ABSTRACT: The Pterocephaliid–Ptychaspid biomere boundary is an Upper Cambrian extinction horizon that can be used to correlate mixed carbonate–siliciclastic strata in northwestern Wyoming. The biomere boundary is defined by the change between the Elvinia and Taenicephalus Zones of the standard North American trilobite zonation. It is also signaled chemostratigraphically by a positive trend in carbon-isotope ratios. Thus, both biostratigraphic horizons and carbon-isotope data are used to constrain the position of the biomere boundary in northwestern Wyoming. Study of 21 stratigraphic sections reveals the presence of distinct facies belts for the middle and upper Elvinia Zone and lower Taenicephalus Zone. These are interpreted to reflect deposition in three distinct paleoenvironments: (1) oolitic shoal, (2) intrashef basin, and (3) inner platform-margin.

Following a major hiatus in sedimentation in Wyoming during much of Pterocephaliid biomere time, a healthy shallow-water carbonate factory was reestablished during Elvinia Zone time. This produced abundant deposits of oolitic and skeletal packstone in the shoal environment and an areally extensive turbidite-boundstone facies in the intrashef basin setting. Approximately coincident with the biomere boundary extinction event (Irvingella major time), ooid production was terminated and the turbidites were drowned. The correlation of this major change in the pattern of sedimentation allows for the recognition of parts of two depositional sequences in the study area. This drowning event (sequence boundary) may also be coeval with the accumulation of transported shallow-water carbonate deposits in an inner-platform-margin environment. During early Taenicephalus Zone time, sedimentation rates were apparently reduced and the entire study area is marked by thin concentrations of turbidite and brachiopod shells. The non-skeletal carbonate factory was never reestablished in the former shoal environment, where thick successions of homogeneous lime mud accumulated during the rest of Taenicephalus Zone time. Flat-pebble conglomerate and shale accumulated in the storm-influenced intrashef basin during this time. Sea-level change and changes in seawater chemistry likely both played roles in the drowning event across the Elvinia–Taenicephalus boundary.

A positive trend in carbon-isotope values across the Pterocephaliid–Ptychaspid biomere boundary is observed in the six sections analyzed. This is consistent with previous studies that indicated that the mass extinction occurs within a regional shift towards positive values, and this may reflect the development of low-oxygen conditions on the shelf. The global significance of these isotopic trends, changes in relative sea level, and the mass extinction has not been confirmed, although recent study of the extinction event at the base of the Pterocephaliid biomere has revealed a similar trend towards positive carbon-isotope values that is worldwide.

INTRODUCTION

The Upper Cambrian of North America is punctuated by three mass-extinction horizons, known as biomere boundaries, that are sharply defined in field localities across the continent (Lochman and Duncan 1944; Palmer 1965, 1984; Thomas 1993). These biomere boundaries represent quasi-isochronous surfaces (Palmer 1979) and are critical elements in the intra- and interbasinal correlation of Upper Cambrian depositional sequences or cycles (Lohmann 1976; Osleger and Read 1993). In addition, biomere boundaries have been recognized in the full range of depositional paleoenvironments and paleotectonic settings on the ancient continent of Laurentia—from the nearshore siliciclastics of the craton to the outer shelf and slope carbonates of the subsiding passive margin—and thus provide an exceptional opportunity to document basin evolution on a continental scale.

