Facies and sequence controls on the appearance of the Cambrian biota in southwestern Mongolia: Implications for the Precambrian–Cambrian boundary

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Facies and sequence controls on the appearance of the Cambrian biota in southwestern Mongolia: implications for the Precambrian–Cambrian boundary

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Abstract – Neoproterozoic–Cambrian rocks of the Zavkhan Basin (Govi-Altay, western Mongolia) comprise large-scale alternations of siliciclastic- and carbonate-dominated units (cf. 'Grand Cycles'). Analysis of such depositional sequences near the base of the Cambrian confirms that the distribution of trace fossils, small shelly fossils and calcimicrobial structures was strongly controlled by ecology and taphonomy, corresponding to specific points in a sea-level cycle. Evolution of the Cambrian biota is thus viewed through a series of narrow time windows, once only for each depositional cycle. Correlation of the Precambrian–Cambrian boundary level on the basis of the first appearance of the *Phycodes pedum* assemblage is also fraught with difficulty, since stratigraphic resolution may be limited to a single sea-level cycle (*c.* 1–5 Ma). It is suggested that, in many cases, basin analysis will need to be undertaken before this boundary can be drawn.

1. Introduction

The Precambrian–Cambrian boundary has now been defined in southeastern Newfoundland, there recognizable on the first appearance of a trace fossil assemblage containing *Phycodes pedum* above an assemblage characterized by *Harlaniella podolica* (Brasier, Cowie & Taylor, 1994; Landing, 1994). A major problem exists, however, in correlating this boundary into Asia, where early skeletal fossils abound but trace fossils are seldom found in association. A search was therefore made for sections in Asia where these two kinds of assemblage could be found intermixed, and the age of the skeletal fossil assemblages is reasonably well constrained. Participants in the Joint Russian–Mongolian Palaeontological Expedition to southwestern Mongolia in 1991 were impressed by the potential of these sections to provide such calibration. One of the aims of the second expedition to Mongolia in August–September 1993 was, therefore, to attempt integration of the skeletal fossil and trace fossil record in the context of facies analysis and sequence stratigraphy which is the focus of this paper.

Neoproterozoic and Cambrian rocks are well exposed in the Zavkhan Basin of the Govi-Altay region of western Mongolia (Fig. 1). The strata in this area form large-scale alternations of siliciclastic-dominated and carbonate-dominated cycles (Fig. 2) similar to those described from the Western Cordillera of North America which have been referred to as 'Grand Cycles' (Mount & Signor, 1992). These cycles are in fact depositional sequences. In this paper we focus on the more fossiliferous depositional sequences slightly above the putative Precambrian–Cambrian boundary in southwestern Mongolia in order to highlight the factors affecting facies and fossil distribution across the boundary itself.

Many sequence boundaries can be shown to be more widespread than just a single basin, and some at least appear to have a global distribution which has led to suggestions that they are eustatic in origin (see, e.g. Vail, Mitchum & Thompson, 1977*a*,*b*; Vail, Hardenbol & Todd, 1984; Haq, Hardenbol & Vail, 1987, 1988). In spite of these suggestions, data are still insufficient to allow full evaluation of the synchroneity of most sequence boundaries at a global scale. One of the strongest pieces of evidence for global synchroneity of at least some Neoproterozoic sequences (1000 to 544 Ma: Plumb, 1991; Bowring *et al.* 1993) is the global distribution of glacial rocks (Hambrey & Harland, 1985; Lindsay *et al.* 1996, this issue).

2. Basinal setting

The Zavkhan Basin appears to have originally been part of the earlier Zavkhan Rift which formed during the Baykalian (Cryogenian, 850–650 Ma) as a result of the breakup of the Baydrik–Tarbagatay microcontinent. The basin formed on continental crust as part of a vast marginal marine back-arc basin that evolved during pre-Vendian accretion of the Zavkhan Rift and the adjacent fragments

