

Cambrian stratigraphy and depositional history of the northern Indian Himalaya, Spiti Valley, north-central India

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ABSTRACT

Recent work on Himalayan tectonics indicates that prior to the Cenozoic collision of India and Asia, an enigmatic Cambrian–Ordovician event may have strongly influenced the regional geology of the Himalaya. Stratigraphic and sedimentological analyses of well-preserved Cambrian deposits are critical for understanding the nature of this early tectonic event and its influence on the later tectonic evolution of the Himalaya. The Parahio Formation, defined herein, of the Parahio Valley, Spiti region, in the Tethyan Himalaya of India, is the best biostratigraphically resolved section of Cambrian strata in the entire Himalaya. This formation consists of >1350 m of dominantly siliciclastic deltaic deposits. The formation ranges from uppermost Lower Cambrian (Lungwangmiaoan Stage) to middle Middle Cambrian (Hsuehchuan Stage), representing a time span of ~5–10 m.y. It contains numerous medium-scale shoaling cycles that range from storm-influenced offshore deposits to thick trough cross-bedded fluvial facies. Many thin carbonate beds with abundant trilobite fossils directly overlie the fluvial facies and represent transgressive systems tract deposits. The cycles are interpreted to have resulted from delta-lobe switching, based on a lack of systematic stratigraphic changes in cycle or

facies thicknesses. This paleoenvironmental reconstruction contradicts previous interpretations of this unit that range from deep-sea flysch to shallow-marine tidalites. In addition, our paleoenvironmental analysis and paleocurrent data suggest that the uppermost Lower to Middle Cambrian deposits of the Lesser and Tethyan Himalaya are parts of the same ancient northward-prograding, fluvial-deltaic depositional system of the paleo-Tethys margin of India.

An angular unconformity with overlying Ordovician conglomeratic rocks has considerable local relief, with meter-scale scours and a valley fill >100 m thick. The scours have northeast-southwest orientations, which parallel both the paleocurrents in the underlying Parahio Formation as well as published paleocurrent readings from the coarse red beds of the overlying Ordovician strata. The Cambrian–Ordovician unconformity is of regional extent, and our recent biostratigraphic database indicates that the minimum hiatus associated with the unconformity in Spiti is ~15 m.y. Our sedimentological analysis and associated paleocurrent data from the Parahio Formation, along with additional data from units both above and below the unconformity, indicate that published models portraying foreland basin development at this time with southward-directed thrusting are problematic. An alternate possibility—that uplift took place south of the Tethyan Himalaya—is also problematic, because no

published stratigraphic or structural evidence exists for such an uplift to the south for either the Greater or the Lesser Himalaya lithotectonic zones.

Keywords: Cambrian, Parahio Formation, India, Tethyan Himalaya, stratigraphy.

INTRODUCTION

The Himalaya consist of three principal lithotectonic zones, based on internal stratigraphy and bounding faults, which were activated during the Cenozoic collision of India with Asia (Gansser, 1964; LeFort, 1975) (Fig. 1). To the south, the Lesser Himalaya contains thick Neoproterozoic through upper Lower Cambrian (and likely younger Cambrian) strata that are unconformably overlain by Permian and younger rocks. The central zone, the Greater Himalaya, consists of high-grade metamorphic rocks. The northernmost Tethyan Himalaya contains an extensive record of deposition from the Precambrian to the Eocene. The paleogeographic affinities of the three zones are the subject of much debate. Nd isotope and geochronological data have been used to suggest that the Lesser Himalaya has distinctly older sedimentary source materials than the Greater Himalaya and the Tethyan Himalaya, and that the two latter zones together represented a separate terrane and cover succession, respectively, that was accreted to India during the early Paleozoic (DeCelles et al., 2000). Myrow et al. (2003) demonstrated that previously published

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Lesser Himalayan data were skewed by the sampling of only older Precambrian strata and that samples from younger units of similar depositional age (late Early Cambrian) from the Tethyan Himalaya and the Lesser Himalaya are similar to each other and to the protolith of the Greater Himalaya. This eliminated the bulk of evidence for an exotic origin of Greater Himalaya and Tethyan Himalaya rocks. Other tectonic models of the early tectonic history of the Himalaya are based on reconstructions of the northern Indian margin during the Cambrian and Early Ordovician (Garzanti et al., 1986; Gehrels et al., 2003), including an enigmatic tectonic event represented in northern India by a prominent Cambrian–Ordovician boundary unconformity. However, the sedimentological and stratigraphic context of the Cambrian Tethyan sedimentary succession, on which such reconstructions rely, is poorly known.

The Cambrian Parahio Formation of the Spiti Valley region of northern India, which lies within the Tethyan Himalayan fold-thrust belt that formed along the northern Indian passive margin during deformation in the Eocene (Searle, 1986; Steck et al., 1993; Corfield and Searle, 2000; Wiesmayr and Grasmann, 2002), is particularly important in this regard. It is a thick and well-preserved sedimentary Cambrian succession that has never been subjected to a detailed sedimentary analysis. The formation represents the oldest Tethyan sedimentary rocks with abundant and well-preserved fossils and sedimentary structures. Chronostratigraphic data for this section is superior to that currently available for any other Cambrian section within the Tethyan Himalaya. We herein provide the first detailed sedimentological analysis and depositional history of the Parahio Formation and discuss the tectonic implications of its overlying unconformity.

GEOLOGIC SETTING OF FIELD AREA

The Spiti Valley lies in the Lahaul-Spiti district of the state of Himachal Pradesh, northern India. The field area for this study is in the Parahio River Valley (Figs. 1, 2), a subsidiary valley of the Pin River Valley, which joins the Spiti Valley ~40 km southeast of the town of Kaza. Low-grade Neoproterozoic to Eocene sedimentary rocks of the Tethyan Himalayan fold-thrust belt exposed within the Spiti Valley are bounded to the south by the high-grade gneisses of the Greater Himalaya across the Sangla Detachment (Vannay and Grasmann, 1998; Wyss et al., 1999; Vannay et al., 2004). The Sangla Detachment is one of a series of north-dipping, extensional shear zones that are generally referred to as the South Tibetan detachment system (Searle, 1986; Herren, 1987; Burchfiel et

al., 1992; Vannay and Grasmann, 1998; Dezes et al., 1999; Wyss et al., 1999; Vannay et al., 2004). The northern boundary of this belt is the south-dipping Cenozoic Great Counter Thrust, which marks most of the Indus-Tsangpo suture zone, separating rocks of the Indian and Eurasian plates to the south and north, respectively (Yin et al., 1999; Yin and Harrison, 2000).

The sections of the Parahio Formation described in this study (Fig. 3) are on the southwest limb of the Muth Syncline, 10–15 km west of the village of Mikkim, on steep slopes on the north side of the Parahio River, near the confluence of the Khemangar River and the Debsa Khad River (Fig. 2). This section was first studied by Hayden (1904). The uppermost part of the section was measured southeast of the village of Thango on the east side of the Sumna River (also known as the Kinlung Nala) just above where it joins the Parahio River.

The base of the Parahio Formation and its contact with the Batal Formation is not exposed within the area. The units are juxtaposed by the Khemangar River Fault, which is partly covered by the river. The fault strikes north-northwest, dips steeply to the east, and displays a normal sense of stratigraphic offset (Figs. 2, 4). The orientation of the Khemangar River Fault is similar to a series of north-striking Cenozoic extensional faults ~30 km to the southeast in the Sutlej Valley area (Vannay et al., 2004). However, the fault is also parallel to north-northwest-striking outcrop-scale strike-slip faults that we measured in exposures near the confluence of the Khemangar and Debsa Khad Rivers (Fig. 4).

PREVIOUS WORK

Hayden (1904) performed the first and most comprehensive study of the Parahio Formation to date. He established a basic stratigraphy of this dominantly siliciclastic succession and collected a Middle Cambrian trilobite fauna from several zones within the unit (Reed, 1910; Jell and Hughes, 1997). His measured section from the Parahio River area shows several mixed siliciclastic-carbonate cycles in what he called the *Cambrian System*. He described upward-coarsening depositional cycles that consist of a thick lower shale unit and an overlying sandstone unit that are capped by a thick, orange-weathering dolostone bed. The environments of deposition for the Parahio Formation have been variably interpreted. Several workers considered the section as a record of an upward transition from deeper-water euxinic and flysch facies to shallow-water deposits with thick, orange-weathering dolostone beds (Srikantia, 1981; Bhargava et al., 1982; Srikantia and Bhargava, 1983). Fuchs (1982) also conducted work in

the Spiti region and considered the formation to be of flysch origin. An intertidal to tidal-flat environment has been invoked for the upper part of the section (Bhargava et al., 1982; Bhargava and Bassi, 1998; Srikantia and Bhargava, 1998, p. 227), and also for intervals within the lower part (Bhargava et al., 1986).

The thickness of the Parahio Formation has long been debated. Hayden (1904, p. 13) noted that in the Parahio Valley “numerous folds can be seen thus proving that the enormous thickness [of the Parahio Formation] is only apparent.” He concluded that the entire section between the Khemangar River Fault and the Cambrian–Ordovician unconformity is not more than ~600–900 m thick, but noted that this estimate was “merely approximate” (Hayden, 1904, p. 15). Other estimates for the thickness of the entire formation in the Parahio Valley are <400 m (Parcha, 1996; Fig. 2), 1080 m (Bhargava and Bassi, 1998), ~2000 m (Fuchs, 1982), 2700 m (Kumar et al., 1984; Bhargava et al., 1986), and 3350 m (Srikantia, 1981). Our detailed bed-by-bed measurement of the section revealed that structural complications were not an issue and that a nearly complete measured section was possible. Our section of the Parahio Formation (Fig. 3) is 1352 m thick. An additional ~100–200 m of poorly exposed Parahio Formation exists at the base of the section but was not measured. The sedimentology of this interval is directly comparable to that of the rest of the measured section.

Biostratigraphic data help to resolve the discrepancy between Hayden’s thickness estimate and our own. His trilobite collections are now assigned to 16 species from shale beds (Jell and Hughes, 1997). New collections were recovered from some of Hayden’s shale zones, and for the first time from carbonate beds, and both lithologies contain similar taxa. Data from all the zones from which we collected are in the expected stratigraphic order, based on comparison with biostratigraphic databases from sections in China. As the order of stratigraphic occurrence of these taxa is similar in both Hayden’s and our studies, the positions of our collections with respect to height in the section indicate that Hayden (1904) consistently underestimated stratigraphic thickness.

Trace fossil assemblages that include ichnogenera attributed to *Astropolichnus* and *Plagiogmus* occur below our measured section within the poorly exposed basal 100–200 m of the Parahio Formation, and were interpreted to be Early Cambrian in age (Bhargava et al., 1982; Bhargava et al., 1986). This is consistent with the age of the oldest trilobites in our section. We have no firm constraint on the age of the base of the Parahio Formation in the Parahio

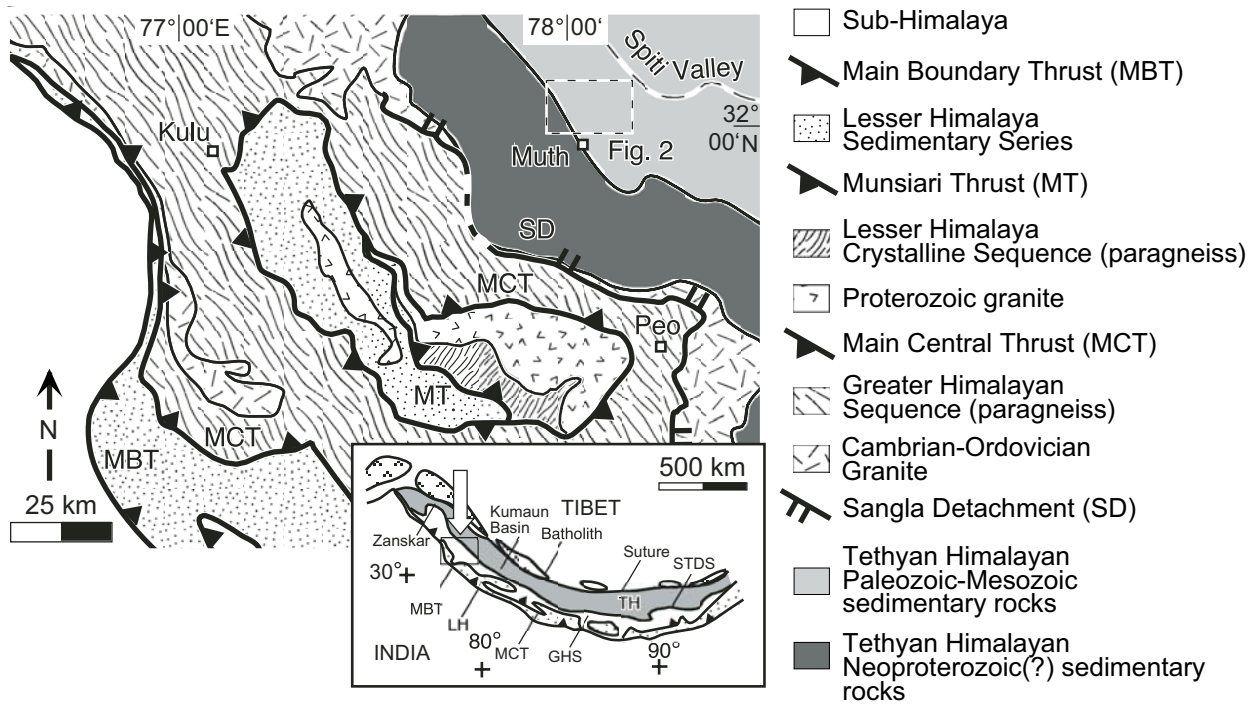


Figure 1. Regional geology, showing major faults and geologic units of the northwestern Himalaya in India (modified from Vannay and Grasemann, 2001). Box delineates region shown in Figure 2. STDS—South Tibetan detachment system; TH—Tethyan Himalaya; GHS—Greater Himalaya sequence; LH—Lesser Himalaya.

