systematic revisions of the benthic faunas have not been done.

Paleomagnetic data for Silurian siliciclastics in the northern Dzhungaria–Balkhash and Predchingiz–North Karaganda TSZs has become available. For both areas, they indicate a paleogeographic position limited to 6–12° N (Didenko et al., 1994). Paleomagnetic data for the cherty and basaltic rocks of the Agadyr TSZ indicate 19° N (Pechersky and Didenko, 1995). These data are only consistent with the Silurian tectonostratigraphic zones and the modern geological structure of central Kazkhstan if a subsequent 180° rotation of the Kazakhstan Plate is considered. At the same time, they support the subsequent displacement of the Agadyr cherty and basaltic rocks by ca. 300–400 km.

Didenko et al. (1994) proposed that during the Silurian, the Kazakhstan Plate was situated with the Salair and Altai fold belts of the Siberian Plate on one side and with eastern Gondwana, including the South China Plate, on the other. The Zajsan and Turkestan Oceans separated Kazakhstan from those large continents. The location of these oceans is consistent with earlier paleogeographic maps (e.g., Scotese and McKerrow, 1990). However, these maps show the Kazakhstan Plate close to the East European and Siberian paleocontinents.

Biogeographic data outlined above support the spatial relationships between all of these Silurian continents and their location in equatorial latitudes. The available data suggest close faunal affinities between Kazakhstan and South China during the Ashgillian–Llandovery, with a significant decrease towards the end of the Silurian. Faunal similarities between Kazakhstan and Siberia are relatively weak, which suggests that wide oceans limited benthic faunal exchange (Fig. 9).

Discussion

We follow a traditional approach to Silurian paleogeographic and tectonic reconstruction of Kazakhstan (e.g., Bandaletov, 1969; Ruzhentsev and Samygin, 1969; Zaitsev and Velikovskaya, 1980; Mossakovsky et al., 1993). Some modifications are introduced on the basis of new sedimentological, paleomagnetic, and biogeographic data from Kazakhstan and adjacent areas.

A tectonic model of the Paleozoic Altai has been proposed (Sengor et al., 1993) and discussed in Russian-language reports (e.g., Ruzhentzev and Massakovsky, 1995). According to Sengor et al., development of the Urals, Tien Shan, Kazakhstan and other Altai Paleozoic structures took place along the Kipchak migratory island-arc system.

This model cannot be fully evaluated herein as it requires a general survey of Kazakhstan's tectonics and stratigraphy through the Cambrian–Devonian. The Si-

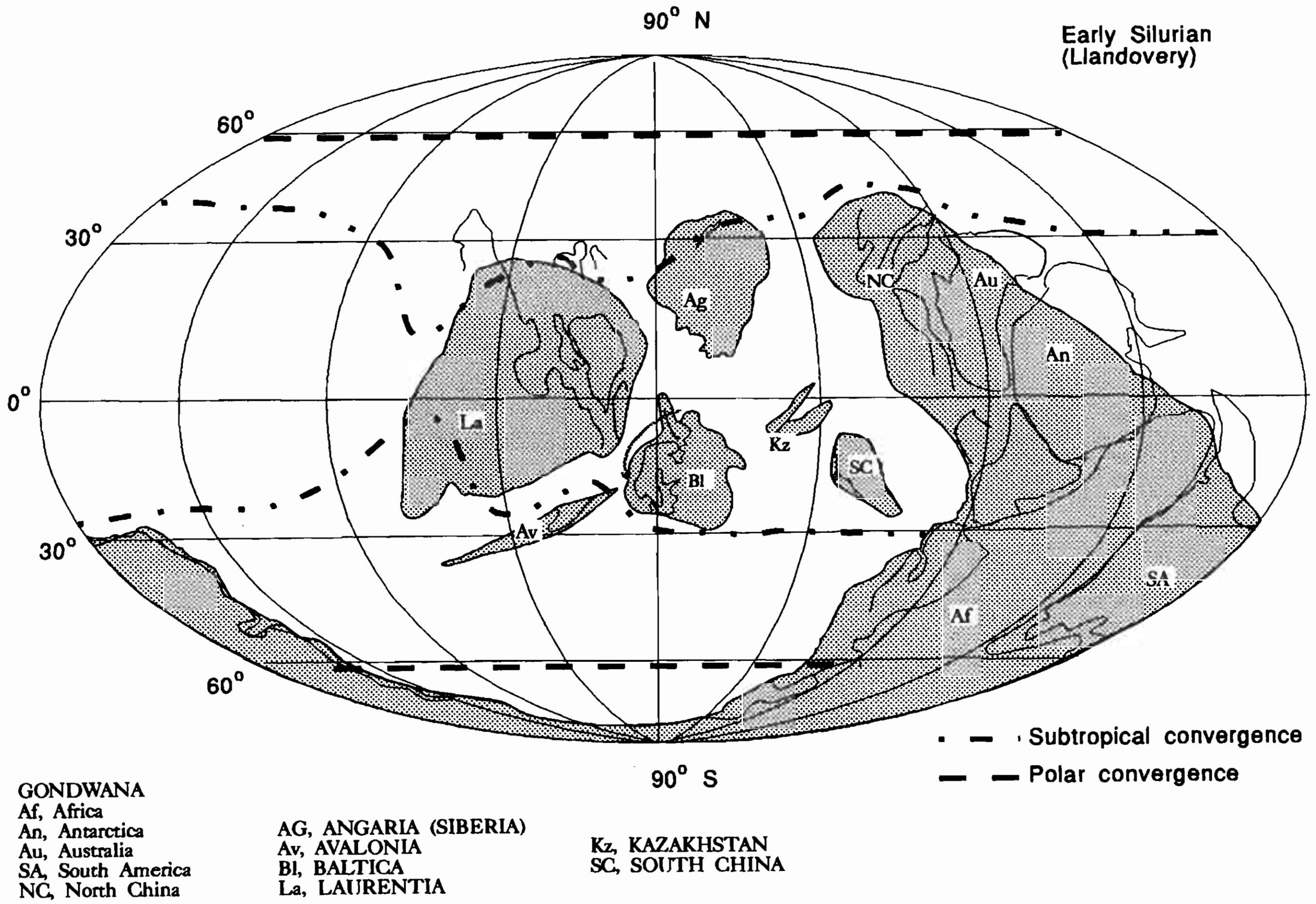

FIGURE 9 — Early Silurian position of Kazakhstan (after E. Popov, L. E. Holmer, and M. E. Bassett, unpublished data). Position of the major continents, with reinterpretation of location of Kazakhstan and South China, after Scotese and McKerrow (1990).

lurian is too short for a comprehensive comparison of the new and traditional paleotectonic reconstructions.

CONCLUSIONS

The Western and Eastern Kazakhstan terranes were not parts of a single island arc, and the polarity of the subduction zone on the northwest (recent coordinates) during the Ordovician–Silurian was opposite to that suggested by Sengor et al. (1993). The major features of the Silurian tectonic development of Kazakhstan resulted from oblique collision of the western crustal terrane with an island arc at the beginning of the period. The relic oceanic basin in south-central Kazakhstan was completely subducted before the end of the Silurian. Biogeographic affinities of the Silurian benthic faunas of the Kazakhstan paleocontinent show its location in low latitudes between the East European (Baltica) and South China Plates, and a considerable separation from the Siberian (Angaran) continent.

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STRATIGRAPHY AND PALEOGEOGRAPHY OF THE SILURIAN OF EAST SIBERIA

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ABSTRACT — A complete review of the Silurian stratigraphy and paleogeography of East Siberia (Siberian platform, Taymyr) is presented. In this region, the Silurian occurs in five subregions (North Taymyr, North Priyenisey, Pritunguska, Nyuya-Berezovo, and Irkutsk) and fifteen districts (Middendorf, Norilsk, Turukhansk, Igarka, Kochumdek, Vorogovo, South Taymyr, Ledyanka, Maymecha, Moyero, Morkoka, Vilyuy, Nyuya-Berezovo, Ilimsk, and Balturino). Regional chronostratigraphic units termed "horizons" (regional stages) include the Moyerocanian (Rhuddanian-lower Aeronian), Khaastyrian (middle-upper Aeronian), Agidyian (Telychian), Khakomian (Wenlock), Tukalian (Gorstian), and Postnichian (Ludfordian-Pridoli). These divisions include thirteen subhorizons and 54 regional biostratigraphic zones. Local series and suites (formations, local stages) are described for each district. Regional biostratigraphic zones are defined by the first appearance and extinction of species. The Silurian of East Siberia features 23 local series and 61 formations. Strato- and hypostratotype sections are described in each district and are correlated at the level of subhorizons. Three profiles show the Silurian sequence and structure in the East Siberian Basin. The ranges of 536 species from 24 biotic groups from the East Siberian Silurian are reviewed and referred to regional biostratigraphic zones. The evolution of the East Siberian Basin is shown in six successive paleogeographic maps. The maps show communities (biogeocoenoses) in the early Rhuddanian, middle Aeronian, Sheinwoodian, Gorstian, Pridoli, and Lochkovian.

INTRODUCTION

Silurian East Siberia between the Yenisey and Lena Rivers (Siberian platform) and on the Taymyr Peninsula featured an epicontinental sedimentary basin with continuing cyclic sedimentation. Deep-water graptolitic mudstones to foreshore-lagoonal facies with fish are recorded along a north-south transect. The most important Silurian exposures are confined to a belt that includes Gorny Taymyr, the western and northeastern margins of the Tunguska syneclise, the western and southern Vilyuy syneclise, and the Irkutsk Amphitheatre. In the central Tunguska and Vilyuy syneclises and Yenisey-Khatanga Depression, the Silurian is overlain by thick sequences.

Silurian work in East Siberia is reviewed in Nikiforova (1955), Nikiforova and Andreeva (1961), Tesakov et al. (1979), and Tesakov and Obut (1984). Early work on general geology was published in the late nineteenth and early twentieth centuries. This work included the first recognition of the Silurian of East Siberia and described some of the fossils. A second stage of work ended in the late 1960s, and featured a survey of Silurian exposures along waterways with detailed descriptions of important sections and their fossils. Modern study began during the 1970s, with detailed lithologic and paleontologic investigation of the Silurian across East Siberia. Subsurface information from bore holes drilled during mineral prospecting was used in the modern synthesis. Bed-by-bed descriptions and correlations have now been done for almost all sections and wells. Regional and local stratigraphic units have been defined for all East Siberian districts. An emphasis is been placed on development of global, regional, and local biostratigraphic schemes and their application to paleogeographic maps (e.g., Tesakov et al., 1979, 1980, 1985, 1986, 1992, 1995, 1996a-c; Tesakov, 1981a, 1981b, 1996; Sokolov, 1982; Tesakov and Obut, 1984).

EAST SIBERIAN SILURIAN OUTCROP REGIONS

The Silurian of East Siberia is divided into "subregions" and "districts" on the basis of persistent lithofacies (Fig.

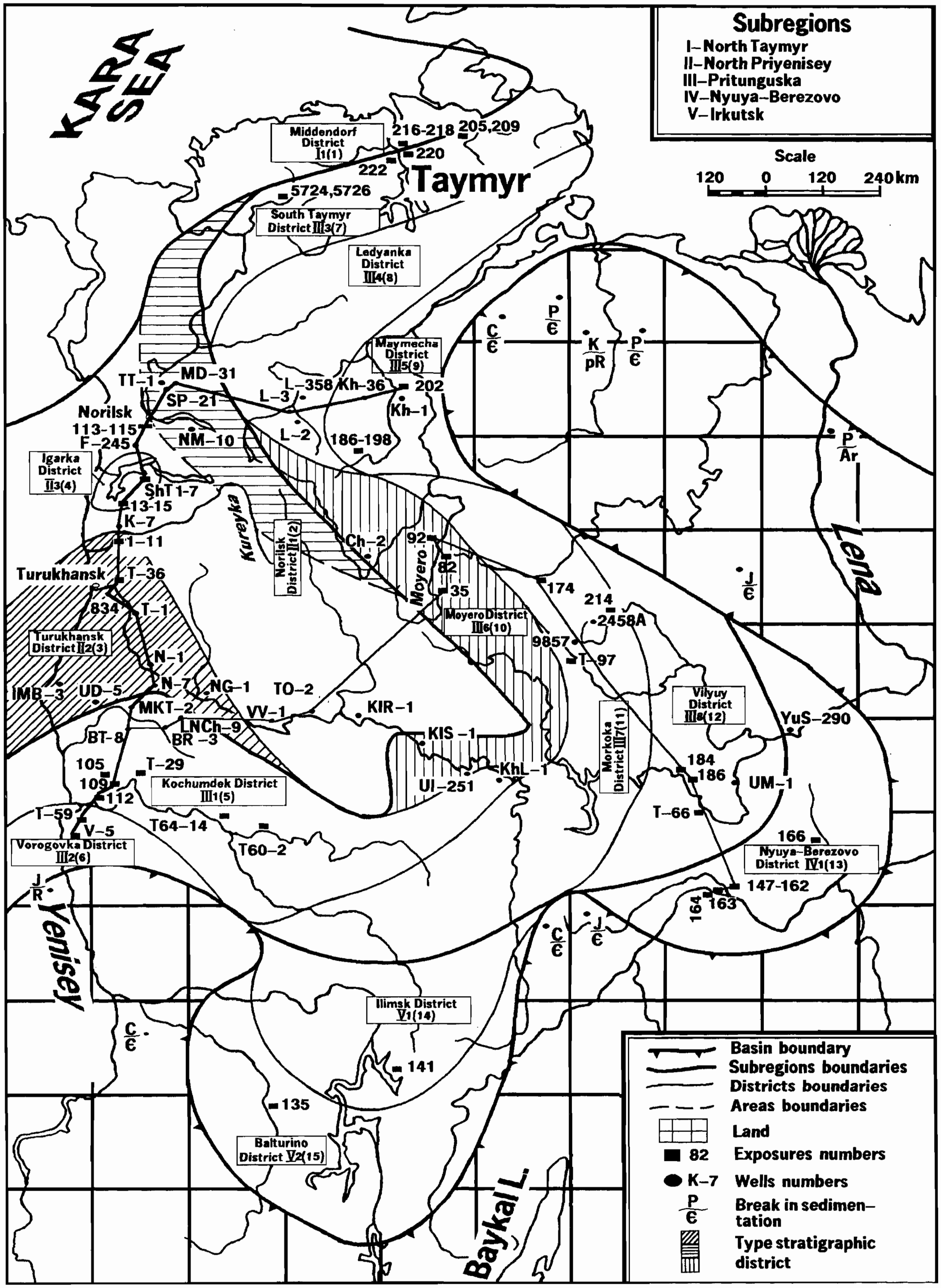

FIGURE 1 — Stratigraphy of the Silurian of East Siberia. I1(1)-V2(15): I-V: Roman numerals indicate subregions; 1, 2-Arabic numbers designate districts in subregion; (1)-(15), stratigraphic districts.

1). Five subregions are recognized: North Taymyr, North Priyenisey, Pritunguska, Nyuya-Berezovo, and Irkutsk. The North Taymyr Subregion is characterized by deep-shelf facies with graptolitic mudstone and occasional limestone. In the Priyenisey Subregion, deep- and shallow-shelf facies (graptolitic marl and brachiopod limestone) are developed. In the Pritunguska Subregion, open shallow-shelf facies with diverse benthic faunas are widespread. The Nyuya-Berezovo Subregion has lagoonal-type facies (dolomitic marl with fish and eurypterids). The Irkutsk Subregion is characterized by shallow-water, gypsum-bearing, dolomitic marl and sandstone with fish.

These subregions are further subdivided into fifteen lithostratigraphic districts in East Siberia. These districts are numbered herein: 1, Middendorf; 2, Norilsk; 3, Turukhansk; 4, Igarka; 5, Kochumdek; 6, Vorogovo; 7, South Taymyr; 8, Ledyanka; 9, Maymecha; 10, Moyero; 11, Morkoka; 12, Vilyuy; 13, Nyuya-Berezovo; 14, Ilimsk; and 15, Balturino. Regional stratigraphy of the East Siberia Silurian is consistent with this demarcation and shows columns with global, regional, and local stratigraphic units.

Global Chronostratigraphy

Global chronostratigraphic series (Fig. 2) include the Llandovery (with the Rhuddanian, Aeronian, and Telychian Stages), Wenlock (Sheinwoodian and Homerian Stages), Ludlow (Gorstian and Ludfordian Stages), and Pridoli (not yet divided into stages) (Bassett, 1985; Holland, 1989). In Russian nomenclature, global series are generally referred to as "superstages." The stages are subdivided into substages, which in turn are divided into global chronozones. These substages and chronozones are preliminary units, and their ranges are interpreted according to Tesakov (1985, 1996). The Rhuddanian is divided into lower (Bronid), middle (Crychan), and upper substages; the Aeronian into lower, middle (Fron), and upper substages; the Telychian into lower, middle, and upper substages; the Sheinwoodian into lower and upper substages; the Homerian into the Whitwell and Glidon Substages; the Gorstian into Elton and Bringewood substages; the Ludfordian into the Leintwardine and Whitcliffe substages; and the Pridoli into lower and upper substages. These substages include 54 global chronozones. The paleontological characteristics of global chronozones are based on global biointervals (biochrons) that coincide in number and age range with global chronozones (Tesakov, 1985, 1996; Tesakov et al., 1996c). A global biointerval is defined by the origin and extinction of the eponymous species. In the majority of reports, they are defined as faunal- and floral-based biozones rather than global chronozones. Their latest versions are summarized in Silurian Times (1993, 1994, 1995) and Koren' et al. (1995). The time of appearance and disappearance of the species widely employed in biostratigraphy, such as graptolites, brachiopods, conodonts, and tabulates, are shown in the stratigraphic chart (Fig. 3).

Regional Chronostratigraphic Units

Regional chronostratigraphic units (Fig. 4) for the Silurian of East Siberia include the Prianabarian Superhorizon (Moyerocanian, Khaastyrian, and Agidyian Horizons) and Priyeniseyian Superhorizon (Khakomaian, Tukalian, and Postnichian Horizons). They correspond to regional series and regional stages. Most of these horizons are divided into two subhorizons. The sole exception is the Postnichian Horizon, with three subhorizons. The subhorizons are divided into 54 regional chronozones, the stratigraphic range of which is consistent with that of global chronozones. The Silurian of East Siberia is a standard that involves the stratigraphic nomenclature of type districts in East Siberia (Moyero, Turukhansk, and Norilsk) and the stratotypes for the horizons, subhorizons, and regional chronozones located in these type districts (Tesakov et al., 1996a, c). Regional biointervals (biophases) are defined by the paleontological characteristics of regional chronozones, the range of which coincides with that of regional chronozones. The sole exceptions are regional biointervals 47–49 and 53–54, which correspond to several global chronozones. Regional biointervals are defined by the species that appeared and became extinct in regional chronozones. Before the biointerval scale applied to the subdivision of these sections was developed, biozones were based on graptolites, tabulates, brachiopods, ostracodes, conodonts, and other complicated regional zones (Tesakov and Obut, 1984). In the stratigraphic chart, the regional stratigraphy is scaled against the ranges of East Siberian species that are widely used in regional stratigraphy (Fig. 5, parts 1–5). The sections below include a description of the regional superhorizons and horizons in East Siberia.

Prianabarian Superhorizon — Abbreviated S_1prb, this regional series (Fig. 4) is equivalent to the Llandovery (Tesakov et al., 1992, p. 81). It is named for the Anabar anteclise. The local series of the same name in the Moyerocan District is considered as its type section. Laterally, it includes the Dvoynaya, Kaybat, Chatan, Koraz, Okhuk, Akher, Prianabar, Kuonda, Meik, Meut, and lower Andrey Local Series. It includes the Rassokha, Balturino, Tugulan, and lower Toyba Formations. This superhorizon is divided into three horizons.

Global Chronostratigraphic Scale for the Silurian System

Global chronostratigraphic units: System	Subsystem (Otdel)	Global series	Stage	Substage		Global chronozone	Global biointerval (biochron)	Preference scale, Ma (Interpolation after McKerrow, 1980)	Standard for the global chronostratigraphic units (English and Czech sections): Regional series	Formation (in range of stratotype)		Subformation	Local chronozone	Local biointerval	Stratotype	Particular units (beds), sites of observation	Thickness, *m*	Correlation level
Accepted by IUGS (Bassett, 1985)				Tesakov, 1985, 1996														
Devonian			Lochkovian	—		—	D	*411*		Lochkov				D		159 = K/20		
Silurian	Upper		Přídoli (*p*)	Bitov	Upp.	*2*	54		Požary	Požary 38.0		*II*	*2*	54	Požary	150-158	10.0	54
						1	53	*412*					*1*	53		141-149	8.0	53
					Lower	*3*	52					*I*	*3*	52		127-140	8.5	52
						2	51	*413*					*2*	51		106-126	7.5	51
				Jarov		*1*	50						*1*	50		96-105	4.0	50
		Ludlow (*ld*)	Ludfordian (*ldf*)	Whitcliffe		*2*	49	*414*	Ludlow	Whitcliffe 53.0		*II*	*1*	49	B'	II.1	30.0	49
						1	48					*I*	*1*	48		39-46	23.0	48
				Leintwardine		*1*	47	*415*		Leintwardine 33.0		*I,II*	*1*	47	C	9-31	33.0	47
			Gorstian (*gor*)	Bringewood		*3*	46			Bringewood 54.0		*II*	*1*	46	B	7-20	17.0	46
						2	45	*416*				*I*	*2*	45		1-6	20.0	45
						1	44						*1*	44	A	29-31	17.0	44
				Elton		*7*	43	*417*		Elton 174.0		*III*	*2*	43		24-28	20.0	43
						6	42						*1*	42		III.1	25.0	42
						5	41	*418*				*II*	*3*	41		16-23	28.0	41
						4	40						*2*	40		12-15	28.0	40
						3	39	*419*					*1*	39		7-11	28.0	39
						2	38					*I*	*2*	38		6	23.0	38
						1	37	*420*					*1*	37		1-5=P.C./F	22.0	37
	Lower	Wenlock (*v*)	Homerian (*hom*)	Gleedon		*2*	36		Wenlock	Much Wenlock 25.0		*I*	*1*	36	"Wenlock area"	47-50	25.0	36
						1	35	*421*		Coalbrookdale 215.0 (Farley)		*III*	*1*	35		24-25; 46	22.0	35
				Whitwell		*4*	34					*II*	*4*	34		19-23	22.0	34
						3	33	*422*					*3*	33		43-45	22.0	33
						2	32						*2*	32		34-42	22.0	32
						1	31	*423*					*1*	31		66-68; 31-33	22.0	31
			Sheinwoodian (*she*)	Upper		*3*	30					*I*	*4*	30		63-65	21.0	30
						2	29	*424*					*3*	29		61-62	21.0	29
						1	28						*2*	28		10	21.0	28
				Lower		*2*	27	*425*					*1*	27		58-60	21.0	27
						1	26			Buildwas 26.0		*I*	*1*	26		X-X'/G-I	26.0	26
		Llandovery (*l*)	Telychian (*tel*)	Upper		*1*	25	*426*	Llandovery	Cerig (*Cc*) 180.0	C_6	*II*	*1*	25	d4	C_6	45.0	25
				Middle		*3*	24				C_5	*I*	*3*	24		107	45.0	24
						2	23	*427*					*2*	23	d	$C_5$2	45.0	23
						1	22						*1*	22	d3	163	45.0	22
				Lower		*1*	21	*428*		Wormwood (*Cb*) 116.0	C_4	*II*	*1*	21		187	32.0	21
			Aeronian (*aer*)	Upper		*2*	20						*2*	20		160-162	42.0	20
						1	19	*429*				*I*	*1*	19		158-159	42.0	19
				Middle (Fron)		*5*	18			Rhydings (*Ca*) 210.0	C_3	*III*	*1*	18	f3	5a	42.0	18
						4	17	*430*			C_2	*II*	*2*	17		103	42.0	17
						3	16						*1*	16		102	42.0	16
						2	15	*431*			C_1	*I*	*2*	15		4	42.0	15
						1	14						*1*	14		101-139	42.0	14
				Lower	Idv	*2*	13	*432*		Trefawr (*B*) 170.0	B_3	*II*	*2*	13	h1	77-40-76	40.0	13
						1	12						*1*	12		72-39	40.0	12
			Rhuddanian (*rhu*)	Upper		*2*	11	*433*			B_2	*I*	*2*	11		70-71	40.0	11
						1	10				B_1		*1*	10		38-38a	50.0	10
				Middle (Crychan)		*6*	9	*434*		Crychan (*Ac*) 252.0	A_4	*II*	*3*	9	i2	34-35	42.0	9
						5	8						*2*	8		65	42.0	8
						4	7	*435*					*1*	7		186-64	42.0	7
						3	6					*I*	*3*	6		32; 63	42.0	6
						2	5	*436*					*2*	5		62	42.0	5
						1	4						*1*	4		28	42.0	4
				Lower (Bronydd)		*3*	3	*437*		Bronydd (*Ab*) 120.0		*III*	*1*	3		27	40.0	3
						2	2				A_3	*II*	*1*	2		61-26-26a	40.0	2
						1	1	*438*			A_2	*I*	*1*	1		88=B/46	40.0	1
Ordovician			Ashgillian	—		–				Scrach (*AA*)	A_1			O		24		

FIGURE 2 — Global chronostratigraphy of the Silurian.

Correlation level (biochron)	Silurian fauna and flora Global levels of species first appearance (+) and extinction (-) Graptoloidea	Brachiopoda	Conodonta	Tabulata
D	(+) Monograptus uniformis	—	—	(+) Favosites kozlowskii
54	(-) Monograptus transgrediens	—	(+) Icriodus woschmidti	(+) Riphaeolites sokolovi P85/6
53	(+) Monograptus perneri	(+) Dayia bohemica Brit	—	(+) Roemeria infundibulifera P219/4
52	—	—	—	(+) Scalites tschernovi P43/25
51	(+) Monograptus bouceki	—	—	(+) Mesosolenia reliqua P35/116
50	(+) Monograptus parultimus	(+) Gracianella graciosa Brit	—	—
49	(+) Monograptus formosus	—	(+) Ozarkodina crispa	—
48	(+) Neocucullograptus kozlowskii	—	(+) Ozarkodina snajdri	—
47	(+) Saetograptus leintwardinensis	—	(+) Polygnathoides siluricus	(+) Squameofav. incredibilis P27/37
46	—	(+) Conchidium knighti Brit	—	(+) Scalites prostratus P44/15
45	—	—	—	—
44	—	—	—	(+) Laceripora cribrosa P186/13
43	(+) Monograptus tumescens	—	(+) Lonchodina greilingi 10/16	—
42	—	(+) Chonetes lepisma Brit	—	(+) Mesofavosites bonus P186/1a
41	(+) Monograptus scanicus	—	(+) Ancoradella ploeckensis	(+) Barrandeolites bowerbanki P186/1
40	—	(+) Conchidium biloculare 218/86	(+) Spathognathodus primus 3/14	—
39	—	(+) Dayia navicula Brit	—	—
38	(+) Lobograptus progenitor	—	—	—
37	(+) Monograptus nilssoni	(+) Atrypa reticularis Brit	(+) Kockelella staurodus	—
36	—	—	—	(+) Desmidopora alveolaris P19/15
35	(+) Pristiograptus ludensis	—	—	—
34	(+) Monograptus deubeli	—	—	—
33	(+) Gothograptus nassa	—	—	(+) Thecia minor spinosa P16/1
32	—	—	(+) Ozarkodina bohemica	—
31	(+) Cyrtograptus lundgreni	—	(+) Ozarkodina sagitta sagitta	—
30	(+) Cyrtograptus ellesae	—	(+) Neoprioniodus excavatus 64/15	—
29	(+) Cyrtograptus linnarssoni	—	(-) Trichonodella symmetrica 58/16	(+) Subalveolites panderi P96/11a
28	(+) Cyrtograptus rigidus	—	(+) Kockelella variabilis 58/10	(+) Thecia minor minor P96/11
27	(+) Monograptus riccartonensis	—	(+) Ozarkodina sagitta rhenana	—
26	(+) Cyrtograptus centrifugus	(+) Eocoelia angelini Brit	(+) Huddlella johni 60/3	—
25	—	(+) Omnutakhella bazenovae 114/64	—	(+) Sapporipora favositoides 114/64
24	(+) Monograptus crenulatus	(+) Eocoelia sulcata Brit	—	—
23	(+) Monograptus greistoniensis	(+) Costistricklandia lirata Brit	(+) Ozarkodina gaertneri SP21/97	(+) Mesosapporipora porosa 87/38
22	(+) Monograptus crispus	(+) Eocoelia curtisi Brit	(+) Spathognathodus ozarkodini 86/30	(+) Mesosolenia festiva 87/35
21	(+) Monograptus turriculatus	(+) Stricklandia laevis Brit	(+) Pterosp. amorphognathoides 84/11	—
20	—	(+) Anabaria rara 77/39	(+) Trichonodella symmetrica 14/101	—
19	—	(+) Septatrypa antiquata 77/31	(+) Carniodus carnulus 178/17	(+) Subalveolites subulosus 113/30
18	—	(+) Plectatrypa wenlockiana 13/16	(+) Icriodella sandersoni 114/17	(+) Tuvaelites hemisphaericus 77/22
17	(+)Monograptus elegans	(+) Pentlandina subcostatula 77/13	(+) Pterospathodus celloni 14/75	—
16	—	(+) Cryptothyrella norilica 77/4	(+) Oulodus kentuckyensis 113/7	—
15	(-) Demirastrites triangulatus	—	(+) Icriodella inconstans 78/26	(+) Parastriatopora rhizoides TT1/28a
14	(+) Monograptus sedgwicki	(+) Kulumbella biconvexa FT3/14	(+) Icriodella deflecta 78/16	(+) Cystihalysites mirabilis LNCh9/11
13	(+) Cephalograptus cometa	(+) Pentamerus oblongus 13/11	(+) Apsidognathus tuberculatus 13/6	—
12	(+) Monograptus convolutus	(+) Strickl. lens intermedia Brit	(+) Kockelella ranuliformis 216/36	—
11	(+) Monograptus leptotheca	(+) Alispira tenuicostata 82/21	(+) Hadrognathus staurognathoides 184/42	(+) Paleofavosites asper 82/23
10	(+) Demirastrites delicatulus	(+) Eocoelia hemisphaerica 82/19	—	(+) Multisolenia tortuosa 82/20
9	(+) Diplograptus magnus	(+) Septatrypa magna 82/17	—	(+) Halysites catenularius LNCh9/8
8	(-) Rastrites norilskensis	(+) Stricklandia salteri 82/16	—	(+) F. gothlandicus gothlandicus 82/16
7	(-) Lagarograpt.inexpeditus	(+) Meifodia recta 82/13	—	—
6	(+) Demirastrites triangulatus	(+) Coolinia gracilis BT8/5	(+) Pterospathodus tenuis	—
5	(+) Coronograptus gregarius	—	(+) Decoriconus fragilis 184/11	—
4	(+) Coronograptus cyphus	(+) Stricklandia lens lens Brit	—	—
3	(+) Metabologr.moyeroensis	(+) Clorinda undata Brit	—	—
2	(+) Cystograptus vesiculosus	(+) Stricklandia lens prima Brit	(+) Distomodus kentuckyensis 79/6	—
1	(+)Parakidograptus acuminatus	(+) Alispira gracilis SP21/14	(+) Oulodus nathani 147/2	—
0	(-) Glyptograptus persculptus	(-) —	—	(-) Tollina keiserlingi

FIGURE 3 — First and last appearances of key taxa used in Silurian inter-regional correlation.

Regional Chronostratigraphic Scale for the Silurian of East Siberia													
Regional chronostratigraphic units					Standard for the regional chronostratigraphic units								Correlation level
Superhorizon	Horizon (Regional stage)	Subhorizon	Regional chronozone	Regional biointerval	Regional series	Formation (Local stage)	Subformation	Local chronozone	Local biointerval	Stratotype	Particular link (bed)	Thickness, *m*	
	Yampachtian					Yampachta					141-144	8.2	
Priyeniseyian (*prn*)	Postnichian (*ps*)	Upper	*2*	53-54	Priyenisey	Postnich 81.0	Upper 20.5	*2*	47 - 54	TT-1	136-140	11.2	54
			1					*1*			129-135	9.1	53
		Middle	*3*	52			Middle 31.5	*3*			125-128	11.2	52
			2	51				*2*			120-124	11.8	51
			1	50				*1*			116-119	8.5	50
		Lower	*3*	47-49			Lower 29.0	*3*			1146-115	7.9	49
			2					*2*			112-114a	11.2	48
			1					*1*			106-111	9.8	47
	Tukalian (*tk*)	Upper	*6*	46		Tukal 67.5	Upper 40.5	*6*	46	10	28-40	7.7	46
			5	45				*5*	45	4	13-14	5.1	45
			4	44				*4*	44		4-12	7.9	44
			3	43				*3*	43	1	10-16; 2-3	6.7	43
			2	42				*2*	42	2	16-18; 1-9	6.0	42
			1	41				*1*	41		29-30; 4-15	6.8	41
		Lower	*4*	40			Lower 27.0	*4*	40	3	14-28	6.4	40
			3	39				*3*	39		2-13	6.2	39
			2	38				*2*	38	11	14-17	6.4	38
			1	37				*1*	37		3-13	7.9	37
	Khakomian (*hk*)	Upper	*6*	36		Khakoma 82.0	Upper 40.0	*6*	36	66	26-31	8.0	36
			5	35				*5*	35		19-25	8.2	35
			4	34				*4*	34		15-18	4.8	34
			3	33				*3*	33		9-14	5.1	33
			2	32				*2*	32		7-8	6.1	32
			1	31				*1*	31		16-6	8.0	31
		Lower	*5*	30			Lower 42.0	*5*	30	64	12-18	9.8	30
			4	29				*4*	29	58	16-22	8.5	29
			3	28				*3*	28		7-15	6.2	28
			2	27				*2*	27	58A	4-8; 2-6	8.3	27
			1	26				*1*	26	60	3-4; 2-3	9.2	26
Prianabarian (*prb*)	Agidyian (*ag*)	Upper	*3*	25	Prianabar	Agidy 84.5	Upper 42.0	*3*	25	87	53-56	9.4	25
			2	24				*2*	24		41-52	14.8	24
			1	23				*1*	23		35-40	17.6	23
		Lower	*2*	22			Lower 42.5	*2*	22	86	23-34	20.9	22
			1	21				*1*	21	84;85	8-22	21.6	21
	Khaastyrian (*hs*)	Upper	*3*	20		Khaastyr 129.5	Upper 50.0	*3*	20	77	36-43	20.0	20
			2	19				*2*	19		27-35	13.6	19
			1	18				*1*	18		20-26	16.5	18
		Lower	*4*	17			Lower 79.5	*4*	17		11-19	21.0	17
			3	16				*3*	16	90	22-23; 1-10	18.0	16
			2	15				*2*	15	78	25-31; 18-21	20.3	15
			1	14				*1*	14		18-24	20.3	14
	Moyerocanian (*mr*)	Upper	*5*	13		Moyerocan 111.0	Upper 57.0	*5*	13	83	27-28	13.0	13
			4	12				*4*	12		24-26	11.4	12
			3	11				*3*	11	82	21-23	10.2	11
			2	10				*2*	10		19-20	10.5	10
			1	9				*1*	9		17-18	11.8	9
		Lower	*8*	8			Lower 54.0	*8*	8		14-16	13.7	8
			7	7				*7*	7		13	11.0	7
			6	6				*6*	6		11-12	8.3	6
			5	5				*5*	5		9-10	8.5	5
			4	4				*4*	4		8	5.0	4
			3	3				*3*	3		4-7	5.0	3
			2	2				*2*	2		2-3	1.8	2
			1	1				*1*	1		1	0.5	1
Keto	Burian					Bur							

FIGURE 4 — Regional chronostratigraphy of the Silurian of East Siberia.