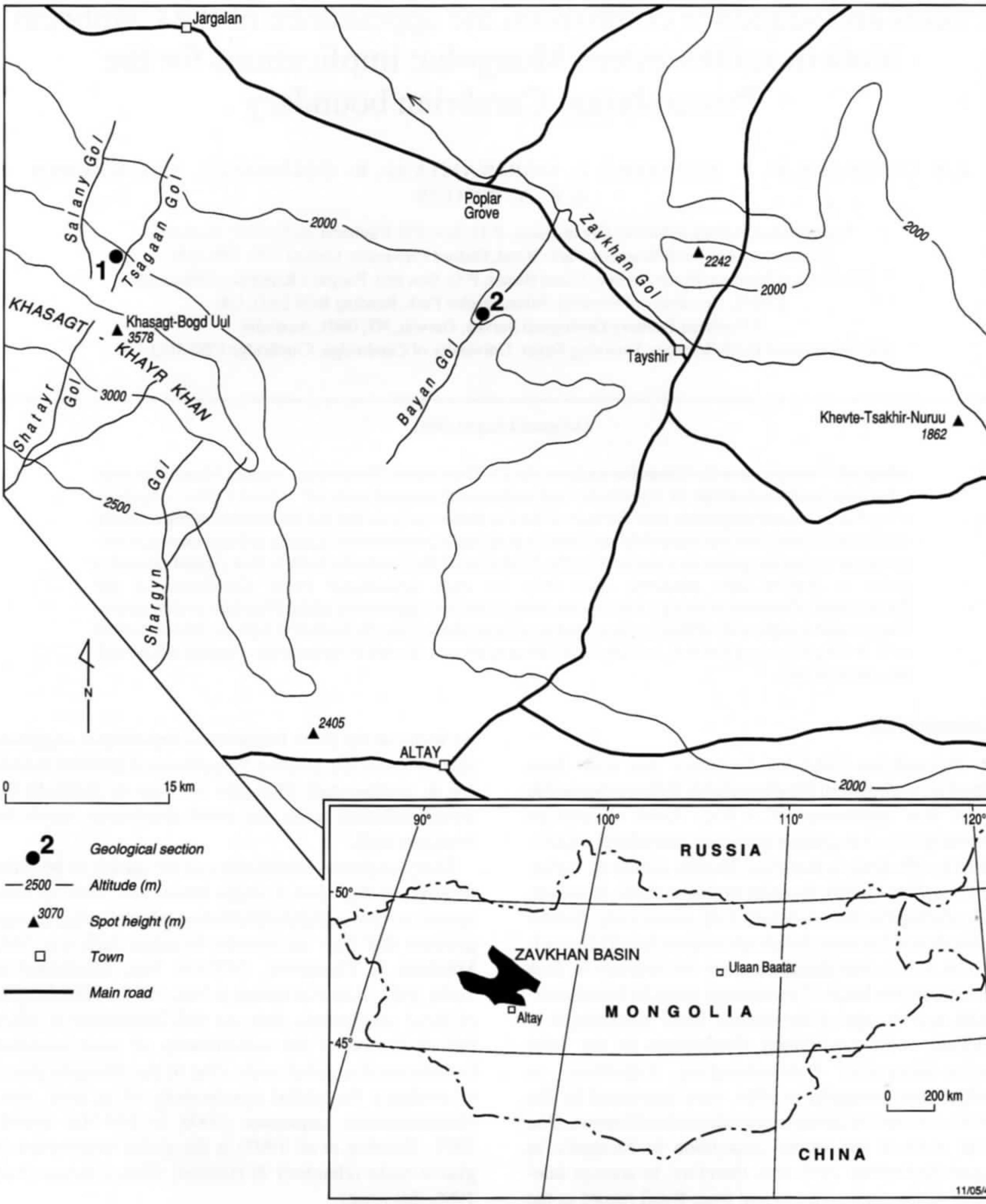

Figure 1. Map of the Altay area showing in detail the location of the two sections examined. (1) Tsagaan Gol section at the foot of the Khasagt-Khayr Khan Range, and (2) Bayan Gol section.

of Baydrik–Tarbagatay microcontinent (Khomentovsky & Gibsher, 1996, this issue). The back-arc basin appears to have been stable over a long period and accumulated a relatively uninterrupted Neoproterozoic and earliest Palaeozoic succession. However, with the exception of a recently discovered outcrop 100 km to the east of Tayshir (Dorjnamjaa *et al.* 1993) where the Neoproterozoic basinal succession can be seen to rest unconformably on the crystalline basement, the boundaries of the basin are almost entirely tectonic.

The Zavkhan Basin sedimentary fill has been subdivided into five formations. The earliest unit, the Dzabkhan Formation, is largely of andesitic volcanic origin and has been shown to rest on crystalline continental

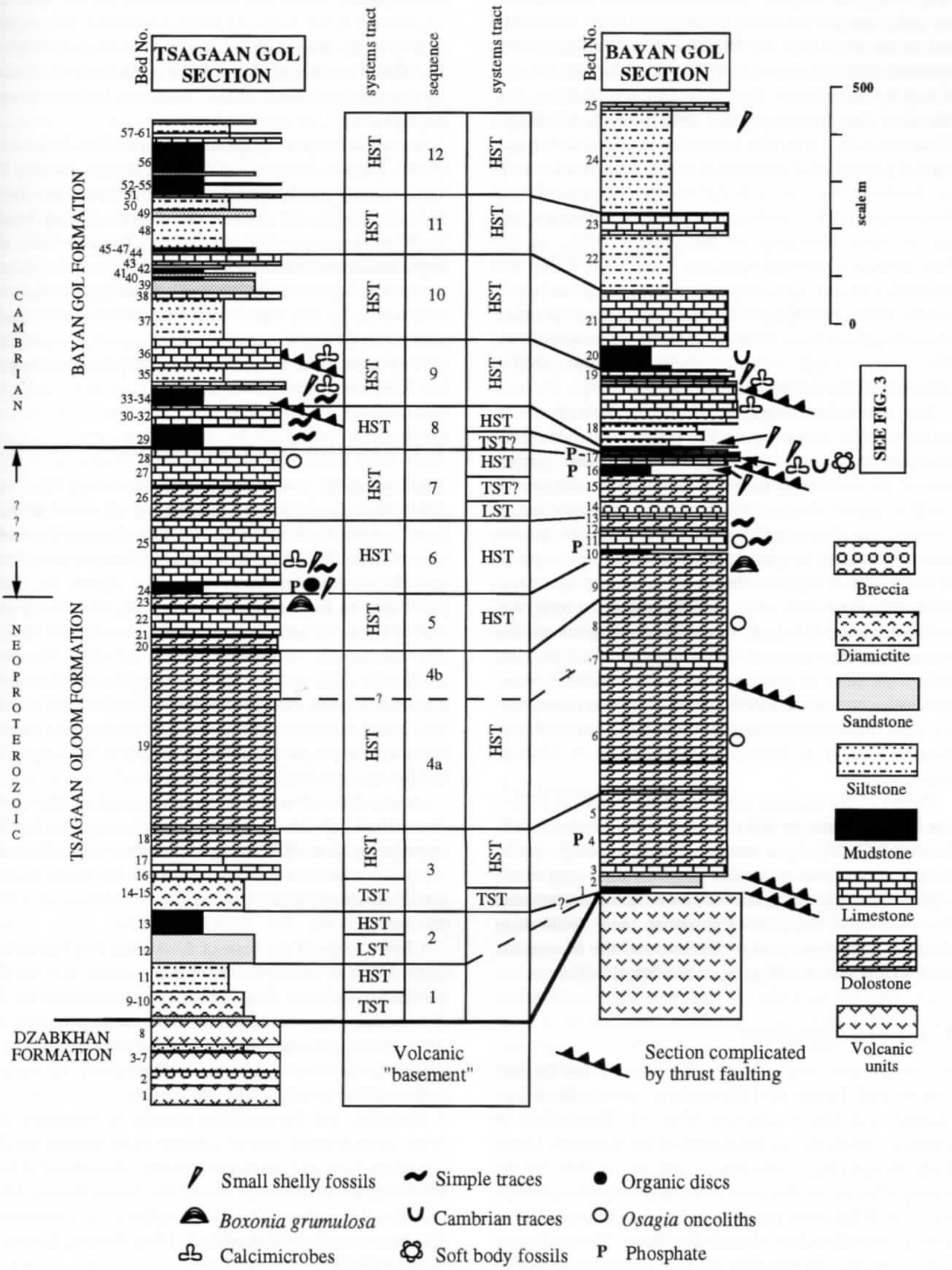

Figure 2. Stratigraphic column for the Tsagaan Gol section on the north flank of the Khasagt-Khayr Khan Range and the Bayan Gol section near Tayshir. Sequence stratigraphy and the associated systems tracts are outlined in the central columns (HST – highstand systems tract, LST – lowstand systems tract, TST – transgressive systems tract). Stratigraphic columns adapted from Gibsher *et al.* (1991).