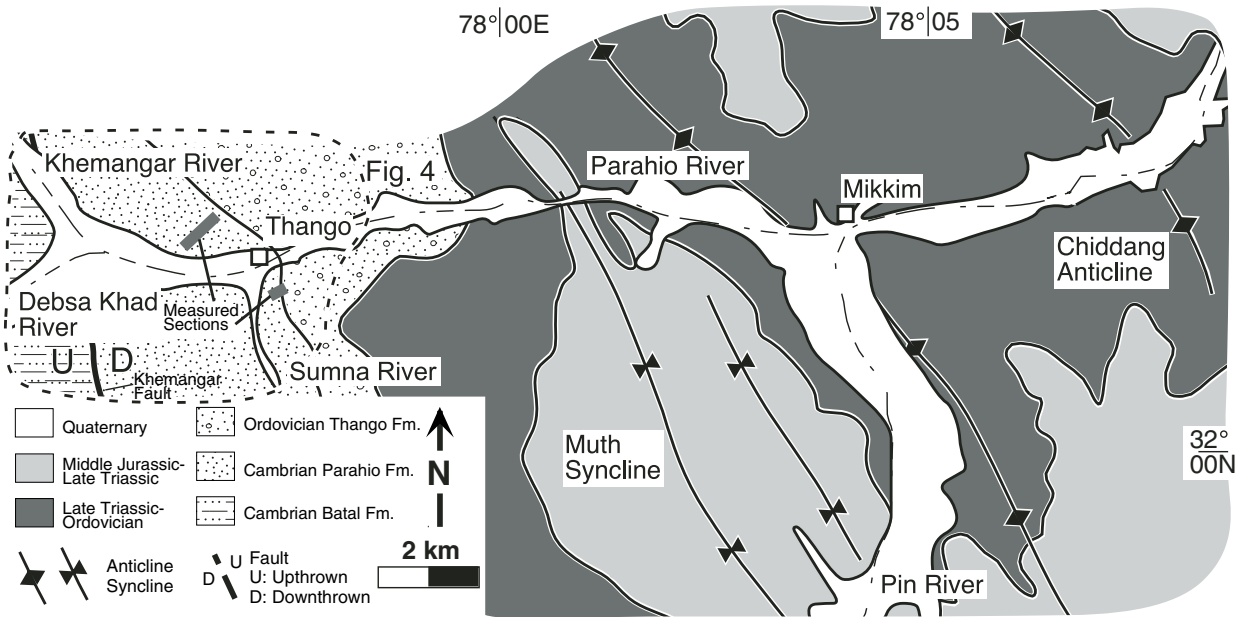


Figure 2. Simplified geologic map of the Tethyan Himalayan fold-thrust belt in the area of the Parahio and Pin Valleys (compiled and modified from Fuchs, 1982; Bhargava and Bassi, 1998; Wiesmayr and Grasemann, 2002). Map area shown as boxed region in Figure 1.

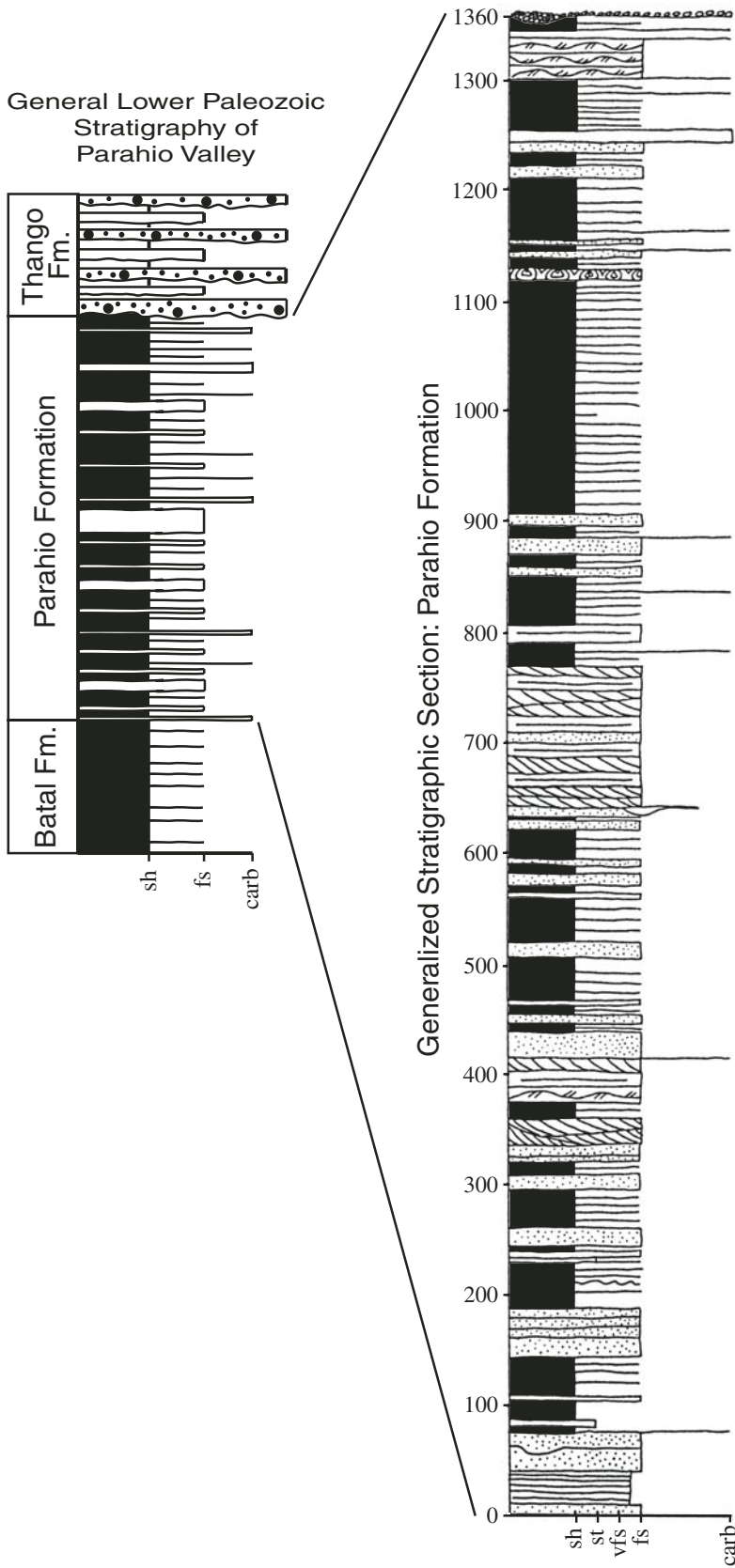


Figure 3. Generalized lowermost Paleozoic stratigraphy of the Parahio Valley and stratigraphic section of the Parahio Formation, measured in meters. Explanation: sh—shale; st—siltstone; vfs—very fine sandstone; fs—fine sandstone; carb—carbonate.

Valley but note that the late Tsanglangpuan trilobite *Redlichia noeltingi* has been recorded in the adjacent Pin Valley (Reed, 1910; Jell and Hughes, 1997), and it is likely that the trace fossils are late Early Cambrian in age.

On the basis of our new collections of trilobites, the Parahio Formation in the Parahio Valley ranges from uppermost Lower Cambrian (Lungwangmiaoan Stage) to middle Middle Cambrian (Hsuehuangian Stage). The base of the formation is almost coeval with the youngest dated Cambrian strata of the Lesser Himalaya, the upper Lower Cambrian of the Tal Group (Hughes et al., 2005).

STRATIGRAPHY

The stratigraphic nomenclature for the oldest sedimentary rocks of the Tethyan Himalaya is confusing (Fig. 5), owing to the paucity of well-described stratigraphic sections and the lack of chronostratigraphic constraints. Revision of the lithostratigraphy and chronostratigraphy of the Parahio Valley is a critical first step toward clarifying regional relationships. Griesbach (1891), following an initial investigation of the geology of the Spiti region, applied the term *Haimanta series* to a lowermost package of sedimentary rocks above the metamorphic rocks of the Greater Himalaya (Fig. 5). This term was modified by Hayden (1904, p. 14–15), who suggested that all of Griesbach’s (1891) *Haimanta series* lies below an angular unconformity that is now known to be no younger than Late Ordovician, and he referred to the rocks beneath the unconformity collectively as the *Cambrian System* (Fig. 5). We concur with Bhargava and Bassi (1998, p. 19) that all sedimentary rocks below the unconformity in the Parahio Valley section should be assigned to the *Haimanta Group*. This group is also considered to contain sedimentary rocks older than those known in the Parahio Valley, such as the Chamba Formation and Manjir Formation in Khasmir (Frank et al., 1995), which are likely Neoproterozoic in age, consistent with radiometric dates on crosscutting granitic intrusions from the Tethyan Himalaya (Miller et al., 2001). Our use of the term *Haimanta Group* excludes beds above the unconformity at the top of the Parahio Formation, in contrast to the usage of others (e.g., Srikantia, 1981; Bhargava et al., 1982) but in accordance with recent practice.

Pascoe (1959) applied the term *Parahio series* to Cambrian sedimentary rocks described by Hayden (1904) from the Parahio Valley (Fig. 5). This unit comprised “grey and green micaceous quartzites and thin-foliated slates and shales, with narrow bands of light grey dolomite” (Pascoe, 1959, p. 581) and repeated upward-coarsening shale-sandstone cycles with

carbonate caps. Hayden's Cambrian trilobite-bearing rocks from the Parahio Valley are within this unit, the upper boundary of which was defined by the unconformity. The base of the Parahio series was considered conformable with the underlying red and black slate, with quartzite beds assigned by Pascoe (1959) to the upper part of the Haimanta series. Pascoe's lithologi-

cal description of the Parahio series provides an accurate description of these beds, and the term has historical precedence as a lithostratigraphic name. Accordingly, we formally herein designate Pascoe's Parahio series as the Parahio Formation, with the type section described in this paper. The Parahio Formation includes the entire section of the Parahio Valley rocks described

herein, with its top defined here by the unconformity and its base defined by the first occurrence of trace fossil-bearing strata. Rocks of the underlying Batal Formation are devoid of carbonate beds and trace and body fossils, although an acritarch assigned to *Angulopanina* has been recorded from the uppermost part of this formation, according to Kumar et al. (1984).