Silurian fauna and flora of East Siberia

Regional levels of species first appearance (+) and extinction (-) in East Siberia

Correlation level (biophase)	*Graptoloidea*	*Chitinozoa*	*Acritarcha*	*Cephalopoda*
D	—	—	—	—
54	—	—	(-) *Leiosphaeridia plicata* SP21/228	—
53	—	—	—	—
52	—	—	—	—
51	(-) *Monograptus rarus* 218/22	—	—	—
50	(+) *Monograptus rarus* 218/20	—	—	(-) *Hemicosmorth. semiannulatum* 218/18
49	—	—	—	—
48	—	—	—	—
47	(-) *Monograptus priodon* 220/36	—	—	—
46	(-) *Bohemograptus bohemicus* 218/11	(-) *Desmochitina densa* 10/38	(-) *Favososphaer. kozlowskii* 10/28	—
45	—	—	(-) *Leiosphaeridia cerina* 115/44	—
44	—	—	(-) *Leiosphaeridia laevigata* 4/4	—
43	—	(-) *Sphaerochit. sphaerocephala* 115/39	(+) *Trachysphaer. raryplicatum* SP21/160	—
42	—	—	(-) *Cymatiosphaera pavimenta* 2/18	—
41	—	—	(-) *Cymatiosphaera nebulosa* 43/2	—
40	—	(-) *Eisenackitina bohemica* 115/25	—	(+) *Hemicosmort. semiannulatum* 218/86
39	(+) *Monograptus uncinatus* 218/8	—	—	—
38	—	(-) *Conochitina conulus* SP21/132	(-) *Lophosphaeridium citrinum* 10/28	—
37	(+) *Bohemograptus bohemicus* 218/6	(+) *Sphaerochit. sphaerocephala* 114/98	(-) *Favososp. heterobrochatum* SP21/131	—
36	—	—	—	(-) *Armenoceras bachtense* 66/31
35	—	—	—	—
34	—	—	—	—
33	(-) *Monograptus testis* 217/15г	(-) *Eisenackitina lagenomorpha* 66/12	—	—
32	(+) *Monograptus testis* 217/15в	—	—	—
31	—	—	(-) *Comasphaeridium williereae* 114/89	(-) *Huroniella inflecta* 114/89
30	(+) *Cyrtograptus lundgreni* 217/15a	—	—	(-) *Armenoc. sauthamptonense* MD31/104
29	—	—	—	—
28	—	(+) *Desmochitina densa* 58/7	—	(-) *Sastoceras richteri* 58/9
27	(+) *Pristiograptus dubius* 217/11	(+) *Ancyrochitina pachyderma* 114/82	(-) *Lophosphaeridium turulosum* 114/85	(+) *Kochoceras cunciforme* TT1/7
26	(+) *Monogr. riccartonensis* 217/9	—	(-) *Comasphaerid. sequestratus* 114/72	(+) *Sastoceras richteri* 60/3
25	(+) *Monograptus ludensis* 218/5a	(-) *Linochitina longa* 178/40	(+) *Favososphaer. polybrochatum* 87/55	(+) *Armenoceras bachtense* 178/42
24	(-) *Streptograptus nodifer* SP21/100	(-) *Conochitina rossica* 114/46	(-) *Leiosphaeridia voigti* 114/53	(+) *Kulinnia hyperborea* 114/48
23	(-) *Monograptus spiralis* 216/6	(-) *Calpichitina acollaris* 178/41	(-) *Trachyspaerid. universalum* PE43/84	(-) *Hiregiriceras costalatum* PE43/86
22	—	(-) *Conochitina edjelensis* PE43/77	(+) *Nucellosphaeridium deunffii* 86/25	(-) *Rizoceras acutum* 114/37
21	(+) *Streptograptus nodifer* SP21/71	(+) *Conochitina conulus* SP21/71	(+) *Multiplicisphaer. oligofurcatum* MD31/62	—
20	—	(-) *Eisenackitina conica* PE43/46	(+) *Dactylofusa estilis* SP21/65	—
19	—	—	(+) *Baltisphaerid. granuliferum* PE43/36	—
18	—	—	(+) *Trachysphaerid. formosum* PE43/26	—
17	(+) *Monograptus elegans* 216/5a	—	(+) *Favososphaerid. kozlowskii* 14/78	(-) *Armenoceras clarum* 177/20
16	—	(+) *Conochitina elegans* 178/4	(+) *Comasphaeridium williereae* 77/9	(-) *Geisonoceras kureikense* 113/4
15	(-) *Demirastr. triangulatus* 216/3д	(+) *Eisenackitina lagenomorpha* SP21/47	(+) *Leiosphaeridia plicata* MD31/20	(+) *Wadeoceras sibiricum* TT1/28a
14	(+) *Monograptus sedgwicki* 13/12	—	(+) *Micrhystridium coronatum* PE43/1	(-) *Armenoceras clarum* 177/12
13	—	(+) *Eisenackitina protracta* MD31/11	(+) *Trachysphaerid. universalum* SP21/34	—
12	(+) *Pernerogr. praecursor* SP21/19	—	(+) *Leiofusa granulacutis* 83/26	(+) *Huroniella inflecta* 83/26
11	—	(+) *Calpichitina acollaris* 175/3	—	(-) *Kentronites conulus* TT1/21
10	(+) *Demirastr. delicatulus* TT1/19г	—	—	—
9	(+) *Monograptus distans* NM10/4	(+) *Eisenackitina oviformis* SP21/33	(+) *Lophosphaerid. parverarum* MD31/5	(+) *Rizoceras acutum* 175/5
8	(-) *Rastrites norilskensis* SP21/26	(+) *Linochitina longa* 174/9	—	(+) *Armenoc. sauthamptonense* 82/16
7	(-) *Lagarogr. inexpeditus* TT1/19a	—	(+) *Leiosphaeridia voigti* SP21/28	—
6	(+) *Demirastr. triangulatus* TT1/13	(+) *Eisenackitina conica* SP21/27	(+) *Baltisphaer.echinodermum* MD31/2	(+) *Hiregiroceras costalatum* 174/3
5	(+) *Coronograp. gregarius* TT1/9	—	—	—
4	(+) *Coronograptus cyphus* 82/8	—	—	—
3	(+) *Metabologr. moyeroensis* 82/4	(+) *Ancyrochitina ancyrea* 85/2	—	—
2	(+) *Cystograptus vesiculosus* 216/1в	—	—	(+) *Geisonoceras kureikense* 82/2
1	(+) *Parakidogr. acuminatus* 216/1a	(+) *Conochitina edjelensis* SP21/14	—	—
0	—	—	—	—

FIGURE 5 — First and last occurrences of key taxa in the Silurian of East Siberia. Parts 1 (graptolites, chitinozoans, acritarchs, cephalopods), 2 (trilobites, brachiopods, scolecodonts, conodonts), 3 (bryozoans, tentaculitids, ostracodes, gastropods), 4 (crinoids, rugosan and tabulate corals, stromatoporoids), 5 (algae, pelecypods [i.e., bivalves], vertebrates, and hyoliths, and conularids).

Silurian fauna and flora of East Siberia

Regional levels of species first appearance (+) and extinction (-) in East Siberia

Correlation level (biophase)	*Trilobita*	*Brachiopoda*	*Scolecodonta*	*Conodonta*
D	—	—	—	—
54	—	—	—	—
53	—	—	—	—
52	—	—	—	—
51	—	—	—	—
50	—	—	—	—
49	—	—	—	—
48	—	—	—	—
47	—	—	—	—
46	—	(-) *Morinorhynchus proprius* 115/55	—	(-) *Ozarkodina typica* 10/38
45	—	(-) *Eohowellella yadrenkinae* 4/14	—	(-) *Spathognathodus primus* 10/28
44	—	(-) *Howellella elevataeformis* MD31/148	—	(-) *Oulodus siluricus* MD31/143
43	—	(-) *Dalejina ribnayaensis* LNCh9/39	—	(-) *Lonchodina greilingi* 10/16
42	—	—	—	
41	—	(-) *Protatrypa lepidota* 115/39	(-) *Xaniprion sibiricus* SP21/133	
40	—	(+) *Conchidium biloculare* 218/86	(-) *Polychaetaspis latus* 115/26	(+) *Spathognathodus primus* 3/14
39	—	(-) *Anabaria rara* 3/13	—	—
38	—	—	(-) *Mochtyella angelini* 115/20	—
37	—	—	(-) *Kettnerites aspersus* 114/102	(+) *Oulodus siluricus* TT1/79
36	—	—	—	(-) *Kockelella variabilis* 52/10
35	—	(-) *Stegerhynchus mojeroensis* 66/22	—	(-) *Huddlella johni* 66/20
34	(-) *Encrinurus creber* 114/94	(-) *Alispira rotundata* 62/8	—	—
33	—	—	—	—
32	—	—	—	—
31	—	(-) *Plectatrypa wenlockiana* MKT2/52	—	(+) *Ozarcodina excavata* 64/20
30	—	—	—	(+) *Neoprioniodus excavatus* 64/15
29	—	—	—	(-) *Trichonodella symmetrica* 58/16
28	—	—	(-) *Vistulella kozlowskii* 114/87	(+) *Kockelella variabilis* 58/10
27	—	—	—	—
26	—	(-) *Eoplectodonta pumila* 217/6	—	(+) *Huddlella johni* 60/3
25	(+) *Encrinurus creber* 114/66	(+) *Omnutakhel. bazhenovae* 114/64	—	(-) *Distomodus kentuckyensis* 114/65
24	—	(-) *Alispira tenuicostata* 87/50	—	(-) *Icriodella sandersi* 114/51
23	—	(+) *Morinorhychus proprius* 114/38	—	(+) *Ozarkodina gaertneri* SP21/97
22	(-)*Eophacops quadrilineatus* FT3/24	(-) *Pentamerus oblongus* FT3/24	—	(+) *Spathognath. ozarkodini* 86/30
21	(-) *Acernaspis incerta* 14/117	(+) *Protatrypa lepidota* TT7/54	—	(+) *Pter. amorphognathoides* 84/11
20	(-) *Acernaspis nanus* 178/24	(+) *Anabaria rara* 77/39	—	(+) *Trichonodella symmetrica* 14/101
19	—	(-) *Septatrypa antiquata* 77/31	(-) *Mochtyella fragilis* 178/15	(+) *Carniodus carnulus* 178/17
18	(+) *Bumastus barriensis* 77/24	(+) *Plectatrypa wenlockiana* 113/16	(+) *Mochtyella angelini* 115/20	(+) *Icriodella sandersi* 114/17
17	(+) *Encrinurus punctatus* TT7/35	(+) *Pentlandina subcostatula* 77/13	(+) *Kozlowskipr.brevialatus* 113/11	(+)*Pterospathodus celloni* 14/75
16	(+) *Stenopareia bowmanni* 177/19	(+) *Cryptothyrella norilica* 77/4	(+) *Multiprion trapezoideus* MD31/26	(+) *Oulodus kentuckyensis* 113/7
15	(+) *Eophacops quadrilineatus* TT7/24	(-) *Brevilamnul. undatiformis* FT3/16	(+) *Polychaetaspis latus* MD31/20	(+) *Icriodella inconstans* 78/26
14	(+) *Eobronteus norilskensis* 78/23	(+) *Kulumbella biconvexa* FT3/14	(+) *Kettnerites aspersus* SP21/37	(+) *Icriodella deflecta* 78/16
13	(+) *Acernaspis pulcher* 175/8	(+) *Pentamerus oblongus* 13/11		(+) *Apsidognath. tuberculatus* 13/6
12	(-) *Acernaspis superciliexcelsis* 83/24	(+) *Howellel. elevataeformis* 216/3d		(+) *Kockelella ranuliformis* 216/36
11	(+) *Unguliproetus enodis* 78/8	(+) *Alispira tenuicostata* 82/21		(+) *Hadrogn. staurognathoides* 184/42
10	(+) *Pseudoproetus tertius* 82/19	(+) *Borealis nanus* 82/20	—	—
9	(+) *Stenopareia angulata* 82/18	(+) *Septatrypa magna* 82/17	—	—
8	—	(+) *Stricklandia salteri* 82/16	—	—
7	—	(+) *Clorinda undata* 83/13	—	—
6	(+) *Acernaspis superciliexcelsis* 82/11	(+) *Coolinia gracilis* BT8/5	—	(+) *Icriodella discreta* 174/2
5	—	—	—	(+) *Decoriconus fragilis* 184/11
4	—	(+) *Strophomena sibirica* FT3/3	—	—
3	(+) *Pseudoproetus bellus* FT3/2	(+) *St. decemplicatus duplex* 82/4	—	—
2	(-) *Cyphoproetus externus* 82/3	(+) *Eridorthis siluriensis* LNCh9/1	—	(+) *Distomodus kentuckyensis* 79/6
1	(+) *Cyphoproetus externus* 82/1	(+) *Alispira gracilis* SP21/14	(+) *Vistulella kozlowskii* SP21/14	(+) *Oulodus nathani* 147/2
0	—	—	—	—

FIGURE 5 *continued* (part 2).

Silurian fauna and flora of East Siberia

Regional levels of species first appearance (+) and extinction (-) in East Siberia

Correlation level (biophase)	*Bryozoa*	*Tentaculitida*	*Ostracoda*	*Gastropoda*
D	—	—	—	—
54	—	—	—	—
53	—	—	—	—
52	—	—	—	—
51	—	—	—	—
50	—	—	—	—
49	—	—	—	—
48	—	—	—	—
47	—	—	—	—
46	(-) *Monotrypa benjamini* 10/34	—	(+-) *Schrenkia multa* 4/5	(-) *Prosolarium cirrhosa* 10/28
45	—	—	(-) *Beyrichia parva* 115/44	—
44	(-) *Lioclema crustulum* 4/9	—	(-) *Beyrichia kureikiana* F245/140	—
43	(+) *Lioclema crustulum* 1/11	—	(+) *Beyrichia parva* 115/39	—
42	(+) *Mesotrypa alashensis* 1/8	—	(+) *Healdianella inornata* 1/16	—
41	—	—	—	—
40	—	—	(+) *Eukloed. kureikensis* 115/25	—
39	(+) *Monotrypa benjamini* 3/7	—	(+) *Signetopsis cardinata* 115/23	—
38	—	—	—	—
37	—	—	—	—
36	—	—	—	—
35	(-) *Homotrypa hondelensis* 6/24	—	(+) *Leperditia lumaea* 66/21	(-) *Lophospira sinuosa* 66/20
34	—	—	—	—
33	—	—	(+) *Beyrichia kureikiana* 66/9	(-) *Trochonema transformis* 114/94
32	(-) *Eridotrypa callosa* 114/90	—	—	(-) *Lophospira alta* 114/90
31	—	—	—	—
30	—	—	(-) *Cytherellina oviformis* K1010/26	—
29	—	—	—	—
28	(-) *Batostoma microcellata* 114/86	—	—	(+) *Lophospira alta* 114/86
27	—	—	(-) *Sibiritia wiluiensis* LNCh9/41	(+) *Oriostoma varvara* 114/83
26	(+) *Monotrypa pediculata* 60/3	—	—	—
25	—	—	—	—
24	(+) *Eridotrypa callosa* 114/50	—	(-) *Bollia cardinis* K1010/21	(-) *Holopea transversa* 114/48
23	—	—	(-) *Imangdites ambiquus* TT7/68	(-) *Pararaph. qualteriatum* 114/43
22	(+)*Batostoma microcellata* 114/36	(-) *Costatulites corniformis* 114/37	(-) *Norilskinia cuspidata* 86/31	(-) *Murchisonia insignis* TT7/55
21	(-) *Moyerella stellata* MKT2/26	—	(+) *Herrmannina nana* 85/20	(+) *Murchisonia cingulata* 114/18
20	(-) *Stictopora markhensis* 178/26	—	(+) *Beyrichia mirabilis* PE43/40	(-) *Bucanopsis sibiricus* 178/27
19	(-) *Hennigopora florida* 14/96	(-) *Costatulites homogenus* 89/1	(+) *Sibiritia eurina* F245/18	(-) *Poleumita mashkovae* 178/9
18	(-) *Ptilodictya lanceolata* 77/25	—	(+) *Norilskinia norilica* SP21/59	(-) *Poleumita anabarica* 83/28
17	(+) *Stictopora markhensis* 14/74	—	(+) *Hatangeus armatus* TT7/33	(+) *Trochonema transformis* 113/9
16	(+) *Stictoporella lamellata* 180/4	(+) *Evenkyites rarus* 77/1	(+) *Cytherellina oviformis* K1010/21	—
15	(+) *Hennigopora florida* 113/1	—	(+) *Daleiella decorata* SP21/47	(+) *Loxoplocus vostokovae* 177/20
14	(+) *Rhinidictya bifurcata* 13/12	(-) *Costatulites undatus* 78/23	(+) *Costaegera hastata* 78/24	(+) *Gyronema quadricarina* 175/15
13	(+) *Helopora spiralis* 83/27	—	—	(+) *Poleumita anabarica* 83/28
12	(+) *Ensipora erecta* 83/24	—	—	(+) *Ruedemannia lirata* 177/10
11	—	(+) *Costatulites homogenus* 78/11	—	—
10	(+) *Monotr. amplexoformis* 82/20	—	—	(+) *Umbonellina infrasilurica* 82/19
9	—	—	(+) *Eurychilina fragilis* 82/17	(+) *Cyclonema bilix* 82/17
8	(+) *Moyerella stellata* 82/15	—	(+) *Thrallella alveolata* LNCh9/7	(+) *Straparollus alacer* UD5/7
7	—	—	(+) *Sibiritia wiluiensis* 82/13	(+) *Eotomaria supracingulata* 175/2
6	—	—	(+) *Silenis sibiricus* 174/1	—
5	—	—	—	(+) *Gyronema multangulata* TT1/16
4	—	(+) *Costatulites undatus* 82/8	—	(+) *Murchisonia insignis* UD5/4
3	—	—	—	—
2	—	—	—	(+) *Prosoptychus globulus* LNCh9/4
1	—	—	—	—
0	—	—	—	—

FIGURE 5 *continued* (part 3).

Silurian fauna and flora of East Siberia

Regional levels of species first appearance (+) and extinction (-) in East Siberia

Correlation level (biophase)	Crinoidea	Rugosa	Tabulata	Stromatoporoidea
D	—	—	(+) Tiverina vermiculata ShT27/11	—
54	—	—	—	—
53	—	—	—	—
52	—	—	—	—
51	—	—	—	—
50	—	—	—	—
49	—	—	—	—
48	—	—	—	—
47	—	—	—	—
46	(-) Bazaricrinus parvulus 10/34	—	(-) Parastriat. kureikiana 10/39	(-)Clathrodictyon mohicanum 10/36
45	—	—	(+) Parastriat. kureikiana 4/13	—
44	(-) Bystrow. quinquelobatus 2/11	—	(-) Syringop. fascicularis 115/40	(-) Densastroma astroites 4/12
43	(-)Bystrowicrinus bilobatus 115/44	—	—	—
42	—	(-) Tryplasma flexuosum 1/6	—	(+ -) Stromat. dzvenigorodensis 1/6
41	—	—	—	—
40	—	(-) Miculiella annae 3/13	—	—
39	—	(-) Entelophyllum articulatum 3/5	(-) Subalveol. subulosus 3/13	(-) Ecclimadictyon sibiricum 115/106
38	—	—	—	(+) Ecclimadictyon sibiricum 115/7
37	—	(+) Tryplasma flexuosum 11/6	(-) Multisolenia tortuosa SP21/127	(-) Ecclimadictyon fastigiatum 114/104
36	(-) Egiasar. egiasarowi 114/96	(-) Yassia enormis 66/31	(-) Mesosolenia festiva 87/35	(+) Plexodictyon savaliense 66/31
35	—	—	—	—
34	—	—	—	(+) Yavorskiina njuilensis 163/3
33	—	(-) Cyathactis typus TT1/69	—	(-) Rosenella hakomiense 53/7
32	—	(+) Cystiphyllum cylindricum 66/8	—	(-) Clathrodictyon muriei BT8/61
31	—	(+) Miculiella compacta TT1/69	—	(-) Labechia conferta BT8/60
30	—	(+) Kymocystis papillaris 54/1	(-) Halysites catenularius K1010/26	(+) Yabeodictyon crispatum UD5/35
29	(-) Crotalocrinites borealis FT3/23	(+) Neocystiphyl. holtedahli 58/16	—	—
28	—	(+) Miculiella crassiseptata 58/7	(+) Parastriatopora mutabilis 114/86	—
27	—	(+) Neocystiphyll. mac\`coyi 58A/4в	(+-) Desmidopora alveolaris 114/85	—
26	(-) Turuchan. turuchanensis 114/80	(+) Protopiloph. cylindricum 60/3	—	(+)Ecclimadictyon fastigiatum 114/75
25	(+) Dastaricr. petaloideus 114/67	—	(+) Sapporipora favositoides 114/64	(+) Clavidictyon cylindricum 114/62
24	(-) Tajmirocr. tajmirensis 78/21	—	(-) Mesofavosites dualis SP21/26	(+) Neobeatricea nikiforovae BT8/44
23	(+) Sibiricrinus helenae 114/44	(+) Cyathactis typus TT1/58	(-) Mesosolenia festiva 87/35	(+) Stromatopora lenensis 85/35
22	(+)Megalocrinus latilobatus 112/7	(+) Yassia enormis 86/22	(+) F.gothland. moyeroensis MKT2/30	—
21	(+) Turuchan. turuchaensis 114/24	(-) Entelophyllum medius TT1/46	(+) Mesofavosites planus 84/12	(+) Pachystylostroma sibiricum 84/22
20	(+) Scalaricr. scalariformis 77/43	(-) Crass. crassiseptatum 178/26	(+) Placocoenites orientalis 77/39	(+) Plümatalinia densa 77/39
19	(+) Bystrowicrinus bilobatus TT1/34	(-) Streptelasma whittardi 178/16	(+) Vaenopora kaljoi 77/29	(-) Labechia obrouchevi 112/3
18	(+) Megalocrinus pentalobatus TT1/31	(+) Streptelasma whittardi 178/9	(+) Tuvael.hemisphaericus 77/22	—
17	—	(+) Cystilasma sibiricum TT1/92	(+) Subalveolites subulosus 114/30	(+) Plectostroma tenuipalum UD5/16
16	(+) Crotalocrinites borealis SP21/48	(+) Cyathactis euryone 77/6	(+) Coenites juniperinus UD5/13	—
15	—	(+) Helicelasma whittardi TT1/28a	(+) Subalveolitella repentina L3/22	(+) Labechia obrouchevi 101/1
14	(+) Myelodactylus flexibilis 90/3	(+) Entelophyl. articulatum 78/21	(+) Cystihalysites mirabilis LNCh9/11	(+) Clathrodictyon variolare 9857/17
13	—	—	—	—
12	—	—	(+) Angopora hisingeri L3/13	—
11	(+) Glyptocrinus elegans TT1/21	(+) Crassilasma completum 82/19	(+) Paleofavosites asper 82/23	—
10	—	(+) Brachyelasma sibiricum 175/4	—	—
9	—	—	—	(+) Clath. microundulatum BT8/116
8	—	(+) Kodonoph. complanatum 82/16	—	—
7	—	—	(+) Multisolenia tortuosa LNCh9/86	—
6	—	—	(+) F.goth.gothlandicus LNCh9/8a	(+) Labechia venusta LNCh9/8
5	—	—	—	—
4	—	—	—	—
3	—	—	—	—
2	(+) Dentiferocr. dentiferus LNCh9/1	—	—	—
1	—	—	—	—
0	—	—	(-) Tollina keyserlingi 205/24	(-) Aulacera nodulosa

FIGURE 5 *continued* (part 4).

Correlation level (biophase)	Silurian fauna and flora of East Siberia — Regional levels of species first appearance (+) and extinction (-) in East Siberia: *Algae*	*Pelecypoda*	*Vertebrata*	*Hyolitha & Conulata*
D	—	—	—	—
54	—	—	—	—
53	—	—	—	—
52	—	—	—	—
51	—	—	—	—
50	—	—	—	—
49	—	—	—	—
48	—	—	—	—
47	—	—	—	—
46	—	—	—	—
45	*(-)Hedstroem. halimedoidea* 115/46	—	—	—
44	—	—	—	—
43	—	—	—	—
42	—	—	—	—
41	—	—	—	—
40	—	*(+-) Cardinia signata* 118/86	—	—
39	—	—	—	—
38	—	—	—	—
37	—	—	—	—
36	—	*(-) Megalomus sp.*	—	—
35	—	—	—	—
34	—	—	—	—
33	—	—	—	—
32	—	—	—	—
31	—	—	—	—
30	—	*(+) Megalomus sp.*	—	—
29	—	—	—	—
28	—	—	—	—
27	—	—	—	—
26	*(+-) Solenopora concentrica* 114/75	—	*(-) Loganellia scotica* 135/64	—
25	*(+)Rothpletzella gotlandica* 114/70	—	—	—
24	*(+) Hedstroem. halimedoidea* 114/58	—	—	—
23	—	—	*(+) Helenolepis trifurcata* 135/57	—
22	—	—	—	—
21	—	—	—	—
20	—	—	*(-) Tubia bergeri* 135/48	—
19	—	—	*(+) Udalepis forata* 135/44	—
18	—	—	*(+-) Tesakoviasp. concentrica* 135/42	*(-) Conulata* 9857/41
17	—	—	—	—
16	—	—	*(+) Loganellia scotica* 140/24	—
15	—	—	—	*(-) Hyolitha* SP21/44
14	—	—	*(+) Elegestolepis conica* 156/30	—
13	—	—	—	—
12	—	—	—	—
11	—	—	—	—
10	—	*(-) Actinopteria pumila* FT3/12	*(+) Ilimia predtechenskii* 135/12	—
9	—	—	—	—
8	—	—	*(+) Loganellia moskalenkoae* 140/8	—
7	—	—	—	—
6	—	—	*(+) Loganellia sibirica* 156/2	*(+) Conulata* 82/11
5	—	—	—	—
4	—	—	—	—
3	—	—	—	—
2	—	*(+) Actinopteria pumila* FT3/1	—	*(+) Hyolitha* SP21/20
1	—	—	—	—
0	—	—	—	—

FIGURE 5 *continued* (part 5).

The lowest is the Moyerocanian Horizon (or regional stage, Rhuddanian–lower Aeronian), which is abbreviated S_1mr (Tesakov et al., 1979, p. 16; Fig. 4). It is named after the type formation, with which it has a common stratotype. From northwest to southeast, it ranges from graptolite mudstone to limestone with abundant fossils to dolostone and restricted marine rocks. The lower part of this horizon has *Parakidograptus acuminatus;* the middle has *Demirastrites triangulatus,* and a *Demirastites convolutus* Zone assemblage occurs in the upper part.

The second is the Khaastyrian Horizon (regional stage, middle–upper Aeronian), which is abbreviated S_1hs (Tesakov et al., 1979, p. 19; Fig. 4). It is named after the type formation with which it shares a common stratotype. It ranges from northwest to southeast, and from graptolitic mudstone to limestone with abundant fossils (on the central Siberian Platform) and to dolostone associated facies. *Monograptus sedgwicki, Pristiograptus regularis, Eocoelia hemisphaerica, Pentamerus oblongus, Subalveolites volutus,* and *Cystihalysites mirabilis* occur in the succession.

The upper horizon is the Agidyian Horizon (regional stage, Telychian), which is abbreviated S_1ag (Tesakov et al., 1979, p. 22; Fig. 4). It is named after the type formation with which it shares a stratotype. To the northwest, it is represented by mudstone and limestone with a normal marine fauna. In the central region, it is mainly dolomitic marl and dolostone. In the Irkutsk Amphitheatre, it includes variegated silt- and sandstone. *Streptograptus nodifer, Monograptus spiralis, M. turriculatus, Stomatograptus grandis, Pentamerus oblongus,* and *Pteraspathodus amorphognathoides* occur in this horizon.

Priyeniseyian Superhorizon — Abbreviated S_{1-2}prn, this regional series (Fig. 4) is equivalent to the Ludlow–Pridoli (Tesakov et al., 1992, p. 81). It is named for the Yenisey River. The local series of the same name in the Norilsk District is the type section for this superhorizon. The superhorizon includes the following local series: Middendorf, Priyenisey, Diavolskaya, the lower Nimda, Mukon, the lower Pankagir, Uskhur, the upper Andrey, Mukir, Dokir, Khakhol, Yartom, and Kurung. It includes the upper Toyba, Khurichi, Bunge, Nyuya, Neryuktey, Deshyma, and Barmo Formations. The superhorizon is subdivided into three horizons.

The lower is the Khakomaian Horizon (or regional stage, =Wenlock), which is abbreviated S_1hk (Tesakov et al., 1979, p. 24; Fig. 4). The horizon is named after the type formation with which it shares a stratotype. It is represented virtually across the platform by massive limestone or dolostone, coral-stromatoporoid biostromes and bioherms, and only in the Middendorf and Norilsk Districts by normal marine marls. The fauna includes *Monograptus riccartonensis, Pristiograptus dubius, Cyrtograptus lundgreni, Testograptus testis, Clathrodictyon boreale, Clavidictyon cylindricum, Sapporipora favositoides, Cystiphyllum siluriense, Herrmannina nana, Pteraspathodus amorphognathoides, Anabaria rara,* and *Protatrypa lepidota.*

The middle division is the Tukalian Horizon (or regional stage, =Gorstian), which is abbreviated S_2tk (Tesakov et al., 1979, p. 25; Fig. 4). This horizon is named after the type formation in the Turukhansk District with which it shares a stratotype. In the Middendorf District, it is made up of graptolitic mudstone, but on the southeast platform it includes variegated dolostone, marl, and gypsum. In the rest of East Siberia, this horizon features grey limestone and dolostone with abundant stromatolites. The fauna includes *Bohemograptus bohemicus, Saetograptus colonus, Monograptus haupti, Didymothyris didyma, Ozarkodina excavata, Stromatopora malinovzyensis, Schrenkia multa,* and *Parastriatopora kureikiana.*

The upper is the Postnichian Horizon (or regional stage, =Ludford–Pridoli), which is abbreviated S_2ps (Tesakov et al., 1992, p. 81; Fig. 4). The horizon is named after the formation in the Norilsk District with which it shares a stratotype. In the Middendorf District this horizon is made up of black limestone and mudstone. In other East Siberia areas, it is represented by gray dolostones, frequently with stromatolites or gypsum-bearing, green or variegated dolomitic marl. *Bohemograptus bohemicus, Monograptus priodon, M. uncinatus,* and *M. rarus* appear in the Middendorf District. Only recrystallized gastropods, eurypterids, and bivalves are found elsewhere.

Local Stratigraphic Units

Local stratigraphic units are described below in all fifteen stratigraphic districts. The local series, formations (suites), subformations (subsuites), and local chronozones are established in each district. Their stratotypes or hypostratotypes are designated in Fig. 2 (asterisks [*] mark hypostratotypes).

Middendorf Stratigraphic District

Named by Tesakov et al., (1996b), this stratigraphic district is located in the middle Gorny Taymyr (Fig. 1). It includes the Dvoynaya (Zaputannaya and Kosaya Formations) and Middendorf (Golub, Privalnaya, and Shara Formations) Local Series (Fig. 6).