basement. The Tsagaan Oloom Formation rests unconformably on the volcanics of the Dzabkhan Formation and in its lower part, the Maikhan Uul Member (lower member of Khomentovsky & Gibsher, 1996, this issue), is largely siliciclastic. Diamictites of the Maikhan Uul Member (Dorjnamjaa *et al.* 1993) of glacial origin (Lindsay *et al.* 1996, this issue) are included in the unit and are of particular importance in regional correlation. It is, however, the main body of the Tsagaan Oloom Formation and the overlying Bayan Gol Formation that are of most relevance to the present study, as the Precambrian–Cambrian boundary is included within this interval. The two formations are mixed siliciclastic–carbonate units containing a diverse fossil fauna including assemblages of trace fossils (Goldring & Jensen, 1996, this issue) and small shelly fossils (SSFs) (Khomentovsky & Gibsher, 1996, this issue).

The regional stratigraphy of the Zavkhan Basin is poorly known. However, the information available from the Altay area shows that, in general, the clastic component of the succession increases through time (Brasier & Lindsay, unpub. data), although our very preliminary suggestions for depositional sequence boundaries do not show an increase in spacing (see Fig. 2). If this is typical of the region, it suggests that accommodation space was increasing with time and that the clastic source was becoming more proximal, or the erosional gradient was steepening. The sequence stacking pattern is consistent with deposition in response to a compressional event, perhaps related to the development of a destructive margin over the Precambrian–Cambrian transition (see, e.g. Şengör, Natal'in & Burtman, 1993; Brasier & Lindsay, unpub. data).

Dating of the sections using small shelly fossils (SSFs) has been outlined by Khomentovsky & Gibsher (1996, this issue). Independent studies of carbon isotopic variation in whole rock carbonates confirms that, with minor adjustments, good correlations can be achieved between the Nemakit-Daldynian, Tommotian and Atdabanian stages in their type area of Siberia, and the Mongolian sections studied here (Brasier *et al.* 1996, this issue).

3. Sequence stratigraphy

Reconnaissance studies were carried out on the Tsagaan Oloom and Bayan Gol Formations at two localities: Tsagaan Gol and Bayan Gol (Fig. 1). The section at Tsagaan Gol on the northern flank of the Khasagt–Khayr Khan Range (Fig. 1, locality 1 and Fig. 2, left) 50 km west of Tayshir is the most complete over this interval, and there it has been possible to provisionally identify at least 13 complete depositional sequences. The sequences consist largely of shoaling cycles, beginning either in fine-grained clastic sediments or laminated dark carbonates and passing upward into massive, cliff-forming platform carbonates at the top (Fig. 2). The sequences are thus, for the most part, relatively thin highstand systems tracts stacked one on the other. Occasionally, transgressive systems tracts are preserved, as is the case for sequence 7 in the Tsagaan Oloom Formation. If lowstand deposits are preserved in any of the Tsagaan Oloom Formation sequences, they cannot be distinguished readily from the carbonates of the underlying highstands with any certainty.

A similar section is exposed at Bayan Gol, 30 km east of the Tsagaan Gol section (Fig. 1). At this locality the sequences are much more clearly defined, but thrust faulting has complicated parts of the section, making larger-scale observations difficult. As at Tsagaan Gol, the sequences consist mostly of stacked highstand systems tracts which show overall upward-shoaling facies associations (Fig. 2). The clastic–carbonate alternation is much sharper in this section and in some sequences transgressive systems tracts are present. Lowstands are not apparent in most sequences.

4. Precambrian–Cambrian boundary

Khomentovsky & Gibsher (1996, this issue) have outlined the biostratigraphic distribution of small skeletal fossils in the Zavkhan Basin and their suggested correlation with the Siberian succession. *Anabarites trisulcatus* and *Cambrotubulus decurvatus* first appear in strata equivalent to Bayan Gol section unit 11, in sections outside of the study area. Since these fossils mark the base of the *Anabarites trisulcatus* Zone of the Nemakit-Daldynian stage in Siberia, they infer that this boundary lies here. Carbon isotopic studies by Brasier *et al.* (1996, this issue) confirm that these fossils appear at a similar level with respect to negative 'Anomaly W' in both northern Siberia and southwest Mongolia.

At the base of unit 18, faunas typical of the upper Nemakit-Daldynian *Purella antiqua* Zone appear. Its first appearance across Siberia is diachronous, but carbon isotopic data confirm a position well below the likely base of the Tommotian stage in Mongolia (Brasier *et al.* 1996, this issue).

Clastic rocks at the base of the Bayan Gol Formation (unit 18) also clearly lie within the upper part of the Nemakit-Daldynian Stage. Faunas characteristic of the Tommotian stage do not appear until the base of unit 19, although we note that this fossil datum lies some 100 m below the position suggested independently by carbon isotopes (Brasier *et al.* 1996, this issue).

Globally, the Precambrian–Cambrian boundary has been defined such that it is there recognizable on the appearance of the *Phycodes pedum* ichnofossil Zone, which lies immediately above the latest Precambrian *Harlaniella podolica* Zone in southeast Newfoundland (Narbonne *et al.* 1987; Landing, 1992; Brasier, Cowie & Taylor, 1994).

The 1991 expedition discovered a Cambrian type trace fossil (*Diplocraterion* isp.) below exposures of the Bayan Gol Formation on the slope of Salany Gol. Small tubular burrows in units 11 to 13 of the Tsagaan Gol Formation at Bayan Gol, northern block (Fig. 2) were indeterminate. A study of trace fossils was therefore one of our priorities during the expedition.