In previous studies, significant changes in the pattern of sedimentation across the base of the Pterocephaliid biomere (Palmer 1984; Osleger and Read 1993; Thomas 1993) and top of the Ptychaspid biomere (Miller 1992; Ludvigsen et al. 1988) have been recognized. These changes have been variously interpreted to reflect local or regional changes in climate, sea level, subsidence, and sediment supply. This study focuses on the stratigraphic record across the Pterocephaliid–Ptychaspid biomere boundary (Steptoean–Sunwaptan boundary of Ludvigsen and Westrop 1985), which previous investigations have recognized as a relatively stable time in basin evolution across Laurentia. For example, Brady and Rowell (1976) demonstrated a remarkable uniformity of sediment type over an area of more than 40,000 km² during this time interval in the eastern Great Basin. Osleger and Read (1993) recognized a similarly stable pattern of sedimentation in sections in the Appalachians, Texas, and Oklahoma and, accordingly, their inferred Late Cambrian sea-level curve shows no significant sea-level events across the Pterocephaliid–Ptychaspid biomere boundary. However, recent study of well exposed strata in northwestern Wyoming by Saltzman et al. (1995) revealed significant facies changes across the Pterocephaliid–Ptychaspid biomere boundary. This evidence would appear to have implications for the ongoing discussion of Late Cambrian eustasy (Osleger 1995), although clearly the extent to which the relative-sea-level history in Wyoming should be considered representative of the continent as a whole remains problematic.

Here, expanding on the initial results of Saltzman et al. (1995), a more detailed analysis of the evolution of the Cordilleran passive margin across the Pterocephaliid–Ptychaspid biomere boundary (during mid-late Elvinia and early Taenicephalus Zone times) in northwestern Wyoming is presented. The primary objectives are to: (1) develop a paleogeographic framework for strata deposited in the northwestern Wyoming basin; (2) integrate new biostratigraphic and chemostratigraphic data to better constrain the timing of significant stratatal surfaces; and (3) interpret these stratatal surfaces within a sequence stratigraphic framework.

TECTONIC AND STRATIGRAPHIC FRAMEWORK

The Cordilleran passive margin on the western edge of the North American craton formed in response to the breakup of a Late Proterozoic–Early Cambrian supercontinent (Bond et al. 1984). In Middle to Late Cambrian time, the drift stage of passive-margin sedimentation was characterized by the development of an extensive carbonate platform (Armstrong 1968; Palmer 1971; Stewart and Poole 1974; Levy and Christie-Blick 1991). In the western United States, the carbonate platform extended approximately 300 km (palinspastic restoration by Levy and Christie-Blick 1989) from near the cratonal–miogeoclinal hinge zone in Utah and Wyoming (“Wa- satch line” of Kay 1951) to the distally steepened ramp edge in Nevada (Osleger and Read 1993).

The Sauk transgression reached Wyoming and Montana by Middle Cambrian Glossopleura Zone time (Sloss 1963; Lochman-Balk 1971) in response to a combination of eustatic rise and flexural bending of the cratonal
edge (Bond et al. 1989). This initial transgression was marked by deposition of coarse clastic sediments of the Flathead Sandstone on Precambrian basement rocks. Subsequent deposition in the inner flexural wedge (summarized in Lochman-Balk 1971) is best characterized as three large-scale siliciclastic–carbonate couplets, or grand cycles (Aitken 1966). The third of these grand cycles is the focus of this study.

The basal carbonate–siliciclastic transition of this third grand cycle corresponds to the abrupt contact between the massive cliff-forming Pilgrim Limestone and the overlying recessive Dry Creek Shale. The lowest carbonate beds within the Dry Creek Shale carry an Elvinia Zone fauna (Shaw and Deland 1955; Grant 1965). The Pterocephaliid–Ptychaspid biomere boundary, corresponding to the transition between the Elvinia and Taenicephalus Zones, lies within the overlying carbonate succession of the Snowy Range and Open Door Formations (Shaw and Deland 1955; Grant 1965). The Pterocephaliid–Ptychaspid biomere boundary is here recognized as an interval of extinction (upper Deadwood Formation; Lochman-Balk 1971). The Open Door Formation passes seaward into a succession of deep subtidal wackestones (Sneakover Limestone and equivalents; Brady and Rowell 1976). All successions contain variable proportions of siliciclastic sediment; these proportions generally increase northeastward toward the cratonal interior.