Dvoynaya Local Series (Llandovery, S_1dv) — First termed by Sobolevskaya et al. (1986) as the "Dvoynaya Beds" (see Tesakov et al., 1995, p. 125), this series is named

Local stratigraphic units for the Silurian of East Siberia

North Taymyr Subregion I

Middendorf District 245.0, I. 1 (1)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D		Dolostone				24	
54	Middendorf *md* 145.1	Shara *sh* 69.7	Upp. 13.5	2	7.0	23в	218
53				1	6.5	23б	
52			Middle 30.0	3	6.5	23а	
51				2	11.2	21-22	
50				1	12.3	15-20	
49			Lower 26.2	3	8.0	14б	
48				2	8.5	14а	
47				1	9.7	12-13	
46		Privalnaya *prv* 38.1	Upper 22.6	6	4.6	10г-11	
45				5	3.5	10в	
44				4	3.5	10б	
43				3	3.5	10а	
42				2	4.0	9б	
41				1	3.5	9а	
40			Lower 15.5	4	3.5	8б	
39				3	3.5	8а	
38				2	3.5	7	
37				1	5.0	6	
36		Golub *gl* 37.3	Upper 21.5	6	4.0	5б	
35				5	3.9	3-5а	
34				4	3.1	15д; 2	217
33				3	3.0	15г	
32				2	3.8	15в	
31				1	3.7	15б	
30			Lower 15.8	5	4.0	15а	
29				4	3.3	12б-14	
28				3	2.5	12а	
27				2	3.3	10б-11	
26				1	2.7	9-10а	
25	Dvoynaya *dv* 99.5	Kosaya *ks* 39.5	Upper 15.5	3	5.0	8в	
24				2	5.0	8б	
23				1	5.5	7-8а	
22			Low. 24.0	2	15.0	6	
21				1	9.0	6-7	216
20		Zaputannaya *zp* 60.0	Upper 36.5	7	6.0	5г	
19				6	6.0	5в	
18				5	6.0	5б	
17				4	5.0	4-5а	
16				3	4.5	3е	
15				2	4.5	3д	
14				1	4.5	3г	
13			Lower 23.5	13	3.0	3в	
12				12	2.9	3б	
11				11	2.6	3а	
10				10	2.2	2г	
9				9	2.0	2в	
8				8	2.1	2б	
7				7	2.0	1ж-2а	
6				6	1.9	1е	
5				5	1.8	1д	
4				4	1.0	1г	
3				3	1.0	1в	
2				2	0.5	1б	
1				1	0.5	1а	
0		Barkov					

North Priyenisey Subregion II

Norilsk District 645.5 (here) - 772.0 (F-245), II. 1 (2)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D		Yampakhta				141	
54	Priyenisey *prn* 269.1	Postnich *ps* 80.7	Upp 20.3	2	11.2	136-140	TT-1
53				1	9.1	129-135	
52			Middle 31.5	3	11.2	125-128	
51				2	11.8	120-124	
50				1	8.5	116-119	
49			Lower 28.9	3	7.9	114б-115	
48				2	11.2	112-114а	
47				1	9.8	106-111	
46		Makus *mkk* 117.8	Upper 64.4	6	4.9	52-55	115
45				5	13.9	40б-51	
44				4	13.5	38-40а	
43				3	13.5	30-37	
42				2	16.9	22б-29	
41				1	15.2	16б-22а	
40			Lower 53.4	4	13.2	12б-16а	
39				3	11.4	9б-12а	
38				2	12.9	2б-9а	
37				1	15.9	97-103;1-2а	
36		Khyukta *khk* 70.6	Upper 38.0	6	5.4	96	114
35				5	7.2	94б-95	
34				4	5.7	92-94а	
33				3	6.4	90б-91	
32				2	6.5	89б-90а	
31				1	6.8	87-89а	
30			Lower 32.6	5	7.7	86б-86в	
29				4	6.3	85б-86а	
28				3	5.1	82-85а	
27				2	7.2	76б-81	
26				1	6.3	72-76а	
25	Kaybat *kb* 376.4	Omnutakh *om* 191.9	Upper 64.7	3	21.1	56-71	
24				2	23.5	44б-55	
23				1	20.1	38-44а	
22			Mid. 85.5	2	37.5	29б-37	
21				1	48.0	17а-29а	
20			Lower 41.7	3	9.9	62б-64	SP-21
19				2	15.9	59б-62а	
18				1	15.9	15-16а	113
17		Talikit *tl* 75.9	Upp. 36.0	2	17.4	52-57	SP-21
16				1	18.6	48-51	
15			Low. 39.9	2	21.7	42б-47	
14				1	18.2	36-42а	
13		Chamba *č* 108.6	Upper 56.9	5	13.9	34-35	SP-21*
12				4	12.2	33е2-33з	
11				3	10.0	33г2-33е1	
10				2	10.0	33б2-33г1	
9				1	10.8	33а-33б1	
8			Lower 51.7	8	10.7	30-32	
7				7	11.9	28-29	
6				6	8.8	26в2-27	
5				5	8.7	26б-26в1	
4				4	4.6	24-26а	
3				3	4.8	20-23	
2				2	1.6	18-19	
1				1	0.6	14-17	
0		Zagorny					

Turukhansk District 340.9, II. 2 (3)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D		Mudstone				25	
54	Nimda *nm*	Postnich *ps* 61.2	Upp. 15.3	2	8.8	22-24	KhL-8*
53				1	6.5	20-21	
52			Middle 23.9	3	8.7	17-19	
51				2	9.0	13-16	
50				1	6.2	12	
49			Lower 22.0	3	6.2	10-11	
48				2	8.3	8-9	
47				1	7.5	7	
46	Diavolskaya *dvl* 143.0	Tukal *tk* 61.7	Upper 40.2	6	7.7	28-40	10
45				5	5.1	13-14	4
44				4	7.9	4-12	
43				3	6.7	10-16; 2-3	1
42				2	6.0	16-18; 1-9	2
41				1	6.8	29-30; 4-15	3
40			Lower 26.9	4	6.4	14-28	
39				3	6.2	2-13	
38				2	6.4	14-17	11
37				1	6.9	3-13	
36		Uragdan *ur* 75.9	Upper 37.4	6	7.6	25б-28	8
35				5	5.8	24д-25а	
34				4	6.6	22б-24г	
33				3	5.3	20б2-22а	
32				2	5.5	17-20б1	
31				1	6.6	15-16	
30			Lower 38.5	5	8.6	12-14	
29				4	8.4	8в-11	
28				3	6.2	6-8б	
27				2	7.7	1в-5	
26				1	7.6	1-4: 1а-б	8A
25	Kaybat *kb* 136.7	Omnutakh *om* 59.5	Upper 21.0	3	5.5	24-25	Z-834*
24				2	7.5	23	
23				1	8.0	21-22	
22			Mid. 21.5	2	10.5	19б-20	
21				1	11.0	18б-19а	
20			Lower 17.0	3	6.2	16-18а	
19				2	5.1	13б-15	
18				1	5.7	10-13а	
17		Talikit *tl* 28.6	Upp. 14.6	2	7.9	15-18	Z-835*
16				1	6.7	12б-14	
15			Low. 14.0	2	7.3	11б-12а	
14				1	6.7	10-11а	
13		Chamba *č* 48.6	Upper 21.8	5	5.0	9в	
12				4	4.0	9б	
11				3	5.0	9а	
10				2	4.3	8в	
9				1	3.5	8б	
8			Lower 26.8	8	8.0	7в-8а	
7				7	5.6	7б	
6				6	4.8	6б-7а	
5				5	4.4	6а	
4				4	2.2	5б	
3				3	1.1	5а	
2				2	0.7	4б	
1				1	0.5	4а	
0		Neruchanda					

Igarka District 635.0, II. 3 (4)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D		Tur				55	ShT-7
54	Pankagir *pn*	Postnich *ps* 97.2	24.3	8	13.4	?	ShT72-7A*
53				7	10.9		
52			37.6	6	13.4		
51				5	14.1		
50				4	10.1		
49			35.3	3	9.4		
48				2	13.4		
47				1	12.5	1; 1	T64-30
46	Mukon *mks* 199.3	Kongda *kn* 128.2	Upper 76.1	6	8.7	4-2	
45				5	13.5	19б-35;5	T64-31
44				4	13.0	18-19а	
43				3	12.2	16-17	
42				2	15.9	36-38; 14-15	ShT72-8
41				1	12.8	32-35	
40			Lower 52.2	4	12.1	23-31	
39				3	12.2	12-22	
38				2	12.3	63б; 1-11	
37				1	15.1	56-63а	ShT72-5
36		Mukte *mk* 71.1	Upper 36.5	6	5.4	55е	
35				5	6.5	55д	
34				4	6.0	55г	
33				3	6.5	55в	
32				2	6.5	55б	
31				1	5.6	54-55а	
30			Lower 34.6	5	7.8	53в	
29				4	7.6	53б	
28				3	5.9	51б-53а	
27				2	6.4	48б-51а	
26				1	6.9	45-48а	
25	Chatan *čt* 338.1	Tanimen *tn* 81.8	Upper 41.5	3	12.9	42б-44	
24				2	13.2	37-42а	
23				1	15.4	31-36	
22			Low. 40.3	2	21.6	18б-30	
21				1	18.7	10-18а	
20		Ugiyuk *ug* 132.1	Upper 56.7	3	24.4	97-114	14
19				2	15.1	93-96	
18				1	17.2	84-92	
17			Lower 75.4	4	18.2	69-83	
16				3	17.0	54-68	
15				2	21.0	25-53	
14				1	19.2	12-24	13
13		Chamba *č* 124.2	Upper 53.7	5	12.9	5в-11	
12				4	11.5	3б-5б	
11				3	9.4	2б-3а	
10				2	9.9	1б-2а	
9				1	10.0	7; 1а	15
8			Lower 70.5	8	17.7	5-6	
7				7	14.0	3-4	
6				6	12.0	2	
5				5	10.0	1б	
4				4	7.0	2; 1а	T73-1
3				3	7.0	21б-23;1	
2				2	2.0	21а	K74-43
1				1	0.8	20	
0		Zagorny					

FIGURE 6 — Silurian stratigraphy of the North Taymyr and North Priyenisey Subregions.

for the Dvoynaya River (Fig. 6). The hypostratotype is along two unnamed tributaries on the right bank of the Nizhnaya Taymyra River. These tributaries are 2 km and 0.4 km below the Middendorf Caves. The series has the Llandovery graptolites *Cyrtograptus vesiculosus, Pristiograptus cyphus, Demirastrites triangulatus, D. convolutus, Monograptus turriculatus,* and *Oktavites spiralis,* among others (Obut et al., 1965). It includes three formations.

The lowest (Zaputannaya Formation, Rhuddanian–Aeronian, abbreviated S_1zp) is named for the Zaputannaya River, the left tributary of the Trautfetter River (Tesakov et al., 1995, p. 126). The stratotype is on the right tributary of the Nizhnyaya Taymyra River, 2 km below the Middendorf Caves. It includes two subformations and comprises black mudstone with black and grey limestone interbeds. In the southern sections of the district, the thickness and number of carbonates increases considerably. The lower subformation is dominated by *Neodiplograptus modestus, Cystograptus vesiculosus,* and *Demirastrites triangulatus.* Siliciclastic lenses with *Isorthis neocrassa* and lenses of shaly limestone appear in the upper subformation. It is dominated by *Pristiograptus regularis* and *Monograptus elegans.*

The middle formation (Kosaya Formation, Telychian, abbreviated S_1ks) is named for the Kosaya River, the right tributary of the Shrenk River (Tesakov et al., 1995, p. 127). The stratotype is along the right tributary of the Nizhnyaya Taymyra River, 2 km below the Middendorf Caves. It has two subformations. The lower subformation is black mudstone with rare, dark-grey limestone interbeds. The upper subformation features black mudstone with a 3 m-thick unit of dark-grey limestone at the base. The fauna of this formation is dominated by *Monograptus spiralis, M. turriculatus, Monoclimacograptus crenulata, Stomatograptus grandis, Alispira gracilis, Isorthis neocrassa, Bystrowicrinus quinquelobatus,* and *Multisolenia tortuosa.*

MIDDENDORF LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}MD) — Noted by R. F. Sobolevskaya as a formation-rank (suite) unit and reported by Tesakov et al. (1995, p. 128), this series (Fig. 6) is named for the Middendorf Caves. The nearby stratotype is located on the right bank of the Nizhnyaya Taymyra River. It includes black and grey mudstone and limestone and is divided into three formations.

The lower (Golub' Formation, Wenlock, abbreviated S_1gl) (Tesakov et al., 1995, p. 129) is named for Lake Golub' at the source of Snezhnaya River. The stratotype is along the right bank of the Nizhnyaya Taymyra River near the Middendorf Caves. It has two subformations. The lower subformation is mainly dark-grey limestone and interbedded mudstone with the graptolites *Monograptus riccartonensis, M. flemingi,* and *Pristiograptus dubius,* among others. The upper subformation is made up of dark mudstone with intercalations of limestone that increase proportionally upward. The fauna includes *Cyrtograptus lundgreni, Testograptus testis,* and *Monograptus flemingi,* among others.

The middle formation in the Middendorf Local Series is the Privalnaya Formation (Gorstian, abbreviated S_2prv) (Tesakov et al., 1995, p. 129). It is named for the Privalnaya River, the right tributary of the Shrenk River. The stratotype is located on the right bank of the Nizhnyaya Taymyra River near the Middendorf Caves. The formation consists of black mudstone with rare intercalations of thin-bedded, platy, dark-grey limestone. It is subdivided into two subformations. The lower subformation is dominated by large limestone concretions. The fauna is dominated by *Bohemograptus bohemicus, Pristiograptus dubius, Saetograptus colonus,* and *Hemicosmorthoceras semiannulatum.*

The upper formation (Shara Formation, Ludford–Pridoli, abbreviated S_2sh) (Tesakov et al., 1995, p. 130) is named for the Shara River, a tributary of the Shrenk River. The stratotype is on the right bank of the Nizhnyaya Taymyra River near the Middendorf Caves. It includes three subformations, the lower of which is made up of medium-dark gray, laminated to massive limestones without a known fauna. The middle and upper subformations are represented by alternating black mudstone and medium-dark gray limestone of about the same thickness. The boundary between the subformations is drawn by the appearance of grey, clayey dolostone. Fossils are present only in the middle subformation. They include *Bohemograptus bohemicus, Linograptus posthumus, Monograptus uncinatus,* and *M. rarus.* Devonian grey limestones and dolostones overlie the upper subformation.

NORILSK STRATIGRAPHIC DISTRICT

Defined by Tesąkov et al. (1979, p. 12), this district is in the northwest Siberian Platform. It includes the Rybnaya River drainage basin and the middle and upper reaches of the Kureyka and Nizhnyaya Tunguska Rivers (Fig. 1). It includes the Kaybat (Chamba, Talikit, and Omnutakh Formations) and Priyenisey (Khyukta, Makus, and Postnich Formations) Local Series (Fig. 6).

KAYBAT LOCAL SERIES (LLANDOVERY, S_1KB) — First defined by Skobelin (1978, p. 155) as a formation (suite), the series' name comes from the word "kaybat," or "gravelly spit" in the local Keto language (Fig. 6). The series comprises dark-grey mudstone and medium-gray limestone. The lower formation (Chamba Formation, Rhuddanian–lower Aeronian, abbreviated S_1) (Tesakov et al., 1979, p. 70) is named for the Evenk-language word for "family," or "Chamba." The stratotype is on the Gorbiachin River. Well SP-21, drilled 8 km northeast of the town of Talnakh, is considered the hypostratotype. The Chamba includes two subformations. The lower subformation features black mudstones with rare, thin limestone and marl lenses. The fauna is dominated by *Glyptograptus tamariscus, Hedrograptus janischewskyi, Lagarograptus inexpeditus,* and *Pernerograptus praecursor.* The upper subformation is made up of grey mudstone and marl with thin limestone intercalations that are clustered into packets; *Clyptograptus tamariscus* and *Pernerograptus tenuipraecursor* are the dominant fossils. A Rhuddanian-early Aeronian age is demonstrated by the presence of *Pribylograptus sandersoni, Coronograptus gregarius, Demirastrites triangulatus, Clorinda undata,* and *Kulumbella kulumbensis.*

The middle formation in the Kaybat Local Series is the Talikit Formation (lower and middle middle Aeronian, abbreviated S_1tl). It was defined and named by N. N. Predtetchensky and Yu. I. Tesakov in 1979 (Tesakov, 1981b; Sokolov, ed., 1982, p. 13) for the Talikit River. The stratotype is a core from well SP-21, 8 km northeast of the town of Talnakh. The formation is subdivided into two subformations composed of gray marl and mudstone with gray nodular limestone interbeds. The fauna is dominated by *Favosites gothlandicus gothlandicus, Multisolenia tortuosa, Crassilasma crassiseptatum, Eocoelia hemisphaerica, Clorinda undata, Costaegera costata, Monograptus distans,* and *Pernerograptus revolutus.*

The third and upper formation in the Kaybat Local Series (Omnutakh Formation, uppermost middle Aeronian–Telychian, abbreviated S_1om) was first recognized by Yu. I. Tesakov in 1977 (Tesakov et al., 1979, table 1; Sokolov, 1982, p. 16) and named for the Omnutakh River. The stratotype is located on the Levyi Omnutakh River. It includes three subformations, and generally consists of gray and greenish-gray marl that alternates with gray, platy and nodular limestone. The fauna is dominated by *Favosites gothlandicus gothlandicus, Mesofavosites dualis, Cystiphyllum densum, Anabaria rara, Septatrypa magna, Morinorhynchus proprius, Beyrichia quadricornuta, Monograptus distans,* and *Streptograptus nodifer.* The lower subformation is uppermost middle–upper Aeronian; the middle and upper subformations are Telychian.

PRIYENISEY LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}PRN) — This series is proposed herein (Fig. 6). It corresponds to the Priyeniseyian Superhorizon (regional series), which was originally defined to group the Khakoma and Tukal Horizons (Tesakov et al., 1992, p. 81). The name is derived from the Yenisey River. The series is generally represented by gray, platy to nodular and biohermal limestone and dolostone with gray and variegated dolomitic marl. It is subdivided into three formations.

The lower of these formations (Khyukta Formation, Wenlock, abbreviated S_1khk) was defined by Yu. I. Tesakov in 1977 (Tesakov et al., 1979, table 1; Sokolov, 1982, p. 17) and named for the Khyukta Ridge. The stratotype is on the Levyi Omnutakh River. The formation includes two subformations and comprises gray, platy to nodular limestone, marl, and dolostone. Coral-stromatoporoid biostromes and bioherms occur throughout, and reef structures are widespread in the lower subformation. Dominant forms include *Labechia regularis, Clathrodictyon fastigiatum, Favosites gothlandicus moyeroensis, Sapporipora favositoides, Parastriatopora tebenjkovi, Yassia enormis, Bystrowicrinus bilobatus, Morinorhynchus proprius, Sibiritia kotelnyensis,* and *Pterospathodus amorphognathoides.*

The middle formation (Makus Formation, Gorstian, abbreviated S_2mkk) was defined by N. N. Predtetchensky and Yu. I. Tesakov in 1979 (Tesakov, 1981b, p. 55; Sokolov, ed., 1982, p. 18) and named for the Makus River. The stratotype is located on the Levyi Omnutakh River. The formation, which includes two subformations, is composed of gray marls with interbeds of gray, platy and nodular limestone and dolostone. Stromatolitic interbeds are common, especially in the lower subformation. The fauna is dominated by *Stromatopora malinovzyensis, Densastroma astroites, Favosites gothlandicus moyeroensis, Parastriatopora kureikiana, Hyattidina parva, Leperditia lumaea, Schrenkia multa,* and *Herrmannina nana.*

The upper formation (Postnich Formation, Ludford–Pridoli, abbreviated S_2ps) was defined by N. N. Predtetchensky and Yu. I. Tesakov in 1979 (Tesakov, 1981b, p. 55; Sokolov, ed., 1982, p. 19), and named for Postnichny Creek, the right tributary of the Imangda River. The stratotype is a core from well TT-1 on the east shore of Lake Pyasino. The formation includes three subformations, and is generally composed of gray dolomitic marl with interbeds of grey dolostone and anhydrite. The remains of brachiopods and ostracodes are poorly preserved. The lower subformation is given a preliminarily assignment to the Ludfordian, with the middle and upper subformations referred to the Pridoli. It is overlain by Lower Devonian, platy, gray dolostone.

TURUKHANSK STRATIGRAPHIC DISTRICT

Named by Tesakov et al. (1979, p. 14), this district is in the western Siberian Platform in the lower reaches of the Kureyka and Nizhnyaya Tunguska Rivers (Fig. 1). Three local series, Kaybat, Diavolskaya, and lower Nimda, are exposed in this district (Fig. 6).

KAYBAT LOCAL SERIES (LLANDOVERY, S_1KB) — This interval (with the Chamba, Talikit, and Omnutakh Formations) extends across the Turukhansk and Norilsk Districts. The series is detailed above under the description of the Norilsk District. The thicknesses of the series and its formations in the Turukhansk District (Fig. 6) are much thinner than in the Igarka and Norilsk Districts.

The Chamba Formation (Rhuddanian–lower Aeron, abbreviated S_1r) has a stratotype on the Gorbiachin River. Its hypostratotype in the Turukhansk District is near the mouth of the left bank of the Tenna-Ses' River, a tributary of the Letnyaya River. The formation is discussed above in the description of the Norilsk District.

The overlying Talikit Formation (lower–middle middle Aeronian, abbreviated S_1tl) has as a stratotype the core of well SP-21. The hypostratotype is in the Turukhansk District near the mouth of the Tenna-Ses' River. The formation is described above in the discription of the Norilsk District.

The highest unit of this local series (Omnutakh Formation, uppermost middle Aeronian–Telychian, abbreviated S_1om) has a stratotype along the Levyi Omnutakh River. The hypostratotype is in the Turukhansk District on the right bank of the Ales-Ses' River (at the river mouth), a tributary of the Letnyaya River. The formation is discussed above in the description of the Norilsk District.

DIAVOLSKAYA LOCAL SERIES (WENLOCK–GORSTIAN, S_{1-2}DVL) — This series (Kirichenko (1940, p. 60; Fig. 6) was first defined as a formation (suite). It is named after the Diavolskaya River in the Sukhaya Tunguska River drainage. The stratotype has not been designated. The series features medium-dark-gray limestone and dolostone with occasional interbeds of gray marl in the upper part. It is subdivided into two formations.

The lower is the Uragdan Formation (Wenlock, abbreviated S_1ur) (Tesakov et al., 1980, p. 81), with a name derived from the Evenk word "uragdan." The stratotype is on the Kureyka River, 6 km above the mouth of the Pelyadka River. The formation is divided into two subformations that consist of medium-dark-gray limestone and biostromes with rare interbeds of nodular to laminated, platy limestone and lower reefs. The fauna is dominated by *Labechia condensa, Neobeatricea nikiforovae, Clathrodictyon boreale, Sapporipora favositoides, Mesosolenia festiva, Parastriatopora tebenjkovi, Cystiphyllum siluriense*, and *Plectatrypa wenlockiana.*

The upper is the Tukal Formation (Gorstian, abbreviated S_2tk) (Tesakov et al., 1980, p. 83), named after the Evenk family "Tukal." The stratotype is along the Kureyka River near Verkhniye Shcheki (Cheeks). The formation, with two subformations, is made up of gray, oolitic to stromatolitic limestone, gray marl, and rarer nodular limestone. The lower subformation is mainly stromatolitic. The fauna is dominated by *Clathrodictyon mohicanum, Stromatopora dnestriensis, Leperditia lumaea, Eohowellella minimus, Hyattidina parva, Morinorhynchus proprius*, and *Favosites gothlandicus moyeroensis.* The upper subformation is dominated by *Schrenkia multa* and *Parastriatopora kureikiana.*

NIMDA LOCAL SERIES (LUDFORDIAN–LOWER LOCHKOV, S_2-D_1NM) — This series was first defined as a formation (suite) by Melnikov (1979, p. 11), and is named for the Nimda River (Fig. 6). The Silurian part (Ludfordian–Pridoli) of the series is the Postnich Formation (abbreviated S2ps), which has the core of well TT-1 as the stratotype. The hypostratotype for the Turukhansk District is well KhL-8, located at the confluence of the Nimda and Muchema rivers in the Nizhnyaya Tunguska River drainage basin. The formation is discussed above in the description of the Norilsk District.

IGARKA STRATIGRAPHIC DISTRICT

Named by Tesakov et al. (1979, p. 14), the Igarka Stratigraphic District lies on the western Siberian platform in the lower drainage basin of the Khantayka, Kulumbe, and Gorbiachin Rivers (Fig. 1). It includes the Chatan, Mukon, and lower Pankagir Local Series (Fig. 6).

CHATAN LOCAL SERIES (LLANDOVERY), S_1CT — Proposed herein, the name of this local series is derived from the first syllables of the Even families, "Chamba and Tanimen." It consists of clay-carbonate rocks that may be subdivided into three formations (Fig. 6).

The lower formation (Chamba Formation, Rhuddanian–lower Aeronian, abbreviated S_1ch) has a stratotype along the Gorbiachin River. The formation is discussed above in the description of the Norilsk District.

The middle formation (Ugiyuk Formation, middle–upper Aeronianian, abbreviated S_1ug) (Tesakov et al., 1979, pp. 59, 72; 1980, p. 80) is named after the old Evenk lineage "Ugiyuk." The stratotype is on the Gorbiachin River 1 km below Olen' Creek. The formation has two subformations that consist of alternating beds of gray marl and nodular or laminated limestone. The fauna is dominated by *Favosites gothlandicus gothlandicus, Subalveolites volutus, Pentamerus oblongus, Kulumbella kulumbensis, Clorinda undata, Monograptus sedgwicki*, and *Pterospathodus celloni.*

The upper formation (Tanimen Formation, Telychian, abbreviated S_1tn) (Tesakov et al., 1979, p. 61; Sokolov, 1982, p. 48) is named after the old Evenk lineage "Tanimen." The stratotype is on the Kulyumbe River near the mouth of Nadporozhny Creek. The formation has two subformations. The lower consists of variegated marl, clayey dolostone, and gray nodular and platy limestone. The upper subformation consists of gray limestone and marl with nodular limestone and breccia lenses. The fauna is dominated by *Favosites gothlandicus gothlandicus, F. gothlandicus moyeroensis, Daleiella ariadnae, Morinorhynchus proprius, Protatrypa lepidota,* and *Cytherellina oviformis.*

MUKON LOCAL SERIES (WENLOCK–GORSTIAN, S_{1-2}MKS) — Proposed herein (Fig. 6), the name of this unit is derived from the first syllables of the Evenk Mukte Tribe and the Kongda River. The stratotype is located on the Kulyumbe River. The Mukte Formation (Wenlock, abbreviated S_1mk) (Tesakov et al., 1979, p. 65; Sokolov, 1982, p. 54) is named for the Mukte Tribe of the Evenk people. The stratotype is on the Kulyumbe River, 0.5 km above the mouth of Nadporozhny Creek. The formation has two subformations. The lower features beds and lenses of gray peloidal limestone. The upper subformation includes gray and cream-colored, fine-grained, nodular and burrowed limestone. *Favosites gothlandicus moyeroensis* and *Hyattidina parva* dominate.

The overlying Kongda Formation (Gorstian, abbreviated S_2kn) (Tesakov et al., 1979, p. 65; Sokolov, 1982, p. 55) is named for the Evenk word "kongda," or "bend." The stratotype is on the Kulyumbe River, 1 km above Nadporozhny Creek. The formation has two subformations and includes medium-dark-gray, argillaceous, peloidal limestone, dolostone, and marl. Stromatolites, sedimentary breccias, coquinite, nodular limestone, and gypsum are frequent. The biota is dominated by *Morinorhynchus proprius, Favosites gothlandicus moyeroensis, Parastriatopora kureikiana, Eukloedenella kureikensis, Schrenkia multa, Howellella elevataeformis,* and stromatolites.

PANKAGIR LOCAL SERIES (LUDFORDIAN–LOWER LOCHKOVIAN, S_2–D_1PN) — Proposed by Tesakov et al. (1979, p. 67; Sokolov, 1982, p. 61), this series (Fig. 6) is named after the Evenk lineage "Pankagir." The stratotype is on the Kulyumbe River, ca. 2 km above Nadporozhny Creek and 1 km above Turkut Creek. The Silurian part of this local series is the Postnich Formation.

The Postnich Formation (Ludford–Pridoli, abbreviated S2ps) is detailed above in the description of the Norilsk District. In the Igarka District, this formation is poorly exposed and is mainly represented by dolomitic marl and laminated dolostone. The overlying Lower Devonian consists of thick-bedded, platy limestone with *Tiverina vermiculata* and *Favosites kozlowskii.*

KOCHUMDEK STRATIGRAPHIC DISTRICT

Located along the northern Yenisey Ridge, this area was first termed the "Tunguska District" (Tesakov et al., 1979, p. 14). It is now named for the Kochumdek River, a tributary of the Podkamennaya Tunguska River, and includes the drainage basins of the Podkamennaya Tunguska and Bakhta Rivers (Fig. 1). The district includes the Koraz and Uskhur Local Series (Fig. 7).

KORAZ LOCAL SERIES (LLANDOVERY, S_1KR) — Proposed herein (Fig. 7), the name of this interval is based on the first syllables of the Kochumdek and Razvilkar Rivers. The series is largely composed of gray marl with nodular limestone, as well as nodular and wavy-bedded, gray limestone that becomes fossiliferous upward. It is subdivided into three formations.

The lower formation (Kochumdek Formation, Rhuddanian–lower Aeronian, abbreviated S_1kc) was first recognized by L. F. Lungersghauzen (Nikiforova, 1955, p. 83), and named for the Kochumdek River. The parastratotype is on the left bank of the Kulinna River, 0.3 km above the mouth of Bolotnyi Creek. The formation has two subformations and consists of gray nodular limestone with interbeds of grey marl. The marls increase in abundance toward the base and top of the formation, particularly in the upper subformation. The fauna includes *Alispira gracilis, Brevilamnulella undatiformis, Isorthis neocrassa,* and *Calamopora alveolaris.*

The middle formation (Kulinna Formation, middle–upper Aeronian, abbreviated S_1kl) was defined by N. N. Predtetchensky, Yu. I. Tesakov, V. G. Khromych, A.Ya. Berger, et al. in 1980 (Tesakov, 1981b, p. 54, 55). The stratotype is on the Kulinna River and extends about 15.6–23.3 km in a straight line from the mouth. The formation is subdivided into two subformations and consists of variegated mudstone and marl with interbeds of gray, nodular and wavy-bedded limestone. The fauna is dominated by *Mesofavosites dualis, Favosites gothlandicus, Pentamerus oblongus, Eocoelia hemisphaerica,* and *Isorthis neocrassa.*

The overlying Razvilka Formation (Telychian, abbreviated S_1rz), defined by N. N. Predtetchensky, Yu. I. Tesakov, et al. in 1980 (Tesakov, 1981b, p. 54, 55), was named for the Nizhnyaya Razvilka River, a tributary of the Kulinna River. The stratotype is on the Kulinna River, 23.3 km in a straight line from the mouth (site 103), and on the Podkamennaya Tunguska River, near the confluence with the Lebyazhiya River (site 112). The formation includes two subformations. The lower is made up of green marl with nodules and lenses of gray limestone. The upper subformation consists of gray limestone, dolostone, and intercalated marl with nodular limestone. The dominant fossils are *Labechia venusta, Mesofavosites dualis, Crotalocrinites borealis, Beyrichia patagium, Eocoelia hemisphaerica,* and *Pentamerus oblongus.*

USKHUR LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}UH) — Proposed herein (Fig. 7), the name of this local series consists of the first syllables of the names of the Usas and Khurichi Rivers. The series is a transgressive cycle of

Local stratigraphic units for the Silurian of East Siberia

Pritunguska Subregion III

Kochumdek District 230.0, III. 1 (5)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, *m*	Bed (particular link)	Stratotype
D	?					80a	
54	Uskhur *uh* 115.8	Khurichi *khr* 54.1	Upper 19.1	5	4.7	78б-79	BT-8
53				4	3.2	77ж-78a	
52				3	3.7	77e	
51				2	3.8	77д	
50				1	3.7	77г	
49			Middle 12.2	3	3.1	77в	
48				2	4.1	77б	
47				1	5.0	77a	
46			Lower 22.8	6	2.2	76в	
45				5	2.5	76б	
44				4	3.7	76a	
43				3	3.9	75в	
42				2	4.5	75б	
41				1	6.0	75a	
40		Usas *us* 61.7	Upper 19.0	4	5.5	72-74	
39				3	5.0	70-71	
38				2	4.3	69б	
37				1	4.2	68-69a	
36			Middle 19.9	6	3.8	27-29	106
35				5	3.6	25-26	
34				4	2.9	23-24	
33				3	3.1	20-22	
32				2	3.4	15-19	
31				1	3.1	13-14	
30			Lower 22.8	5	5.8	9-12	112
29				4	4.5	34-37a	
28				3	3.6	29б-33	
27				2	4.0	26б-29a	
26				1	4.9	23-26a	
25	Koraz *kr* 113.5	Razvilka *rz* 30.5	Upper 10.4	3	2.7	20-22	
24				2	3.3	16-19	
23				1	4.4	9-15	
22			Low. 20.1	2	10.6	16-8	
21				1	9.5	10-12б	103
20		Kulinna *kl* 46.0	Upper 15.3	3	5.4	8-9	
19				2	5.0	4б-7	
18				1	4.9	3-4a	
17			Lower 30.7	4	7.0	2; 1-2	
16				3	7.5	16; 1	105
15				2	6.3	15	109
14				1	9.9	14	
13		Kochumdek *kč* 37.0	Upper 17.6	5	4.8	13	109*
12				4	3.1	12б	
11				3	3.2	12a	
10				2	3.5	11б	
9				1	3.0	11a	
8			Lower 19.4	8	5.0	9-10	
7				7	5.0	7-8	
6				6	2.7	6б	
5				5	2.8	6a	
4				4	2.4	5	
3				3	0.5	4в	
2				2	0.5	4б	
1				1	0.5	4a	
0	Bur						