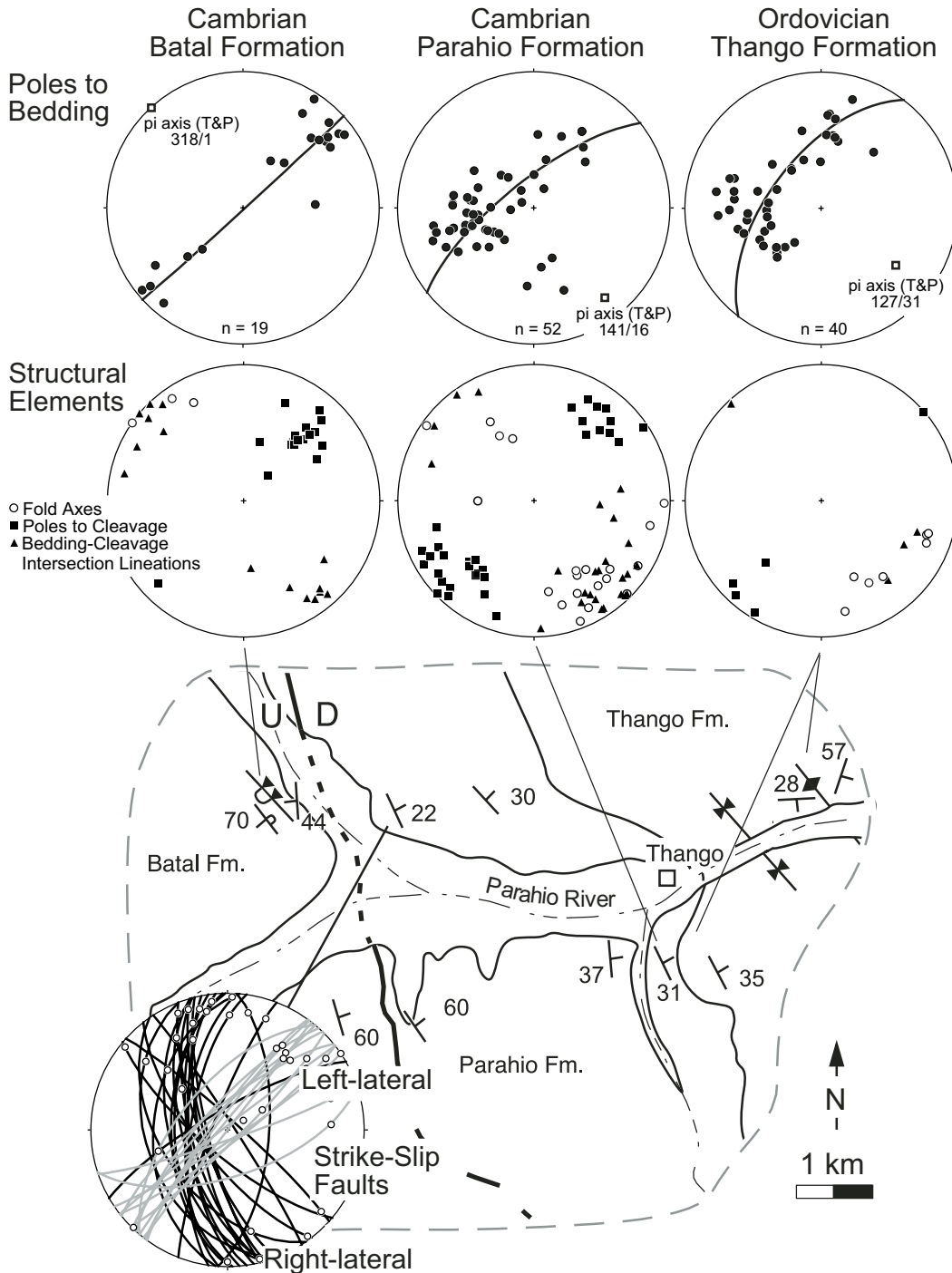


Figure 4. Simplified geologic map of the field area in the Parahio Valley (modified from Bhargava and Bassi, 1998). Map area is shown in Figure 2. Synoptic equal-area plots of structural data measured in the Batal, Parahio, and Thango Formations show little difference in the orientation of structural data from Cambrian units and the Ordovician Thango Formation. The major fault in the area displays a normal stratigraphic offset and is covered by the Khemangar River. Note that the fault is parallel to outcrop-scale, right-lateral strike-slip faults that we measured in exposures near the confluence of the Khemangar and Debsa Khad Rivers (Fig. 2). The right-lateral faults are part of a conjugate strike-slip fault array, but the parallelism of the right-lateral faults and the Khemangar Fault may indicate that faulting included a component of right-lateral strike-slip movement.

Recent authors have commonly referred the Parahio Cambrian rocks to the Kunzam La Formation (Bhargava et al., 1982; Bhargava et al., 1986; Bhargava and Bassi, 1998; Kumar et al., 1984; Parcha et al., 2005), a term first published by Srikantia (1981) (Fig. 5). The type section for this formation was named as the Kunzam La Pass, but this section is considered to be structurally complicated and is a poor choice as a type section (Bhargava and Bassi, 1998, p. 20). Although some trace fossils have been recorded from the Cambrian rocks of the Kunzam La section (Sudan and Sharma, 2001), no body fossils have been described from it, and thus the Kunzam La section lacks the biostratigraphic constraint of the Parahio Valley section. The upper and lower boundaries of the Kunzam La Formation at the type section are also structurally complex and have not been adequately described.

The upper part of the Kunzam La Formation, which contains prominent orange-weathering dolomite beds, was named the Parahio Member (Kumar et al., 1984) (Fig. 5). Our new discoveries show that carbonate beds occur throughout the formation, although the lower beds are limestone instead of orange-weathering dolomite. Our work in the Zanskar Valley (Myrow et al., 2005) indicates that dolomitization of the carbonate beds of the formation there extended to deeper stratigraphic levels. This variable pattern of dolomitization cannot be considered a primary feature for lithostratigraphic correlation, and thus we do not subdivide the newly defined Parahio Formation on the basis of this feature. Kumar et al. (1984) also defined a lower Debsa Khad Member of the Kunzam La Formation on the basis of trace fossils, but our results show that trace fossils occur throughout the section. Accordingly, we also reject this stratigraphic term.

LITHOFACIES

Six lithofacies are defined for the Parahio Formation of the Parahio Valley section. Lithofacies are arranged into cycles tens of meters thick with repeated stratigraphic patterns. The lithofacies are described in order of stratigraphic appearance in these cycles.

Dark Shale Facies

Description

This facies consists of dark-gray to greenish-gray shale. Bed thicknesses range from 2 to 10 m and average ~4 m. Thick units of this facies commonly occur directly over carbonate beds throughout the Parahio Formation and contain minor, discontinuous siltstone laminae (Fig. 6A, B) and centimeter-thick silty sand-

Griesbach, (1891)	Hayden, (1904)	Pascoe, (1959)	Bhargava et al. (1982)	Kumar et al. (1984)	This paper
Haimanta series	Cambrian System	Parahio series	Kunzam La Formation	Kunzam La Fm. Parahio Member Debsa Khad Member	Parahio Formation
		Upper Haimanta series			

Figure 5. Stratigraphic schemes for the Cambrian deposits of the Parahio Valley, including our proposed revision (this paper).

stone beds at facies transitions with either the silty facies or the sandstone and shale facies (described later). Molds of trilobites, typically articulated, exist locally within this facies.

Interpretation

The general lack of sandstone and the presence of articulated olenimorphic (Fortey and Owens, 1990) trilobites in this facies indicate a low-energy, marine-shelf environment well below fair-weather wave base. The depositional environment was distal from the shoreline, at a distance beyond which flows were capable of transporting sand during tempestite deposition. Most shale facies units occur stratigraphically below interbedded fine sandstone and shale deposits, with an upward transition that indicates a change to more proximal environments.

Silty Facies

Description

This facies consists of gray siltstone and silty shale with <1–5 cm beds of very fine brown weathered sandstone. Siltstone and silty shale beds range from 3 to 10 cm thick and average 5 cm. Very fine sandstone beds make up 10% to 60% of this facies (average, ~35%). They pinch and swell laterally over tens of centimeters and generally contain parallel and wavy lamination. Additional sedimentary structures include starved symmetrical ripples and, rarely, sandstone-filled vertical cracks. Trace fossils occur within a few layers in the lower 200 m of our measured section and also occur sporadically higher in the section.

Interpretation

Low sandstone to shale ratios suggest quiet water deposition where mud settled out of suspension. Rare symmetrical ripples indicate infrequent wave reworking of sand. Facies transitions with the shale facies and tempestite-rich sandstone and shale facies (described later) indicate deposition below fair-weather wave

base in an intermediate shelf position. However, the presence of sandstone-filled cracks in a few units is at odds with this interpretation if they are interpreted as desiccation cracks. The simplest interpretation is that these features formed in a low-energy nearshore environment and represent either synaeresis cracks, which form subaqueously from shrinkage associated with strong osmotic gradients (Astin and Rogers, 1991), or diastasis cracks, which form as a result of compaction (Cowan and James, 1992). Bedding plane exposures are lacking, so the plan-view geometry of these structures is unknown.

Sandstone and Shale Facies

Description

This facies is composed of very thin to medium beds of very fine to fine brown-weathering sandstone and grayish-green silty shale and shaly siltstone (Fig. 6D). Sandstone to siltstone-shale ratios are ~1:1, although sandstone content locally reaches 75%. Very fine to fine sandstone beds range in thickness from 1 to 60 cm and average <5 cm in thickness; thicker beds are generally widely spaced. Beds tend to pinch and swell laterally. Sedimentary structures include wave-ripple lamination, hummocky cross-stratification (HCS) (9 cm average thickness) (Fig. 6C), parallel lamination, parting lineation, symmetrical ripples, interference ripples (Fig. 6F), ball-and-pillow structures (Fig. 7A), gutter casts (Fig. 7B), and pot casts. Ball-and-pillow structures are abundant and occur in beds up to 2 m thick throughout the formation in this facies. A few units of this facies contain spindly and polygonal shrinkage cracks (Fig. 7C). Orientations of parting lineation and gutter casts suggest a north-northeast or a south-southwest transport direction (Fig. 8).

Interpretation

Features such as symmetrical ripples, HCS, gutter casts, and pot casts are all common

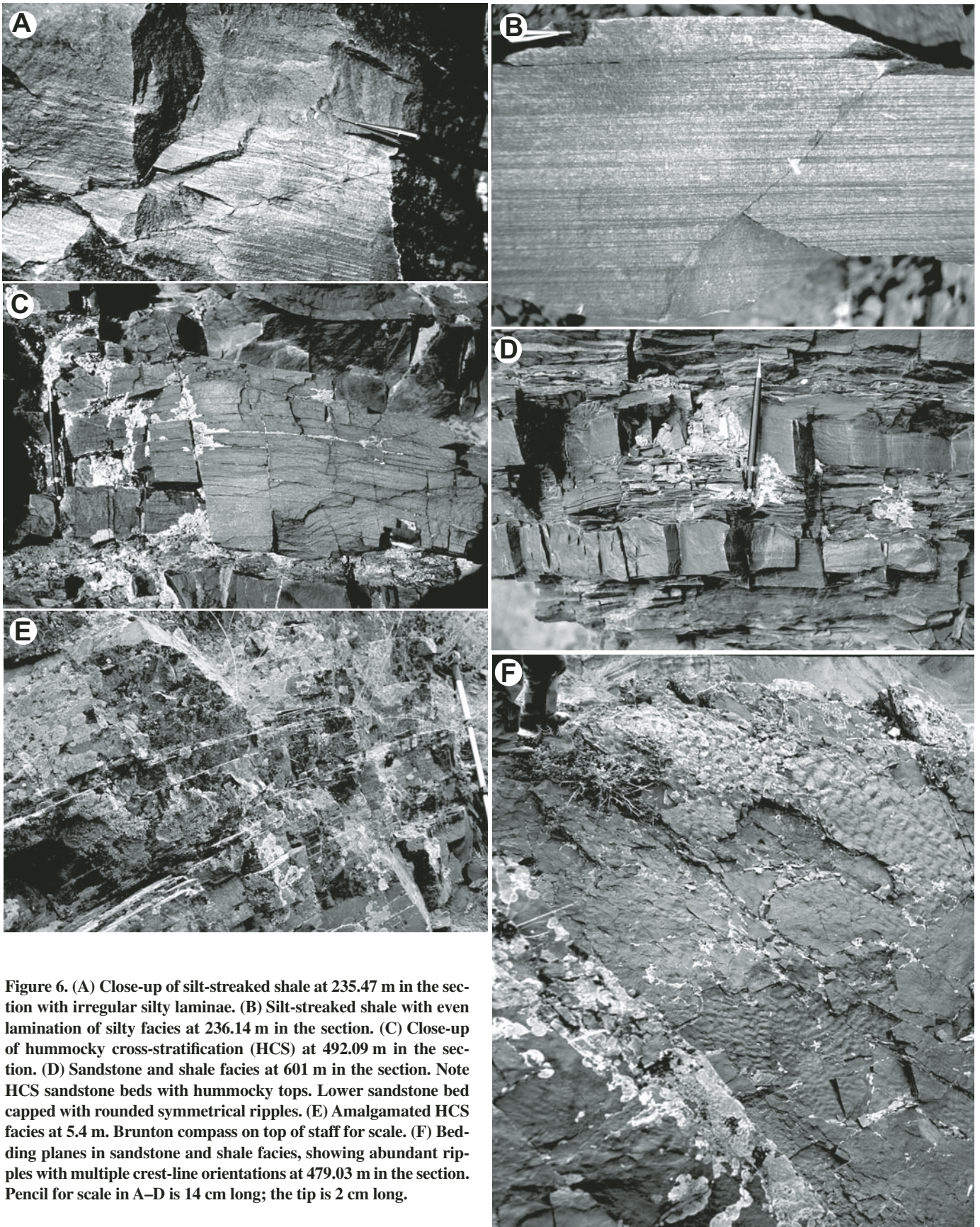


Figure 6. (A) Close-up of silt-streaked shale at 235.47 m in the section with irregular silty laminae. (B) Silt-streaked shale with even lamination of silty facies at 236.14 m in the section. (C) Close-up of hummocky cross-stratification (HCS) at 492.09 m in the section. (D) Sandstone and shale facies at 601 m in the section. Note HCS sandstone beds with hummocky tops. Lower sandstone bed capped with rounded symmetrical ripples. (E) Amalgamated HCS facies at 5.4 m. Brunton compass on top of staff for scale. (F) Bedding planes in sandstone and shale facies, showing abundant ripples with multiple crest-line orientations at 479.03 m in the section. Pencil for scale in A–D is 14 cm long; the tip is 2 cm long.