Several assemblages of trace fossils are confirmed by Goldring & Jensen (1996, this issue). In unit 18, *Didymaulichnus miettensis*, *Helminthoida* cf. *miocenica*, *Planolites* isp. and doubtful *Phycodes* isp., occur together with the psammocorals *Spatangopsis* sp., and *Nemiana*. Float in the upper part of unit 20 contains a more diverse assemblage including *Phycodes pedum*, *Helminthoida* cf. *miocenica*, *Palaeophycus tubularis*, *Rusophycus* cf. *avalonensis*, *Monomorphichnus* isp. *Treptichnus ?bifurcus*, *T.* cf. *triplex.*, *Cochlichnus* isp. *Hormosiroidea* isp., *Planolites* isp. *Didymaulichnus miettensis*. Unfortunately, diagnostic traces could not be found at levels lower than unit 18.

Helminthoida has been reported from the Vendian (Redinko horizon) of Russia (Fedonkin, 1985; 1988) and is associated with an Ediacaran biota in northwestern Canada (Narbonne & Aitken, 1990). Crimes & Anderson (1985) recorded *Didymaulichnus miettensis* from member 5 of the Chapel Island Formation (Newfoundland), still higher in the stratigraphy, and *Didymaulichnus* isp. from members 2 and 3 of the Chapel Island Formation of Newfoundland, although it was not recorded as such by Narbonne *et al.* (1987) and may have been included as *Taphrhelminthopsis circularis*; present in members 2A–5. In China and Australia *Didymaulichnus miettensis* occurs in Lower Cambrian strata only (Crimes & Jiang, 1986; Walter, Elphinstone & Heys, 1989). However, reports of *Didymaulichnus miettensis* from the Miette Group (Young, 1972) and Stelkuz Formation (Fritz & Crimes, 1985), British Columbia, are of uncertain late Proterozoic–earliest Cambrian age. The occurrence of *Spatangopsis* in unit 18 gives the assemblage an early Cambrian aspect (Seilacher & Goldring, submitted). Hence the assemblage in unit 20, at least, is indisputably Cambrian.

Unfortunately, therefore, our preliminary study of trace fossils does little to resolve the question of the Precambrian–Cambrian boundary in Mongolia. The distribution of *Phycodes* in Siberia is reviewed by Brasier *et al.* 1996 (this issue). From this (and Bowring *et al.* 1993) it seems that the taxon appeared contemporaneously with the *Anabarites trisulcatus* fauna. This suggests that the base of the Nemakit-Daldynian Stage in Siberia correlates approximately with the base of the Cambrian System in Siberia (see, e.g. Brasier, 1992*a*,*b*; Brasier, Anderson & Corfield, 1992; Brasier *et al.* 1996, this issue).

The columnar stromatolite *Boxonia grumulosa*, prominent in Bayan Gol unit 9 and correlative strata across the region, is a form typical of the early and middle (that is, apparently pre-Nemakit-Daldynian) Vendian (Sokolov & Fedonkin, 1984, fig. 2; Semikhatov, 1991; Khomentovsky & Gibsher, 1996, this issue). This accepted, *B. grumulosa* provides a biostratigraphic datum that must lie below the Precambrian–Cambrian boundary.

Taking the above evidence into account, in particular that elements of the *Anabarites trisulcatus* Zone appear in Bayan Gol section unit 11, we suggest (*pace* Khomentovsky & Gibsher, 1996, this issue) that the Precambrian–Cambrian boundary lies within units 10 to 11 of the Bayan Gol section, and at equivalent levels elsewhere. This spans the first appearance of the *Anabarites trisulcatus* Zone in both Mongolia and Siberia, the first appearance of *Phycodes* isp. in Siberia, and is characterized in both regions by a distinctive carbon isotopic negative anomaly ('Anomaly W').

5. Sequence analysis

Depositional sequences 8 and 9 (Fig. 2) are among the best exposed and most complete sequences in the Bayan Gol section (Figs 4, 5). Since these two sequences (or parasequences) contain SSFs, trace fossils and carbon isotopic signatures suggestive of the Nemakit-Daldynian to Tommotian transition, an opportunity was taken to study the relationship between fossil occurrences and facies, and relate these as far as possible to sequence stratigraphic concepts.

Sequence 8 is a relatively thin unit consisting only of a thin carbonate transgressive systems tract overlain by a condensed siliciclastic highstand systems tract in the Bayan Gol section, although it is similar in thicknesses to the other sequences at the Tsagaan Oloom section a few kilometres to the northwest. Sequence 9 is a much thicker unit, again interpreted as a thin carbonate transgressive systems tract overlain by a well developed thick highstand systems tract (Fig. 3). A lowstand systems tract cannot be distinguished for either sequence.

5.a. Sequence 8

The transgressive systems tract of sequence 8 is approximately 9 m thick and consists largely of cliff-forming limestone. Bed A (Fig. 5) appears to form the base of the transgressive unit. The lower 4.5 m of the transgressive unit (Beds A–B) consist largely of limestone and dolostone with interbeds of marl and occasional quartz-rich pockets. The lower and upper parts of Bed B tend to be more siliciclastic-rich than the intermediate parts which tend to be nodular carbonates. Trace fossils are present locally as are limonite-coated clasts and calcimicrobial biostromes.

The highstand systems tract is approximately 10 m thick in the Bayan Gol section (Fig. 5). The downlap surface (maximum flooding surface) is a well defined ferruginized surface at the top of Bed C with a relief of approximately 1.5 m. The rocks consist in large part of olive green-brown, thinly bedded shales (Bed D) which are otherwise virtually devoid of sedimentary structures. These shales pass upward into medium-grained, well-bedded, brown quartz sandstone and intraformational sand-clast conglomerate with thin shale interbeds. This sandstone (Bed E) contains the trace fossil *Helminthoida* isp. (*Planolites* isp. having appeared in a thin clastic interval in the uppermost part of the preceding sequence).