The Snowy Range Formation typically is poorly exposed on the flanks of the Beartooth uplift and Horseshoe Hills, where it overlies the cliff-forming Pilgrim Limestone of Crepicephalus and early Aphelaspis Zone age. Four units in the lower half of the formation were studied: (1) a basal shaly interval 5–15 m thick, which is typically covered (Dry Creek Shale Member of the Beartooth Formation; Neal 1955; Grant and Hu 1960). The Snowy Range Formation passes seaward into a succession of shale and minor flat-pebble conglomerate (upper Deadwood Formation; Lochman-Balk 1971). The Open Door Formation passes seaward into a succession of deep subtidal wackestones (Sneakover Limestone and equivalents; Brady and Rowell 1976). All successions contain variable proportions of siliciclastic sediment; these proportions generally increase northeastward toward the cratonal interior.

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY

Twenty-three measured sections of mixed carbonate–siliciclastic strata were logged in northwestern Wyoming and south-central Montana (Fig. 1). The sections can be divided into two distinct mappable units (Fig. 2): (1) a succession of thrombolite boundstone, shale, and flat-pebble conglomerate referred to as the Snowy Range Formation (Dorf and Lochman 1940; Grant 1965), and (2) a succession of calcarenite, shale, and lime mudstone referred to as the Open Door Formation (Miller 1936; Shaw and Deland 1955; Lochman and Hu 1960). The Snowy Range Formation passes seaward into a succession of shale and minor flat-pebble conglomerate (upper Deadwood Formation; Lochman-Balk 1971). The Open Door Formation passes seaward into a succession of deep subtidal wackestones (Sneakover Limestone and equivalents; Brady and Rowell 1976). All successions contain variable proportions of siliciclastic sediment; these proportions generally increase northeastward toward the cratonal interior.

In marked contrast to outcrops of the Snowy Range Formation, the Open Door Formation is spectacularly exposed in the Gallatin, Teton, Gros Ventre, and Wind River Ranges. It consists of four prominent units overlying the cliff-forming Pilgrim (Du Noir) Limestone: (1) a lower unit of...
interbedded shale and limestone 1–5 m thick (lower? to mid-Elvinia Zone), which locally contains beds of glauconitic siltstone, (2) a middle unit of oolitic grainstones and argillaceous wackestones 2–8 m thick, which contains an upper Elvinia Zone fauna, (3) a thin, 1–2 m thick unit dominated by trilobite and brachiopod shell concentrations, and (4) an upper unit (above the highest shell concentration) of burrowed lime mudstones and wackestones 5–15 m thick, which contains a lower and middle Taenicephalus Zone fauna (Miller 1936; Shaw and DeLand 1955; Lochman and Hu 1960). A single section in the Gallatin Range at Three Rivers Peak, although sharing some rock types with the Open Door/Snowy Range successions (Ruppel 1972), differs markedly from the general pattern in that it contains beds of quartz sandstone at the base, an ~1-m-thick carbonate breccia in the middle, and abundant chert in the upper wackestone unit. The top of the Open Door Formation is unconformably overlain by the mid-Ordovician Bighorn Dolomite. Thus, in contrast to the Snowy Range Formation, which is conformable through the uppermost Cambrian Idahoaia and Saukia Zones, the location of the top of the Taenicephalus Zone is poorly known within the Open Door Formation and may be cut out by pre–Middle Ordovician erosion.

LITHOFACIES ASSOCIATIONS

All deposits of the Snowy Range and Open Door Formations reflect subtidal settings (Fig. 3). Lime mudstone is prevalent and may reflect epipelicular precipitation by algae (Robbins and Blackwelder 1992) or disaggregation of skeletal and nonskeletal carbonate grains. Ooids are present only in the middle part of the Open Door Formation. The distribution and composition of the skeletal component is highly variable, as is the degree of preservation of the fragments. Siliciclastic components also show considerable local variability and are generally claystone with minor amounts of sandstone and siltstone. Dolomite is present in minor amounts (<5%) but reaches higher percentages in the oolitic facies in certain sections. Nearly all deposits are burrowed, and primary sedimentary structures are rarely preserved.