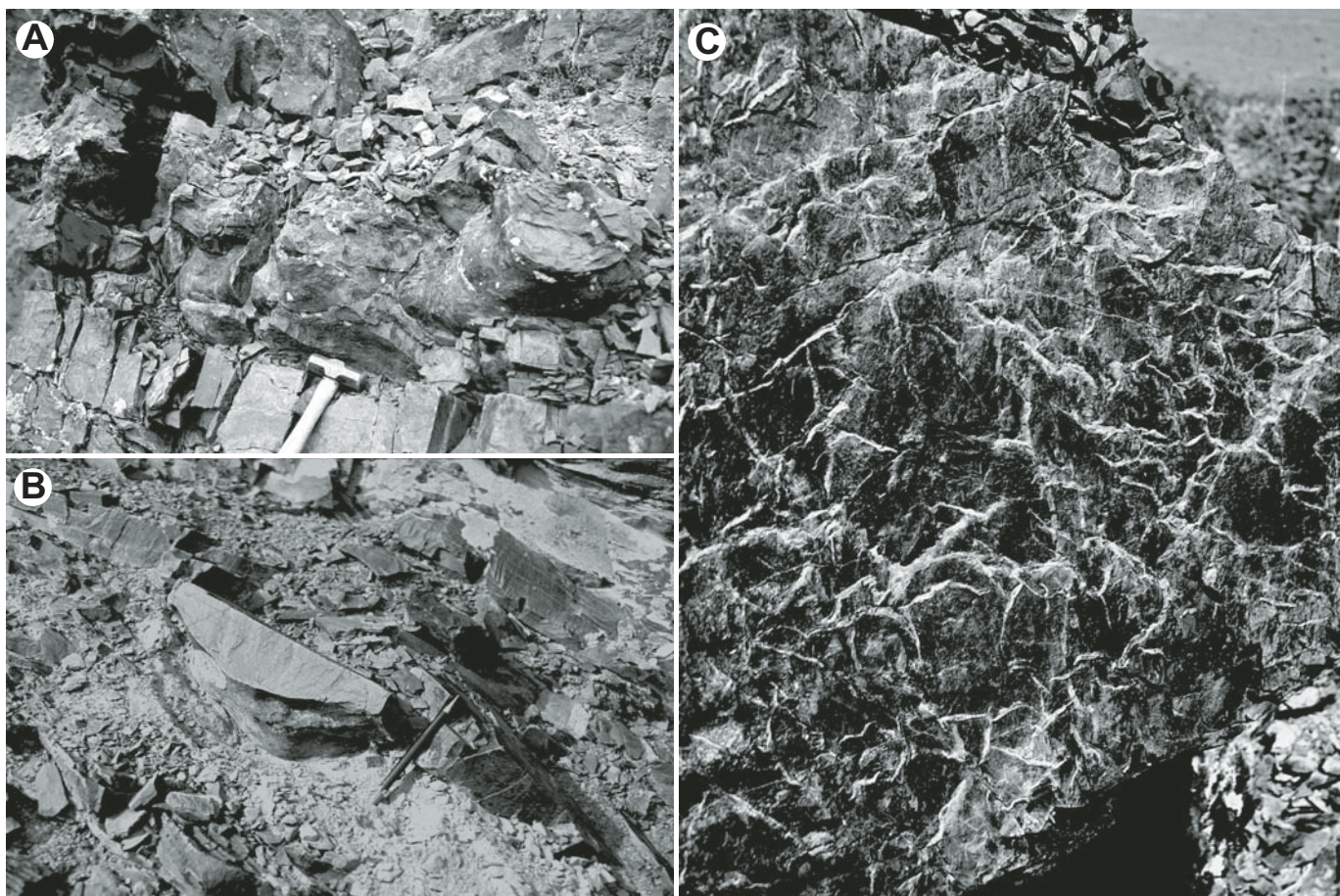


Figure 7. (A) Ball-and-pillow structures at 156.26 m in the section. Such structures are common throughout the formation. Hammer for scale. (B) Oblique view of partly exhumed gutter cast (next to pencil) at 491.49 m in the section; 14-cm-long pencil for scale. (C) Shrinkage cracks at 290.62 m in the section.

storm-generated sedimentary features (Myrow, 1992a, 1992b). HCS is produced during high-energy, storm-generated multidirectional and oscillatory currents (Arnott and Southard, 1990; Dumas et al., 2005). Interbedding of fine sandstone and shale indicates deposition in an inner shelf setting below fair-weather wave base. Gutter casts reflect movement of sand via small-scale channels and are interpreted to form as a result of the unidirectional component of storm currents that erode the substrate, leaving small channel-like structures that are commonly filled by lag or fine sand from the shoreface (Myrow, 1992a). Pot casts are infillings of erosional features that form as a result of vertical vortices within storm-generated flows (Myrow, 1992a). Gutter cast and parting lineation orientations are north-northeast–south-southwest, indicating consistent orientations of flow during erosive and depositional stages of storm flows. Wave-ripple, crest-line orientations are bimodal, with the strongest mode oriented west-northwest–east-southeast. Wave-ripple crest-line orienta-

tions in ancient shallow-water deposits commonly reflect shoreline orientation caused by wave refraction (Potter and Pettijohn, 1977; cf. Myrow, 1992b). The gutter cast and parting lineation data are also parallel to paleocurrent data from inferred fluvial deposits (described later), thus indicating flow offshore toward the north-northeast. This is similar to other studies that document offshore-directed, storm-generated bypass of sand (Myrow, 1992b).

Amalgamated Hummocky Cross-Stratified (HCS) Sandstone Facies

Description

This facies consists of amalgamated beds of gray, fine-grained HCS sandstone (Fig. 6E). The sandstone is reddish gray to steel blue on weathered surfaces and gray on fresh surfaces. Beds range in thickness from tens of centimeters to upward of a meter, with an average bed

thickness of ~30 cm. Beds generally pinch and swell laterally, and are locally separated by centimeter-scale shale beds. Individual hummocks range in spacing from 1 to nearly 4 m in length. This facies is interbedded with the sandstone and shale facies in some areas. Both symmetric and asymmetric forms of HCS occur within this facies.

Interpretation

The abundance of HCS indicates deposition along a storm-dominated shoreline (Leckie and Walker, 1982; Dott and Bourgeois, 1982; and others). The absence of intervening shale beds reflects deposition within the shoreface where mud is removed by fair-weather processes and amalgamation of sand beds results from storm deposition. Deposition of thick beds of HCS accompanied reworking of the shoreface and elimination of stratification generated during fair-weather conditions. The existence of this facies stratigraphically above the sandstone and shale facies reflects shoaling from

nearshore and transition zone environments into shoreface settings.

Trough Cross-Bedded Sandstone Facies

Description

This facies consists of large-scale, trough cross-bedded, gray-green fine sandstone (Fig. 9A, B) that is light gray to white on weathered surfaces. Bed thicknesses range from 15 to 120 cm, with an average of 54 cm. Units of this facies range up to 30 m thick. The bases of these units show small-scale cutout of underlying beds laterally in the section. Decimeter-scale, parallel-laminated sandstone beds locally separate trough cross-bed sets. This facies generally overlies the amalgamated HCS facies and underlies the limestone facies (described later) and/or shale facies. Paleocurrent readings suggest north-northeastward flow (Fig. 8).

Interpretation

The thick beds of trough cross-bedded fine sandstone record the migration of three-dimensional dunes in high-energy depositional systems (Harms et al., 1982). The scale of the cross-bedding, unimodal paleocurrents, lack of body or trace fossils, and facies transitions indicate deposition in a riverine environment (Potter and Pettijohn, 1977; Walker and Cant, 1984). Stratigraphic transitions with underlying shoreface deposits indicate shoaling to nonmarine environments associated with progradation of the shoreline.

Carbonate Facies

Description

This facies consists of limestone and dolostone beds, many of which serve as marker beds within the formation. The limestone consists of greenish-gray carbonate mudstone to coarse grainstone with abundant trilobites, phosphatic linguliformid brachiopods, and sponge spicules in grainstone. Limestone beds range in thickness from 3 cm to 1.86 m, with an average thickness of ~60 cm. The limestone is thinly laminated (millimeter scale) and is pink to rusty orange on weathered surfaces. Some of the thinner beds are discontinuous over tens of meters. Thin-bedded homogeneous fine-crystalline limestone beds occur below grainstone throughout the section (e.g., 78.7 m). Much of the coarse and fine grainstone is thinly interbedded. Limestone beds are partially dolomitized and silicified locally (Fig. 9C, F). Some beds contain abundant intraclasts at their base and locally contain thin lag deposits on their upper bedding contacts.

Orange-weathering dolostone beds occur in the uppermost 200 m of the formation (Fig. 9D). They consist of extremely well-sorted, massive

dolosiltite (Fig. 9E). Bed thicknesses range from 23 cm to 12.8 m, with an average thickness of ~4 m. Internal sedimentary structures and biogenic structures are absent except where phosphatic brachiopods are concentrated along individual laminae. Diagenetic nodules of orange-weathering dolostone occur locally within underlying sandstone units.

In general, beds of the carbonate facies rest stratigraphically above trough cross-bedded facies of presumed fluvial origin, and, in a few places, shoreface deposits of amalgamated HCS sandstone. These beds in turn are overlain by the dark shale or silty facies.

Interpretation

The presence of minor carbonate units within the dominantly siliciclastic Parahio Formation suggests intermittent decreases in supply of terrigenous siliciclastic sediment. Trilobite and brachiopod fossils indicate deposition under marine conditions. Micritic beds accumulated in lower energy conditions, whereas the bioclastic grainstone beds were reworked by ocean waves and currents. The homogeneous nature of the orange-weathering dolosiltite suggests a shallow-marine environment of uniform energy. Transitions from underlying trough cross-bedded sandstone to limestone indicate that marine flooding surfaces (Van Wagoner et al., 1988) separate these facies.

PALEOCURRENTS

Paleocurrent readings were measured from parting lineation, trough cross-beds, symmetrical wave ripples, and gutter casts throughout

the Parahio Formation. The long axis directions of parting lineations (Fig. 8A) and gutter casts are south-southwest–north-northeast. Wave ripples (Fig. 8B) display scattered orientations, but the strongest mode is essentially perpendicular to the other features, namely, west-northwest–east-southeast. Large-scale sets of trough cross-beds of the trough cross-bedded sandstone facies indicate a unimodal northeastward transport (Fig. 8C). This is consistent with the reconstructed position of these Tethyan deposits along the northern margin of India at that time. The largest mode in the wave-ripple data, perpendicular to the trough cross-bed sets, could represent refraction of waves parallel to the inferred paleoshoreline, although the data are sparse and somewhat scattered. The scatter in the data is not unusual for shoreline deposits, because variable coastal winds and shallow depths lead to variable oscillatory water movements (Weimer et al., 1982).

The paleocurrent readings do not account for oroclinal bending of the Himalaya and counterclockwise rotation of India during Cenozoic orogenesis, which resulted in a relative movement between India and its northern continental margin (Klootwijk et al., 1985; Schill et al., 2001, 2002). Paleomagnetic studies of the Pin Valley, ~10–15 km south and east of our study area (Schill et al., 2001), suggest a possible 20°–35° clockwise vertical axis rotation of rocks with respect to India. Trial rotations of our paleocurrent data, using 20° and 35°, produce nearly due north mean paleocurrent directions, which are an even better fit with the idea that the Tethyan Himalaya records the northern margin of India during the Cambrian.

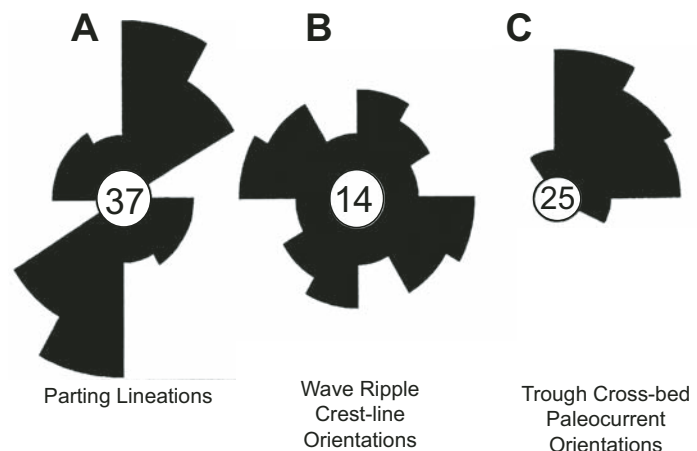


Figure 8. Equal-area paleocurrent rose diagrams of the Parahio Formation (north at top). Parting-lineation orientations and wave-ripple crest-line orientations were taken from the sandstone and shale facies. Cross-bed orientations were taken from the trough cross-bedded sandstone facies.