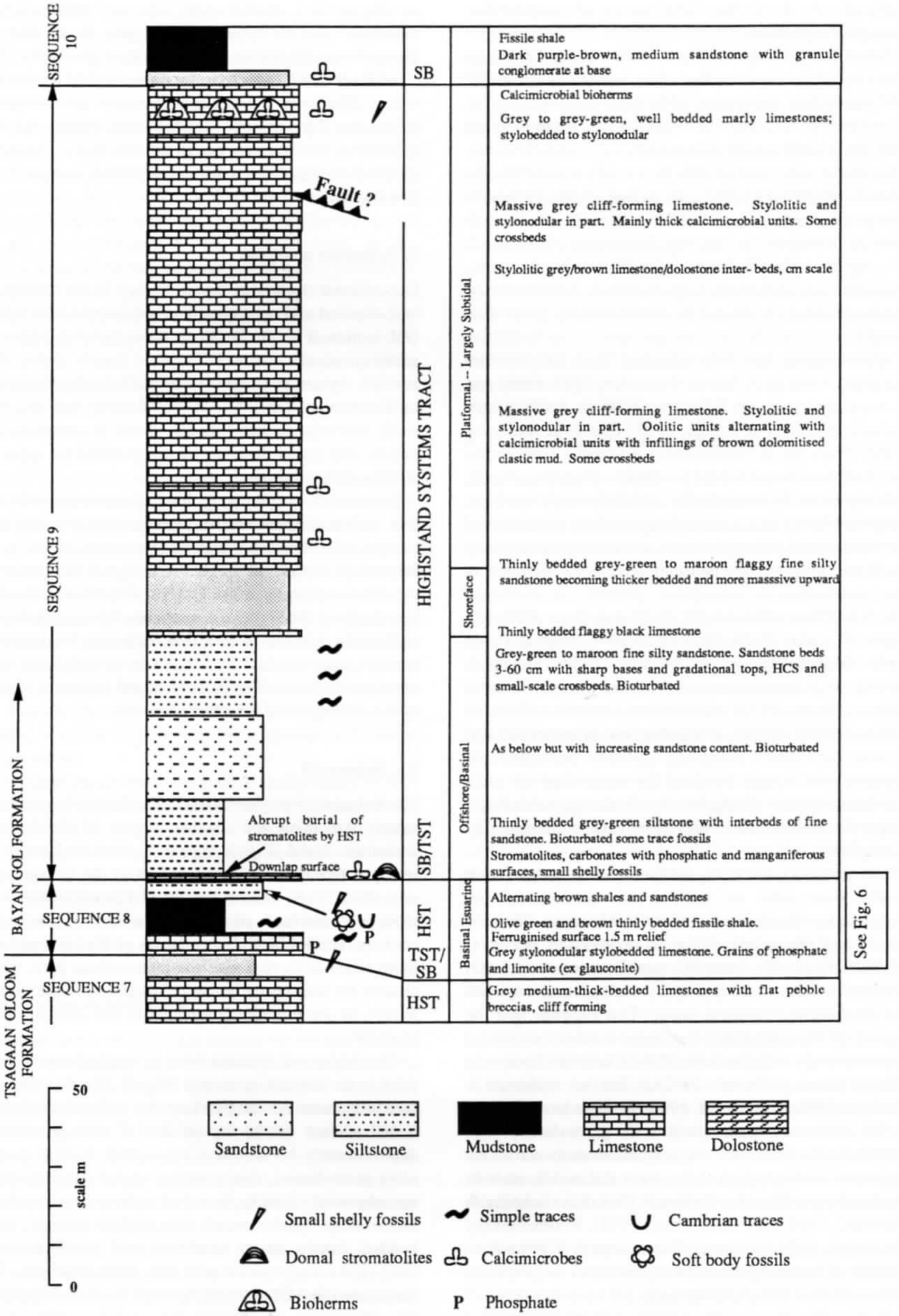

Figure 3. Stratigraphic column for sequences 8 and 9 at the base of the Bayan Gol Formation in the Bayan Gol section (see Fig. 2).

Figure 4. A panoramic view of sequences 8 and 9 showing a stratigraphic interval of approximately 220 m from the floor of the Bayan Gol ravine to the skyline. Sequence 8 is to the lower right. The highstand of sequence 9 to the centre and left can be seen to consist of a lower siliciclastic unit overlain further up the slope by cliff-forming carbonates to form a 'Grand Cycle' (cf. Mount & Signor, 1992). See also Khomentovsky & Gibsher (1996, this issue, fig. 7)

Succeeding interbedded sandstones and shales additionally yield first appearances of *Didymaulichnus miettensis*.

5.b. Sequence 9

There is an abrupt transition at the sequence boundary (base of Bed H, Fig. 5) from largely siliciclastic rocks of the previous highstand to limestone and dolostone interbeds with the first abundant small skeletal fossils (SSFs) of the 4-m-thick transgressive unit. The units are irregularly bedded and often contain a quartzitic sandy component. Tabular rip-up clasts of olive-green shale and phosphate are also common in the grainy carbonates (Bed H) at the base of the unit. Phosphatic crusts and glauconitic–phosphatic sands are common and toward the top (Bed J) are pink carbonates with phosphatic hardgrounds overlain with Fe–Mn crusts. Bedding surfaces throughout the limestone and dolostone unit are irregular and generally appear to be erosional.

Beds J and K represent a complex carbonate interval (Fig. 6). They form a distinctive 3-m-thick unit which rests directly on the phosphatized hardgrounds and appears to consist of two carbonate cycles. The lower cycle (J1 and J2 in Fig. 5) begins in pink and grey stylobedded limestone with calcimicrobial biostromes resting directly on a firm substrate. Small calcimicrobial bioherms are up to 20 cm in diameter and 10 cm in relief. The size of the bioherms decreases upwards, and ministromatolites up to 5 cm in diameter, which can be seen to be growing on the small bioherms, cap the succession. Thin brown dolomitized mudstone interbeds link the bioherms.

The uppermost carbonate cycle (J3, J4 and K in Fig. 5) begins with grey stylobedded–stylonodular limestone (J3) incorporating carbonate sand and coated grains. This is replaced abruptly upwards by grey limestone with columnar-stacked calcimicrobial bioherms up to 30 cm in diameter which extend upward to the top of J4. Reddish, ferruginous, domal to digitate ministromatolites grow on and between the bioherms. Above J4 the grey limestone consists of prominent domal stromatolites (Bed K) up to 30 cm in diameter and slightly elongate in plan (Fig. 6). The stromatolites developed upon a well-defined substrate of stacked calcimicrobial bioherms at the top of J4.