Lithofacies occur in three distinct associations described below. The two dominant facies associations (accounting for 20 of the 21 detailed measured sections) are interpreted as reflecting deposition in intrashelf basin (Snowy Range Formation) and shoal (Open Door Formation) environments. Five lithofacies are recognized in the intrashelf basin environment: thrombolitic boundstone, bioturbated wackestone, flat-pebble conglomerate, shale, and coquinite. Four lithofacies are recognized in the shoal environment: ooid packstone/grainstone, lime mudstone, nodular argillaceous limestone, and coquinite. The third facies association is recognized only in the Gallatin Range, near the western edge of the study area. It contains beds of quartz sandstone, carbonate breccia, and cherty wackestone that likely accumulated in an inner-platform-margin setting.

Intrashelf-Basin Facies Association (Snowy Range Formation)

Thrombolitic Boundstone.—This facies occurs in laterally continuous outcrops as a distinct bioherm that is relatively resistant and forms a marker bed within the Upper Cambrian outcrops in northwestern Wyoming (Fig. 4A). The bioherm consists of vertically stacked columns that locally occur as discrete structures but ultimately become laterally linked (SH-V/LLH-C; Lochman-Balk 1971). Columns are 0.5–2 m high and 10–50 cm in diameter. At some localities, the entire bioherm is one continuous unit of columns, whereas at others three or four units are separated by thin (10 cm) beds of shale and flat-pebble conglomerate. The maximum total thickness of the bioherm, including interbedded shale and conglomerate, is 9.5 m. It is thickest along a NW–SE axis from the Clark Fork area (4.5 m at Clark Fork) to the Horseshoe Hills (6.4 m at Nixon Gulch) and thins away from this axis to 1.2 m at Mill Creek and 1.1 m at Wind River Canyon (Fig. 1). The bioherm tapers to zero thickness in the Big Horn Mountains.

The columns are composed of micritic limestone that exhibits a clotted fabric and colonies of Epiphyton (Collenia magna of Grant 1965). Notable features within the columns are clayey partings or microstylolites. Traces of quartz silt and carbonate sand are seen in thin section. Irregular pockets of fill (<10 cm across) between columns locally contain skeletal debris, shale, carbonate conglomerate, or material eroded from the bioherm itself. Where the base of the bioherm could be located, it is in contact with a bed of flat-pebble conglomerate. This is commonly the lowest stratigraphic occurrence of flat-pebble conglomerate. At some localities, the upper surface of the bioherm is in contact with a bed of flat-pebble conglomerate, whereas at others, it is overlain by shale or wackestone.

The thrombolitic unit formed a linear organosedimentary structure approximately 50 km wide and 250 km along strike, approximately parallel to the shoreline, in the Late Cambrian. The geographic extent of this buildup is exceptional for Upper Cambrian strata, where bioherms are typically discontinuous on scales of meters (Lohmann 1976) to kilometers (Westrop 1989).

Flat-Pebble Conglomerate.—This facies occurs above and at the base of the thrombolitic boundstone facies (Fig. 5A). Flat-pebble conglomerate beds alternate with beds of clay shale containing abundant pods of lime mudstone. Beds of flat-pebble conglomerate typically range in thickness
from 5 to 20 cm, although beds up to 50 cm thick are present. The thicker beds are most commonly amalgamations of thinner beds separated by stylolitic clay seams. Bases of beds are commonly erosional and in places exhibit load casts, whereas the tops are planar to irregular. Many flat-pebble conglomerate beds can be traced laterally for tens of meters along outcrops.