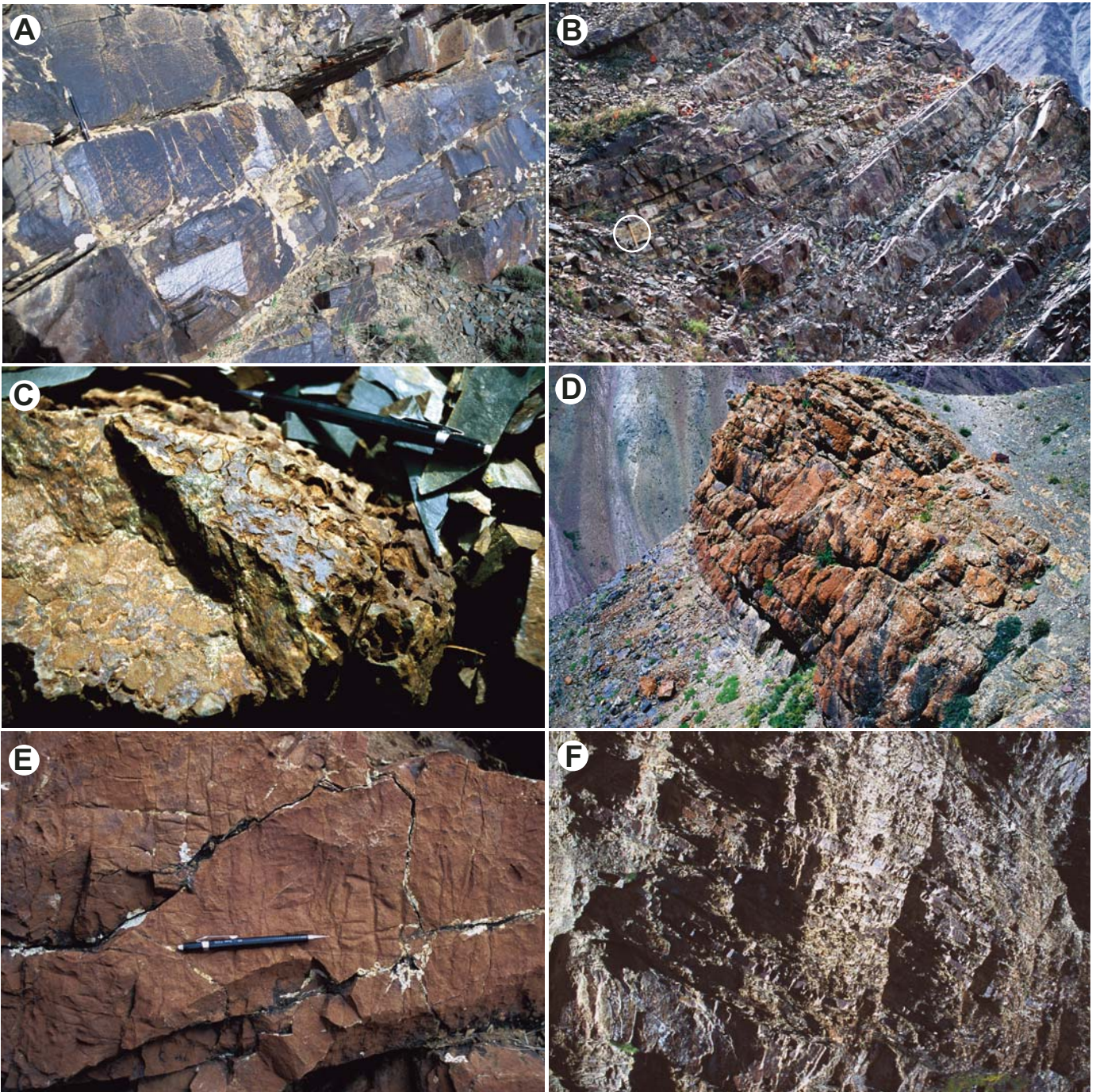


Figure 9. Trough cross-bedded facies and carbonate facies. (A) Compound trough cross-bedding at 57.3 m in the section. Pencil (top left) for scale. (B) Large-scale trough cross-bedding with sets up to 1.5 m thick at 760.5 m in the section. Hammer for scale (circled). (C) Partly silicified limestone bed, showing nodular weathering pattern at 765.14 m in the section. (D) Thick (12.8 m) orange dolosiltite bed at 1242.4 m in the section. (E) Close-up of D, showing well-sorted, homogeneous orange dolosiltite. (F) Nodular gray limestone unit, 1.3 m thick, with abundant chert nodules at 779.35 m. Pencil for scale is 14 cm long.

PALEOENVIRONMENTAL SYNTHESIS

The foregoing sedimentological analysis indicates that the dominantly siliciclastic strata of the Parahio Formation range from offshore-marine to trough cross-bedded fluvial deposits. Overall, the strata are relatively fine grained, ranging from shale to fine sandstone, with a few scattered medium sandstone beds. The presence of thick, well-developed deposits of the fluvial trough cross-bedded sandstone facies indicates a strong influence of fluvial processes and earmarks these strata as deltaic deposits. The fluvial deposits are also relatively fine grained and reasonably well sorted, indicating low local stream gradients and a position relatively distal to sediment sources during the Cambrian in this region. However, the general character of the section indicates that rivers deposited large volumes of sediment on the shoreline and inner shelf on the ancient northern passive margin of India.

Extensive ball-and-pillow features occur throughout these deposits in several facies. These features represent gravitational synsedimentary deformation during sediment accumulation. This kind of deformation results from partial liquefaction caused by high pore-fluid pressures and unstable packing (Seed, 1968). Such conditions are common in environments with rapid sediment accumulation rates, particularly in deltas with abundant silt to very fine sand. The thickness of Middle Cambrian strata in the Parahio Formation indicates high sediment accumulation rates, which supports our interpretations of deltaic deposition. Trilobite biostratigraphic data indicate that most of the ~1350-m-thick measured section is of early and middle Middle Cambrian age. On the basis of recent geochronological work (Landing et al., 1998; Davidek et al., 1998), the entire Middle Cambrian was ~10 m.y. in duration. A conservative accumulation rate estimate has been calculated on the basis of (1) the assumption that the lower and middle parts of the Middle Cambrian were 7 m.y. in duration, (2) an estimate of ~50% shale and 50% sandstone for the formation, and (3) compaction estimates of 50% for shale and 15% for sandstone with >5 km burial (Houseknecht, 1971; Bond and Kominz, 1984). The estimate of sediment accumulation rate is 30.4 cm/1000 yr, similar to time-averaged accumulation rates in deltaic settings (Stanley and Hait, 2000). The interpretation of the Parahio Formation as a storm-influenced deltaic succession contradicts previous paleoenvironmental interpretations. Srikantia (1981) and Fuchs (1982) considered the succession to be dominated by flysch deposits. Similarly, Parcha (2005) suggested that the Parahio Formation accumulated in an outer detrital belt setting. Srikantia (1981), Bhargava

et al. (1982, 1986), Bhargava and Bassi (1998, p. 146), and Parcha et al. (2005) all argued that the Parahio Formation records overall shallowing from a euxinic basin toward a tidally dominated nearshore setting.

We find no sedimentological evidence for deposition in an oxygen-stratified basin, e.g., biofacies patterns of anaerobic or dysaerobic environments (Byers, 1977; Myrow and Landing, 1992), let alone evidence for a long-term euxinic basin. In addition, trilobite taxa that occur in the shallow-water limestone beds are identical to those in overlying dark shale facies, suggesting normal-marine settings for both paleoenvironments (i.e., similarly well-oxygenated conditions). Fuchs (1987) and Garzanti et al. (1986) both consider time-equivalent strata exposed in the Zanskar Valley to the north-northwest to represent tidal flat, coastal sand, and peritidal carbonate deposits. Our sedimentological analysis, as well as our work on the Zanskar strata (Myrow et al., 2006), reveals little evidence for tidal processes in these deposits, such as diagnostic sedimentary structures (e.g., herringbone cross-bedding, reactivation surfaces) or facies successions.

SEDIMENTARY CYCLES

The strata of the Parahio Formation range from storm-influenced shelf deposits to thick, trough cross-bedded fluvial facies, many of which are arranged in medium-scale, upward-coarsening shoaling cycles (Fig. 10A). The cycles range from 2.5 m to 61 m thick and differ slightly in facies sequences from cycle to cycle. An idealized cycle begins with sand-starved deposits of the dark shale facies (Fig. 10B, C) and grades upward sequentially into the silty facies, sandstone and shale facies, amalgamated HCS facies, and finally the trough cross-bedded sandstone facies. Within the cycles, beds of sandstone become thicker and increasingly more abundant upward until all shale is missing in the amalgamated HCS facies. The trough cross-bedded sandstone facies generally overlies the amalgamated HCS facies across an erosional surface. Carbonate-rich sandstone or beds of the carbonate facies overlie the trough cross-bedded sandstone facies and are succeeded by another unit of the dark shale facies (Fig. 11A).

Many of these shoaling cycles are incomplete. Some lack the dark shale facies and instead commence with the silty facies, the sandstone and shale facies, or shoreface deposits (Fig. 11B). One cycle commences with amalgamated HCS sandstone deposits. Other cycles are abbreviated at their tops, and the carbonate facies is absent, or both the trough cross-bedded sandstone facies and carbonate facies are absent

(Figs. 11B, 12). The latter type of cycle ends with the amalgamated HCS facies (Fig. 12) or with thick, bioturbated sandstone beds, both of which are of shoreface origin (Fig. 11A). In a few cycles, the trough cross-bedded sandstone facies directly overlies the sandstone and shale facies.

Interpretation of Cycles

The succession of lithofacies in a cycle is interpreted to record upward shoaling along a prograding storm-dominated shoreline (Fig. 13). Complete cycles record changes from offshore-marine environments to fluvial settings, although incomplete cycles begin in nearshore facies or end in shoreface deposits, the latter being due to renewed transgression before shoaling culminated in deposition above sea level. The upper cycle boundaries represent marine flooding surfaces. The marine parts of the shallowing cycles record progradation of a storm-dominated shoreline with upward increases in tempestite bed thickness and decreasing shale content.

In these cycles, the transitions from fluvial cross-bedded sandstone to carbonate facies are interpreted as marine flooding surfaces and subsequent deposition in shallow-marine environments that were essentially flooded coastal plains (Fig. 13). Fully marine conditions are supported by the presence of trilobite and brachiopod fossils, as well as trace fossils. Beds of the carbonate facies are sharply overlain by deep-water shelf deposits of the dark shale facies, the transition representing another flooding surface. The flooding surface at the top of the carbonate represents deepening to the point of shutdown of carbonate production, as was the case with some ancient carbonate platforms (Schlager, 1981). The carbonate facies are thus essentially thin transgressive deposits. The rest of these cycles, from the base or slightly above the base of the dark shale facies to the top of the cycles, formed during shoreline progradation (Fig. 13A). Given the deltaic setting of these strata, and the fact that the cycles do not form regular stacking patterns (i.e., progradationally or retrogradationally stacked parasequences; Lehrmann and Goldhammer, 1999), we interpret the cyclicality to be mostly autocyclic, driven by changes in sediment input associated with switching of delta lobes. Changes in eustasy and subsidence may have played roles in the development of these cycles, but it is not possible to determine their relative influence. The few places where fluvial deposits rest directly on inner shelf deposits of the sandstone and shale facies may point to forced regression (Posamentier et al., 1992) and rapid seaward shift of environments. Avulsion and deltaic lobe shifting would have been enhanced