Strata attributed to the highstand systems tract of sequence 9 consist of two distinctive lithologic units: a siliciclastic-dominated unit and an overlying massive cliff-forming carbonate-dominated unit (Figs 3, 4).

5.b.1. The clastic unit

The stromatolites of the transgressive systems tract were buried abruptly by clastic sediments as the sequence 9 highstand systems tract prograded basinward across the site (Fig. 6). Thinly bedded grey-green siltstone with very thin interbeds of fine grey quartz sand rest directly on the domed surface of the stromatolites, which have a relief of up to 40 cm. The downlap surface is thus very sharply defined in terms of both lithology and facies. The siltstone contains small numbers of poorly-preserved trace fossils. The frequency and thickness of sandstone interbeds gradually increases upward in the section. At a point approximately 42 m above the stromatolite surface (downlap surface), sandstone occurs in approximately equal proportions to siltstone in beds 3 to 60 cm thick. The sandstone beds have sharp bases and gradational tops, with clay-rich drapes on their upper surfaces. They are frequently cross-bedded and some contain hummocky cross-stratification. The sandstone units appear to be bioturbated.

At 61 m above the stromatolite unit there is an abrupt change in both facies and rock type, and clastic sediments

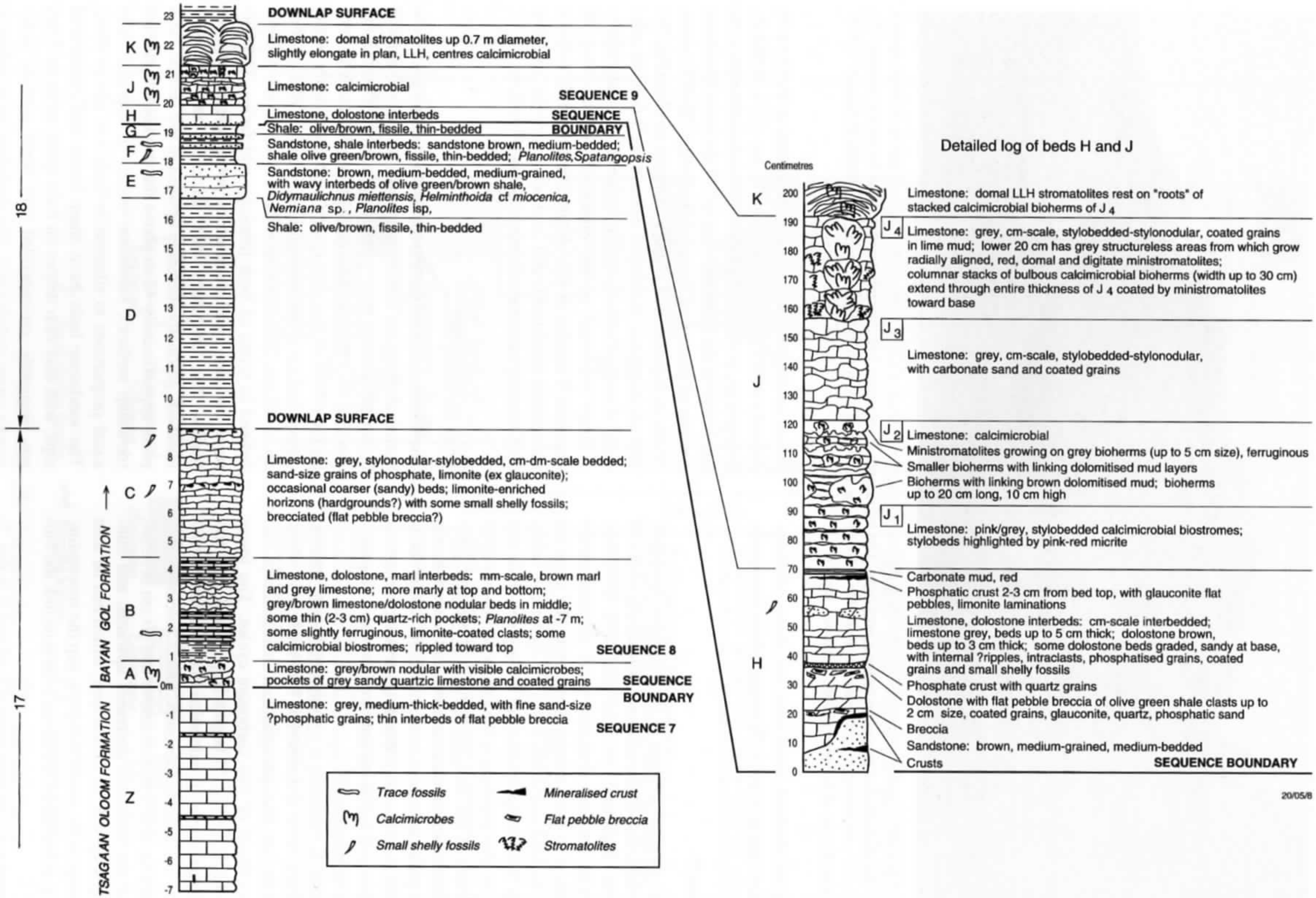

Figure 5. Detailed stratigraphic column of sequence 8 at the Bayan Gol section (see Fig. 3).

Figure 6. The downlap surface of sequence 9 showing siliciclastic sediments of the highstand (left) abruptly overlying the domal stromatolites (Bed K, Fig. 5) at the top of the transgressive systems tract. The stromatolitic domes are typically 30 cm in diameter.

are replaced by a 2-m-thick, thinly bedded, black, tabular, stylolitic limestone. Above this thin limestone there is again an abrupt return to clastic sediments. Grey-green, thinly bedded, tabular, silty sandstone is the dominant rock type. In general, bedding is thicker than in clastic units below and the unit is cross-stratified.

5.b.2. The carbonate unit

The upper 124 m of the highstand consists of massive cliff-forming carbonates, mostly limestone (Fig. 3). At the base, grey, stylonodular, calcimicrobial limestones alternate with interbedded ooid units. These pass upward into a thick interval of medium-bedded grey limestone with thinner brown dolomitized interbeds. Toward the top, recessive, stylobedded to stylonodular, impure limestone units gradually become more common. Calcimicrobial bioherms up to 1 m in diameter reappear in thick beds in the uppermost 20 m. Due to poor exposures and the limited nature of the reconnaissance it was not determined whether the upper carbonate unit includes a transgressive unit belonging to the overlying sequence 10. However, a dark purple-brown, medium sandstone with granule conglomerate at its base which overlies the upper cliff-forming carbonates of sequence 9 is likely to be the transgressive systems tract of sequence 10.