Vorogovo District 300.0, III. 2 (6)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, *m*	Bed (particular link)	Stratotype
D	?						
54		Khurichi *khr* ~ 80.0	11.6	5	7.0		Vorogovka R. (PGS 64-5) and Glotikha R. (T61-57-62) sections
53				4	4.6		
52			16.6	3	5.5		
51				2	5.6		
50				1	5.5		
49			18.1	3	4.7		
48				2	6.0		
47	—			1	7.4	?	
46			33.7	6	3.3		
45				5	3.7		
44				4	5.5		
43				3	5.6		
42				2	6.7		
41				1	8.9		
40		Toyba *tbk* ~ 85.0	22.4	18	6.5		
39				17	6.0		
38				16	5.0		
37				15	4.9		
36			23.5	14	4.5		
35				13	4.2		
34				12	3.7		
33				11	3.4		
32				10	4.0	?	
31				9	3.7		
30			26.8	8	6.8		
29				7	5.3		
28				6	4.2		
27				5	4.7		
26				4	5.8		
25			12.3	3	3.2		
24				2	3.9		
23				1	5.2		
22		Tugulan *tg* ~ 135.0	26.2	22	13.8		
21				21	12.4		
20			19.8	20	7.0		
19				19	6.5		
18				18	6.3		
17			41.0	17	9.2		
16				16	9.8		
15				15	9.6		
14				14	12.4		
13			23.0	13	6.3		
12				12	4.0		
11				11	4.2	?	
10				10	4.6		
9				9	3.9		
8			25.0	8	6.5		
7				7	6.5		
6				6	3.4		
5				5	3.6		
4				4	3.0		
3				3	0.7		
2				2	0.7		
1				1	0.6		
0	Limestone						

South Taymyr District 490.0, III. 3 (7)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, *m*	Bed (particular link)	Stratotype
D	Dolostone					11	
54		Bunge *bn* 121.4	Upper 77.7	8	10.5	8-10	210*
53				7	8.0	5б-7	
52				6	11.0	4-5a	
51				5	10.2	2-3	
50				4	8.0	4б	209*
49				3	8.0	45б	
48				2	11.5	44б-45a	
47	—			1	10.5	43-44a	
46			Middle 19.5	6	6.5	42б	
45				5	4.0	42a	
44				4	7.5	40в-41	
43				3	7.0	40б	
42				2	5.8	38б2-40a	
41				1	6.7	14; 38б1	
40			Lower 24.2	4	5.3	13б	211*
39				3	6.0	11б-13a	
38				2	5.8	6б-11a	
37				1	7.1	4-6a	
36	Andrey *an* 369.2	Trubka *tb* 129.8	Upper 39.6	6	8.1	28б-33	209
35				5	8.2	26в-28a	
34				4	4.5	26б	
33				3	5.0	25б-26a	
32				2	6.0	23-25a	
31				1	7.8	19-22	
30			Middle 39.2	5	8.5	17-18	
29				4	8.0	14-16	
28				3	6.5	12-13	
27				2	8.0	70-72	205
26				1	8.2	67-69	
25			Lower 51.0	3	12.0	66	
24				2	16.0	59б-65	
23				1	23.0	52-59a	
22		Tari *tr* 38.5		2	19.1	14-17;50-51	208
21				1	19.4	2-13	
20		Karaku *kk* 97.1	Upper 36.7	3	14.5	37-38	205
19				2	10.2	33-36	
18				1	12.0	32	
17			Lower 60.4	4	16.0	30-31	
16				3	14.0	29	
15				2	14.0	24-28	
14				1	16.4	20-23	
13		Ust'pryamaya *upm* 103.8	Upper 49.8	5	11.3	1г	222
12				4	10.0	1в	
11				3	9.0	1б	
10				2	9.5	1a	
9				1	10.0	5	227
8			Lower 54.0	8	14.0	4б	
7				7	12.0	3г-4a	
6				6	9.0	3в	
5				5	7.0	3б	
4				4	4.8	2г-3a	
3				3	5.0	2в	
2				2	1.7	2б	
1				1	0.5	2a	
0	Povorotnaya						

Ledyanka District 276.7, III. 4 (8)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, *m*	Bed (particular link)	Stratotype
D	Dolostone					93	
54	Mukir *mkr* 109.8	Kira *kr* 29.5	Upp. 7.5	2	4.2	92	L-3
53				1	3.3	91	
52			Middle 11.5	3	4.1	90	
51				2	4.3	88-89	
50				1	3.1	86-87	
49			Lower 10.5	3	3.0	85	
48				2	4.0	83-84	
47				1	3.5	82	
46		Nerakachi *nr* 50.8	Upper 30.0	6	5.7	80-81	
45				5	3.8	79	
44				4	5.9	78	
43				3	5.0	77	
42				2	4.5	76	
41				1	5.1	73-75	
40			Lower 20.8	4	5.0	71-72	
39				3	5.2	69-70	
38				2	5.2	67-68	
37				1	5.4	66	
36		Munil *mn* 29.5	Upper 14.2	6	2.8	65	
35				5	2.9	64	
34				4	1.7	63	
33				3	1.8	62	
32				2	2.2	61	
31				1	2.8	60	
30			Lower 15.3	5	3.5	59	
29				4	3.1	58	
28				3	2.3	57	
27				2	3.0	56	
26				1	3.4	55	
25	Okhuk *okh* 167.0	Khukelchi *khl* 78.5	Upper 26.9	3	8.5	53-54	
24				2	8.8	51-52	
23				1	9.6	48-50	
22			Mid. 27.4	2	12.4	45-47	
21				1	15.0	40-44	
20			Lower 24.2	3	9.0	37-39	
19				2	7.3	36	
18				1	7.9	33-35	
17		Oran *orn* 94.4	Upper 44.8	4	10.3	27-32	
16				3	11.4	22-26	
15				2	10.6	19-21	
14				1	12.5	16-18	
13			Middle 28.0	5	6.3	14-15	
12				4	5.5	13	
11				3	4.9	12	
10				2	5.1	11	
9				1	6.2	9-10	
8			Lower 21.6	8	5.1	8	
7				7	4.5	7	
6				6	3.4	6	
5				5	3.5	5	
4				4	2.1	4	
3				3	1.7	3	
2				2	0.8	2	
1				1	0.5	1	
0	Kayyalak						

FIGURE 7 — Silurian stratigraphy of the Pritunguska Subregion. Figure continued on Fig. 8.

rocks that vary from gray, biostromal limestone to green and variegated dolomitic marl.

The lower formation (Usas Formation, Wenlock–owest Gorstian, abbreviated S_{1-2}us) was recognized by N. N. Predtetchensky, Yu. I. Tesakov, et al. in 1980 (Tesakov, 1981b, p. 54, 55) and named for the Usas River. The stratotypes are on the Podkamennaya Tunguska River, near the confluence with the Lebyazhiya River (site 112); on the Kulinna River above the mouth of the Usas River; and the core from well BT-8 in the Bakhta River drainage basin. The formation is subdivided into three subformations. The lower parts of the lower and middle subformations are composed of gray dolomitized limestone with biostromes. The middle and upper parts of these subformations are alternations of gray, massive, stromatolitic limestone, greenish-gray mudstone and marl, and interbeds of limestone conglomerate. The dominant fossils include *Labechia condensa, Ecclimadictyon fastigiatum, Mesosapporipora porosa, Favosites gothlandicus moyeroensis*, and *Beyrichia mirabilis*. The upper subformation is composed of gray, massive dolostone with stromatolites.

The overlying Khurichi Formation (uppermost Gorstian–Pridoli, abbreviated S_2khr) is proposed herein. It is named for the Khurichi River in the larger drainage basin of the Bakhta River. The stratotype is the core from well BT-8 in the Bakhta River basin. This formation includes three subformations. The lower and upper subformations at the stratotype are composed of gray, green, and variegated dolomitic, gypsum-bearing marl with thin interbeds of gray dolostones. The middle subformation is made up of grey, thin-bedded, platy dolostone interstratified in its middle part with gypsum-bearing dolomitic marl. Marls occur at the top of the Khurichi Formation, and it is overlain by the Devonian.

Vorogrovo Stratigraphic District

Proposed herein, the Vorogovo stratigraphic district is on the northern Yenisey Ridge and includes the drainage basins of the Vorogovka and Glotikha Rivers. The Tugulan and Toiba Beds and Khurichi Formation are characteristic of this district (Fig. 7).

The Tugulan Beds (Rhuddanian–lower middle Telychian, abbreviated S_1tg) (Tesakov et al., 1996b) are named for the Tugulan River, a tributary of the Yenisey River. A stratotype has not been selected. Isolated sections are located on the Vorogovka River (Podgornaya et al., 1965) and Glotikha River (unpublished field data of Yu. I. Tesakov). The Tulugan Beds include gray marl with nodular limestone. The proportion of nodular and lensing limestone increases upwards. The fauna is dominated by *Alispira gracilis, Eocoelia hemisphaerica, Calamopora alveolaris, Favosites gothlandicus gothlandicus*, and *Multisolenia tortuosa*.

The overlying Toyba Beds (upper middle Telychian–lowermost Gorstian, abbreviated S_{1-2}tbk) (Tesakov et al., 1996b) are named for the Toyba River, a tributary of the Yenisey River. A stratotype has not been selected. The Toyba Beds are made up of gray, platy dolostone and limestone that locally alternate with dolomitic marl with interbeds of conglomeratic limestone. The fossils have not been studied.

The highest unit in this regional series is the Khurichi Formation (uppermost Gorstian–Pridoli, abbreviated S_2khr). It extends from the Kochumdek District into the Vorogovo District (see description of the Kochumdek District).

South Taymyr Stratigraphic District

Earlier termed the "Southern Facial Zone" (Zlobin, 1962; Resolutions, 1983), this district lies in the central Taymyr Peninsula (see Fig. 1). It includes the Andrey Local Series and Bunge Formation (local series) (Fig. 7).

Andrey Local Series (Lower Silurian, S_1an) — This unit was named by M. N. Zlobin (Zhizhina, 1959, p. 155) for Andrey Island, east of the Taymyr Peninsula (Fig. 7). The type sections are in the Nyun'karaku-Tari River drainage basin. The series is composed of gray limestone and dolostone and includes four formations.

The lowest (Ust'pryamaya Formation, Rhuddanian–lower Aeronian, abbreviated S1upm) (Tesakov et al., 1995, p. 131) is named for the Pryamaya River. The stratotype is along the Pryamaya River and near its mouth. The lower subformation comprises alternating black, medium-bedded, platy limestone and siliciclastic mudstone. The upper subformation is composed of black and grey, wavy-bedded limestone with thin interbeds of black chert.

The overlying Karaku Formation (middle–upper Aeronian, abbreviated S_1kk) (Tesakov et al., 1995, p. 132) is named for exposures along the middle Nyun'karaku-Tari River. The stratotype is on the Parnaya River near the mouth of Dvukhvershynnyi Creek. The formation, with two subformations, consists of gray, thick-bedded, platy limestone with abundant corals, pentamerid beds, and chert. Dominant fossils include *Clathrodictyon regulare, Favosites gothlandicus gothlandicus, Multisolenia tortuosa, Tungussophyllum densum, Pentamerus oblongus, Isorthis neocrassa*, and *Icriodella discreta*.

The Tari Formation (lower–lower middle Telychian, abbreviated S_1tr) (Tesakov et al., 1995, p. 134) is named after the Nyun'karaku-Tari River. The stratotype is along the Parnaya River, near the mouth of Dvukhvershynnyi Creek. The formation is divided into two subformations. The lower consists of greenish-gray and gray platy and laminated dolostone. The upper subformation is dominated by gray replacement dolostone with platy and nodular limestone and nodular dolomitized limestone. Dominant taxa include *Clathrodictyon regulare, Favosites gothlandicus gothlandicus, Subalveolites volutus, Entelophyllum articulatum, Sibiritia norilskensis,* and *Borealis borealis.*

The next overlying unit (Trubka Formation, upper middle Telychian–Wenlock, abbreviated S_1tb) (Tesakov et al., 1995, p. 136) is named for the Trubka River, a tributary of the Nyun'karaku-Tari River. The stratotype is along the Parnaya River, near the confluence with Dvukhvershynnyi Creek, 0.4 km from the river's mouth. The formation has three subformations. The lower subformation consists of gray and brown-gray dolostone with lower coral-stromatoporoid biostromes. The middle subformation is composed of alternating gray replacement dolostone and biostromes, and the upper subformation includes brown-gray replacement dolostone with local biostromes at the top. The fauna is dominated by *Clavidictyon cylindricum, Ecclimadictyon fastigiatum, Labechia conferta, Yavorskiina aspectabilis, Favosites gothlandicus moyeroensis, Mesosolenia festiva, Parastriatopora tebenjkovi,* and *Sapporipora favositoides.*

The Bunge Formation (Ludlow–Pridoli, abbreviated S_2bn) was named by M. N. Zlobin (Zhizhina, 1959) for the Bunge River. The stratotype is on the right bank of the Nizhnyaya Taymyra River, 5–10 km above the mouth of the Bunge River. The hypostratotypes are on the left bank, 100–400 m from the mouth of the Parnaya River, on the right bank near its mouth. The formation has three subformations. The lower comprises brown-gray and gray replacement dolostone with rare stromatoporoids and tabulates. The middle subformation is composed of alternating black and gray platy dolostone and limestone with stromatolites. The upper subformation consists of alternating gray, thin- to medium-bedded, platy dolostone without macrofossils. The overlying Devonian is brownish-grey, replacement dolostone.

Ledyanka Stratigraphic District

Proposed by Tesakov et al., (1996b), this district includes the upper Khatanga River drainage basin (Fig. 1), and is is named for the Ledyanka geological structure where the Okhuk and Mukir Local Series are exposed. The local series and formations in the Ledyanka District (Fig. 7) were proposed by Yu. I. Tesakov and T. A. Divina in 1994 based on cores from wells L-3, L-2, and L-358 near the confluence of the Ayan and Ayakli Rivers.

Okhuk Local Series (Llandovery S_1okh) — The name for this local series (Fig. 7) is based on the first syllables in the names of the Oran and Khukelchi Rivers. The stratotype is the core from well L-3. The series is composed of nodular limestone at the bottom and marl with nodular limestone at the top. It includes two formations.

The lower (Oran Formation, Rhuddanian–lower middle Aeronian, abbreviated S1orn) is named for the Oran River. The stratotype is a core from well L-3. The formation consists of grey nodular limestone with interbeds of grey marl in the uppermost part. It is subdivided into three subformations. The lower subformation is dominated by *Mesofavosites dualis* and *Calamopora alveolaris;* the middle subformation by *Septatrypa magna, Cryptothyrella lacrima,* and *Leptostrophia talikitensis;* and the upper by *Cystihalysites mirabilis, Favosites gothlandicus gothlandicus,* and *Pentamerus oblongus.*

The upper formation (Khukelchi Formation, upper middle Aeronian–Telychian, abbreviated S_1khl) is named for the Khukelchi River. The stratotype is a core from well L-3. The formation has three subformations, and consists of green marl with isolated, irregular gray limestone nodules and lenses of fossiliferous limestone with intraclasts. The lower and middle subformations have interbeds of grey nodular limestone, and the upper subformation is comprised of rather thick units of massive limestone with corals and stromatoporoids. The lower subformation is dominated by *Eocoelia hemisphaerica, Daleiella decorata, Leptostrophia talikitensis,* and *Pentamerus oblongus;* the middle subformation by *Daleiella ariadnae and Cytherellina oviformis;* and the upper one by *Hyattidina parva, Multisolenia tortuosa, Plümatolinia densa,* and *Favosites gothlandicus moyeroensis.*

Mukir Local Series (Wenlock–Pridoli, S_{1-2}mkr) — The name of this local series (Fig. 7) is composed of the first syllables of the names of the Munil and Kira Rivers. The stratotype is the core of well L-3. In general, this series is composed of lower coral and stromatoporoid limestone, middle stromatolitic dolostone, and upper laminated dolostone. It has three formations.

The lower formation (Munil Formation, Wenlock, abbreviated S1mn) is named for the Munil River. The stratotype is the core of well L-3. The formation, with two subformations, consists of gray, medium- and thick-bedded, platy, dolomitic limestone with numerous coral-stromatoporoid biostromes and bioherms. The fauna is dominated by *Ecclimadictyon fastigiatum, Favosites gothlandicus moyeroensis, Mesosolenia festiva, Sapporipora favositoides, Subalveolites subulosus, Parastriatopora tebenjkovi, Labechia condensa,* and *Clavidictyon cylindricum.*

The middle formation (Nerakachi Formation, Gorstian, abbreviated S_2nr) is named for the Nerakachi River.

The stratotype is the core of well L-3. The formation, with two subformations, consists of grey, thin-bedded, platy and laminated dolostone with stromatolites, coral-stromatoporoid biostromes, and dolostone lenses with intraclast breccia. The coral-stromatoporoid biostromes are characteristic only of the lower subformation, and the breccias only of the upper. The dominant fossils include *Favosites gothlandicus moyeroensis, Sapporipora favositoides, Parastriatopora tebenjkovi, Plectatrypa wenlokiana*, and *Strophodonta omnutakhensis.*

The upper formation (Kira Formation, Ludfordian–Pridoli, abbreviated S_2kr) is named for the Kira River. The stratotype is the core of well L-3. The formation, with three subformations, consists of gray, dolomitic marl and thin-bedded, platy, laminated dolostone, frequently with gypsum. The lower and upper subformations are marl, but the middle consists of dolostone. In the north part of the district, this formation is largely dolostone. Only recrystallized brachiopod valves have been found. The overlying massive gray Devonian dolostones contain *Tiverina vermiculata.*

MAYMECHA STRATIGRAPHIC DISTRICT

Proposed by Tesakov et al. (1979, p. 14), this district lies in the drainage basin of the Maymecha River and the lower reaches of the Kotuy River (Fig. 1). The district embraces the Akher and Dokir Local Series. The regional series and formations of the district (Fig. 8) were defined by Yu. I. Tesakov, N. N. Predtetchensky, et al., in 1986 (Tesakov et al., 1990, p. 72) with stratotypes in well Kh-36. This well was drilled on the left bank of the Pravyi Atyrdyak River, 0.7 km to the northeast of site 512.

AKHER LOCAL SERIES (LLANDOVERY, S_1AKH) — Proposed herein (Fig. 8), the name of this local series derives from the first syllables of the names of the Atyrdyak and Kherkimi Rivers. The series is generally composed of gray, laminated, nodular, wavy-bedded limestone and, at the top, gray marl and dolostone. It is subdivided into three formations.

The lower formation (Pravoatyrdyak Formation, Rhuddanian–lower Aeronian, abbreviated S_1pv) is named for the Pravyi Atyrdyak River. The formation, with two subformations, consists of cream-gray, laminated limestone (lower subformation) and gray, wavy-bedded limestone with marly intercalations (upper subformation). The fauna is dominated by *Alispira gracilis, Isorthis neocrassa, Distomodus kentuckyensis, Calamopora alveolaris*, and *Multisolenia tortuosa.*

The middle formation (Orachi Formation, middle–upper Aeronian, abbreviated S_1or) is named for the Orachi River. The formation, with two subformations, consists of alternating brownish- and greenish-gray, wavy-bedded and nodular limestone units and greenish-gray marls with abundant nodular limestone. In the east part of the district, the upper subformation is dominated by marls with interbeds of limestone intercalated with intraclast beds. The fauna is dominated by *Cystihalysites mirabilis, Favosites gothlandicus gothlandicus, Multisolenia tortuosa, Entelophyllum articulatum, Streptelasma whittardi, Panderodus unicostatus, Clathrodictyon boreale, Labechia venusta, Pentamerus oblongus*, and *Daleiella decorata.*

The upper formation (Kherkimi Formation, Telychian, abbreviated S_1hrk) is named for the Kherkimi River. The formation is divided into two subformations. The lower subformation is composed of gray, platy and nodular limestone and greenish-gray marl with nodular limestone; the upper subformation is made up of gray, thick-bedded, platy, locally laminated limestone and dolostone with numerous stromatoporoids and tabulates. The lower subformation fauna is dominated by *Mesofavosites dualis, Subalveolites volutus, Entelophyllum articulatum, Alispira tenuicostata*, and *Thrallella vermiformis.* The dominant fossils in the upper subformation are *Favosites gothlandicus moyeroensis, Mesosolenia festiva, Subalveolites subulosus, Yavorskiina aspectabilis, Alispira tenuicostata, Distomodus staurognathoides*, and *Imangdites ambiquus.*

DOKIR LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}DK) — Proposed herein, the name of this local series (Fig. 8) is derived from the first syllables of the names of the Dolgotnaya and Kira Rivers. The series is commonly composed of lower replacement dolostone and upper dolomitic marl that alternate with dolostone. It is divided into three formations.

The lower formation (Dolgotnaya Formation, Wenlock, abbreviated S1dl) is named for the Dolgotnaya River. The formation, with two subformations, consists of medium-light grey, thick-bedded, platy replacement dolostone with local biostromes. The fauna is dominated by *Labechia condensa, Ecclimadictyon fastigiatum*, and *Favosites gothlandicus moyeroensis.*

The middle formation (Bakhanay Formation, Gorstian, abbreviated S_2bh) is named for the Bakhanay River. It is subdivided into two subformations. The lower subformation is composed of gray (with rare red interbeds) dolomitic marl and light-gray, thick-bedded, platy dolostone. The upper subformation consists of variegated dolomitic marl with thin, rare interbeds of grey dolostone. In the lower subformation, the dominant fossils are *Hyattidina acutisummitatus, Morinorhynchus proprius, Cyathactis tenuiseptatus*, and *Straparollus alacer.*

The upper formation (Kira Formation, Ludfordian–Pridoli, abbreviated S_2kr) occurs in the western part of the Maymecha District, but extends into the Ledyanka District. In some areas of the Maymecha District, the

Local stratigraphic units for the Silurian of East Siberia

Pritunguska Subregion

III

Maymecha District 272.0, III. 5 (9)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D		Mudstone				24	
54	Dokir *dk* ~ 101.4	Kira *kr* 0 - 20.0 ?	5.1		2.8	?	?
53					2.3		
52			7.8		2.8		
51					2.9		
50					2.1		
49			7.2		2.6		
48					2.8		
47					2.4		
46		Bachanay *bh* ~ 51.0	Upper > 19.4	6	5.0	?	192
45				5	3.6		
44				4	5.5	49-52	
43				3	5.1	45-48	
42				2	4.0	43б-44	
41				1	4.8	42-43а	
40			Lower 23.2	4	5.1	55б	
39				3	5.8	55а	
38				2	5.8	52б-54	
37				1	6.2	51-52а	
36		Dolgotnaya *dl* 30.4	Upper 15.0	6	3.0	50д	
35				5	3.1	50г	
34				4	1.8	50в	
33				3	1.9	50б	
32				2	2.2	49е-50а	
31				1	3.0	49д	
30			Lower 15.4	5	3.6	49г	
29				4	3.1	49в	
28				3	2.3	49б	
27				2	3.0	48б-49а	
26				1	3.4	46-48а	Kh-36
25	Akher *akh* 170.9	Kherkimi *hrk* 53.1	Upper 26.6	3	9.1	44-45	
24				2	7.7	40-43	
23				1	9.8	38-39	
22			Low. 26.5	2	13.3	35-37	
21				1	13.2	28-34	
20		Orachi *or* 69.0	Upper 25.7	3	9.0	26б-27	
19				2	8.0	25б-26а	
18				1	8.7	24-25а	
17			Lower 43.3	4	9.4	23	
16				3	11.9	21б-22	
15				2	10.2	19-21а	
14				1	11.8	15-18	
13		Pravoatyrdyak *pv* 48.8	Upper 27.9	5	6.3	14г	
12				4	5.7	14в	
11				3	5.0	14б	
10				2	5.1	13б-14а	
9				1	5.8	13а	
8			Lower 20.9	8	5.3	11в-12	
7				7	4.2	11б	
6				6	3.2	10в-11а	
5				5	3.3	10б	
4				4	2.0	10а	
3				3	1.6	9	
2				2	0.7	8	
1				1	0.6	7	
O		Kayyalak					

Moyero District 513.0, III. 6 (10)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D	–	Mukdeken				21	27
54	Khakhol *khl* 189.3	Kholyukhan *hl* 68.5	Upp. 20.4	2	9.6	9-20	
53				1	7.3	13; 2-8	18
52			Middle 24.9	3	8.4	1-12	
51				2	10.0	6-7	36
50				1	6.5	6; 1-5	
49			Low. 23.2	3	6.7	5	45
48				2	8.5	2-4	
47				1	8.0	8-9; 1	
46		Yangada *jd* 38.6	Upper 27.4	6	5.5	5; 4-7	44
45				5	4.5	1-4	43
44				4	3.5	20б	67
43				3	4.6	19-20а	
42				2	5.0	18	
41				1	4.0	17	
40			Lower 11.2	4	3.4	14-16	
39				3	2.3	12-13	
38				2	3.4	10-11	
37				1	2.1	8-9	
36		Khakoma *hk* 82.2	Upper 40.2	6	8.0	26-31	66
35				5	8.2	19-25	
34				4	4.8	15-18	
33				3	5.1	9-14	
32				2	6.1	7-8	
31				1	8.0	1б-6	
30			Lower 42.0	5	9.8	12-18	64
29				4	8.5	16-22	58
28				3	6.2	7-15	
27				2	8.3	4-8; 2-6	58A
26				1	9.2	3-4; 2-3	60
25	Prianabar *prb* 324.8	Agidy *ag* 84.3	Upper 41.8	3	9.4	53-56	87
24				2	14.8	41-52	
23				1	17.6	35-40	
22			Low. 42.5	2	20.9	23-34	86
21				1	21.6	8-22	84-85
20		Khaastyr *hs* 129.5	Upper 50.0	3	20.0	36-43	77
19				2	13.6	27-35	
18				1	16.5	20-26	
17			Lower 79.5	4	21.0	11-19	
16				3	18.0	22-23;1-10	90
15				2	20.3	25-31;18-21	78
14				1	20.3	18-24	83
13		Moyerocan *mr* 111.0	Upper 57.0	5	13.0	27-28	
12				4	11.4	24-26	82
11				3	10.2	21-23	
10				2	10.5	19-20	
9				1	11.8	17-18	
8			Lower 54.0	8	13.7	14-16	
7				7	11.0	13	
6				6	8.3	11-12	
5				5	8.5	9-10	
4				4	5.0	8	
3				3	5.0	4-7	
2				2	1.8	2-3	
1				1	0.5	1	
O	–	Dzheromo					

Morkoka District 395.0 ?, III. 7 (11)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D	–	Mudstone				–	
54	Yartom *jat* 182.5	Tomba *tm* 114.6	Upper 64.0	8	8.2	8	Sections in up. reaches of Nizhnyi Yaralin R.
53				7	8.2	7	
52				6	8.3	6	
51				5	8.8	5	
50				4	7.7	4	
49				3	7.8	3	
48				2	8.0	2	
47				1	7.0	1	
46			Lower 50.6	10	5.5	6б	TSh67-4Б
45				9	5.0	6а	
44				8	5.7	4-5	
43				7	3.4	3	
42				6	5.0	2	
41				5	5.0	1д	
40				4	5.0	1г	
39				3	5.0	1в	
38				2	5.0	1б	
37				1	6.0	1а	
36		Yaralin *ja* 67.9	Upper 35.0	6	7.0	10г-11	TSh67-4A
35				5	7.0	10в	
34				4	4.0	10б	
33				3	4.0	9-10а	
32				2	5.5	7-8	
31				1	7.5	6	
30			Lower 32.9	5	8.4	4-5	
29				4	6.0	2-3	
28				3	4.0	1б	
27				2	6.5	2; 1а	3
26				1	8.0	1	
25	Kuonda *knd* 213.0	Nepperende *np* 45.4	Upper 22.6	3	5.0	50	178
24				2	8.0	46-49	
23				1	9.6	41-45	
22			Low. 22.8	2	10.8	36-40	
21				1	12.0	28-35	
20		Mashkova *mš* 106.1	Upper 46.3	3	19.4	17-27	
19				2	12.0	12-16	
18				1	14.9	5-11	
17			Lower 59.8	4	15.4	22;1-4;26-4	180
16				3	14.4	19б-21	177
15				2	15.5	11б-19а	
14				1	14.5	3-11а	
13		Bashnya *bš* 43.6	Upper 23.1	5	5.5	8	175
12				4	5.0	7	
11				3	3.9	4-6	
10				2	5.0	3	
9				1	3.7	1-2	
8			Lower 20.5	3	10.0	9	174
7				2	5.6	5-8	
6				1	4.9	1-4	
5		Baytakh bt 17.0		5	6.0	14г	T63-97
4				4	3.0	14в	
3				3	3.3	14б	
2				2	1.7	14а	
1				1	3.0	13	
O	–	Delingde					

Vilyuy District ~ 388.0, III. 8 (12)

Correlation level	Local series	Formation	Subformation	Local chronozone	Thickness, m	Bed (particular link)	Stratotype
D				?			
54	—	~ 160.0	14.4	29	7.2	?	?
53				28	7.2		
52			21.7	27	7.3		
51				26	7.7		
50				25	6.7		
49			19.9	24	6.8		
48				23	7.0		
47				22	6.1		
46			26.0	21	4.8		
45				20	4.4		
44				19	5.0		
43				18	3.0		
42				17	4.4		
41				16	4.4		
40			18.5	15	4.4		
39				14	4.4		
38				13	4.4		
37				12	5.3		
36			30.6	11	6.1		
35				10	6.1		
34				9	3.5		
33				8	3.5		
32				7	4.8		
31				6	6.6		
30			28.9	5	7.4		
29				4	5.3		
28				3	3.5		
27				2	5.7		
26				1	7.0		
25	Meik *mc* ~ 227.5	Khordogoy *hrd* 60. ?			6.6	?	T63
24					10.6		
23					12.7		
22					14.3		
21					15.8		
20		At-Yuryakh *at* ~ 107.7	Upper ~ 44.0	3	18.6	?	
19				2	11.4		
18				1	14.0	103 - ?	
17			Lower 63.7	4	17.1	92-102	184*
16				3	14.2	83-91	
15				2	16.4	71-82	
14				1	16.0	57-70	
13		Oguguut *og* 59.8	Upper 32.5	5	7.0	53-56	
12				4	6.9	44б-52	
11				3	6.1	41-44а	
10				2	6.2	38-40	
9				1	6.3	30-37	
8			Lower 27.3	8	7.0	22б-29	
7				7	5.8	15-22а	
6				6	4.2	12б-14	
5				5	4.2	10-12а	
4				4	2.6	7-9	
3				3	2.2	6б	
2				2	1.0	5-6а	
1				1	0.3	3-4	
O	–	Oyusut					

FIGURE 8 — Silurian stratigraphy of the Pritunguska Subregion. Figure continued on Fig. 7.

Bakhanay Formation is disconformably overlain by Devonian red beds.

Moyero Stratigraphic District

Proposed by Tesakov et al. (1979, p. 14), the Moyero stratigraphic district lies in the drainage basin of the Moyero River and the upper reaches of the Kotuy and Vilyuy Rivers (Fig. 1). This district features the Prianabar and Khakhol Local Series (Fig. 8). All the formations have been named and described by Tesakov et al. (1979, pp. 44–54; 1985).

Prianabar Local Series (Llandovery, S_1prb) — As noted by Tesakov et al. (1992, p. 81), this unit is named for the Anabar structure. The stratotype is on the Moyero River. The series (Fig. 8), with three formations, is commonly composed of laminated and nodular limestone.

The lower formation (Moyerocan Formation, Rhuddanian–lower Aeronian, abbreviated S_1mr) is named for the Moyerocan River. The stratotype is on the Moyerocan River, 2–3 km from its mouth. The formation has two subformations. The lower subformation is made up of gray mudstone and alternating mudstone and platy, gray limestone (at the bottom) and laminated, grey limestone (at the top). The fauna is dominated by *Alispira gracilis, Coronograptus cyphus, Paraclimacograptus innotatus, Acernaspis superciliexcelsis,* and *Clorinda undata.* The upper subformation features gray nodular limestone. The fauna is dominated by *Sibiritia wiluiensis, Septatrypa antiquata, Zygospiraella duboisi,* and *Isorthis neocrassa.*

The middle formation (Khaastyr Formation, middle–upper Aeronian, abbreviated S_1hs) is named for the Khaastyr River. The stratotype is along the Moyero River, near the mouth of the Khaastyr River. The formation, with two subformations, is composed of alternating thick units of gray nodular limestone with grey marl with thin, lensing fine shell-hash and intraclast limestone. The fauna is dominated by *Quadralites quadratus, Stegerhynchus decemplicatus duplex, Calamopora alveolaris, Favosites gothlandicus gothlandicus, Eocoelia hemisphaerica, Pentamerus oblongus,* and *Parastriatopora rhizoides.*

The upper formation (Agidy Formation, Telychian, abbreviated S_1ag) is named for Lake Agidy. The stratotype is on the Moyero River, between the Moyerocan Rapids and the mouth of the Moyerocan River. The formation has two subformations. The lower subformation is composed of gray, platy and nodular limestone with interbeds of variegated marls (lower part) and green marl with nodules and small lenses of limestone with intraclasts (upper part). It is dominated by *Herrmannina moierensis, Anabaria rara, Bystrowicrinus quinquelobatus,* and *Multisolenia tortuosa* (lowest part) and *Cytherellina oviformis, Mendacella tungussensis,* and *Pentamerus oblongus* (at the top). The upper subformation is represented by gray nodular limestone with rare thin interbeds of gray marl. The dominant fossils are *Multisolenia tortuosa, Favosites gothlandicus moyeroensis, Cytherellina oviformis, Bystrowicrinus bilobatus, Beyrichia mirabilis, Pachystylostroma sibiricum, Alispira rotundata,* and *Stromatopora lenensis.*

Khakhol Local Series (Wenlock–Pridoli, S_{1-2}khl) — Proposed herein, the name of this local series (Fig. 8) is derived by combining the first syllables of the names of the Khakoma and Kholyukhan Rivers. The stratotype is on the Moyero River, and extends from the Smirnov Rapids to the mouth of the Khakoma River. Subdivided into three formations, the series is commonly composed of reef limestone, dolostone, and marl.