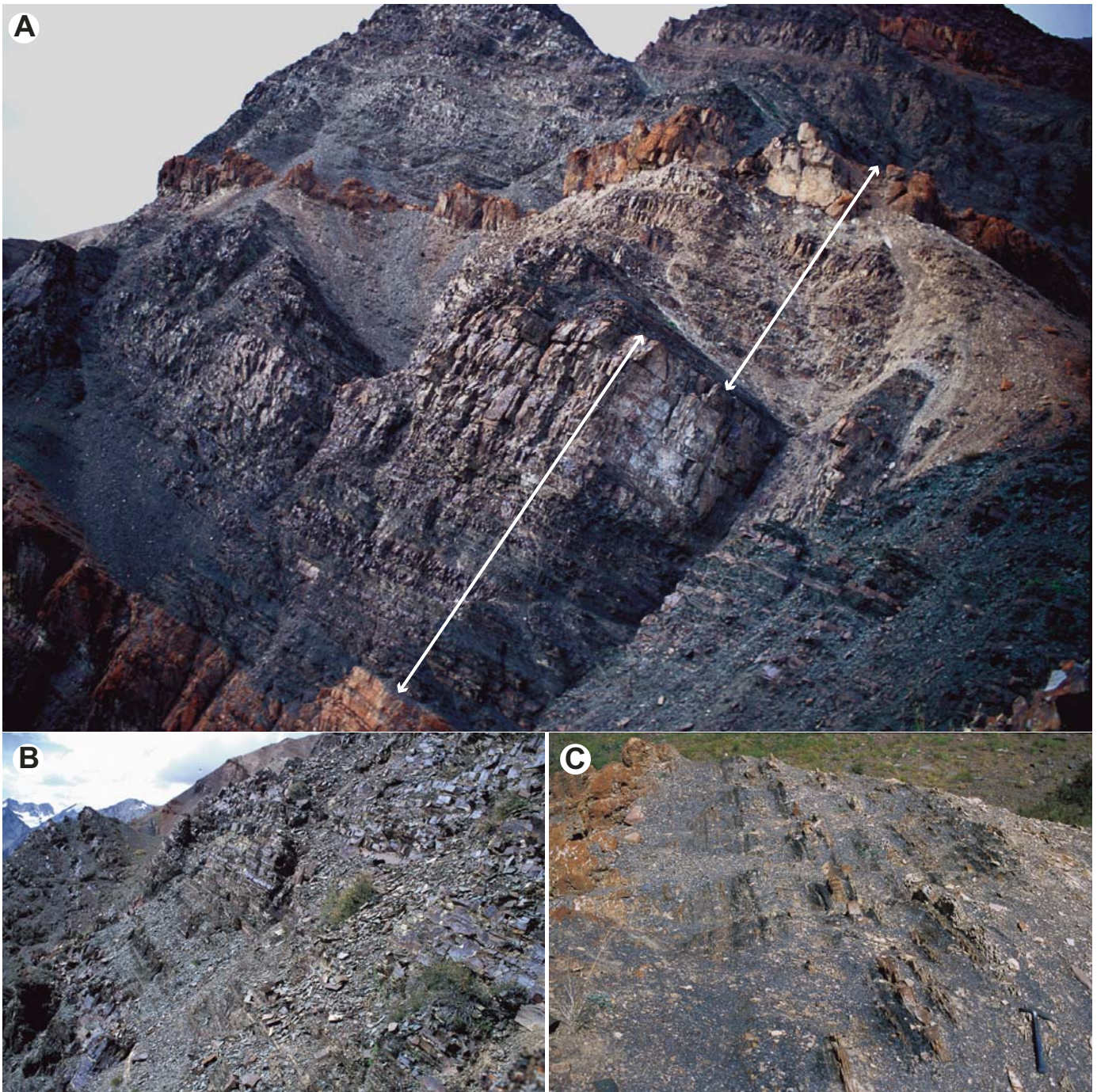


Figure 10. (A) Two well-developed cycles in the upper part of the Parahio Formation. Lower cycle rests on an orange dolostone bed (at top of an underlying cycle) in lower left of photo. Lower cycle begins at 1166.7 m in the section, is 61 m thick, and is capped with an 18-m-thick unit of amalgamated HCS sandstone. Overlying cycle begins at 1227.9 m in the section, is 29.6 m thick, and is capped with a 12.8-m-thick bed of orange dolostone. Sharp upper contact of second cycle is overlain by shale, marking the base of a third cycle. (B) Base of 30-m-thick cycle at 1097.92 m in the section. Note shale (left center) and an upward increase in sandstone thickness and abundance. (C) Base of 18.6 m cycle at 1258.77 m in the section. Note the top of a carbonate bed in the upper left and a dark shale directly above. Note an upward increase in sandstone bed thickness and abundance. Hammer for scale.

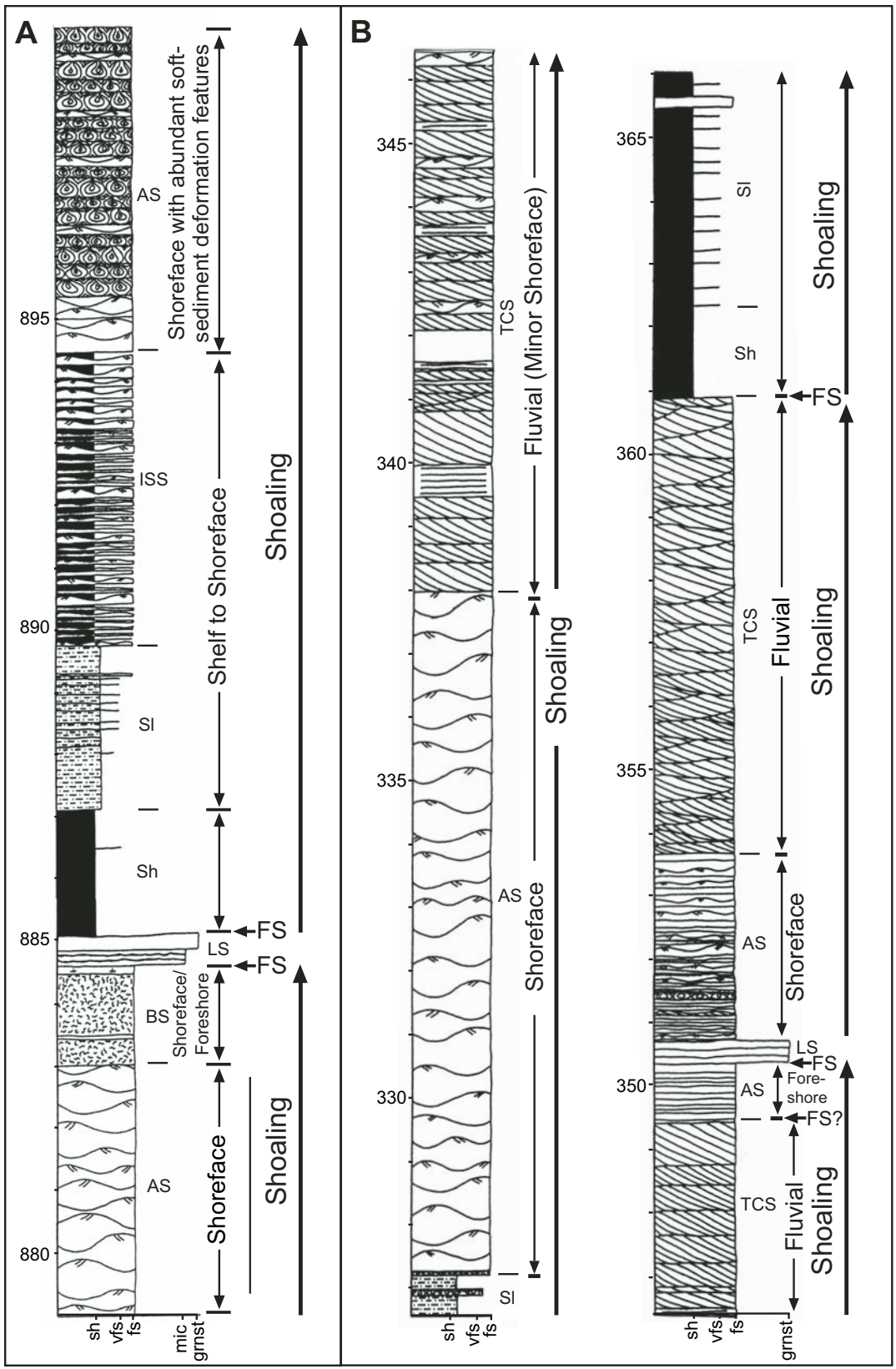


Figure 11. (A) Parts of two shoaling cycles from 879.09 to 899.72 m in the section. Amalgamated HCS shoreface deposits of the lower cycle, overlain by bioturbated upper shoreface or foreshore deposits and then limestone. The base of the limestone is a marine flooding surface, and the top another flooding surface. (B) Parts of three shoaling cycles between 326.59 and 366.42 m in the section. The lower cycle has very thick amalgamated shoreface deposits overlain by trough cross-bedded facies of fluvial origin. Possible foreshore deposits at the top are bracketed by flooding surfaces, and the base of the next cycle begins in lower shoreface deposits directly above the limestone. FS—flooding surface; sh—shale; vfs—very fine sandstone; fs—fine sandstone; mic—micrite; grnst—grainstone; Lithofacies: SH—dark shale; SI—siltstone; ISS—interbedded sandstone and shale; BS—bioturbated sandstone; AS—amalgamated sandstone; TCS—trough cross-bedded sandstone.

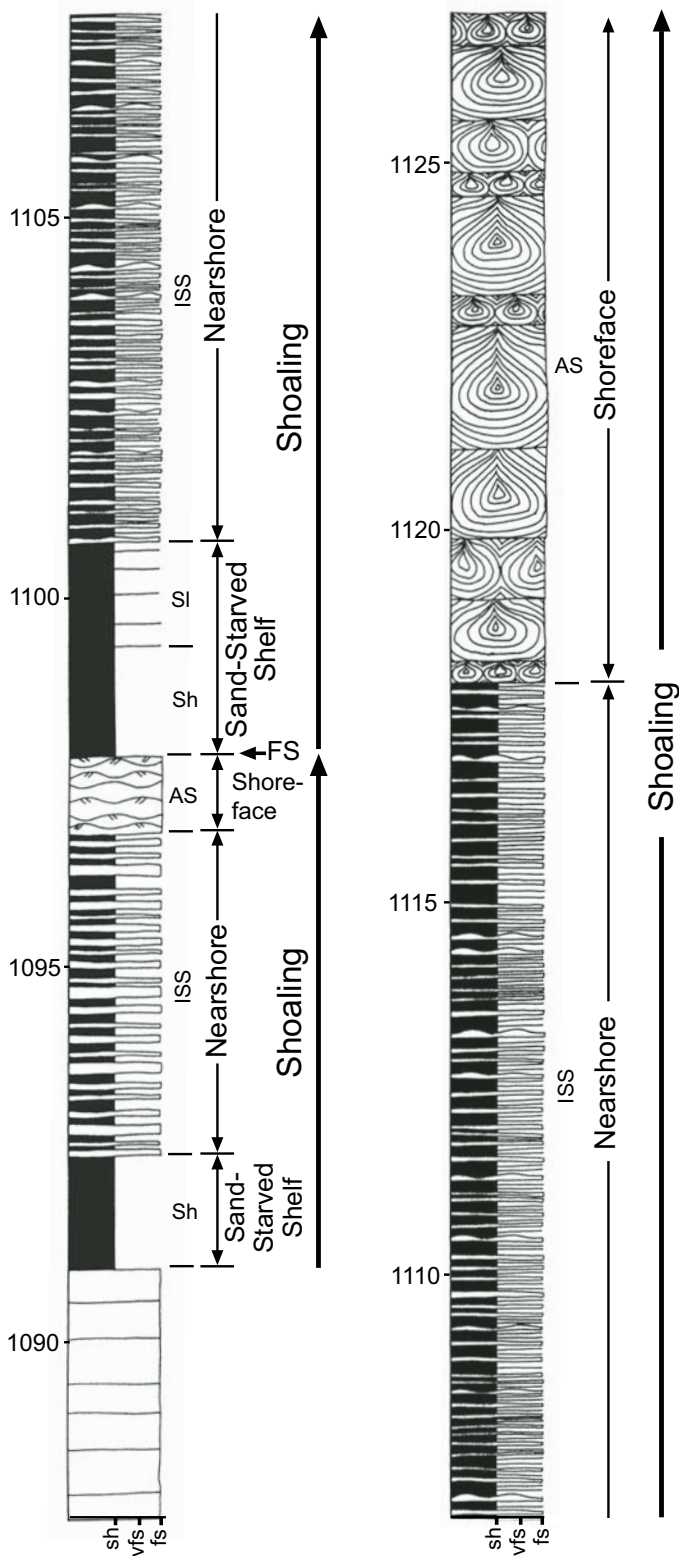


Figure 12. Two shoaling cycles from 1087.67 to 1127.92 m in the section. Note thick ball-and-pillow zones of probable shoreface origin at the top of the second cycle. See Figure 11 caption for explanation of abbreviations.

by regional subsidence resulting from sediment loading within the deltaic setting, which would have been caused by high sediment accumulation rates (Coleman, 1988). This subsidence would have enhanced any thermal subsidence to increase accommodation space. Following avulsion events that shifted sediment supply to adjacent parts of the shoreline, local production of accommodation space would have been rapid. This would have resulted in deepening through the zone of carbonate production, which may have been generally narrow in this deltaic setting, and in deposition of relatively thin, transgressive carbonate deposits.

One stratigraphic pattern of note is the presence of thicker (orange-weathering dolostone) carbonate beds in the upper part of the Parahio Formation. These carbonate beds are more extensive and continuous than the thinner limestone beds in the lower Parahio. Similar dolostone beds also occur in the upper part of the Parahio Formation in the Zanskar Valley to the northwest. Unfortunately, our biostratigraphic information does not allow us to determine whether individual dolostone beds are continuous between the Zanskar and Spiti regions. In Zanskar, the Parahio Formation is overlain by the Thidsi Member of the Karsha Formation (Myrow et al., 2006), which is composed of >150 m of orange-weathering dolostone (Gaetani et al., 1986). The regionally extensive Cambrian–Ordovician unconformity (see following discussion) cuts downsection between Zanskar and the Spiti Valley, so the Thidsi Member and even younger strata are absent in Spiti. The upper Parahio Formation thus records a transition from siliciclastic- to carbonate-dominated strata, representing a long-term decrease in supply of terrigenous detritus, and to some degree this transition might represent the signal of part of a third- (1–2 m.y.) or second-order (up to 15 m.y.) depositional sequence (Vail et al., 1977).