5.c. Environmental interpretation

Subsidence was relatively slow in the Zavkhan Basin during Neoproterozoic and earliest Cambrian times. This, combined with low palaeoslopes and a limited supply of fine-grained clastic sediment, led to a delicate balance between clastic and carbonate sedimentation which, in turn, resulted in the deposition of cyclic sequences consisting of thin highstand systems tracts which alternate in a regular, relatively predictable way between carbonate and siliciclastic sedimentary units. In general, lowstand systems tracts are rarely preserved, and transgressive systems tracts, where present, are extremely thin. Lower in the Bayan Gol section (unit 14, sequence 7), carbonate breccias are interpreted as lowstand deposits, formed by mass movement of underlying platform carbonates of the previous highstand.

Evidence indicative of depositional environment is relatively limited in sequence 8. The downlap surface shows minimal erosional relief and some evidence of iron staining and encrustation. Initially the environment was dominated by fine muds in a low-energy setting. Toward the top, scoured surfaces beneath the sandstone units together with small-scale cross-bedding suggest a relatively high-energy subtidal setting. The environment was likely to have been a broad embayment which gradually shallowed. The fact that the same sequence is much thicker at the Tsagaan Oloom section suggests that the Bayan Gol section is more proximal and perhaps close to the basin margin.

The thin carbonate-dominated transgressive systems tract of sequence 9 was deposited in a higher-energy setting in which SSFs were winnowed and concentrated on sediment-starved hardgrounds which were stabilized by phosphatization. Calcimicrobes were able to flourish on these stable surfaces which led in turn to a succession of stromatolites stacked upon calcimicrobial bioherms.

Calcimicrobial activity at the end of the transgression was, however, abruptly terminated by the rapid progradation of clastic sediment across the downlap surface to form the main body of the lower part of the highstand systems tract. Mount & Signor (1992) have suggested that clastic sediments are released because drowning of the platform develops an open storm-influenced shelf which allows sediments previously trapped in nearshore and marginal depocentres to be eroded. Clastics entering the Zavkhan Basin at this time were predominantly fine-grained and deposition began below the local storm-wave base. Thin sandstones gradually increase in numbers upward from the downlap surface until, at approximately 40 m above the stromatolites, the influence of the storm wave base is clearly seen in sandstone units with sharp bases, cross-bedding and locally some hummocky cross-stratification. Each storm event produced a thin, graded sandstone unit. The upper surfaces of these units are coated with a thin clay-rich drape deposited from the water column by the waning storm event. The catastrophic burial of the fauna may have reduced population density abruptly, so that overlying drapes provided a setting suitable for the preservation of surviving trace-makers. Since the sediment is fine grained and clay rich and the density of trails is relatively low, they can be easily seen. Unfortunately, this kind of top-surface preservation does not yield many diagnostic features. This unit is approximately 20 m thick. Above this, in the more massive shoreface sandstones, individual animal traces are generally not preserved (Fig. 3).

The rapid transition from clastic to carbonate sedimentation at a point about 90 m above the downlap surface may reflect the fine balance between clastic supply and carbonate production. It may be that, with low palaeoslopes and slow erosion rates, the site had become too distal from the clastic source to allow effective transport of clastics. In general, the platform carbonates which form the upper 124 m of the sequence appear to have resulted from relatively shallow, increasingly low-energy settings. The association of calcimicrobial with ooid limestones in the lower part (Fig. 3) implies initial shallow subtidal, higher-energy conditions. Cross-bedding is present but not common. Lower sequences in the Tsagaan Oloom Formation also became supratidal, with collapse breccias indicative of evaporites. The setting shares many similarities with Grand Cycle carbonates of western North America (see, e.g. Mount & Signor, 1992). We infer that carbonate banks were able to nucleate and expand as the rate of sea-level rise decreased towards the end of the highstand.

Overall, the facies associations in both sequences suggest a fine balance between siliciclastic influx and carbonate production in a setting where regional subsidence was relatively steady over an extended period (cf. Mount, 1984; Mount & Signor, 1992). Some evidence of erosional surfaces near the top of the carbonate indicates that in its latest stages the environment may have been intertidal. This suggests that a thin transgressive systems tract may be present.

6. Discussion and conclusions

Preservation of organic remains is dependent upon the nature of the material to be preserved and the depositional environment in which it accumulates. Fossil occurrences are thus facies dependent. While these observations are perhaps obvious they are critically important to the development of a precise time scale, especially in a setting such as the Zavkhan Basin.

In Neoproterozoic times, subsidence rates in the Zavkhan Basin were relatively low and clastic sediment supply appears to have been minimal, leading to the accumulation of relatively thin carbonate units. Clastic supply then increased progressively, from the latest Neoproterozoic to late Early Cambrian interval, along with a markedly logarithmic increase in the subsidence rate (Brasier & Lindsay, unpub. data). Over the Precambrian–Cambrian boundary interval discussed herein, there was therefore a fine balance between clastic supply and carbonate production. As a consequence, depositional patterns tended to be controlled by sea-level cycles and the facies patterns preserved in sequences were highly cyclic, regularly switching from siliciclastic- to carbonate-dominated settings. Under these conditions, fossils can be found at specific and predictable points within each cycle. Below we give examples of this effect upon trace fossils and small skeletal fossils.

The problem of a strong connection between facies and the first appearance of Cambrian faunas was forewarned by Brasier (1979, 1982) and it continues to be discussed (e.g. Lipps & Signor, 1992). Latterly, it has tended to be the view that the appearance of early skeletal fossils was controlled by the appearance of suitable carbonate facies (Brasier, 1979; Brasier & Hewitt, 1979; Landing, 1992) and phosphatic taphofacies (Brasier, 1990) whereas trace fossils are much less affected, partly because they had a wider environmental tolerance than at later times (e.g. Crimes, 1992).