The lower formation (Khakoma Formation, Wenlock, abbreviated S_1hk) is named for the Khakoma River. The stratotype is along the Moyero River, near the mouth of the Khakoma River. The formation, with two subformations, consists of medium-dark gray, platy limestone and dolostone with rare nodular limestone. Coral-stromatoporoid bioherms and biostromes are frequent. The fauna is dominated by *Labechia condensa, Ecclimadictyon fastigiatum, Neobeatricea nikiforovae, Yavorskiina membrosa, Clavidictyon cylindricum, Stelodictyon moierense, Beyrichia mirabilis,* and *Bystrowicrinus bilobatus.*

The middle formation (Yangada Formation, Gorstian, abbreviated S_2jd) is named for the Yangada River. The stratotype is located on the Moyero River, 8 km below the Mramorny Rapids and 12.5–14 km below the confluence with the Kholyukhan River. This formation is divided into two subformations. The lower subformation consists of gray, platy and laminated limestone and dolostone. The upper subformation is formed of lower gray marl; middle gray, platy and laminated dolostone; and upper gray, nodular and platy limestone. The fauna includes *Hyattidina acutisummitatus, Beyrichia parva, Murchisonia cingulata,* and *Straparollus alacer.*

The upper formation (Kholyukhan Formation, Ludfordian–Pridoli, abbreviated S_2hl) is named for the Kholyukhan River. The stratotype sections are on the Moyero River, 6 km below the Smirnov Rapids, and 10.5 km farther down the Kholyukhan River. With three subformations, the Kholyukhan consists of variegated dolomitic marl, grey platy dolostone, and variegated gypsum. Fossils are not known. Overlying red siltstones belong to the Devonian Mukdeken Formation.

MORKOKA STRATIGRAPHIC DISTRICT

Defined by Tesakov et al. (1979, p. 14), this stratigraphic district lies in the drainage basins of the upper Olenek, Markha, and Morkoka Rivers (Fig. 1). The district features the Kuonda and Yartom Local Series (Fig. 8).

KUONDA LOCAL SERIES (LLANDOVERY, S_1KND) — Tesakov and Shpunt (1967) initially distinguished this interval (Fig. 8) as a formation (suite) and named it for the Kuonda River. The stratotype is in the middle reaches of the Nizhnyaya Bolshaya Kuonda River. This series is commonly made up of gray laminated and nodular limestone with green marl at the base and biostromes at the top. It has four formations.

The lowest formation (Baytakh Formation, lower–lower middle Rhuddanian, abbreviated S_1bt) (Tesakov, et al., 1992, p. 14) is named for Lake Baytakh. The stratotype is on the Morkoka River, 2 km above the mouth of the Kerekhteekh River. Its lower part is composed of intraclast conglomerates, but the upper consists of grey-green marl. The formation yields *Parakidograptus acuminatus, Hedrograptus scalaris, Glyptograptus tamariscus,* and *Alispira gracilis.*

The next formation (Bashnya Formation, uppermost middle Rhuddanian–lower Aeronian, abbreviated S_1bs) (Tesakov et al., 1990, p. 71; 1992, p. 16) is named for Bashnya Mountain. The stratotype is on the Nizhnyaya Bolshaya Kuonda River (47–48 km along a straight line from the mouth). The formation has two subformations. The lower is composed of cream-colored laminated limestones, and the upper of gray wavy-bedded and nodular limestones. The fauna is dominated by *Alispira gracilis, Septatrypa pentagonalis, Hiregiroceras costalatum, Geisonoceras kureikense, Isorthis neocrassa,* and the tracks of deposit feeders.

The Mashkova Formation (middle–upper Aeronian, abbreviated S_1ms) (Tesakov et al., 1990, p. 71; 1992, p. 26) was named in memory of the paleontologist T. V. Mashkova. The stratotype is on the Nizhnyaya Bolshaya Kuonda River in the permafrost area. The formation, with two subformations, is composed of gray nodular limestone with thin lenses of gray calcarenite. The fauna is dominated by *Calamopora alveolaris, Zygospiraella duboisi, Isorthis neocrassa, Borealis borealis schmidti, Costaegera hastata, Lenatoechia multicostata, Gibberella lenaica, Straparollus scalaris, Stegerhynchus pseudonuculus, Favosites gothlandicus gothlandicus, Tungussophyllum tenuiseptatum, Sibiritia wiluiensis,* and *Panderodus unicostatus.*

The uppermost formation (Nepperende Formation, Telychian, abbreviated S_1np) (Tesakov et al., 1992, p. 55) is named for the Nepperende River. The stratotype is near the mouth of a creek on the right bank of the Nizhnyaya Bolshaya Kuonda River near site 484.0, on the right bank, 0.5 km from the river's mouth. The formation has two subformations. The lower subformation is composed of gray platy to wavy-bedded and nodular limestone with isolated beds of gray marl. The upper subformation is made up of light-gray, thick-bedded, platy limestone and dolostone with coral-stromatoporoid biostromes and bioherms. The fauna includes *Panderodus unicostatus, Holophragma mitrata, Dalejina hybrida, Favosites gothlandicus moyeroensis, Plümatalinia densa,* and *Sibiritia wiluiensis.*

YARTOM LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}J AT) — Proposed herein, the name of this local series (Fig. 8) is derived from a combination of the first syllables of the names of the Yaralin and Tomba Rivers. The stratotype lies at the type sections of the Yaralin and Tomba Formations. The series is represented by lower gray massive dolostone and upper variegated dolomitic marl. It has two formations.

The lower formation (Yaralin Formation, Wenlock, abbreviated S2ja) (Tesakov and Shpunt, 1967, p. 81) is named for the Yaralin River. The stratotype is along the Nizhniy Yaralin River on the slope of Bashnya Mountain. This formation has two subformations. At the base of the unit is light-gray massive dolostone with an interval of dark-grey limestone with black chert and biostromes. The rest of the formation is made up of interbeds of coral dolostone, lenses of sandstone, and intraclast (flat pebbles) conglomerate. The fauna includes *Subalveolites subulosus, Favosites gothlandicus moyeroensis, Strombodes socialis, Sapporipora favositoides, Parastriatopora tebenjkovi, Anabaria rara,* and *Sibiritia kotelnyensis.*

The upper formation (Tomba Formation, Ludlow–Pridoli, abbreviated S_2tm) (Tesakov and Shpunt, 1967, p. 84) is named for the Nizhnyaya Tomba River. The stratotype is in the drainage basin of the Nizhniy Yaralin River. The formation has two subformations. The lower part of the formation is dominated by thin-bedded, platy and laminated dolostone with stromatolites. The upper part is variegated dolomitic marl with rare lenses of gray dolostone. The lower subformation has *Didymothyris didyma, Schrenkia* sp., *Herrmannina* sp., and *Bystrowicrinus quinquelobatus.* The contact with the Devonian is not known. The formation is overlain by Permian.

VILYUY STRATIGRAPHIC DISTRICT

Defined by Tesakov et al. (1979, p. 14), this district lies in the drainage basin of the middle reaches of the Vilyuy River and the lower reaches of the Markha River (Fig. 1). The district contains the Meik Local Series, which forms part of the Prianabar Regional Series. Other units of the Priyenisey Regional Series are not yet known in the

Vilyuy District (Fig. 8).

MEIK LOCAL SERIES (LLANDOVERY, S_1MC) — Described by Arseniev and Ivanova (1954), this interval (Fig. 8) was distinguished as a formation that included the uppermost Ordovician and Lower Silurian. Subsequently, the Meik Formation was restricted to the Llandovery (Mikhailov and Tesakov, 1972). The series is named for the village of Verkhniy Meik. The stratotype is on the Vilyuy River near Verkniy Meik. In general, the series is composed of lower limestone and dolostone with marl intercalations, middle dolomitic marl with thin interlayers of dolostone, and upper light-gray dolostones. It is divided into three formations.

The lower formation (Oguguut Formation, Rhuddanian–lower Aeronian, abbreviated S_1og) was established by the authors during their 1985 field season (Tesakov et al., 1996b). It is named for the Oguguut (Ogogut) River. The stratotype is on the Vilyuy River, just above the mouth of the Kholomolokh-Yuryakh River. The formation, with two subformations, consists of gray, laminated and platy limestone and dolostone and units of gray dolomitic marl with thin lenses of gray limestone and dolostone. Stromatolites are widespread, and dolostones predominate in the upper subformation. The fauna includes *Sibiritia wiluiensis, Mendacella tungussensis, Mesofavosites dualis, Favosites gothlandicus gothlandicus, Lenatoehia elegans, Coolinia gracilis, Dentiferocrinus dentiferus, Myelodactylus flexibilis, Tungussophyllum conulus,* and *Crassilasma electum.*

The middle formation (At-Yuryakh Formation, middle–upper Aeronian, abbreviated S1at) was also established by the authors during the 1985 field season (Tesakov et al., 1996b). It is named for At-Yuryakh Creek. The stratotype is on the Vilyuy River, just above the mouth of the Kholomolokh-Yuryakh River. The two subformations consist of gray and variegated dolomitic marl with thin lenses of gray dolostones, often with stromatolites. The upper subformation is poorly exposed. The lower part is dominated by *Isorthis neocrassa, Strophomena pectenoides, Stegerhynchus decemplicatus duplex, Lenatoechia elegans, Favosites gothlandicus gothlandicus, Bystrowicrinus bilobatus, Entelophyllum articulatum,* and *Streptelasma whittadi.*

The uppermost formation (Khordogoy Formation; Telychian?, possibly Wenlock or Ludlow; abbreviated S_{1-2}hrd) was recognized by Yu. I. Tesakov during the 1963 field season, but is formalized herein for the first time. It is named for the Khordogoy River. The stratotype is on the Vilyuychan River below the mouth of the Stan River. The formation is made up of light-gray, platy dolostone with eurypterids. Higher Silurian rock has not been found in the Vilyuy District.

NYUYA-BEREZOVO STRATIGRAPHIC DISTRICT

Defined by Tesakov et al. (1979, p. 14), this stratigraphic district lies in the middle reaches of the Lena River and its Nyuya, Dzherba, and Biryuk River tributaries (Fig. 1). It includes the Meut and Kurung Local Series (Fig. 9).

MEUT LOCAL SERIES (LLANDOVERY, S_1MT) — Proposed herein (Fig. 9), the name of this local series is composed of the first syllables of the names of the Melichan and Utakan Rivers. The stratotype is along the lower Nyuya River. The series, with two formations, is composed of dolomitic marl with dolostone interbeds.

The lower formation (Melichan Formation, Rhuddanian–lower middle Aeronian, abbreviated S_1ml) (Mikhailov and Ushakov, 1966) is named for the Melichan River. The stratotype is along the lower Melichan River, and the hypostratotype is on the lower Nyuya River. The formation, with two subformations, consists of grey dolostone and dolomitic marl with interbeds of gray, platy dolostone. *Loganellia sibirica, Lenatoechia elegans, Icriodella discreta,* and *Distomodus kentuckyensis* are known in the formation.

The upper formation (Utakan Formation, upper middle Aeronian–Telychian, abbreviated S_1ut) (Mikhailov and Ushakov, 1966) is named for the Utakan River. The stratotype is on the upper Utakan River, and the hypostratotype is on the lower Nyuya River. The formation, with three subformations, consists of variegated dolomitic marl with rare interbeds of platy dolostone that are gypsiferous at the top. Known taxa include *Lenatoechia elegans, Elegestolepis conica, Loganellia sibirica,* and *Panderodus unicostatus.*

KURUNG LOCAL SERIES (WENLOCK–PRIDOLI, S_{1-2}KRN) — Proposed herein (Fig. 9), this local series is named for Kurung village. The stratotypes are along the Nyuya River above Neryuktey village and on the Lena River above Nyuya village. This series is commonly composed of dolomitic marl with interbedded dolostone and biostromes. It has two formations.

The lower formation (Nyuya Formation, Wenlock, abbreviated S_1nsk) (Mikhailov and Ushakov, 1966, p. 31) is named for the Nyuya River. The stratotypes are on the Nyuya River, 1.5 km above Neryuktey village and on the Lena River 6 km above Nyuya village. The formation, with two subformations, is composed of gray dolomitic marl with thin interbeds of platy dolostone and biostromes. It has *Ozarkodina tamashkovae, Panderodus unicostatus, Favosites gothlandicus moyeroensis, Sapporipora favositoides,* and *Subalveolites subulosus.*

The upper formation (Neryuktey Formation, Gorstian, abbreviated S2nrk) was recognized by N. N.

	Local stratigraphic units for the Silurian of East Siberia																					Adjacent Regions		
	Nyuya-Berezovo Subregion IV							Irkutsk Subregion V														Altaj-Sayan Mountain Area		North-East Russia
	Nyuya-Berezovo District ~ 381.0 IV. 1 (13)							Ilimsk District ~ 162.0 V. 1 (14)							Balturino District ~ 170.0 V. 2 (15)									
Correlation level	*Local series*	*Formation*	*Subformation*	*Local chronozone*	*Thickness, m*	*Bed (particular link)*	*Stratotype*	*Local series*	*Formation*	*Subformation*	*Local chronozone*	*Thickness, m*	*Bed (particular link)*	*Stratotype*	*Local series*	*Formation*	*Subformation*	*Local chronozone*	*Thickness, m*	*Bed (particular link)*	*Stratotype*	*Horizon*	*Subhorizon*	*Horizon*
D				?							?							?				–		–
54	Kurung *krn* ~ 104.0	~ 50.0	12.3	8	6.8				~ 40.0	5.6	18	3.1				~ 40.0	5.6	18	3.1			Chernyi-anuy		Mirnyian
53				7	5.5						17	2.5						17	2.5					
52			19.1	6	6.8					8.6	16	3.0					8.6	16	3.0					
51				5	7.1						15	3.2						15	3.2					
50				4	5.2						14	2.4						14	2.4					
49			17.6	3	4.8					8.0	13	2.2					8.0	13	2.2			Kuimovian		
48				2	6.8						12	3.1						12	3.1					
47				1	6.0						11	2.7	?	?				11	2.7	?				
46		Neryuktey *nrk* ~ 45.0	Upper ~ 25.0	6	4.5	?	?			10.6	10	2.2					10.6	10	2.2					Bizonian
45				5	3.1						9	1.4						9	1.4					
44				4	4.8						8	2.2						8	2.2					
43				3	4.0						7	1.9						7	1.9					
42				2	3.6						6	1.4						6	1.4					
41				1	4.1						5	1.5						5	1.5		?			
40			Lower ~ 20.0	4	4.4					7.2	4	1.8					7.2	4	1.8					
39				3	4.4	13 - 14	163				3	1.7						3	1.7					
38				2	5.5	9б - 12					2	1.8						2	1.8					
37				1	5.7	6 - 9a					1	1.9						1	1.9					
36		Nyuya *nsk* 59.4	Upper 29.4	6	5.9	4в - 5			Deshyma *ds* ~ 25.0	12.3	11	2.5	?	Deshyma R. section		Barmo *br* ~ 26.0	12.2	11	2.4	?		Chagyrian	Upper	Sunduganian
35				5	5.7	4б					10	2.5						10	2.5					
34				4	3.5	2г - 4a					9	1.5						9	1.5					
33				3	3.8	2в					8	1.5						8	1.5					
32				2	4.5	2б					7	1.9						7	1.9					
31				1	6.0	18г;1-2a					6	2.4						6	2.4					
30			Lower 30.0	5	7.0	18в	151			12.7	5	3.0					13.8	5	3.0				Lower	
29				4	6.0	18б					4	2.6						4	2.6					
28				3	4.5	15 - 18a					3	1.8						3	1.8					
27				2	6.4	11 - 14					2	2.5						2	3.0	65 - 66				
26				1	6.1	5в; 8-10					1	2.8						1	3.4	63 - 64	135			
25	Meut *mt* 226.2	Utakan *ut* 169.2	Upper 62.5	5	7.0	4 - 5б	158*		Rassokha *rh* ~ 96.9	Upper ~24.0	8	2.9	?	?		Balturino *bl* ~ 104.0	Upper 26.8	8	3.3	61б - 62		Palatyian		Anikinian
24				4	11.0	1в - 3					7	4.0						7	4.4	60 - 61a				
23				3	13.0	1б					6	4.6						6	5.1	57б - 59				
22				2	16.0	7;1a					5	6.5						5	7.2	54 - 57a		Chinetian	Upper	
21				1	15.5	7в-9;1-6	151*			Upper 15.2	4	6.0	48 - 49	141				4	6.8	52 - 53				
20			Middle 50.1	3	20.0	3 - 7б	159*				3	6.1	43в - 47				12.3	3	4.1	46 - 51				
19				2	13.7	40-45;1-2	157*				2	4.2	41б - 43б					2	4.4	43 - 45				
18				1	16.4	24 - 39					1	4.9	39 - 41a					1	3.8	42				
17			Lower 51.1	4	12.7	6 - 23				Middle 21.9	4	6.1	32 - 38				Middle 27.7	4	9.1	35 - 41				
16				3	11.2	39б;1-5					3	4.8	27б - 31					3	7.3	31 - 34				Chamakian
15				2	14.5	36б-39a	156*				2	5.5	23в - 27a					2	6.2	28 - 30				
14				1	12.7	29 - 36a					1	5.5	20 - 23б					1	5.1	23 - 27				
13		Melichan *ml* 57.0	Upper 30.4	5	6.6	23 - 28				Lower 35.8	13	4.7	16д - 19				Lower ~ 37.0	13	3.9	19б - 22			Lower	
12				4	5.7	19 - 22					12	4.1	16г					12	3.5	15 - 19a				
11				3	5.5	18					11	3.7	16в					11	3.4	12б - 14				
10				2	6.3	13б - 17					10	3.8	16б					10	3.4	10в - 12a				
9				1	6.3	13a					9	4.2	16a					9	3.7	10б				
8			Lower 26.6	8	7.0	10б - 12					8	4.0	15в					8	4.2	7 - 10a				
7				7	5.7	8б-9;10a	155*				7	3.2	15б					7	4.1	4б - 6				
6				6	4.3	7 - 8a					6	2.3	14 - 15a					6	2.6	3 - 4a				
5				5	3.9	4 - 6					5	2.3	11 - 13					5	2.6	2в				
4				4	2.3	2б - 3					4	1.4	8б - 10					4	1.6	2б				
3				3	2.0	2a					3	1.4	6 - 8a					3	1.6	1 - 2a				
2				2	0.9	1б					2	0.5	4 - 5					?	2.4	0				
1				1	0.5	1a					1	0.2	3											
O	Krasnokamen								Bratsk							Bratsk								

FIGURE 9 — Silurian stratigraphy for Nyuya-Berezovo and Irkutsk Subregions, East Siberia, and adjacent areas.

Predtetchensky, Yu. I. Tesakov, A. Ya. Berger, V. G. Khromych, and E. O. Kovalevskaya in 1984 (Karatajute-Talimaa and Predtechenskyj, 1995, p. 48). It is named for Neryuktey village. The stratotype is on the Lena River, 6 km above Nyuya village. The formation is composed of largely red-colored siltstone and dolomitic marl that alternate with gray dolostone. Fossils are unknown. The upper boundary is not defined. The Gorstian age of this formation is only a preliminary evaluation. The overlying Silurian is poorly known from isolated exposures of dolomitic marl. The contact with the Devonian is unknown.

Ilimsk Stratigraphic District

This region was first termed the "Angara-Ilimsk District" (Tesakov et al., 1979, p. 14). It lies in the drainage basins of the middle Angara River and the Ilim River (Fig. 1). The district has two formations (Fig. 9).

The lower formation (Rassokha Formation, Llandovery, abbreviated S1rh) was defined by N. N. Predtetchensky, Yu. I. Tesakov, et al. (*in* Tesakov et al., 1990, p. 72). It is named for its stratotype along Nazarovskaya Rassokha Creek near Tushama Station. The formation, with three subformations, consists of variegated siltstone with thin sandstone lenses. It has *Mesofavosites dualis*, *Lenatoechia ramosa*, *Eotomaria gatlense*, *Pseudoproetus bellus*, *Loganellia sibirica*, and *L. scotica.*

The upper formation (Deshyma Formation [Series], Wenlock?, abbreviated S_1ds) (Tesakov et al., 1996b) is named for exposures along the Deshyma River. It includes variegated siltstone with rare interbeds of gray dolostone. Fossils are unknown, and its age is poorly defined. Higher Silurian is unknown in the Ilimsk District.

Baturino Stratigraphic District

This region was first termed the "Prisayan District" (Tesakov et al., 1979, p. 14). It lies in the Uda and Biryusa River drainage basins (Fig 1) and features the Balturino Formation and the Barmo Beds (Fig. 9).

The lower unit (Balturino Formation, Llandovery, abbreviated S_1bl) was defined by N. N. Predtetchensky, Yu. I. Tesakov, et al. in 1982 (*in* Karatajute-Talimaa and Predtechenskyj, 1995, p. 44) and named for Balturino village. The stratotype is on the right bank of the Chuna River, 3 km above Staroye Balturino village. The formation, with three subformations, is composed of gray sandstone and green, red, and rarely gray laminated siltstone. The upper subformation contains *Eotomaria gatlense*, *Tesakoviaspis concentrica*, *Loganellia scotica*, and *Elegestolepis conica.*

The higher unit (Barmo Beds, Wenlock, abbreviated S_1br) (Tesakov et al., 1996b) is named for the Barmo River. The stratotype is on the right bank of the Chuna River, 3 km above Staroye Balturino village. The Barmo Beds include red and green mudstone and silstone, sometimes calcareous and with rare interbeds of gray sandstone. The unit has *Loganellia asiatica.* Higher Silurian in the Balturino District is not yet known.

Correlation of Stratotypes

Correlation of the east Siberian Silurian is based on virtually all available sections and wells. All petrographic aspects of these rocks, including primary depositional structures, mineralogy, and texture, have been noted. This allows use of standardized symbols in the sections and paleogeographic maps (Fig. 10). As detailed sections that include global, regional, and local biostratigraphic zones cannot be figured in this report, we provide only small-scale stratotype and hypostratotype correlation charts (Figs. 11–13). In all of these figures, the smallest correlation unit is the subhorizon. Geological sections (lithological columns) are plotted at different scales to show the lithology. In districts with thick Silurian sequences, the sections are figured at a smaller scale. The lithological column for each district is a composite standard section with successive formations (suites) and subformations (subsuites). If a formation stratotype is in another stratigraphic district, the formation's hypostratotype is illustrated in the stratigraphic column for the Silurian district that is figured. The stratigraphic units in the lithological columns are described above.

Structure and Facies

The East Siberian epicontinental basin was a stable, persistent structure in the Silurian. This is demonstrated by three profiles that extend across the Siberian platform and Taymyr Peninsula. The first profile (Fig. 14) extends from the Yenisey Ridge in the western Siberian platform into the Norilsk District (well MD-31) and east into the western Anabar massif. The second profile (Fig. 15) crosses the central Siberian platform and extends from the southwest (Bakhta River Basin) to the northeast (Moyero River Basin) and back to the southeast (Nyuya River Basin). The third profile (Fig. 16) extends along the Nizhnyaya Taymyra River and crosses the central Taymyr Peninsula. Two depressions are clearly recognized in the East Siberian Basin. One extends southwest-northeast through the NorthTaymyr Subregion (Fig. 1).

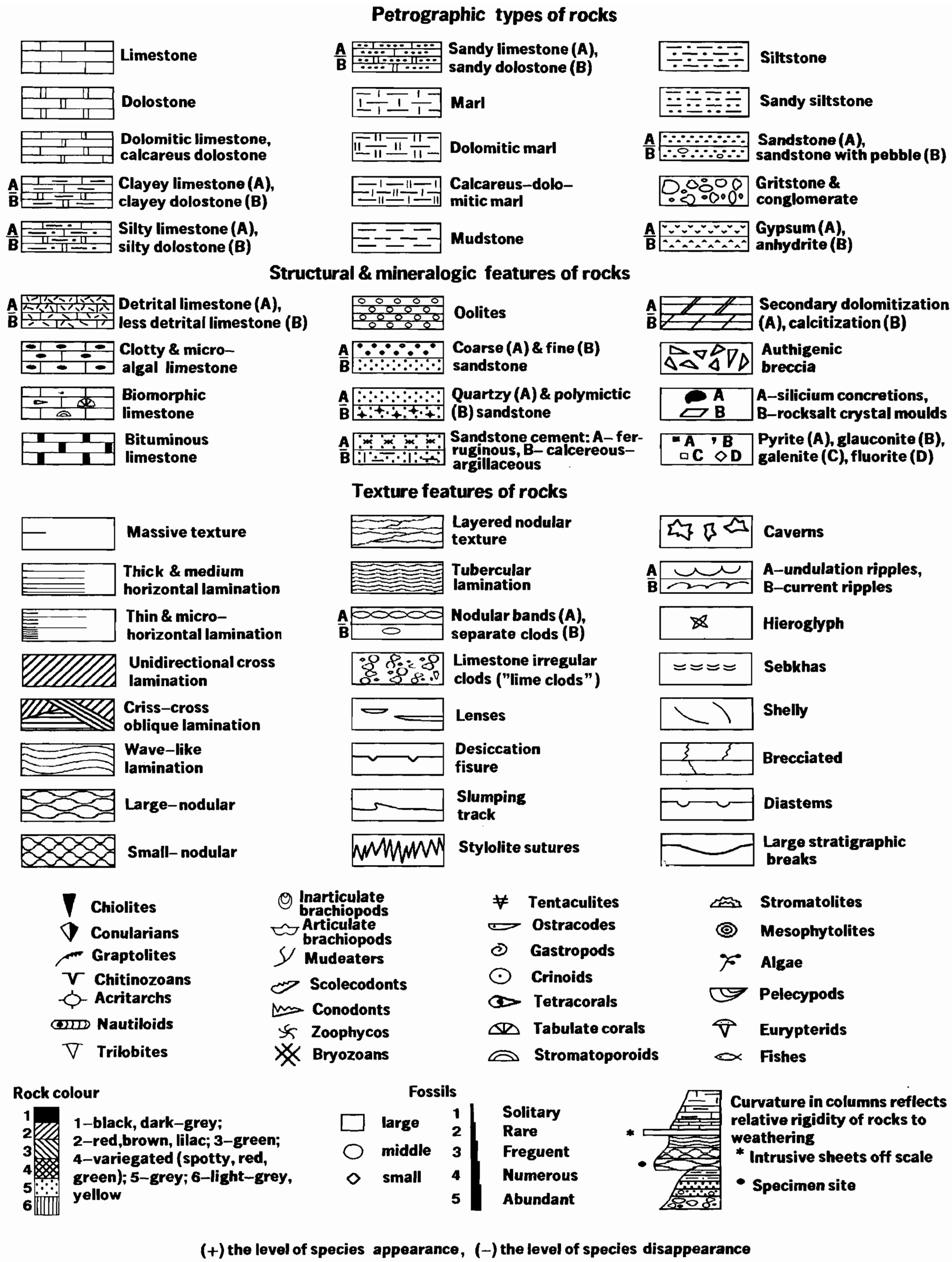

FIGURE 10 — Standard symbols for the Silurian of East Siberia.

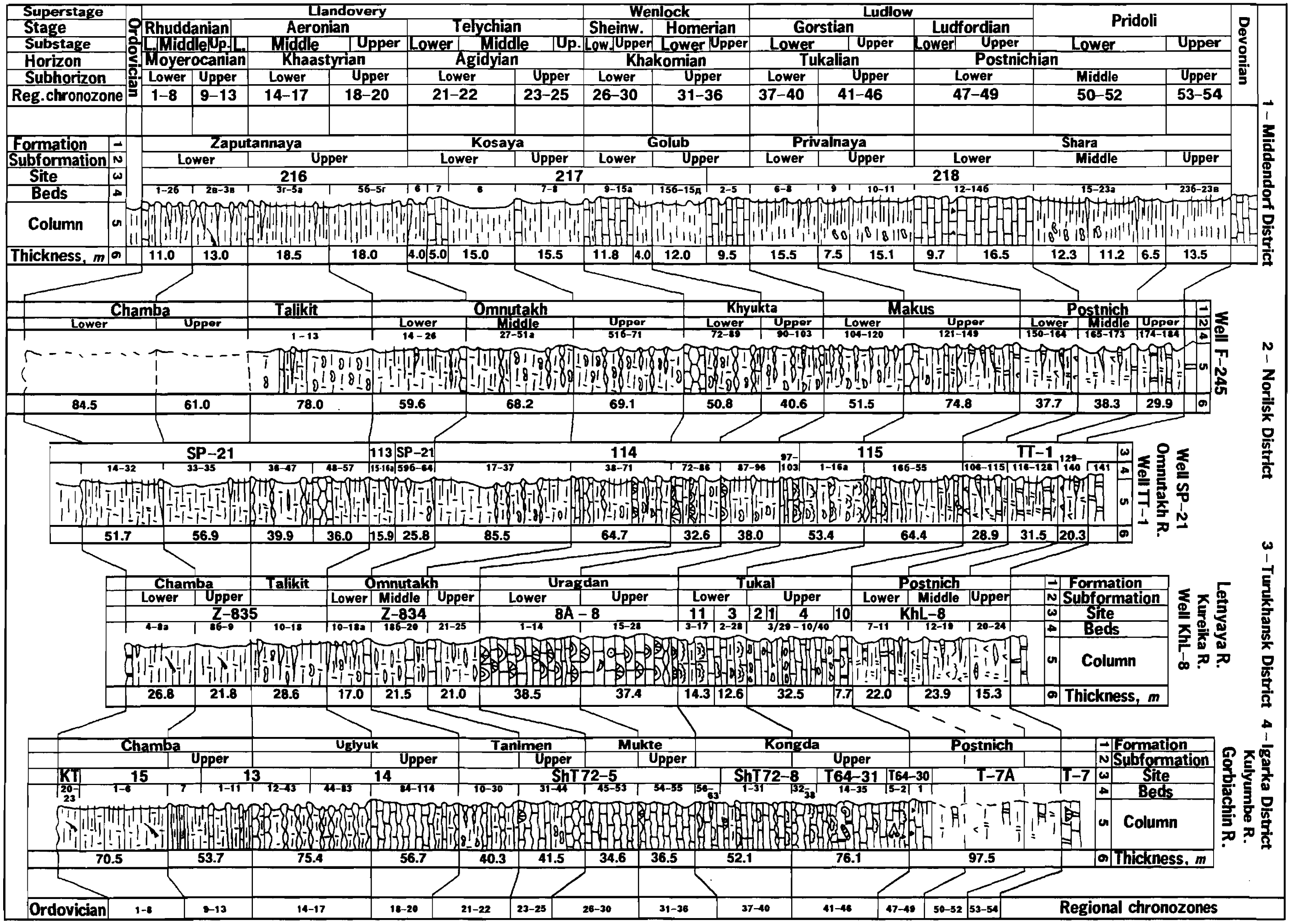

FIGURE 11 — Silurian correlations of the North Taymyr and North Pryenisey Subregions. Compare Fig. 6.

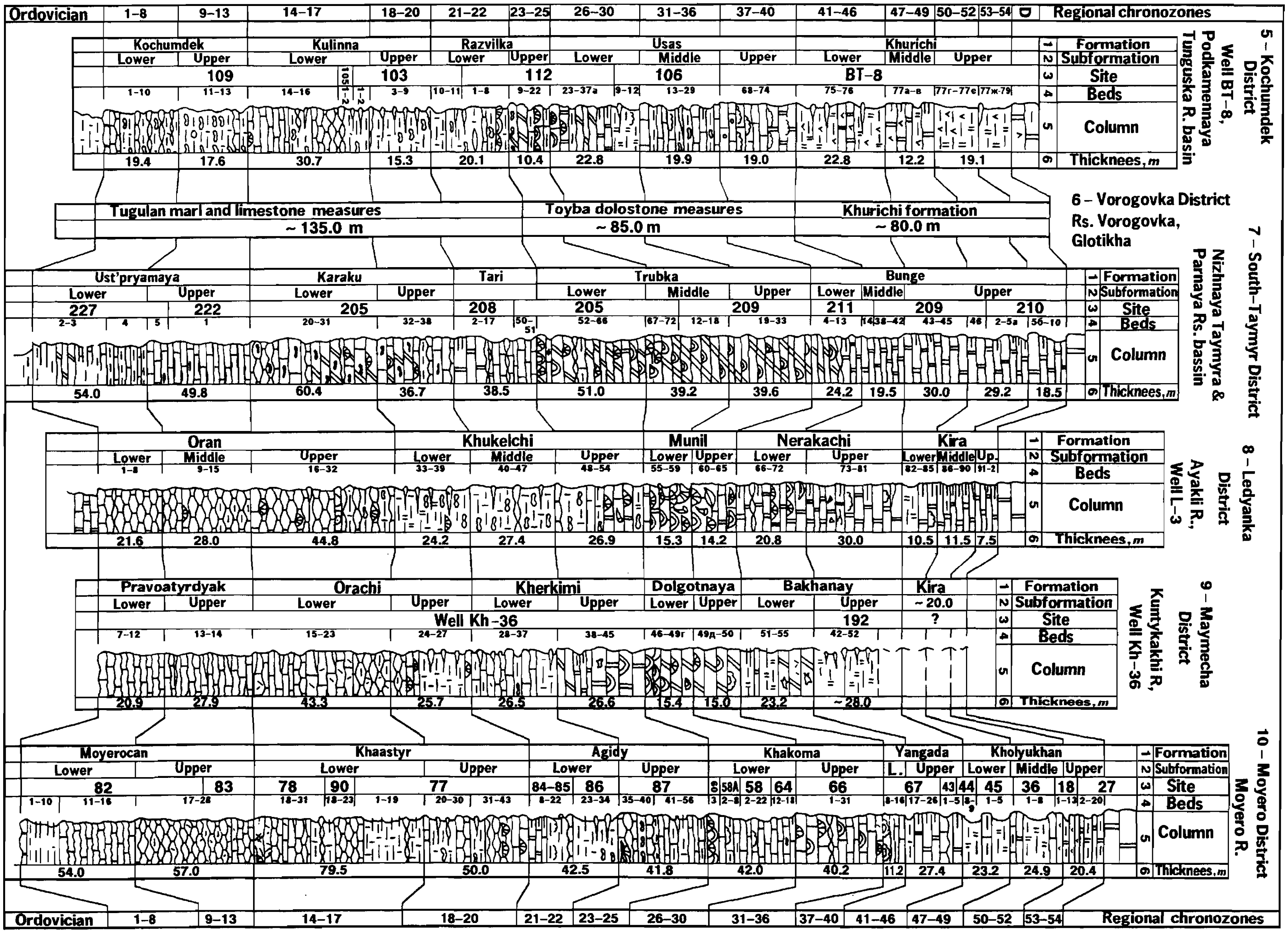

FIGURE 12 — Silurian correlations of the Pritunguska Subregion, continued on Fig. 13. Compare Figs. 7, 8.