CAMBRIAN–ORDOVICIAN UNCONFORMITY

The unconformity between the Cambrian Parahio Formation and the overlying Thango Formation is poorly understood, but it has important implications for the early tectonic history of the Himalaya. The age of the unconformity is in part constrained by trilobites from the basal part of the Takche Formation, which overlies the Thango Formation and which is no older than Middle Ordovician and no younger than latest Ordovician (Patterson, 2004; R.A. Fortey, 2004, personal commun.). Trace fossils from the Thango Formation (Bhargava and Bassi, 1998) are consistent with this estimate, because they also suggest an Ordovician age.

The Cambrian–Ordovician unconformity in the Parahio Valley is a low-angle truncation surface (Fig. 14A). Small-scale relief at the base of the cobble conglomerate of the Thango Formation consists of large decimeter- to meter-scale elongate scours (Fig. 14C, D). These scours are generally oriented northeast-southwest. Southeast of the confluence of the Parahio and Sumna Rivers, at one of Hayden's (1904) sections, the unconformity shows considerable local relief. Boulder conglomerate in the Thango makes up an incised valley fill that cuts nearly vertically through hundreds of meters of upper Parahio deposits (Fig. 14B, E–G). The valley fill includes large meter-scale boulders of dolostone derived from thick dolostone beds that are cut out by the unconformity (Fig. 14F).

A Cambrian–Ordovician unconformity occurs along the length of the Himalaya west of Nepal, although in western Kashmir, Lower Ordovician rocks rest without angular discordance on Upper Cambrian rocks (Reed, 1934; Jell, 1986). In the Peshawar Basin of Pakistan, north of the Panjal-Khairabad Fault, the Ordovician Misra Banda Quartzite unconformably overlies the poorly dated Tanawai Formation (Pogue et al., 1992), which is presumably of Cambrian age. In the Zaskar Basin the unconformity caps latest Middle Cambrian deposits (Gaetani et al., 1986; Jell and Hughes, 1997). In the Kumaon Himalaya (LeFort, 1975), adjacent to the western border of Nepal, the Ralam Conglomerate overlies the Martoli Group, from which Early Cambrian trilobites have been collected several hundred meters below the unconformity (Kacker and Srivastava, 1996). These relationships suggest uplift and differential erosion of sedimentary successions (Hayden, 1904) along the length of the orogen. The event is roughly correlative with the timing of late Pan-African granitoid emplacement and associated fold-thrust deformation and high-grade metamorphism (Garzanti et al., 1986). Granite intrusions of this age with high $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios occur throughout the Himalaya (LeFort et al., 1986), but their tectonic significance is a subject of debate. Gehrels et al. (2003) suggest that the intrusions were associated with a major fold and thrust belt that developed within the Tethyan Himalaya during the Late Cambrian and Ordovician. Other authors relate the granites to extensional, rather than contractional, tectonics (Miller et al., 2001).

In the Parahio Valley, the Cambrian–Ordovician contact (Fig. 14A) shows truncation of a fold in the underlying Parahio Formation (Fig. 15A; plate 6 in Hayden, 1904). Structures that formed during the Cambrian–Ordovician deformation episode are important in the debate over the tectonic significance of the related unconformity

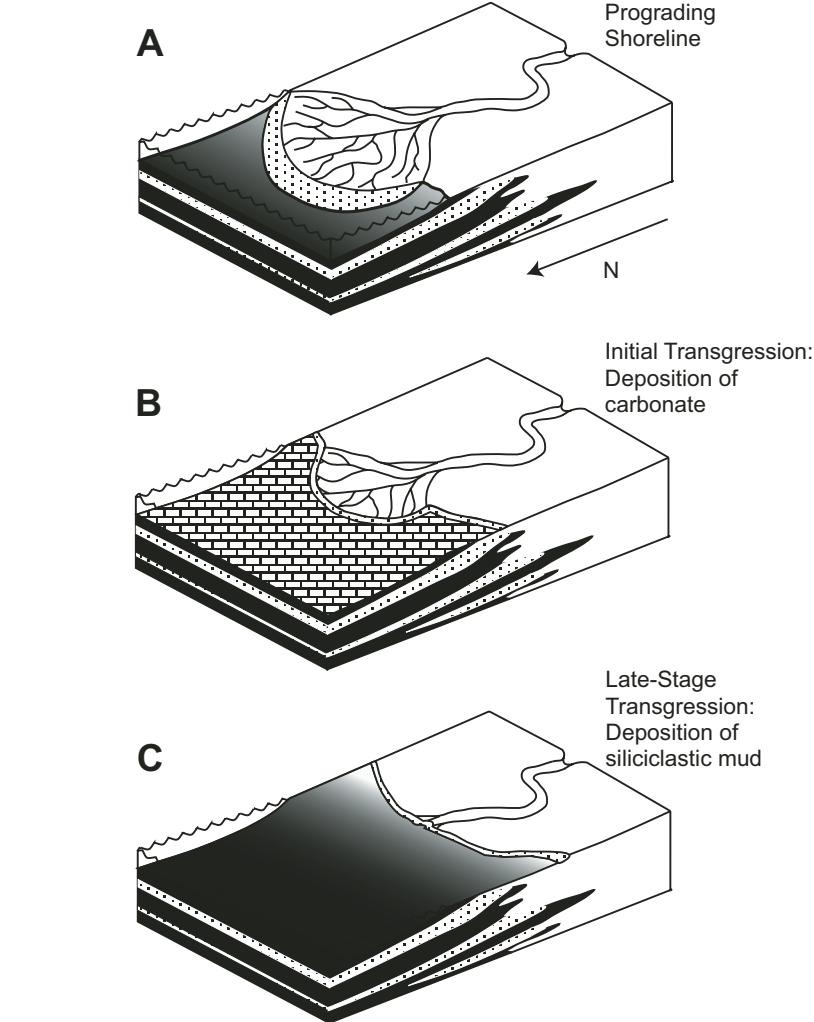


Figure 13. Block diagrams illustrating development of large-scale cycles within the Parahio Formation. North arrow for orientation. (A) Progradation of a deltaic shoreline. (B) During initial transgression, sediment supply is reduced and carbonate deposition begins in shallow water. (C) Continued transgression results in deposition of siliciclastic mud.

(Wiesmayr and Grasemann, 2002; Gehrels et al., 2003). However, such structures are difficult to document, owing in part to the dominance of Cenozoic deformation and high-grade metamorphism. Normal faulting and minor folding have been related to Cambrian–Ordovician deformation south of the Parahio Valley in the Pin Valley (Wiesmayr and Grasemann, 2002). However, the fold and unconformity surface reported by Hayden (1904) were never given a modern structural analysis.

Our analysis shows that the axis of the fold in Figure 15A plunges 14° to the east-southeast and that bedding is offset by a fault with a throw of ~ 5 m. This fault strikes east-west and dips 70° to the south. The fault surface contains quartz slip fibers with steps that indicate left-oblique, top-

to-the-northeast reverse movement. The fault is truncated by the Cambrian–Ordovician contact. However, local evidence for shearing along this contact suggests that interpretation of the fold and fault as Cambrian in age is at best dubious. Approximately 10 m south of the area shown in Figure 15A, the contact is accessible and is marked by a 40-cm-thick shear zone with evidence of anomalously high strain in comparison to rocks directly above and below (Fig. 15B). The shear zone contains a strong, northeast-dipping foliation, the deflection of which suggests top-to-the-southwest thrusting. This interpretation is substantiated by mesoscopic fault surfaces spaced 6–30 cm apart that locally crosscut foliation within the shear zone. These faults are inclined at a low angle to the trace of the shear

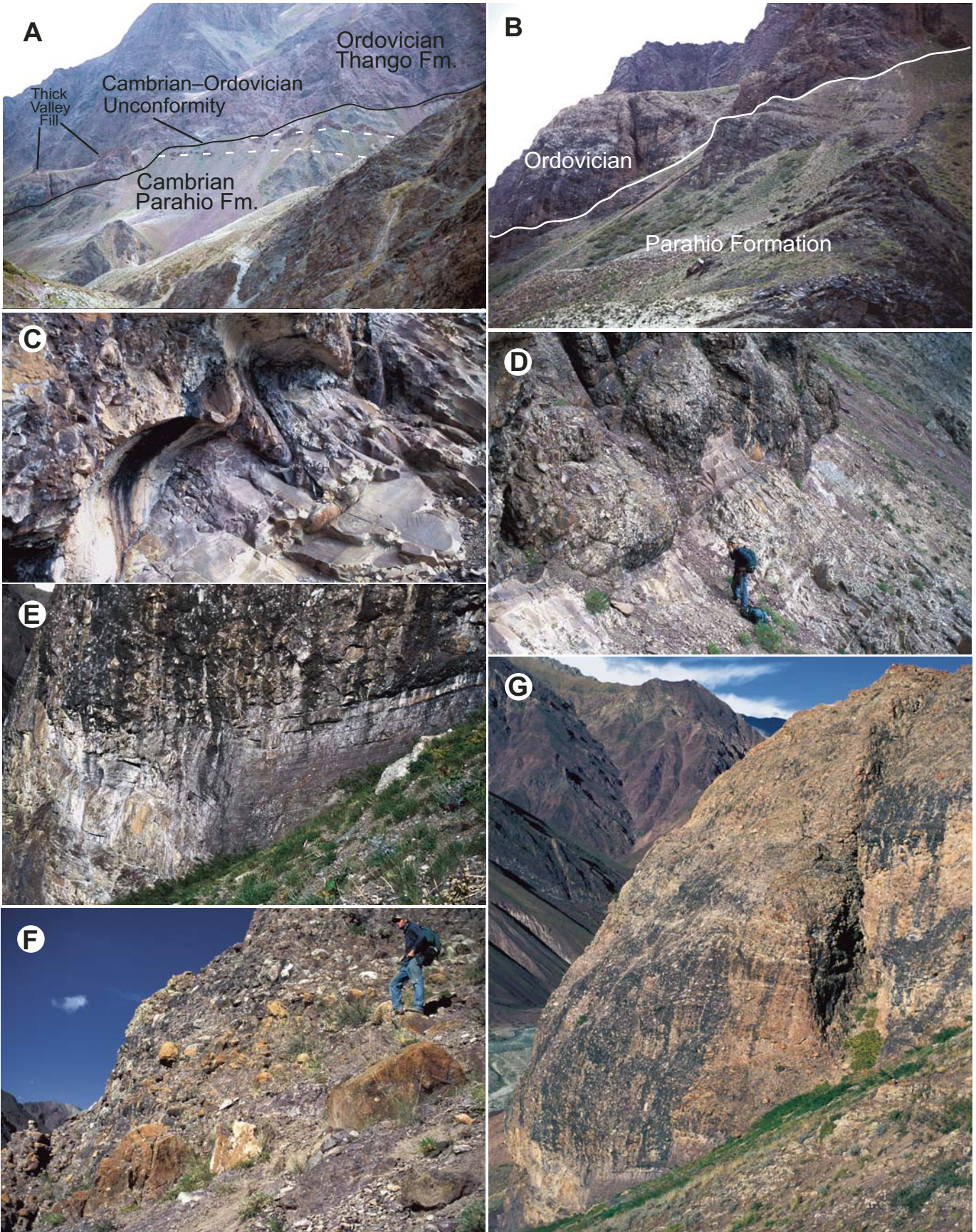


Figure 14. Cambrian–Ordovician unconformity. (A) View of the upper Parahio Formation above the Sumna River. Note cutout of orange dolostone beds below the sub-Ordovician unconformity (white dashed lines) and thick valley fill on left. (B) View of the upper Parahio Formation in the foreground and conglomeratic Ordovician valley fill in the background at left at the same stratigraphic level. (C) Close-up of oriented meter-scale scours along the unconformity at the base of Ordovician deposits. (D) Irregular unconformity at the base of the Ordovician, including oriented scours as shown in C. (E) Close-up of the unconformity at the base of thick valley fill, with sandstone and shale of the Parahio Formation at the base and coarse conglomerate of the Thango Formation above. (F) Close-up of coarse, thick valley fill with large boulders of orange dolostone that were cut out below the Cambrian–Ordovician unconformity. (G) Thick Ordovician valley fill. Photo shows >100 m of conglomeratic strata.

zone, and deflection of the foliation by the faults, along with stepped quartz slip fibers on their surfaces, suggests that they represent riedel shears related to top-to-the-southwest thrusting. Interpretation of the contact as a thrust also serves to explain the stair-step shape of the contact in profile as it cuts upsection to the southwest in the footwall (Fig. 15A; plate 6 in Hayden, 1904); this resembles a footwall ramp-on-flat geometry that is commonly displayed by thrust faults (Boyer and Elliot, 1982). The southwest-directed sense of movement along the surface is consistent with the direction of thrusting for the region (Wiesmayr and Grasemann, 2002) and with minor out-of-the-syncline thrusting related to the folding of the Devonian Muth Formation to the north in Parahio Valley (Fig. 4).