The distribution of small skeletal fossils in Mongolia conforms to this picture; they occur as phosphatic lags and concentrations within carbonates that lie in the upper, carbonate half of sea-level cycles, some of which may be regarded as highstand systems tracts. Here, however, we suggest that the first appearance of a *Phycodes pedum* trace fossil assemblage is, in practice, affected by problems of facies changes. We note that diagnostic trace fossils, such as *Phycodes*, *Rusophycus* and *Helminthoida*, appear in the stratigraphic section at a point in the shoaling cycle where sands, probably of storm origin, alternate with muds, for example, just below the base of the shoreface. In deeper-water muds of underlying strata, the lack of sediment grain size contrasts means that traces may go undetected. In the carbonates of the overlying half cycle, the lack of suitable lithological variation invariably conceals traces.

As mentioned above, the Precambrian–Cambrian boundary has now been defined to coincide with the appearance of the *Phycodes pedum* assemblage above the *Harlaniella podolica* assemblage in southeastern Newfoundland (Landing, 1994). In this study, we have found that the preservation of diagnostic traces is very likely to be controlled by facies especially within storm-generated sand–mud intercalations in the highstand systems tract, not far below the carbonates. We therefore suspect that the Precambrian–Cambrian boundary cannot be resolved with any greater accuracy using trace fossils than it can be using small skeletal fossils.

The nature of the problem is shown, somewhat diagrammatically, in Figure 7. Successions like those of the stratotype region (Avalonia) and Laurentia are shown to

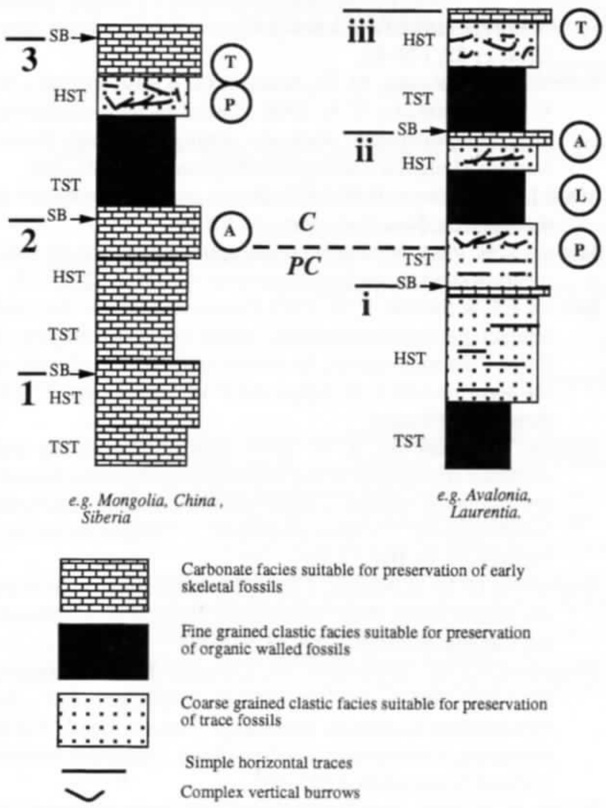

Figure 7. Diagram to illustrate the problem of correlating biological events across the Precambrian–Cambrian boundary between different regions. Cartoon sequences are shown for two kinds of section: low latitude, carbonate-dominant at left (cf. Mongolia, Siberia, China, India, Iran, Pakistan); and mid- to high latitude and/or clastic dominant on the right (cf. Avalonia, Laurentia, Baltica, Australia). Sequence boundaries (SB) are here assumed to relate to regional basin history, and not to coincide. TST – transgressive systems tract; HST – highstand systems tract. Fossil first occurrences are circled: A – first appearance of *Anabarites trisulcatus* and related skeletal faunas, related to occurrence of suitable carbonate facies, often with phosphatic preservation, in all areas. L – first appearance of *Ladatheca* skeletal fauna is related to occurrence of pyritic taphofacies in southeast Newfoundland. T – first appearance of Tommotian skeletal fauna, related to suitable carbonate facies. P – first appearance of *Phycodes pedum* ichnofossil assemblage, related to occurrence of sand–mud interfaces in Mongolia. The Precambrian–Cambrian boundary is here drawn at the first occurrence of *Phycodes pedum* in Avalonia.

the right of the figure. In these two regions, siliciclastic sediments predominate in the Ediacarian to Tommotian interval. Such sediments are suitable for the preservation of trace fossils but not for early skeletal fossils. Earliest indications of the Cambrian biota are usually given by the *Phycodes pedum* ichnofossil assemblage (Fig. 7, level P). Early skeletal fossils tend to occur at higher levels associated with suitable taphonomic conditions such as pyritization in the transgressive systems tract (level L), or phosphatization and carbonate preservation in the highstand systems tract (levels A, T; see, e.g. Landing, 1992; Myrow & Hiscott, 1993).

Successions of Mongolian type are shown at the left in Figure 7. Here, massive carbonates predominated from late Ediacarian to Nemakit-Daldynian times. Lithologies have potential for the preservation of SSFs but not complex trace fossils. Earliest indications of the Cambrian biota are therefore of the *Anabarites trisulcatus* fauna (Fig. 7, level A) and, in higher carbonates, the *Purella antiqua* and Tommotian faunas (Fig. 7, level T). In Mongolia, an ichnofauna of Cambrian type first appears at a high level, in the upper part of the Nemakit-Daldynian (Fig. 7, level P). *Phycodes pedum* appears higher, in the Tommotian. Both assemblages coincide with intervals of sand–mud deposition, associated with upward shoaling.

If sequences were globally synchronous, this would imply that trace fossils (or any other facies-bound group) could seldom achieve resolution better than a single cycle (perhaps *c.* 1–5 Ma in duration). There is, however, no convincing evidence for such global synchroneity of sequences over this interval. Indeed, quite the reverse would be expected from the greatly varying rates of sediment accumulation, and of presumed rates of subsidence (Brasier & Lindsay, unpub. data). If so, then biostratigraphic resolution of the Precambrian–Cambrian in a given basin is likely to depend on the chance matching between four independent processes: organic evolution, faunal migration, sediment supply and accommodation space creation. We infer, therefore, that basin analysis will need to be undertaken before the Precambrian–Cambrian boundary can meaningfully be drawn.

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