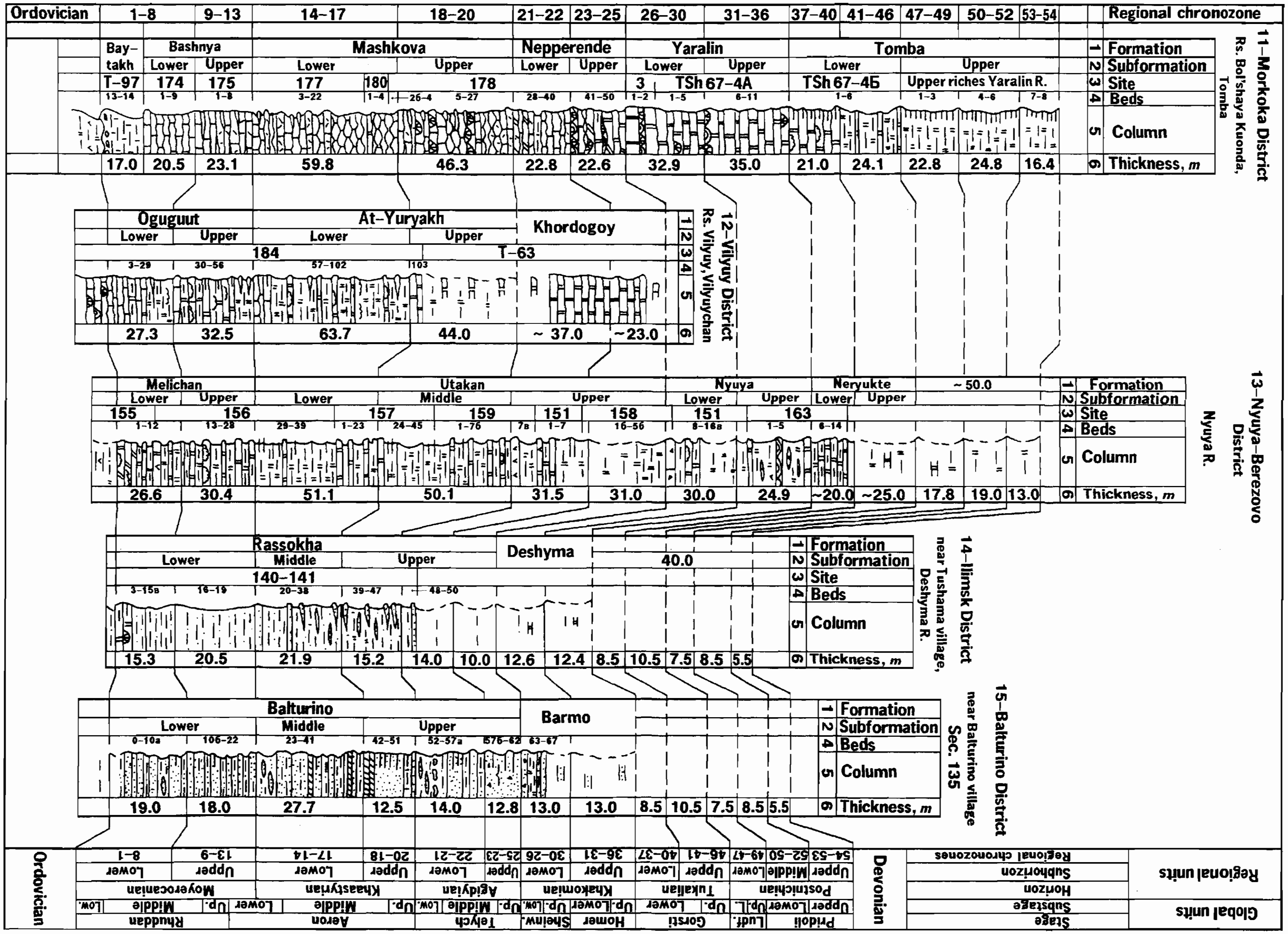

FIGURE 13 — Silurian correlations of the Pritunguska Subregion, continued on Fig. 12. Compare Figs. 7, 8.

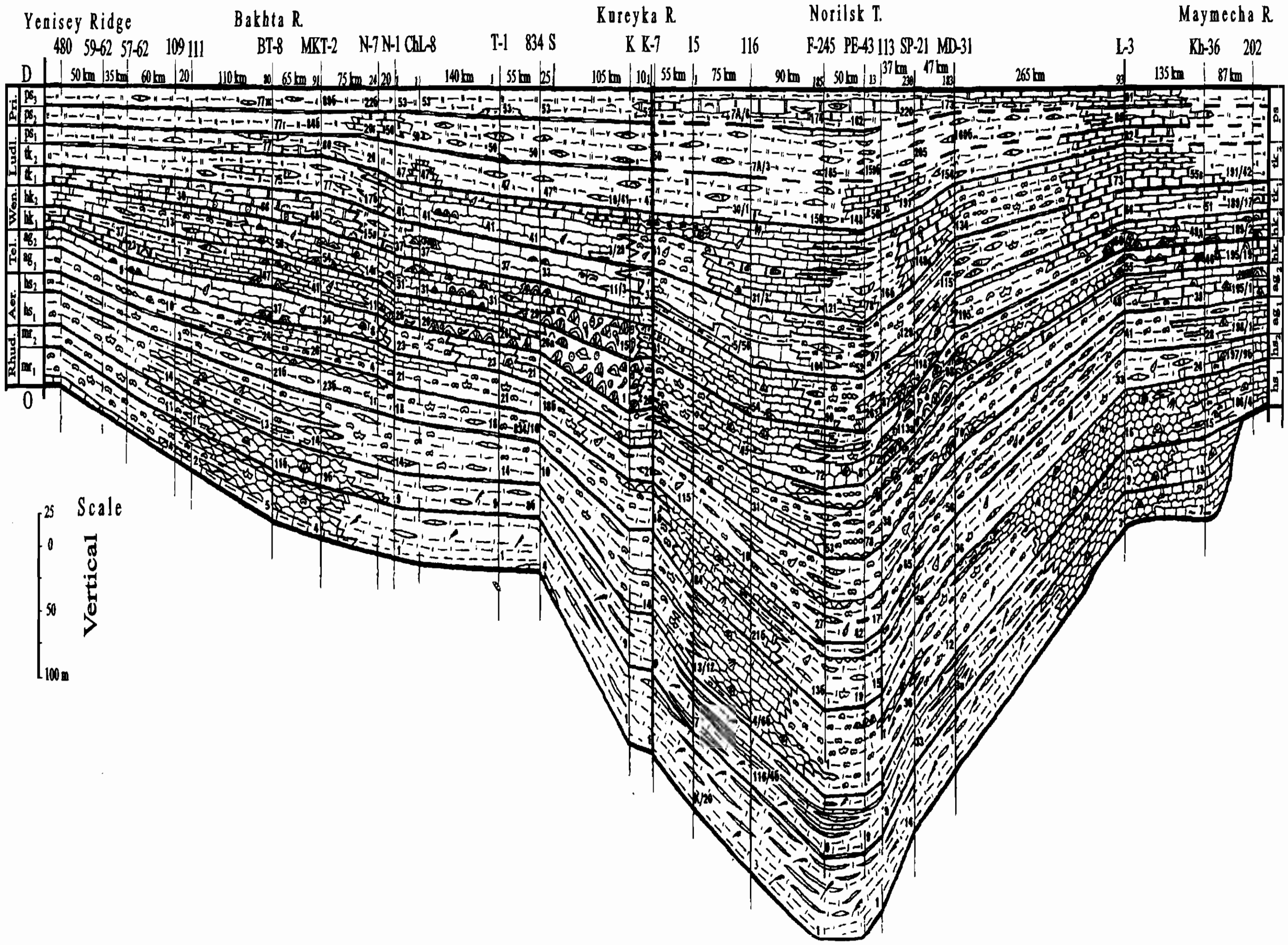

FIGURE 14 — Silurian cross-section of East Siberia along a line extending from the Yenisey Ridge, Bakhta River, Kureyka River, and from Norilsk to the Maymecha River (Fig. 1). For symbols, see Fig. 10.

The other stretches northwest–southeast along the central Norilsk District. In the North Taymayr Depression, graptolitic siliciclastics accumulated through the Silurian. In the Khantayka-Tura Depression, graptolitic shales accumulated in the Early Silurian; marls with brachiopods and ostracodes accumulated in the Middle Silurian, and gypsiferous dolomitic marl in the Late Silurian. Carbonates accumulated on the fringes of both depressions. Early Silurian (Llandovery) brachiopod-bearing, argillaceous carbonates formed. In the Middle Silurian (Wenlock), carbonates with stromatoporoid-coral biostromes, bioherms, and even reefs appeared. The late Silurian featured unfossiliferous dolomite facies. In the southeastern part of the East Siberian Basin (Nyuya-Berezovo and Ilimsk Districts), virtually the entire Silurian is composed of gypsiferous argillaceous dolostone. Unrestricted marine conditions associated with the deposition of carbonates with corals and stromatoporoids occurred only episodically (in the Wenlock of the Nyuya-Berezovo District). In the southern basin (Balturino District), the Silurian is composed of variegated sandstone and siltstone.

Fauna and Flora

Silurian biotas in East Siberia are abundant and taxonomically diverse. The Silurian paleontology of East Siberia is described in many reports (e.g., Abushik, 1960; Yeltyscheva, 1960; Nekhoroshev, 1961; Nikiforova and Andreeva, 1961; Vostokova, 1962; Maksimova, 1962;

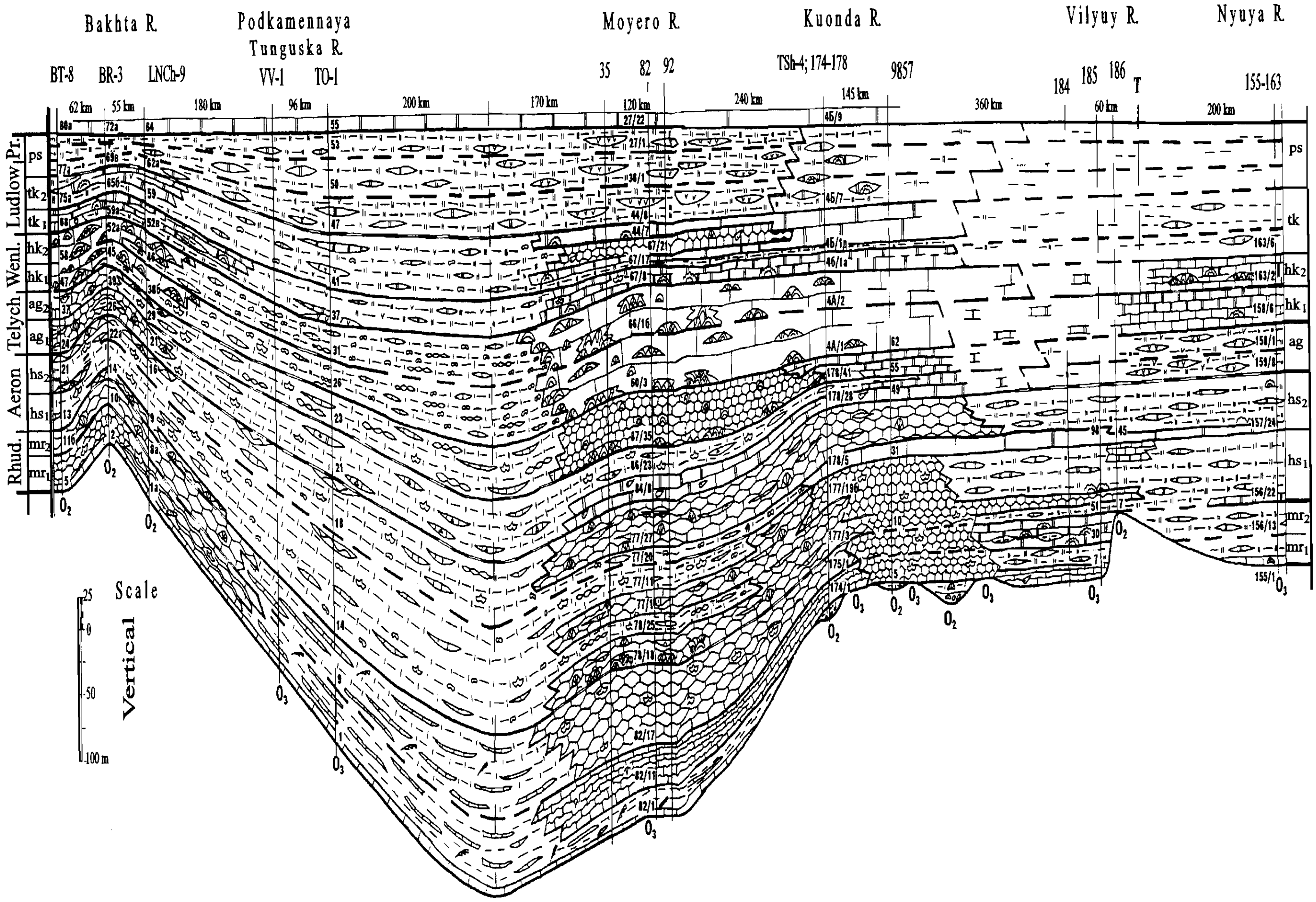

FIGURE 15 — Silurian of East Siberia along a line extending from the Bakhta River, Nizhnyaya Tunguska River, Moyero River, Kuonda River, and the Vilyuy River to the Nuya River (Fig. 1). For symbols, see Fig. 10.

Ivanovsky, 1963; Sokolov and Tesakov, 1963; Obut et al., 1965, 1968; Miagkova, 1967; Lopushinskaya, 1976; Latypov, 1977; Tesakov et al., 1979, 1980, 1985, 1986, 1995; Tesakov, 1980; Sokolov, 1982; Zaslavskaya, 1983; Sheshegova, 1984; Karatajute-Talimaa and Predtechenskyj, 1995). There are over 600 known Silurian species. Figs. 17–26 show selected species and their ranges by regional biostratigraphic zone. Symbols indicate the lower (+) and the upper (-) limits of species ranges. The figures also show the relative abundance of specimens: 1, isolated specimen report; 2, rare; 3, frequent; 4, numerous; 5, abundant. A square indicates large-sized specimens of a given species; a circle indicates mid-sized specimens, and a hexagon indicates small-size specimens. The stratigraphic distribution of Silurian taxa (Figs. 17–26) is related to 54 regional biostratigraphic zones that can be correlated with the global zonation (Fig. 2).

All the species are confined to particular lithofacies belts. The following sections note the facies association and regional distribution of Silurian taxa in the East Siberian Basin.

ALGAE AND STROMATOLITES — Diversity of calcareous algae (Fig. 17) is extremely low throughout the Silurian of East Siberia. Algae occur only in buildups, where they are abundant. Their greatest abundance is in Wenlock reefs in the North Yenisey, Pritunguska, and part of the Nyuya-Berezovo Subregions. They are rather numerous in the Ludlow of the North-Yenisey Subregion, where stromatolites and coral-stromatoporoid build ups are widespread.

SPONGES AND FORAMINIFERANS — These forms are rare and have not been studied.

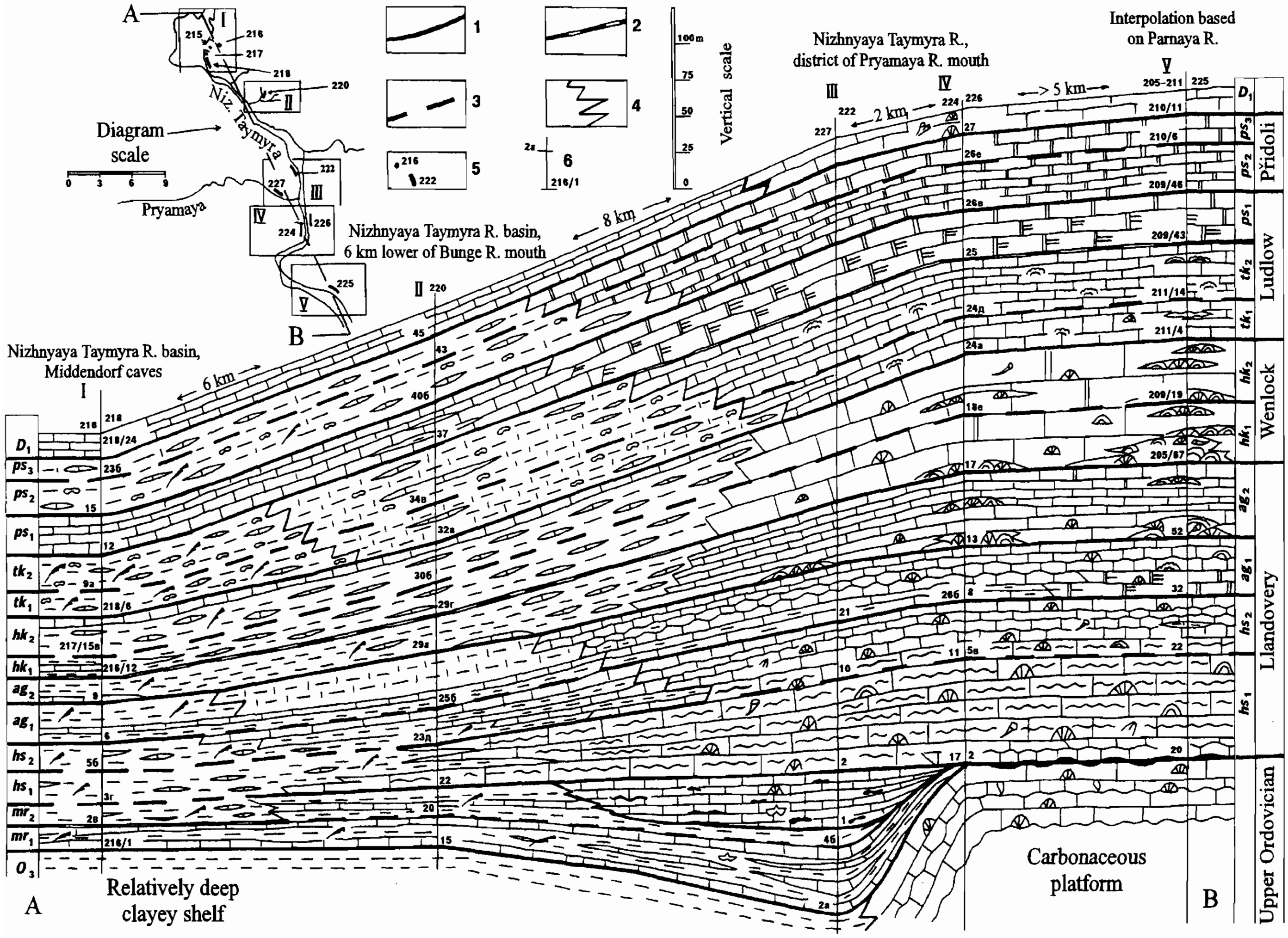

FIGURE 16 — Silurian of the Taymyr Peninsula along the Nizhnyaya Taymyra River. 1–3 indicate isochronous boundaries: 1 — change of lithological and biological dominants; 2 — change of lithological or biological dominant only; 3 — arbitrary; 4 — formation lateral boundaries; 5 — section localities; 6 — bed numbers. For other symbols, see Fig. 10.

STROMATOPOROIDS — These sponges occur in East Siberia in shallow-shelf facies of the North-Yenisey and Pritunguska Subregions. They are most abundant and diverse in, and dominate, the coral-stromatoporoid shoaly facies of the Wenlock. In the Llandovery and Ludlow, stromatoporoids are common (Figs. 17, 18), but less diverse. They are not known in the Pridoli of East Siberia.

TABULATE CORALS AND HELIOLITIDS — These groups (Fig. 18) are largely found largely in shallow-shelf environments of the North-Yenisey and Pritunguska Subregions. They are rare in deep-shelf and lagoonal facies, but numerous and diverse in Aeronian and Telychian shallow-shelf environments. These groups are often dominant in Wenlock shallow-water facies, but show lower species diversity. Tabulates are most numerous in nodular limestone, marl with nodules, and bioherms.

RUGOSE CORALS — These shallow-shelf taxa are most numerous and diverse in the Aeronian, Telychian, and Wenlock of the North Yenisey and Pritunguska Subregions. The Ludlow and particularly the Pridol feature low abundance and diversity (Fig. 19). In the Nyuya-Berezovo District, rugose corals are found only in the Wenlock. Rugose corals are unknown in the Irkutsk Subregion.

Crinoids — This group is found throughout east Siberia in the upper range of the deep-shelf, as well as shallow-shelf, and lagoonal settings. They are most numerous and diverse in the shallow-shelf and shoaly facies of the North-Yenisey and Pritunguska Subregions as benthic and nektic forms. The crinoids are most abun-

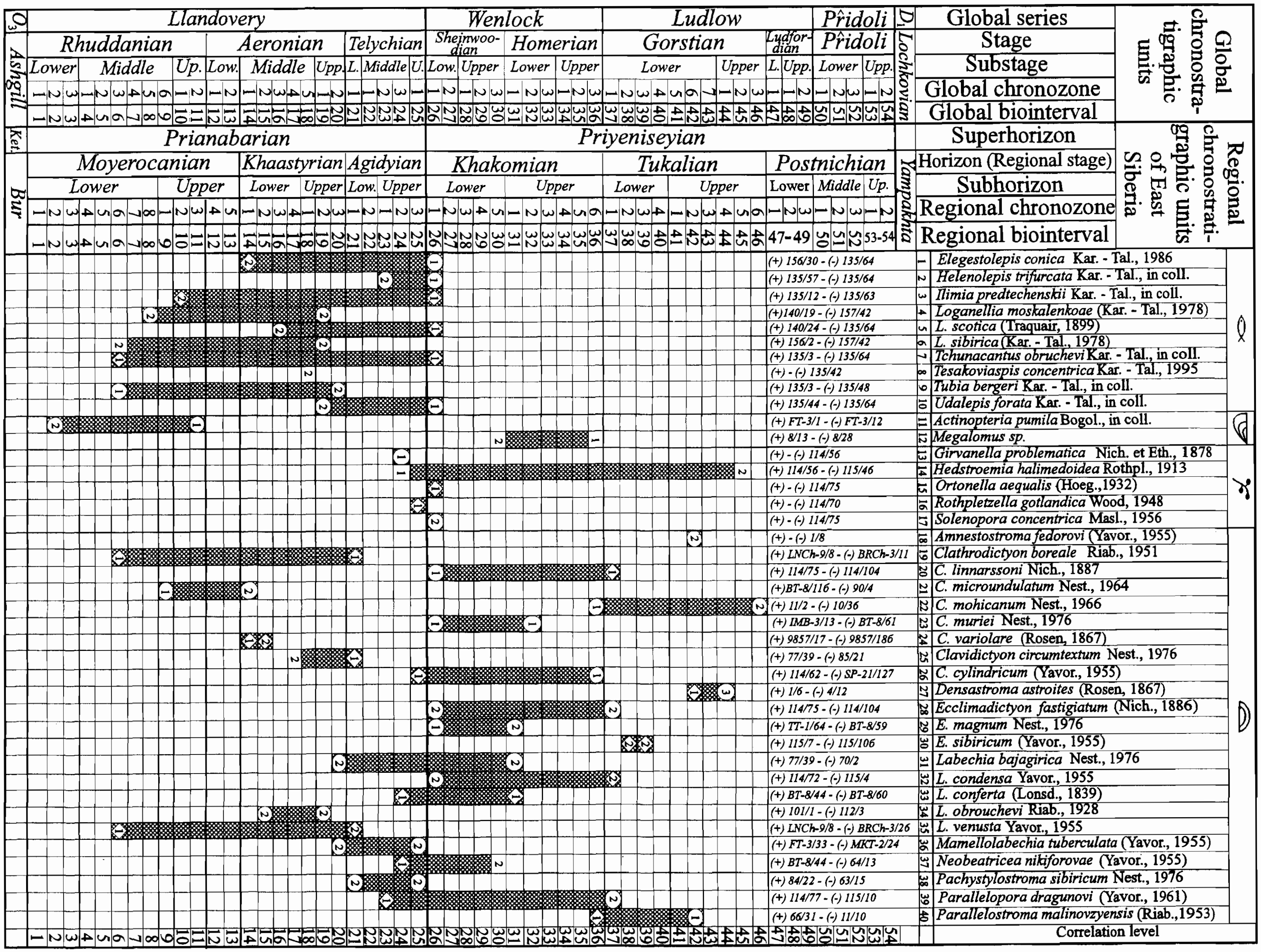

FIGURE 17 — Distribution of fish (left side), bivalves (left-middle), algae (center), and stromatoporoids (right side, continued on Fig. 18) in the Silurian of East Siberia. For symbols, see Fig. 10.

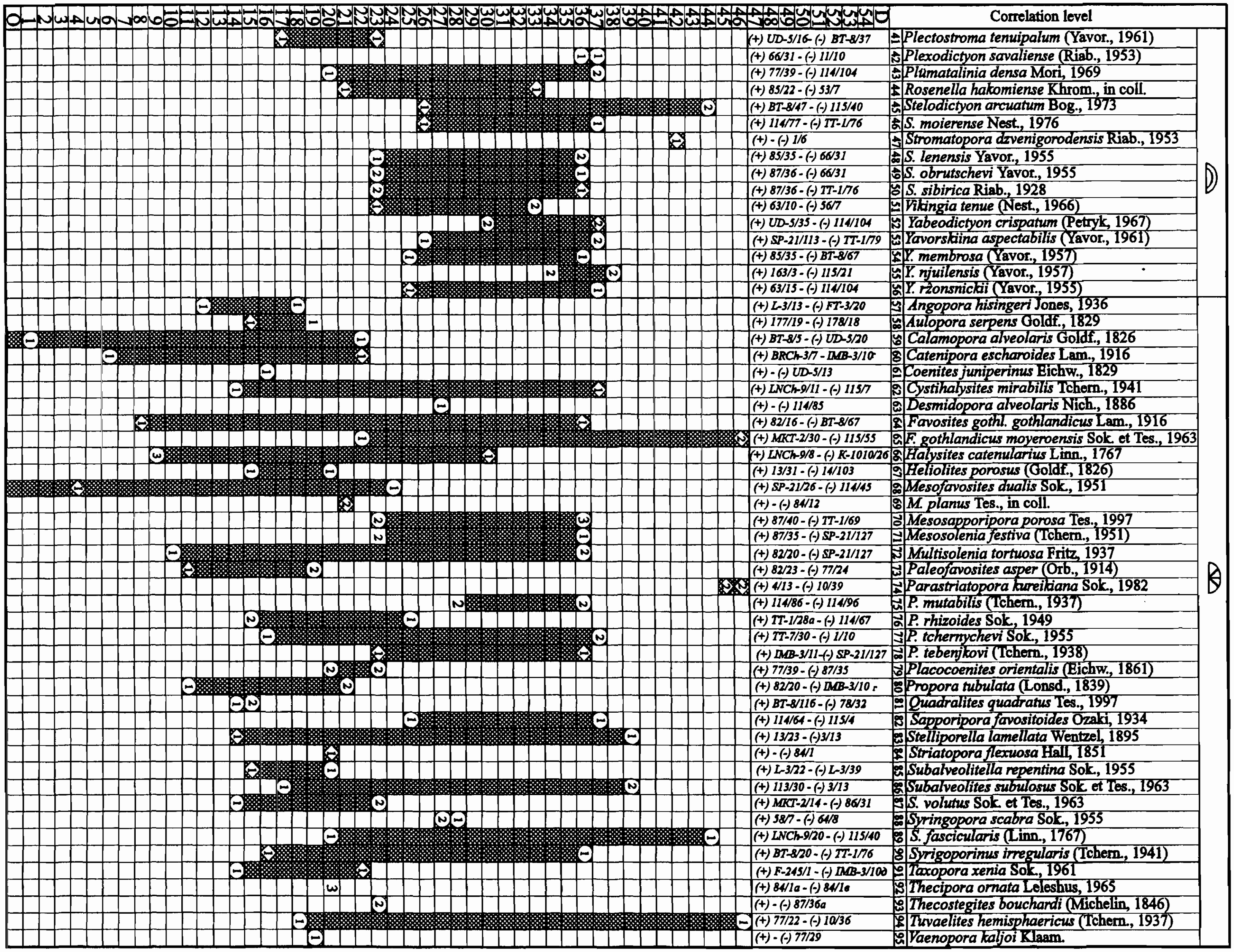

FIGURE 18 — Distribution of stromatoporoids (left side, continued on Fig. 17) and tabulate corals (right side) in the Silurian of East Siberia. For symbols see Fig. 10.

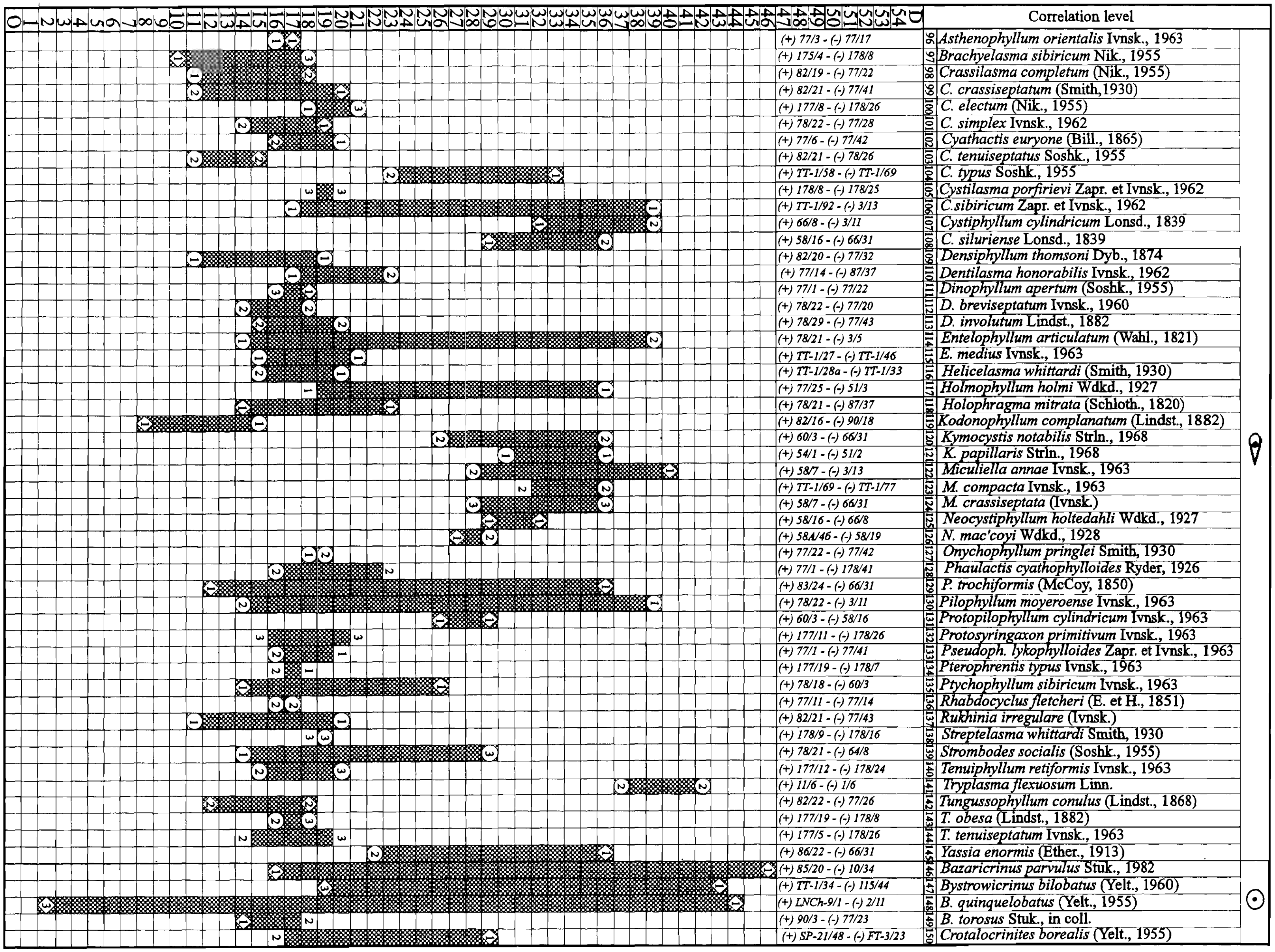

FIGURE 19 — Distribution of rugose corals(left side) and crinoids (right side, continued on Fig. 20) in the Silurian of East Siberia.
For symbols see Fig. 10.