To further evaluate whether Cambrian rocks contain evidence for pre-Ordovician deformation, we also examined and compared structures above and below the Cambrian–Ordovician unconformity. Folds occur locally within all of the stratigraphic units within the study area and are upright to inclined, cylindrical (Fig. 4), and have amplitudes and wavelengths that range from a few meters to a kilometer. Bedding-parallel slip surfaces occur on the limbs of folds, and these contain quartz and epidote slip fibers with steps, indicating that folds formed predominantly by flexural slip. We did not observe refolded folds indicative of polyphase deformation. Bedding poles define great-circle girdles that have horizontal or shallow, southeast-plunging pi axes (Fig. 4), similar to the orienta-

tion of fold axes in the area (Fig. 4). Disjunctive slaty to spaced cleavage occurs within all stratigraphic units in the study area and varies from being axial planar to slightly fanning around folds. Cleavage strikes northwest-southeast, and cleavage-bedding intersection lineations plunge $\sim 20^\circ$ to the southeast (Fig. 4), similar to the orientation of the pi axes and fold axes. A comparison of structures between Cambrian and Ordovician stratigraphic units reveals little difference in orientation between fold axes, cleavage, and cleavage-bedding intersection lineations (Fig. 4). Thus, we conclude that no clear evidence exists that the Cambrian rocks have been deformed in a separate phase of thrust-related folding and faulting during the Cambrian–Ordovician interval.

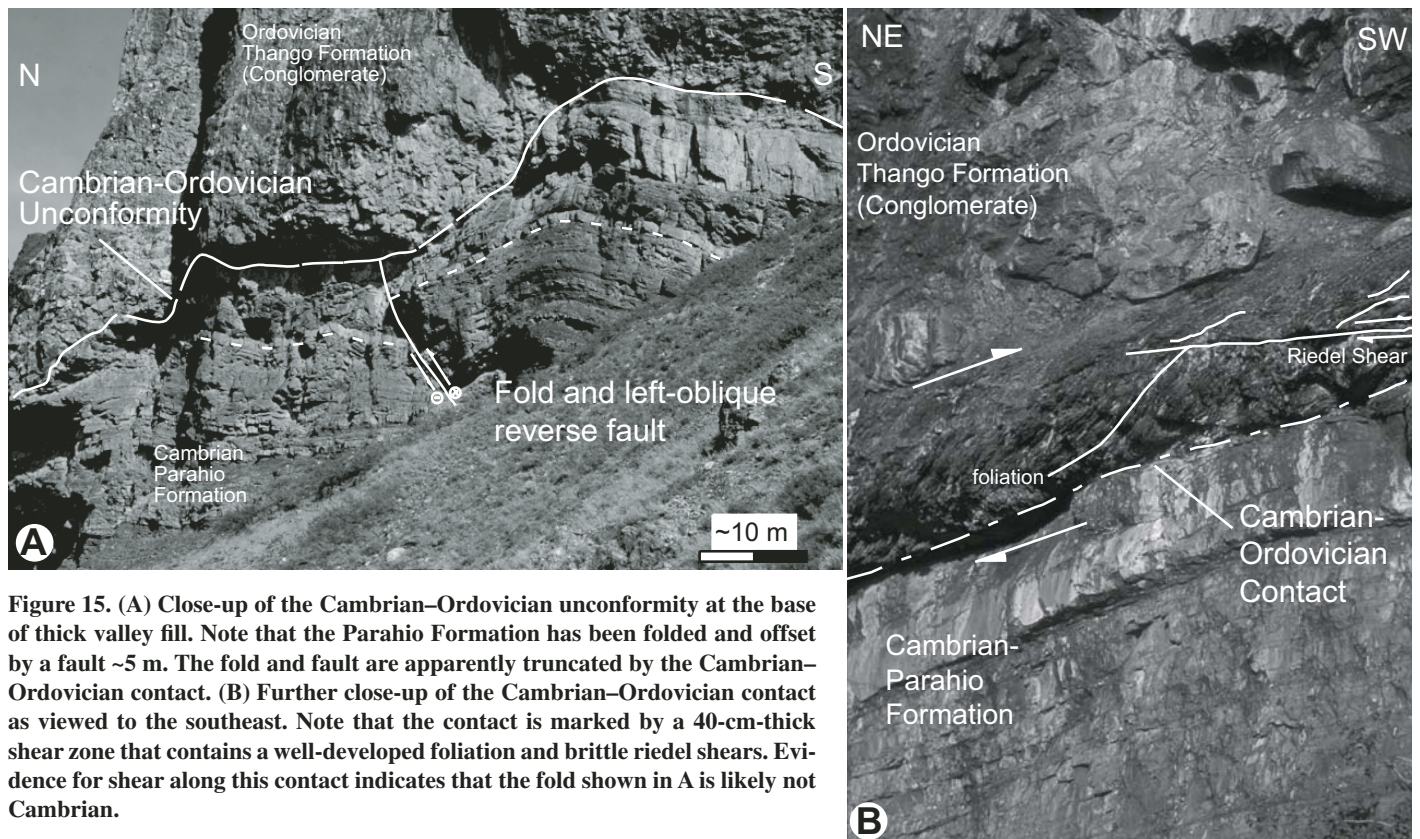


Figure 15. (A) Close-up of the Cambrian–Ordovician unconformity at the base of thick valley fill. Note that the Parahio Formation has been folded and offset by a fault ~ 5 m. The fold and fault are apparently truncated by the Cambrian–Ordovician contact. (B) Further close-up of the Cambrian–Ordovician contact as viewed to the southeast. Note that the contact is marked by a 40-cm-thick shear zone that contains a well-developed foliation and brittle riedel shears. Evidence for shear along this contact indicates that the fold shown in A is likely not Cambrian.

DISCUSSION

As outlined earlier, the paleoenvironments and depositional history of the Parahio Formation have been radically reinterpreted as recording a series of progradational cycles of a storm-dominated shoreline. Thin limestone beds represent transgressive systems tract deposits, which were shown in this study to contain abundant trilobite fossils. New collections from these beds allow for detailed correlation of this important section with those elsewhere in the Himalaya and globally, as well as to define the duration of deposition of the formation and the hiatus of the overlying Cambrian–Ordovician unconformity.

The results of this study have bearing on the continued debate concerning the early depositional and tectonic history of the Himalaya. Part of the debate concerns the relationship between the three lithotectonic zones of the Himalaya—the Lesser Himalaya, the Greater Himalaya, and the Tethyan Himalaya. Paleocurrent data, particularly those from the fluvial trough cross-bedded facies of the Parahio Formation, show a parallelism to those of the uppermost Lower Cambrian Tal Group of the Lesser Himalaya (Ganesan, 1975), namely, with transport to the north-northeast. Myrow et al. (2003) suggested that the uppermost Member E of the Tal, which consists of more than 1000 m of unfossiliferous cross-bedded sandstone, is generally age equivalent to the Parahio and is of possible fluvial origin. However, Hughes et al. (2005) suggested a revised lithostratigraphic correlation in which the carbonate of Member D of the upper part of the Tal Group equates to the late Middle Cambrian Karsha Formation of Zanskar. In this case the Parahio Formation would be correlative with Member C of the Tal Group, a 300-m-thick unfossiliferous cross-bedded sandstone also likely of fluvial origin. Member B of the Tal Group is nearly contemporaneous (± 1 –2 m.y.) with the earliest dated beds in the Parahio Formation.

According to this model, the upper Tal is the record of Lesser Himalaya rivers that fed the deltaic Tethyan Himalaya environments of the Parahio along the northern Indian margin. It is difficult to evaluate the original distance between these two zones because of uncertainties in the Cenozoic displacement along two faults: (1) the Main Central Thrust, with top-to-the-south thrusting; and (2) the South Tibetan detachment system, with top-to-the-north extensional movement along with an older phase of top-to-the-south thrusting. The Main Central Thrust is estimated to have a displacement of 150–200 km or more (Schelling and Arita, 1991; Schelling, 1992); southwest of our study area the displacement is estimated as a minimum of 100 km

(Vannay and Grasemann, 2001). The recent top-to-the-north displacement along the South Tibetan detachment system is estimated to be a minimum of 35 km in both Nepal (Burchfiel et al., 1992; Hodges, 2000) and Zanskar (Dezes et al., 1999). The displacement magnitude of the earlier south-directed phase of thrust movement along the South Tibetan detachment system is also unclear. Detailed lithofacies analysis of the upper Tal Group is needed as a test of the link between Lesser Himalaya and Tethyan Himalaya ancient environments, although the similarities in detrital zircon age spectra (Myrow et al., 2003) and paleocurrents are strong evidence for contiguous depositional systems of the same northern Indian margin, with no evidence for an intervening landmass separating the two regions (cf. Saxena, 1971; Bhargava and Bassi, 1998).

Our structural and stratigraphic analysis indicates that although the unconformity in the Spiti Valley is generally a low-relief surface, considerable local relief of several hundred meters is represented by a thick valley fill. The duration of the hiatus in the Spiti Valley was a minimum of 15 m.y. and possibly as much as 20–30 m.y. (Myrow et al., 2006). A variety of data, including the spatial extent of the unconformity, the difference in the level of erosion along the length of the Himalaya, and evidence for early structural deformation in the adjacent Pin Valley, all indicate that the Cambrian–Ordovician unconformity had a tectonic, as opposed to a purely eustatic, origin.

The unconformity and bounding strata in Zanskar were considered by Garzanti et al. (1986) to have recorded development of a deep-water foreland basin and adjacent subduction complex. Gehrels et al. (2003) extended this model to include southward-directed thrusting and development of coarse molasse wedges shed from north to south. Our paleocurrent data from Spiti and additional data from Zanskar (Myrow et al., 2006) indicate flow toward the north during the Cambrian. Paleocurrents taken from Ordovician molasse deposits of northern India also indicate transport to the north and northeast (Garzanti et al., 1986; Bagati et al., 1991), parallel to the orientation of the meter-scale scours along the Cambrian–Ordovician unconformity described earlier. These are inconsistent with the models of foreland basin development as just described (see also Myrow et al., 2006). Given the northward-directed Ordovician paleocurrents, it is possible that uplift may have taken place south of the present position of the South Tibetan detachment system to provide a source of sediment for Ordovician strata, but features that would clarify the nature of this purported uplift are obscured owing to Cenozoic high-grade

metamorphism and deformation as recorded in the Greater Himalaya.

A southern uplift is also problematic in that no well-developed Cambrian-age structural fabrics are preserved in strata from either the Tethyan Himalaya or the Lesser Himalaya. In addition, clasts in the Ordovician molasse of Spiti are coarse (boulder size) and of local origin, so they provide no evidence of long distance transport. Ultimately, detailed integrative analyses will be required along the length of the Himalaya to decipher, even to the first order, the nature of this enigmatic Cambrian–Ordovician event.

CONCLUSIONS

The Parahio Formation of the Spiti region represents the northern margin of India during the Cambrian. Facies analysis and northeast-directed paleocurrents in this >1350-m-thick succession indicate that these strata record prograding river- and storm-influenced deltaic deposits. This is in striking contrast to previous paleoenvironmental interpretations that ranged from a deep-water flysch setting to a tidally dominated shoreline (Srikantia, 1981; Fuchs, 1982). These strata contain well-defined decameter-scale cycles that record high sediment accumulation rates and likely resulted from delta-lobe switching, based on a lack of systematic stratigraphic changes in cycle or facies thicknesses. Trilobite-bearing limestone beds represent thin transgressive systems tract deposits developed over marine flooding surfaces. The fauna indicate that the formation ranges from uppermost Lower Cambrian (Lungwangmiaoan Stage) to middle Middle Cambrian (Hsuchuangian Stage). Paleocurrent data from the Parahio Formation indicate northward flow, and, combined with geochemical and geochronological data for these strata, they suggest continuity of depositional systems between the Lesser Himalaya and the Tethyan Himalaya in the Cambrian.

A prominent Cambrian–Ordovician unconformity at the top of the Parahio Formation is marked by irregular meter-scale scours and a valley-fill succession >100 m thick. The meter-scale scours are oriented northeast-southwest, parallel to the inferred north- and northeast-directed transport of sediment, as measured from paleocurrents in the overlying Ordovician deposits. The unconformity is of regional extent and, within northern India, shows progressive downcutting from Zanskar in the north-northwest toward the Spiti Valley. Based on our trilobite data, the minimum hiatus associated with the unconformity in Spiti is 15 m.y.

The nature of this Cambrian–Ordovician boundary unconformity remains an enigma. Stratigraphic and sedimentological data from

the Parahio Formation and overlying units from both the Spiti and Zaskar Valleys are not consistent with published models of foreland basin development with southward-directed thrusting. The possibility exists that uplift took place south of the present southern limit of the Tethyan Himalaya, namely, the South Tibetan detachment system, but there is no published evidence for such an uplift in either the Greater or Lesser Himalaya lithotectonic zones to the south.

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