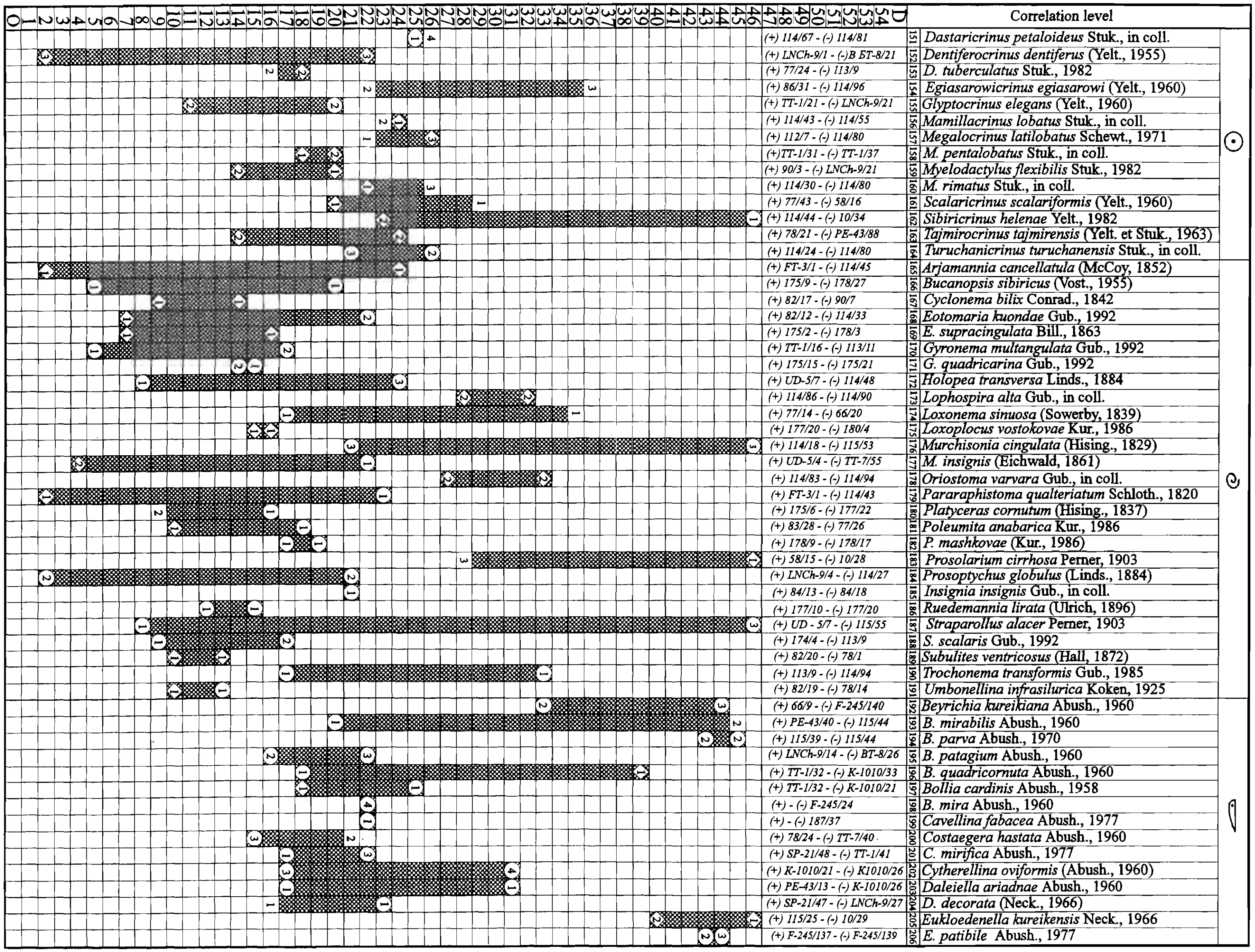

FIGURE 20 — Distribution of crinoids (left side, continued on Fig. 19), gastropods (middle), and ostracodes (right side, continued on Fig. 21) in the Silurian of East Siberia. For symbols see Fig. 10.

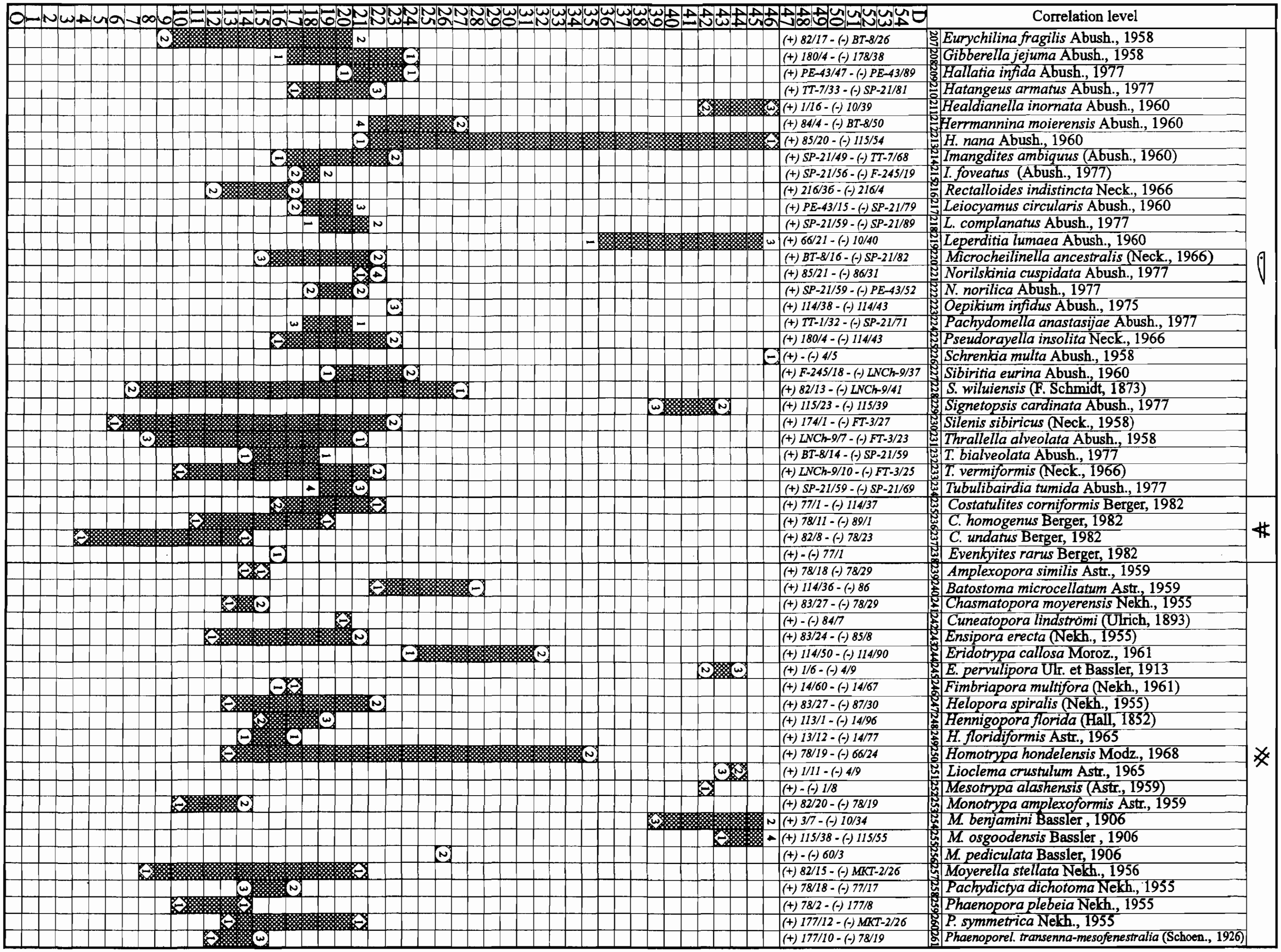

FIGURE 21 — Distribution of ostracodes (left side, continued on Fig. 20) and bryozoans (right side, continued on Fig. 22) in the Silurian of East Siberia. For symbols see Fig. 10.

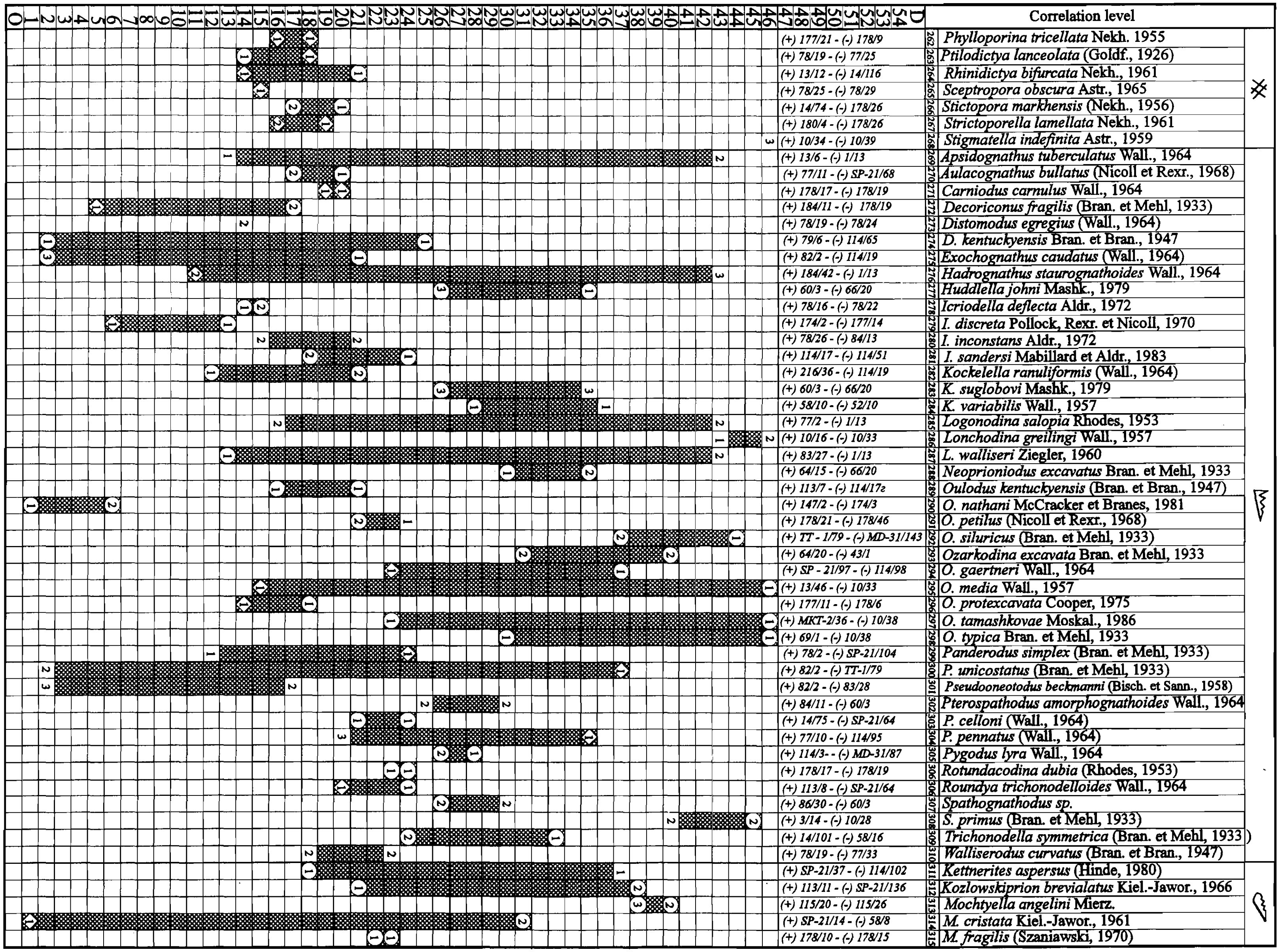

FIGURE 22 — Distribution of bryozoans (left side, continued on Fig. 21), conodonts (middle), and scolecodonts (right side, continued on Fig. 23) in the Silurian of East Siberia. For symbols see Fig. 10.

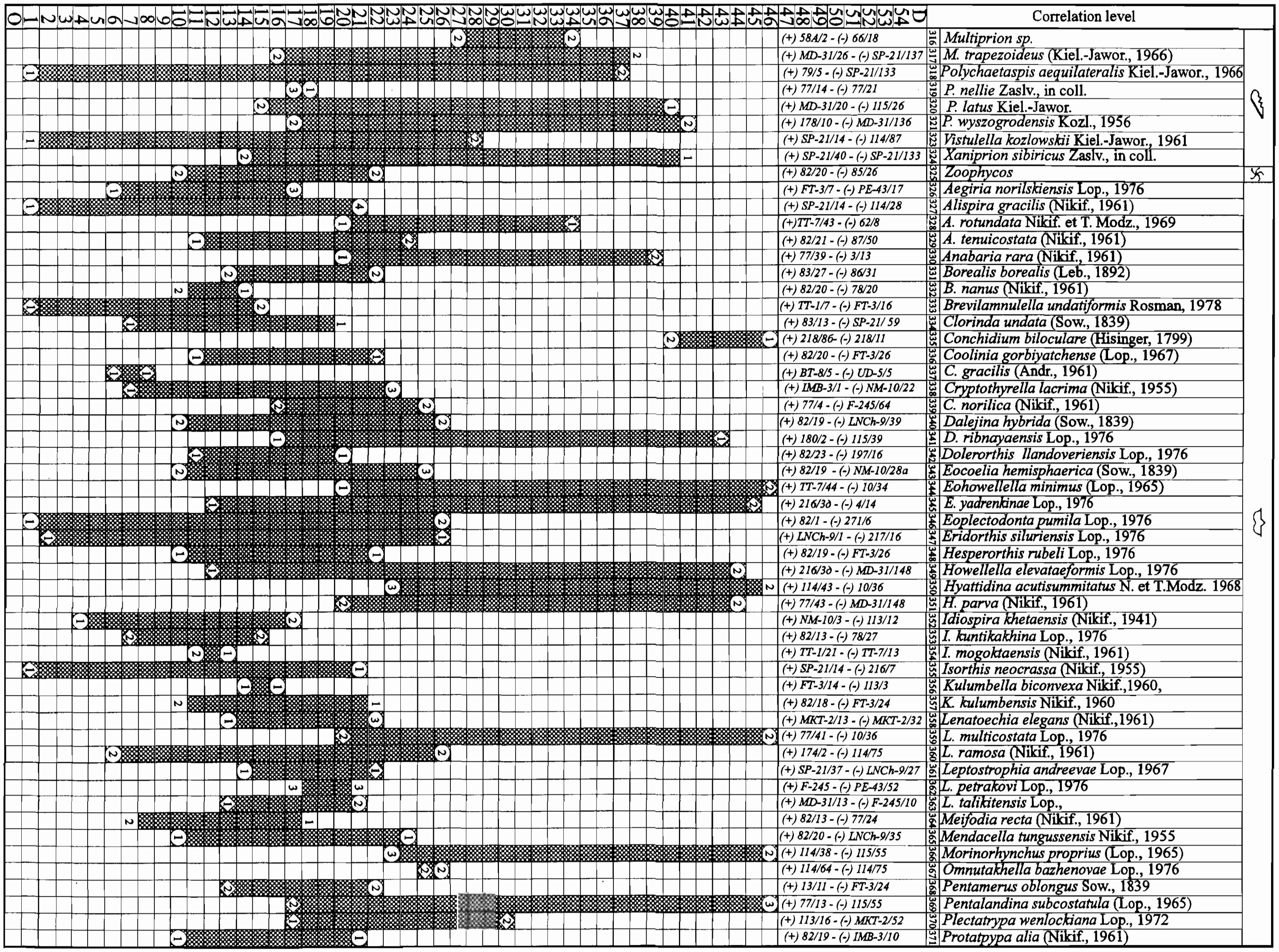

FIGURE 23 — Distribution of scolecodonts (left side, continued on Fig. 22), trace fossils (*Zoophycos*), and articulate brachiopods (right side, continued on Fig. 24) in the Silurian of East Siberia. For symbols see Fig. 10.

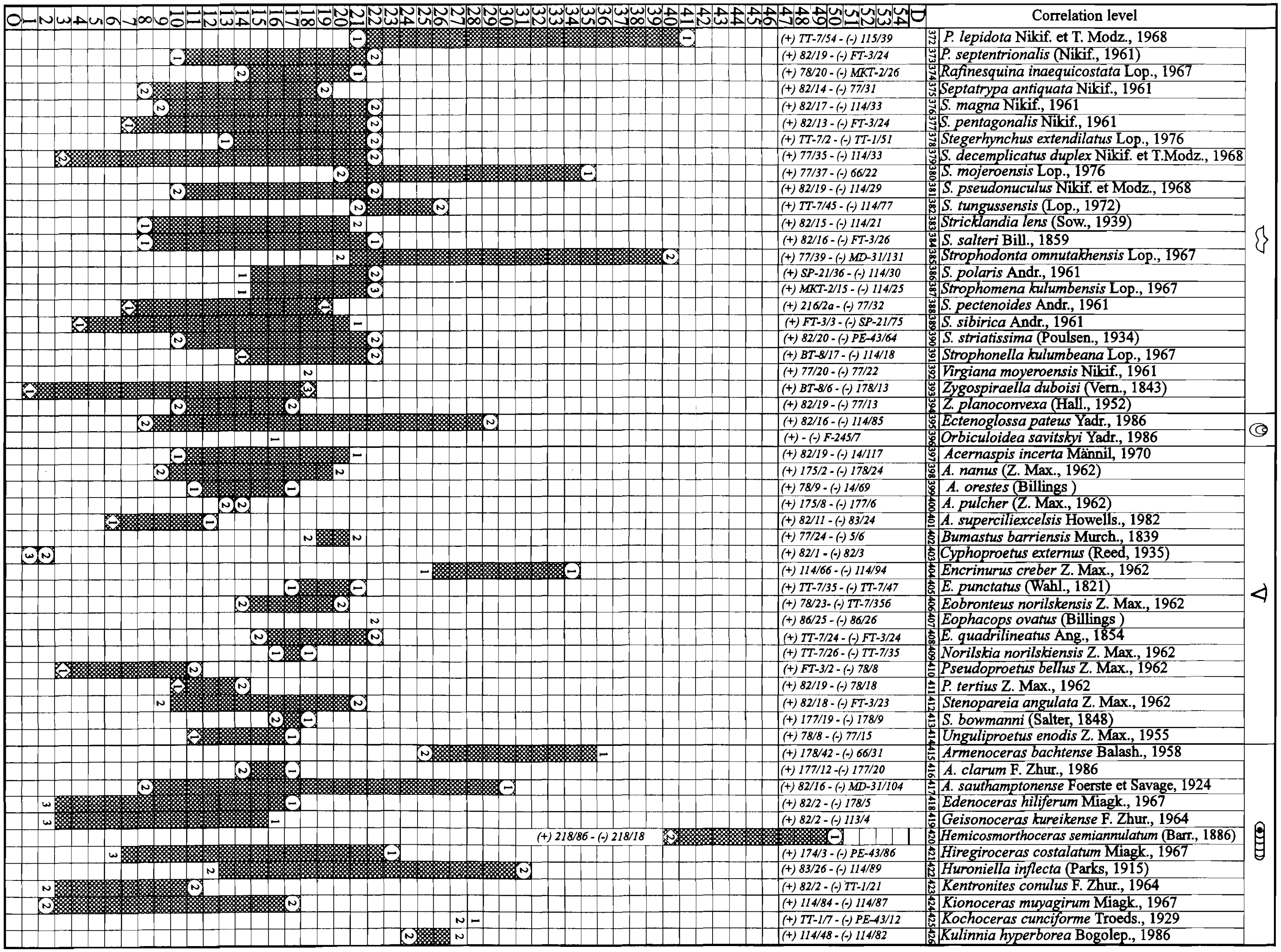

FIGURE 24 — Distribution of articulate brachiopods (left side, continued on Fig. 23), inarticulate brachiopods, trilobites, and cephalopods (right side, continued on Fig. 25) in the Silurian of East Siberia. For symbols see Fig. 10.

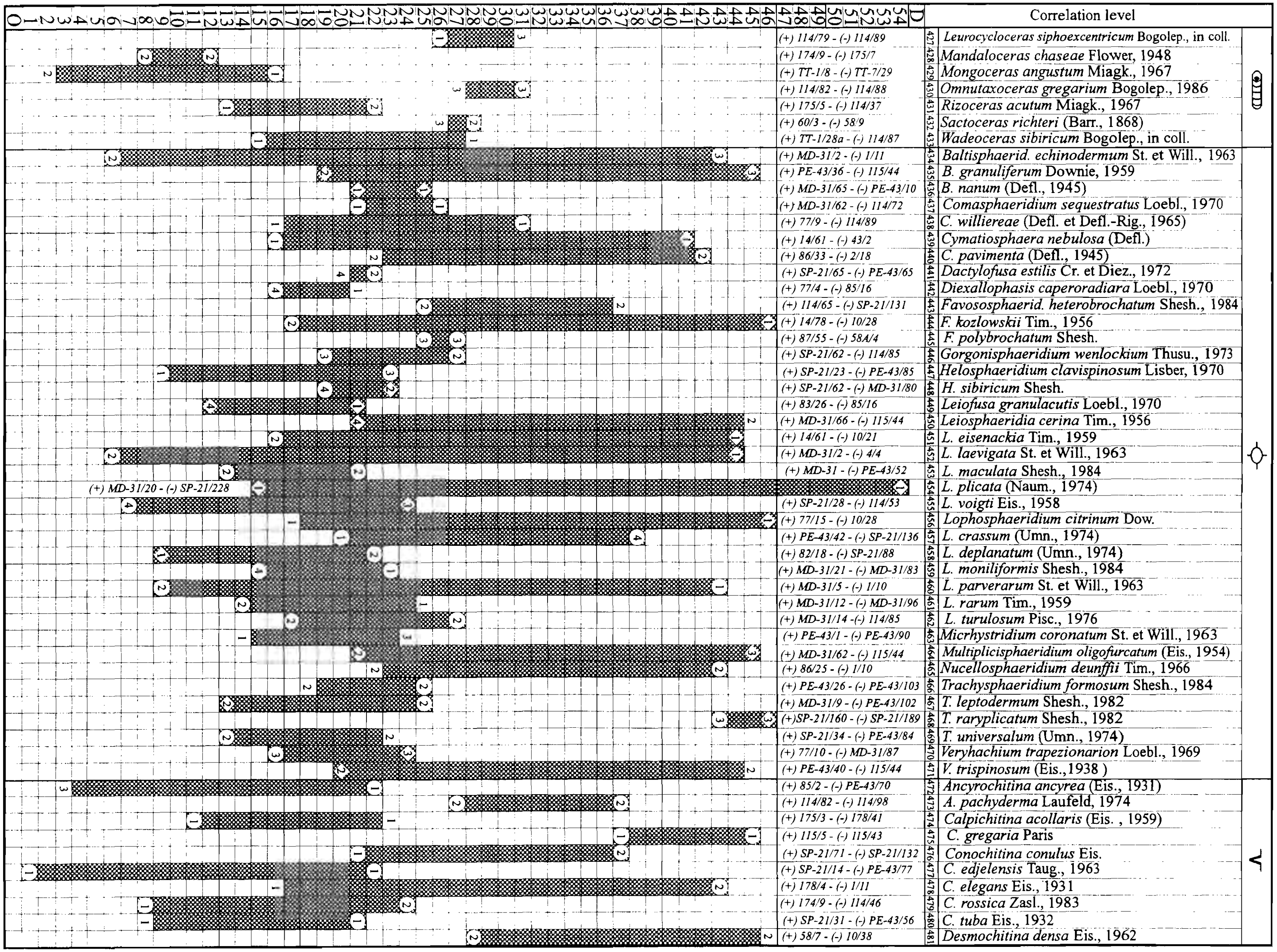

FIGURE 25 — Distribution of cephalopods (left side, continued on Fig. 24), acritarchs (middle), and chitinozoans (right side, continued on Fig. 26) in the Silurian of East Siberia. For symbols see Fig. 10.

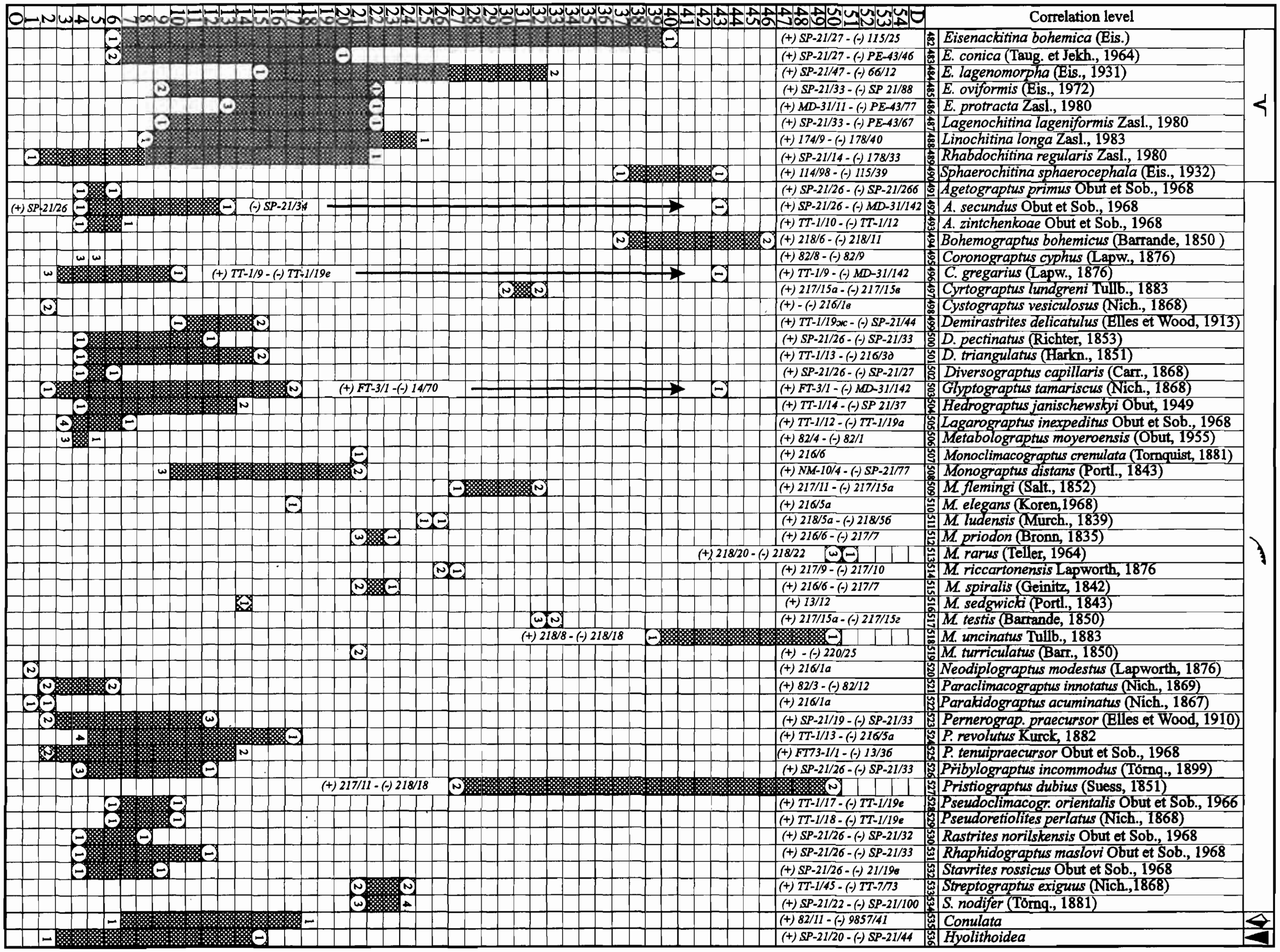

FIGURE 26 — Distribution of chitinozoans (left side, continued on Fig. 25), graptolites, conularids, and hyoliths (right side) in the Silurian of East Siberia. For symbols see Fig. 10.

dant in the Aeronian–Wenlock, and become rare in the Ludlow and very rare in the Pridoli (Figs. 19, 20).

PROBLEMATICA — Among the most interesting problematical forms are rhombic, flat carbonate plates with parallel ribs on the upper surface. These plates are likely derived from "cystoid" echinoderms. They are abundant but confined to Wenlock reefs.

OSTRACODES — This group is dominant in Silurian facies in East Siberia. They are most abundant and diverse in shallow shelf and lagoonal facies of the North-Yenisey and Pritunguska Subregions. Shallow-shelf facies with siliciclastic mudstones have numerous specimens in monotonous, low-diversity assemblages. Shallow carbonate-shelf settings have higher diversities. Deep-shelf, reef, and gypsiferous lagoonal facies have rare ostracodes. They are most abundant and varied in the Aeronian, Telychian, and Ludlovian (Figs. 20, 21).

BRYOZOANS — Bryozoans are abundant throughout East Siberia in shallow, open-shelf facies. They are less common and diverse in lagoonal facies. Higher diversity assemblages are present in the Aeronian and Telychian of the Pritunguska Subregion and in Wenlockian and Ludlovian reef and shoal facies in the North-Yenisey and Pritunguska Subregions (Figs. 21, 22).

BRACHIOPODS — These are one of the most numerous Silurian groups in East Siberia (Figs. 23, 24). They are very abundant and diverse throughout the sequence, except for the Pridoli and in the Nyuya-Berezovo and Irkutsk Subregions. Brachiopods are highly lithofacies-specific in East Siberia. They are smaller and less diverse in siliciclastic mudstones than in carbonate. They frequently form coquinite lenses and beds. Monospecific accumulations are characteristic of the Ludlow.

CEPHALOPODS — This group (Figs. 24, 25) is diverse and abundant only in the lowermost Silurian (lower–middle Rhuddanian), where conchs occur in clusters in black limestones and concretions in shales in the North-Priyenisey and Pritunguska Subregions. Concentrations of specimens are found in the Wenlock of these subregions; the diversity is rather low, and the specimens are poorly preserved (mainly siphuncles and other fragments). They are rare at other Silurian levels in these subregions. In the North Taymyr Subregion, generally rare and low-diversity cephalopod assemblages occur throughout the Silurian in siliciclastic mudstone and black limestone. In the Nyuya-Berezovo and Irkutsk Subregions they are solitary, of poor preservation, and can be found only at separate levels.

GASTROPODS — These are common in shallow-shelf, shoaly, and lagoonal facies. They are most numerous and diverse in the Aeronian, Telychian, and Ludlovian of the North-Yenisey and Pritunguska Subregions. Their preservation is often poor, but most specimens can be identified at the level of the species (Fig. 20).

BIVALVES — Clams occur throughout the Silurian across East Siberia, although they are not common and show low diversity (Fig. 17). They are not adequately studied.

HYOLITHS — Abundant hyoliths occur only in the lower and middle Rhuddanian of the North-Priyenisey Subregion (Fig. 26). They show low diversity in black shale and calcareous concretions.

GRAPTOLITES — This group (Fig. 26) is abundant and diverse and dominates fossil assemblages through the Silurian of the North-Taymyr Subregion. In the North-Priyenisey Subregion, they are abundant and diverse only in the Rhuddanian; the Aeronian and Telychian have rare specimens. Graptolites are common, though not diverse and found only in the lower–lower middle Rhuddanian and lowermost Telychian in the Pritunguska Subregion. These fossils are not found in other subregions. They occur in black and green shale but are rare in black limestone.

CONULARIDS — Rare specimens occur in the Llandovery of the North-Priyenisey and Pritunguska Subregions. They occur in black shale and gray nodular marl (Fig. 26).

CHITINOZOANS — Relatively diverse assemblages occur in the Silurian of East Siberia (Figs. 25, 26). They are abundant in the North-Priyenisey and Pritunguska Subregions, but the local Wenlock and Pridoli assemblages are somewhat limited. Chitinozoans are not yet studied from the Taymyr Silurian. They are absent in the remaining subregions of East Siberia. They are common in black and green shales, but are most abundant in marl, marl with limestone nodules, and occasionally in clayey nodular limestones. They are rare and poorly preserved in non-argillaceous nodular and platy limestone.

ACRITARCHS — This group (Fig. 25) is abundant and diverse in Llandovery black shale and marl of the North-Priyenisey Subregion. They are rather numerous in the Pritunguska Subregion at a few levels in Llandovery siliciclastic mudstone, marl, and marl with limestone nodules. The Wenlock of these subregions has rare acritarchs. They are virtually unknown in the Ludlow and Pridoli. Acritarchs have not been studied from the Taymyr Peninsula. The Nyuya-Berezovo and Irkutsk Subregions have not yielded acritarchs.

TRILOBITES — This group (Fig. 24) is abundant and diverse only in the Llandovery of the Priyenisey and Pritunguska Subregions, where they are found in laminated and lensing calcarenite. In the rest of the sections in these subregions and in the North-Taymyr Subregion, they are rare and less diverse. Trilobites are particularly rare in the Nyuya-Berezovo and Irkutsk Subregions.

TRACE FOSSILS — Deposit-feeder burrows are widely distributed in most Silurian facies of East Siberia, except for graptolitic and gypsiferous facies. Three basic groups of trace fossils are common. The first group includes straight and dichotomously branching, small, vertical-to-horizontal burrows. These burrows characterize practically all facies. The second group includes very large burrows that frequently cross and occur on lower bedding surfaces from the shallow shelf. The third group is of the *Zoophycos* type (Fig. 23).

SCOLECODONTS — Rare specimens (Figs. 22, 23) that form low-diversity assemblages occur in shallow-shelf and shoal facies in the North-Yenisey and Pritunguska Subregions.

CONODONTS — This group (Figs. 22) is widespread in shallow-shelf facies in the North-Yenisey and Pritunguska Subregions. They are practically absent in graptolite mudstone, reef, and gypsum-bearing lagoonal facies. They have not yet been studied from the Taymyr Peninsula.

TENTACULITIDS AND CORNULITIDS — These are uncommon groups in the North-Yenisey and Pritunguska Subregions, where they occur in shallow-shelf, argillaceous carbonates mainly in the Aeronian and Telychian. Rare specimens occur in Wenlock and Ludlow shoaly facies. Their species diversity is low.

EURYPTERIDS — These arthropods occur largely in the Ludlow–Pridoli of the North Yenisey and Pritunguska Subregions and in the Llandovery–Wenlock of the Nyuya-Berezovo Subregion. They have not been studied, although complete specimens with fully preserved exoskeletons are known.

FISH (ACANTHODIANS AND TELEODONTS) — Fish (Fig. 17) are widespread in Llandovery lagoonal and in-shore facies of the Nyuya-Berezovo and Irkutsk Subregions, but uncommon in the Pritunguska Subregion. Many new local forms are known.

PALEOGEOGRAPHY

Pre-plate tectonic paleogeographic maps that include east Siberia were first published in the *Atlas of Lithologo-Paleogeographic Maps of the USSR* (Vinogradov, 1967-1969) and the *Atlas of Litho-Paleogeographic Maps of the World* (Ronov et al., 1984). More recent paleogeographic maps that feature Silurian reconstructions are available (Tesakov, 1981a; Predtetchensky et al., 1985).

The East Siberian epicontinental basin developed on the Siberian paleocontinent (Zonenshain et al., 1990), where it is bounded to the south (in modern coordinates) by the Mongol-Okhotsk suture. In the northern Pribaikal area, its boundary runs along the Patom arc. In the southeast, its margin follows the East-Sayan fracture, and on the Yenisey Ridge the border lies in the submeridional thrust zone. Under the cover of the West Siberian lowland in the west, this boundary probably runs along a line to the mouth of Ob' Bay. The northern and eastern boundaries are largely hidden under the overthrusted Taymyr and Verkhoyansk folded complexes. An actual boundary is only found exposed near the mouth of the Bunge River in the Taymyr Peninsula, where Lower Paleozoic shales are in fault contact with coeval carbonates of the Siberian continent. Apparently, the shales were deposited in a relatively narrow (oceanic ?) strait between the Arktida and Siberian continents (Zonenshain and Natapov, 1987). The fault zone is about 10 km wide, but it is difficult to reconstruct the amount of convergence along this fault zone. It has been estimated at 20–50 km. To the east, the boundary of the Siberian paleocontinent is drawn at Verkhoyansk along the Chersky Ridge suture zone, where deep-water shales are in contact with shallow-water carbonates.

There are two principal proposals for the paleogeographic position of the East Siberian epicontinental basin in the Silurian. One view is that the basin was in the northern subtropical-subequatorial zones, and the other suggests that the Siberian continent was rotated relative to its present position. Based on a wide variety of reports (Scotese et al., 1979; Zonenshain and Savostin, 1979; Ronov et al., 1984; Zonenshain and Natapov, 1987; Zonenshain et al., 1990; Khramov et al., 1992) and our own investigations, we believe (exclusive of V. G. Khromych, who is not an adherent of plate tectonics) that the East Siberian Basin was rotated clockwise ca. 156° and lay at 8–40° N and 65–75° E. The basin was on the southern margin of the Siberian paleocontinent and was open on the south and southeast to the Mongolo-Urals Ocean. On the southwest and west, it was bounded by the Anabar Plain, and on the north by a rather high Angara land. To the northeast, the Tobolsk accumulative coastal plain formed the border. In the main, sediments were transported into the basin from Anabar and Angara land. This interpretation is suggested by a decreased thickness of Silurian in the Maymecha and Morkoka Districts, the presence of basal conglomerates with quartz and quartzite pebbles, and the occurrence of thick sandstones in the Balturino District. West and southwest winds were dominant on the Siberian paleocontinent during the Silurian, as suggested by ripple mark orientations in the Irkutsk and Nyuya-Berezovo Subregions. Based on the Ca/Mg ratio in biogenic carbonates, it is evident that a latitudinal shift in paleotemperature from the Irkutsk Subregion towards the North-Yenisey Subregion existed. Throughout the Silurian, the northern Siberian platform was closer to the equator than its southern part. Widespread buildups, large colonies of tabulate and rugose corals and stromatoporoids, thick-walled valves of mollusks, and the seasonal bands of many tabulate corals support the interpretation that the basin was in the subtropical–subequatorial zone. It is not improbable, however, that the Siberian paleocontinent moved during the Silurian from the southern subequatorial zone to the northern subtropical zone. Widespread Early Silurian organic-rich black shales and reefs suggest that the climate was warm and humid and that the basin was tropical. On the other hand, red gypsiferous and gypsum beds are widespread in the Upper Silurian, and this suggests a change to an arid climate and a shift to more temperate latitudes.

The six paleogeographic reconstructions summarized below demonstrate the Silurian development of the East Siberian Basin. These maps are based on three lines of evidence: 1) ecosystems (i.e., seemingly "biofacies;" editors' note) reflected by biogeocoenosis (i.e., faunal assemblage; editors' note) and catena (i.e., the area of a lithofacies belt; editors' note) (Sukachev, 1945; Walter, 1968; Tesakov, 1978, 1981a); 2) lithologic facies, as developed by N. N. Predtetchensky for the Siberian and Russian platforms (Tesakov et al., 1980, p. 120–138); and 3) bionomic zones (i.e., "fossil community assemblage;" editors' note) and their depths (Ziegler et al., 1968, 1977; Boucot, 1975; Johnson, 1987).

EARLY RHUDDANIAN — This interval (Fig. 27) is characterized by ten catenas. The first and second catenas occupied broad areas of the northwest basin. They were confined to the muddy deep shelf that preserved graptolite biocoenoses. The third catena was a shallow-water area with carbonaceous mud and abundant small brachiopods. The fourth catena was apparently the stagnant, deeper part of a shallow shelf with carbonates and abundant small brachiopods. The fifth catena developed on a shallow shelf, where peloidal, carbonaceous mud with abundant small ostracodes was deposited. The sixth catena occupied wide areas in the south and southeast basin, and was a low-relief, shallow-shelf region with carbonaceous mud and with solitary, euryhaline tabulates. The seventh catena was confined to a southern shallow bay with a sandy bottom. The eighth catena was characterized by round-stone cobbles in local Late Ordovician reefs. The ninth catena was probably related to the tidal zone. The tenth catena is treated only in a preliminary way. It was possibly a highland in the southern part of the region. The boundaries of the epeiric sea basin, as well as those of many marginal catenas, are quite arbitrary and run mainly parallel to the Silurian outcrop. The development of the shoreline in the northeast basin has been discussed (Tesakov, 1967; Johnson et al., 1997).

EARLY MIDDLE AERONIAN — This interval (Fig. 28) is characterized by nine catenas. The first and second catenas occupied extensive areas in the northwestern part of the basin. They were confined to deep-shelf clayey mudstone with lenses of carbonaceous mudstone and widespread graptolites. The third catena was associated with shallow-shelf carbonaceous muds and abundant corals and brachiopods. Bioherms occur along the inner part of this catena in the northeast. Large coral banks developed in the central part of the sea. The succession of biocoenoses in this catena has not yet been adequately studied. The fourth catena was related to a shallow-water zone, where carbonaceous-detritic muds with reworked sediments and abundant corals and brachiopods occur. The fifth catena formed in an isolated part of the basin under shallow-water conditions, and was characterized by dolomitic clayey mudstone with lenses of brachiopod-bearing lime mudstone. The sixth catena developed in an isolated shallow bay with dolomitic clayey mud where fish were widespread. The seventh catena was located in a southern shallow bay of the basin, where aleurolitic muds occurred and fish were abundant. The eighth catena was associated with the inshore part of the southern bay, where well-winnowed sands accumulated. The ninth catena is rather arbitrarily defined in the northeast of the region, and was likely related to a slightly elevated land source. In the south, the land was considerably higher, and terrigenous material was eroded from it.

EARLY SHEINWOODIAN — This interval (Fig. 29) is characterized by ten catenas. The first catena developed in the northwest basin, and was associated with deposition of relatively thick, carbonaceous muds and graptolite biocoenoses. The second catena apparently corresponded to the deeper part of the shallow shelf, where peloidal muds and small, low coral bioherms were associated. The third catena was also related to the deeper part of the shallow shelf, and had reefs up to 40 m high in the Khyukta (Norilsk District) and Kureyka (Turukhansk District). The fourth catena was characterized by limestone and locally clayey mudstone with corals. The fifth catena developed across a wide belt with dolomitic lime mudstones with extensive coral-stromatoporoid biostromes and low bioherms. The sixth catena was associated with shallow-water dolomitic mudstone and local stromatolite buildups. The seventh catena was confined to isolated, shallow bays in the southeastern and southern basin, with dolomitic, clayey mudstone with lenses of dolomitic mudstone and stromatolite build ups. The eighth catena formed in an inshore part of the southern shallow bay and included dolomitic, clayey mudstone and abundant fish. The ninth and tenth catenas were confined, respectively, to the low and relatively high surrounding lands.

EARLY GORSTIAN — This map (Fig. 30) is characterized by ten catenas. The first catena lay in the northwest basin. It was confined to the deep shelf and featured clayey mudstone with carbonate nodules, lenses of lime mud,

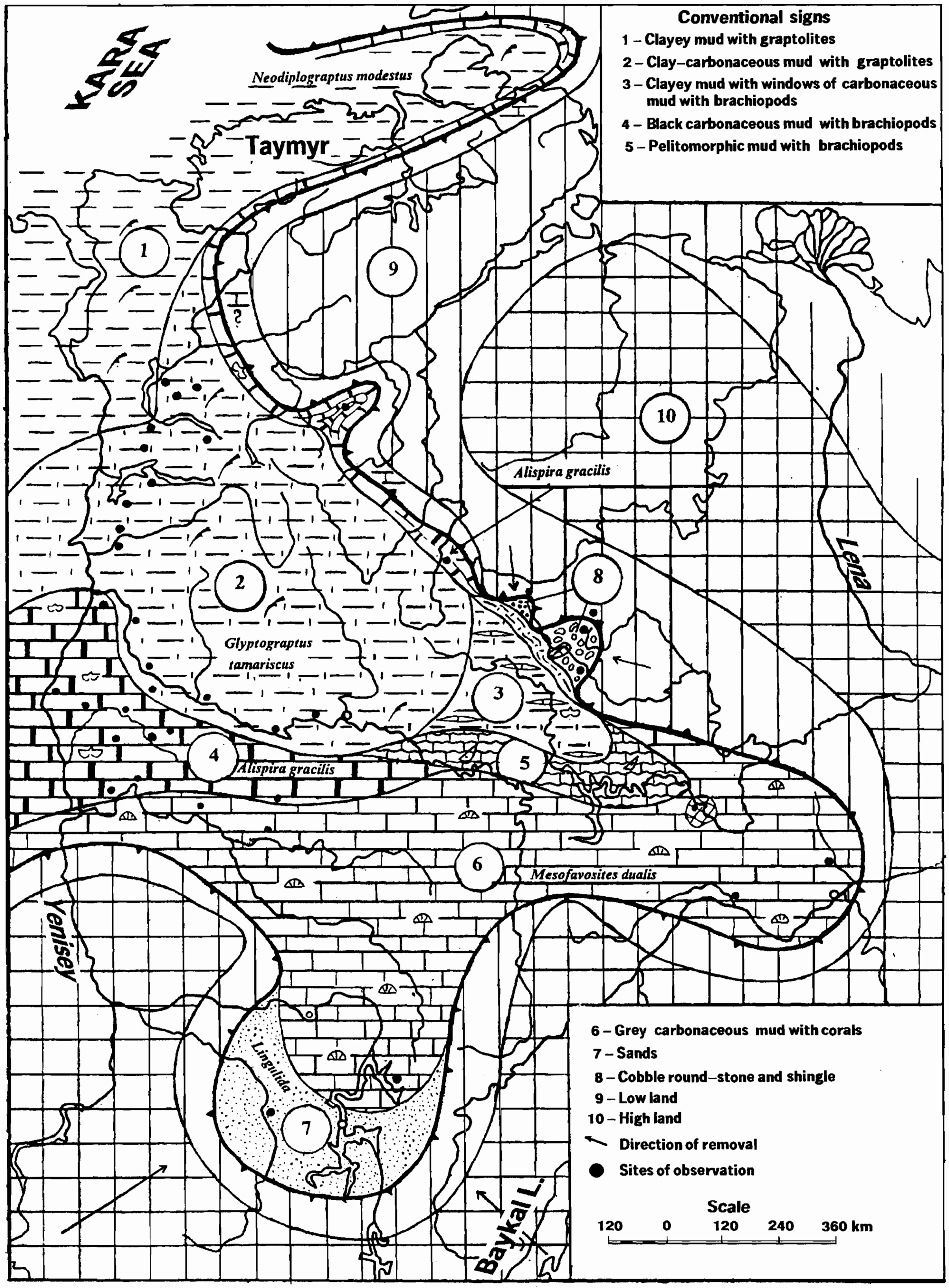

FIGURE 27 — Paleogeographic map of the East Siberian Basin in the early Rhuddanian.

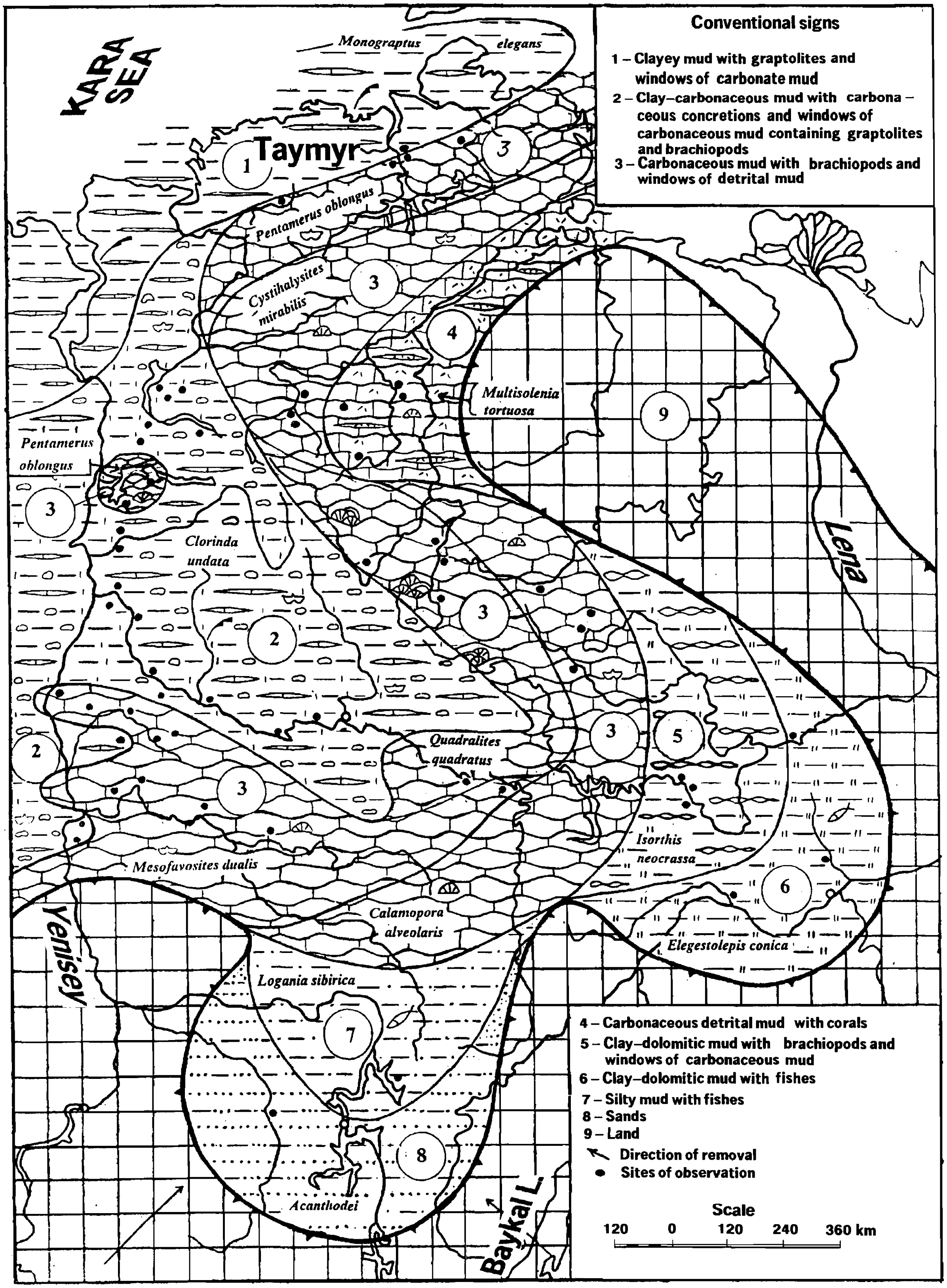

FIGURE 28 — Paleogeographic map of the East Siberian Basin in the early middle Aeronian.

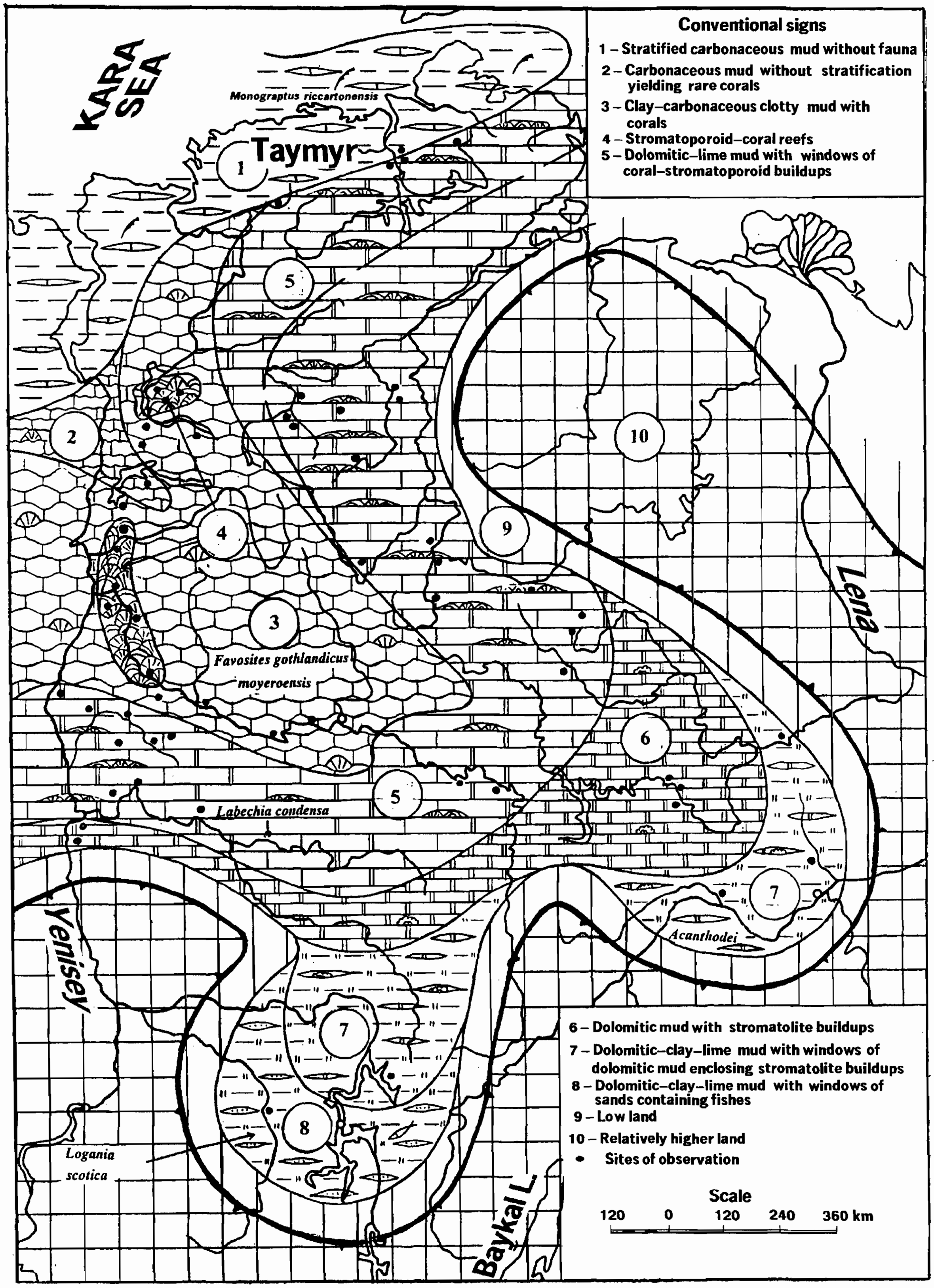

FIGURE 29 — Paleogeographic map of the East Siberian Basin in the early Sheinwoodian.

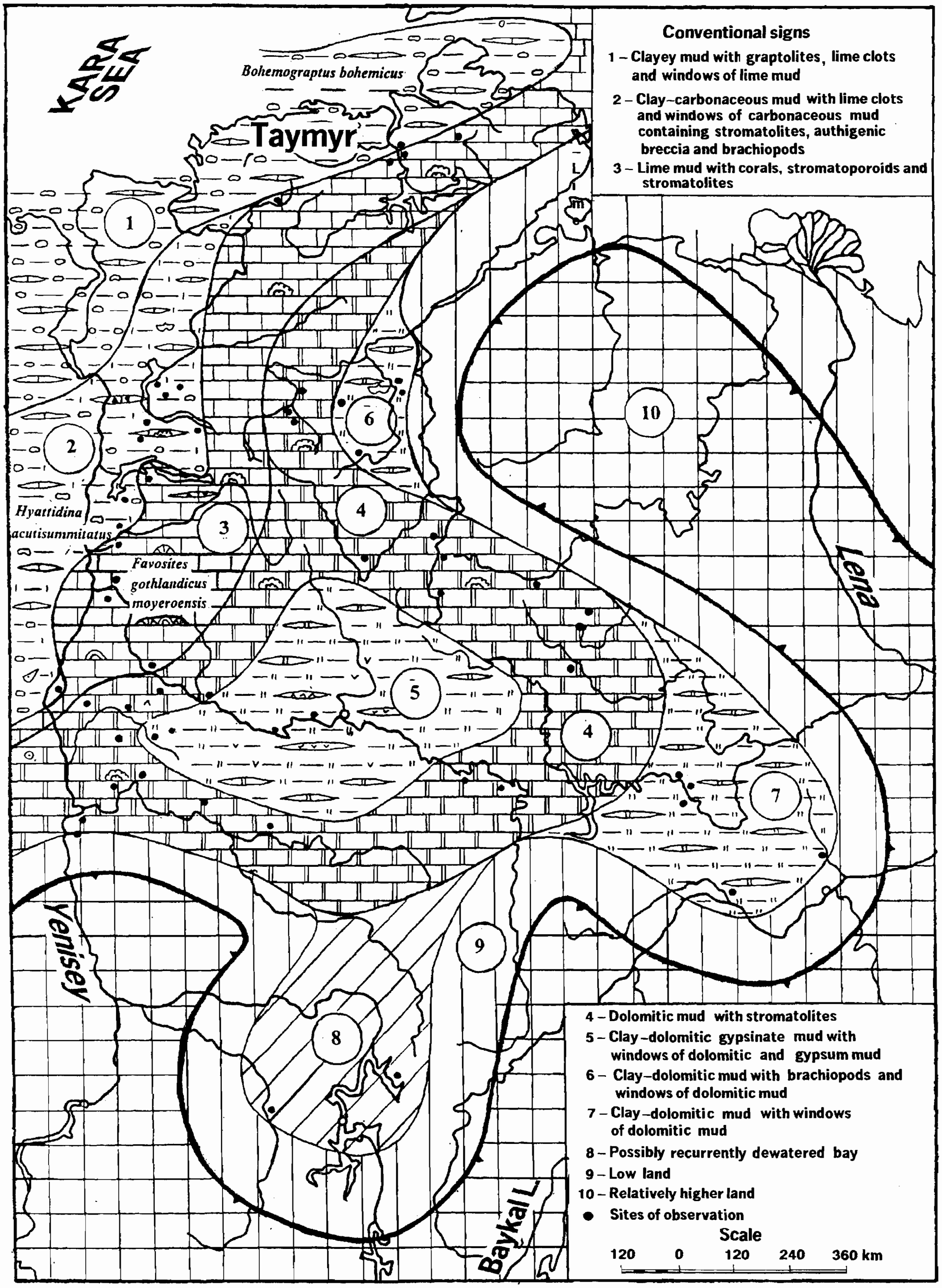

FIGURE 30 — Paleogeographic map of the East Siberian Basin in the early Gorstian.

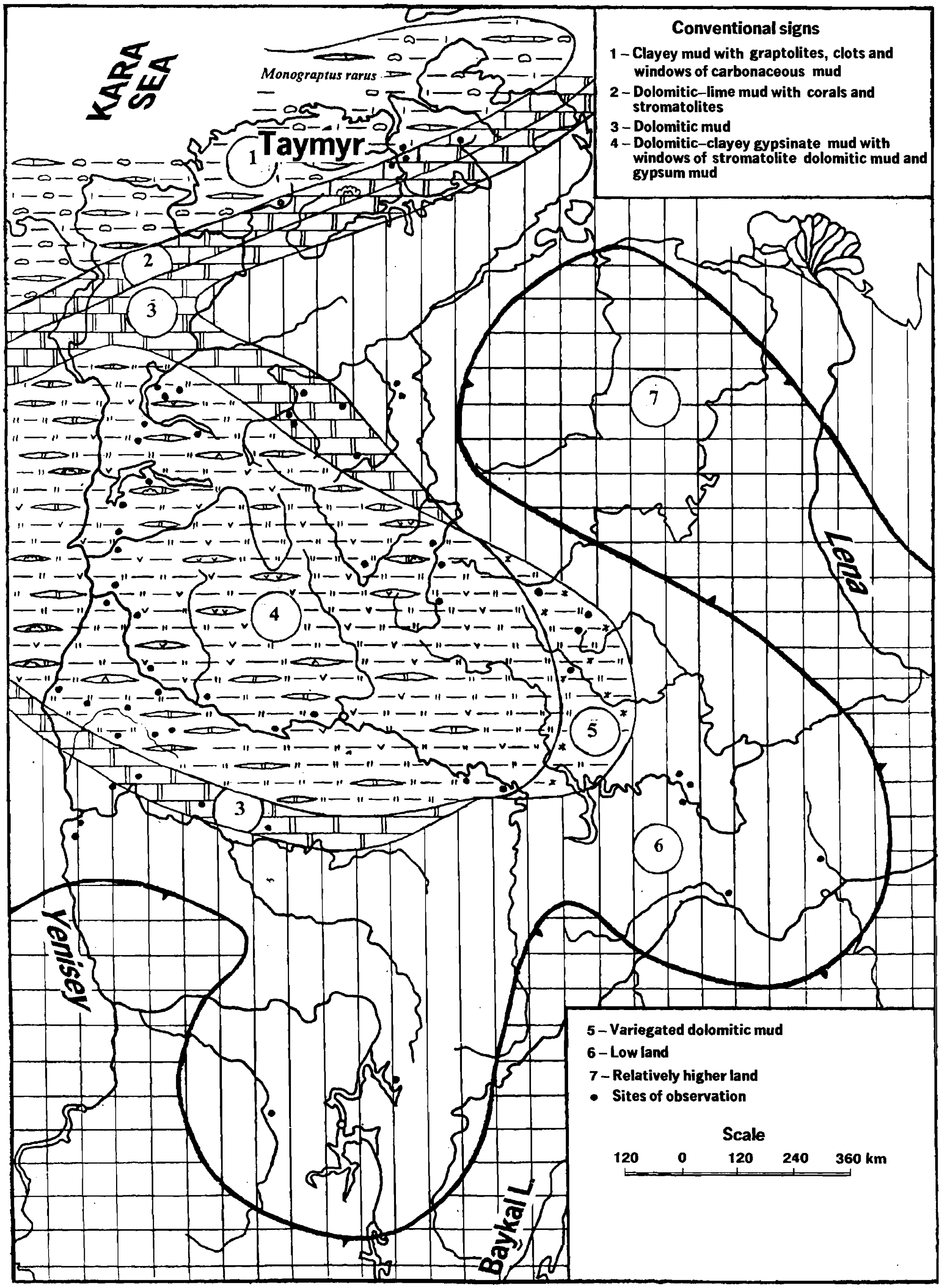

FIGURE 31 — Paleogeographic map of the East Siberian Basin in the early Pridoli.

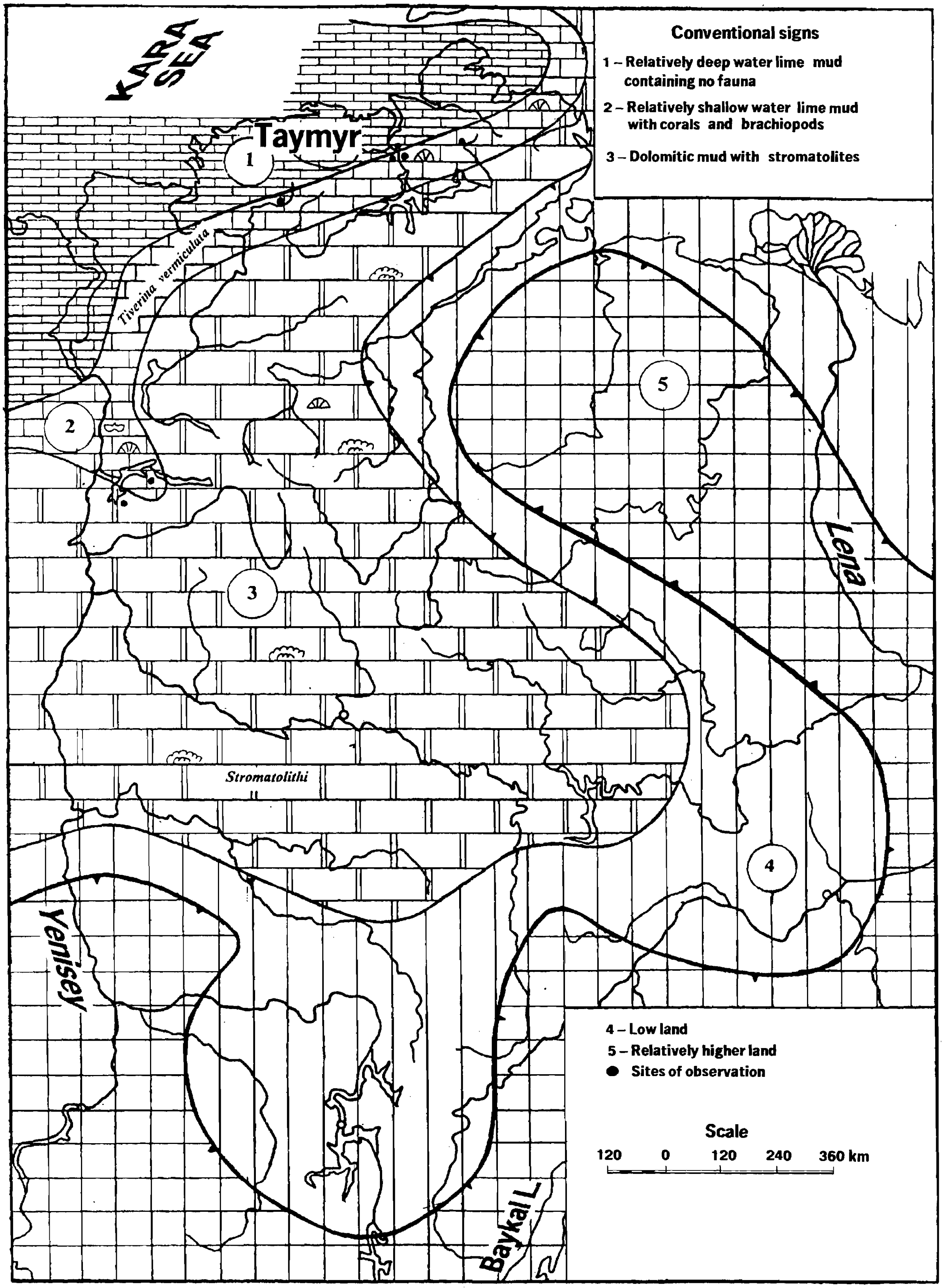

FIGURE 32 — Paleogeographic map of the East Siberian Basin in the early Lochkov.

and graptolites. The second catena was related to shallow-shelf deposition of argillaceous lime mudstone with carbonate nodules, lenses of lime mud, stromatolites, intraclast breccia, and widespread brachiopods. The third catena featured shallow-water, lime mudstone with local low stromatolites. The fourth catena was in a shoaly zone with dolomitic mudstone and stromatolites. The fifth catena was in the shallow-water, highly saline, central part of the basin, where clay-rich, dolomitic mudstone with dolostone, anhydrite, and gypsum lenses were deposited. The sixth catena was inshore, with dolomitic clayey mudstone with rare small brachiopods. The seventh catena was associated with a shallow-water bay in the southeast. Non-fossiliferous, dolomitic clayey mudstones with dolomite lenses were deposited. The eighth-tenth catenas were apparently confined to lands of somewhat different heights.

EARLY PRIDOLI — This interval (Fig. 31) is characterized by seven catenas. The first catena occurred in the northwest basin and was confined to a deep shelf where clay-rich, carbonaceous mudstones with calcareous nodules and lime mudstone lenses were deposited. Graptolites were widespread. The second and third catenas were fairly shallow with calcareous and dolomitic mudstone. In the outer zone (catena 2), rare tabulate corals and stromatolites occurred. The first and second zones are in contact along a thrust zone. The fourth catena was related to an inner, very shallow, hypersaline part in the basin. This catena is represented by non-fossiliferous dolomitic and argillaceous mudstone with lenses of dolostone, anhydrite, and gypsum. The fifth catena featured inshore, variegated (mainly red) dolomitic mudstone. The sixth and seventh catenas were apparently confined to surrounding highlands. The shore of the marine basin paralleled the Silurian outcrop.

EARLY LOCHKOVIAN — This Early Devonian interval (Fig. 32) featured a basin-wide change of facies following a minor transgression. The entire basin in East Siberia was shallow, and gently subsided from the southeast to the northwest. The first and second catenas were characterized by deposition of lime mudstone. The first catena lacks fossils, and the second one contains isolated tabulates and large build-ups. The third catena is represented by dolomitic mudstone with isolated, euryhaline tabulates and stromatolite build-ups. The fourth and fifth catenas corresponded to surrounding highlands.

CONCLUSIONS

A correlation chart for the Silurian of East Siberia (Siberian platform and Taymyr Peninsula) has been developed that outlines regional (horizons) and local stratigraphic units (formations), shows appearance and extinction events, and correlates regional and local units with the global scale and with the horizons in adjacent regions (Altay-Sayan and northeast Russia). The stratotypes and hypostratotypes for all fifteen districts of East Siberia can be correlated at the level of subhorizons (regional substages). The facies across the profile of this Silurian basin can be traced from the deep basin into the near shore. The biostratigraphic ranges of the more common biotic elements have been documented. Paleogeographic changes during development of the East Siberian Basin have also been reconstructed. These data are the basis for synthesizing the history of sedimentation and development of the biologic world during the Silurian in East Siberia.

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SILURIAN LANDS AND SEAS

Drawing from Linnaeus' field notes made at the Upper Silurian section at Kyllaj, Gotland (compare with cover illustration). *From M. Åsberg and W. T. Stern's translation "Linnaeus's Öland and Gotland Journey 1741" (1973, published for the Linnean Society of London by Academic Press).*