Pebbles are rounded to subangular, with diameters in the range of 1–10 cm. They are composed of faintly laminated lime mud/pellet aggregates. The pebbles are intraclasts derived from the pods of lime mudstone that occur in the interbedded shale (Grant 1965). The interstitial material between the pebbles is composed of varying proportions of skeletal debris and terrigenous silt. Pebbles are typically coated with glauconite or limonite and some exhibit reticulate pitting, which may be indicative of submarine corrosion (Brett et al. 1983). The tops of many pebble pavements are hardgrounds, some of which exhibit evidence of encrusting echinoid communities. Pebbles in the thinner beds tend to be flat-lying or show weak bi-modal imbrication. Some of the thicker beds are true edgewise conglomerates, showing random clast orientation.

Shale.—This facies occurs above and below the thrombolite facies but is generally not well exposed. The Dry Creek Shale below the thrombolite boundstone is fissile and somewhat micaceous. Minor amounts of angular quartz and feldspar occur in the lower half of the Dry Creek in a few localities. In contrast, shale (mudstone) above the thrombolite is fissile to blocky and contains only traces of angular quartz or feldspar. These shales contain abundant small pods (~3 cm by 5 cm) of micritic limestone with high percentages of clay. Fossils are very rare in the shales but locally occur along the contacts with overlying or underlying limestone beds.

Bioturbated Wackestone.—This facies occurs as a minor component of the shale and flat-pebble conglomerate facies association. Beds are 10–50 cm thick. They are typically massive deposits of burrowed lime mudstone with minor admixtures of trilobite and brachiopod debris and pellets. Reworked glauconite and rhombs of dolomite also occur.

Coquinite.—This facies occurs above the thrombolite facies as shell concentrations of distinct taxonomic assemblages of trilobites or brachiopods (Fig. 4B). The Irvingella major coquina contains trilobite taxa of the
Irvingella major "faunizone", which Wilson and Frederickson (1950) recognized as a remarkably consistent assemblage across the North American craton. In this study, evidence for the Irvingella major coquina was found in the Snowy Range Formation only in float blocks at the Clark Fork locality. Grant (1965) recognized the Irvingella major coquina at 3 of his 27 Snowy Range Formation collecting localities, where it is ∼ 1–20 cm thick and apparently forms a lens. In contrast to the sporadic I. major coquina, brachiopod shell concentrations occur as continuous beds (10–20 cm thick) in all sections studied. The articulate brachiopods Eoorthis and/or Billingsella dominate these beds, with inarticulate brachiopods and trilobites forming minor components of the assemblage (see Li and Droser 1997 for a thorough discussion of Cambrian shell beds). These brachiopod shell beds are also remarkably widespread across the North American craton (Lochman-Balk 1971). Both the Eoorthis–Billingsella and Irvingella major coquinas locally contain zones rich in glauconite and locally exhibit evidence of reworking and shell dissolution.

Interpretation.—The shale and flat-pebble conglomerate facies of the Snowy Range Formation reflect deposition in a quiet shelf lagoon (intra-shelf basin) that was episodically affected by storms. Low background sedimentation rates between storm events are indicated by synsedimentary lithification of clasts, hardgrounds on the pebble pavements, and common glauconitic coatings on pebbles (Brett et al. 1983). Similar storm-influenced facies described from the Upper Cambrian include those in Aitken (1978) and Westrop (1989) for the southern Canadian Rockies and Markello and Read (1982) for the southern Appalachians. These intrashelf basins can be areally quite extensive. In the case of the Wyoming study area, the intrashelf basin represented by the Snowy Range Formation stretches hundreds of kilometers northward and eastward into the Dakotas (Deadwood Formation; Lochman-Balk 1971; Sepkoski 1982).

The thrombolitic boundstone facies was built on a hard pebbly substrate and contains interbeds of the shale and flat-pebble conglomerate facies association. The boundstone shows no evidence of exposure and must have accumulated in a subtidal setting during mid-late Elvinia Zone time. Water depths were shallow enough for sufficient light penetration, and favorable water-mass conditions (i.e., temperature, oxygen concentration, and nutrient content) characterized this particular region bordering the intrashelf basin. The approximately 50-km-wide and 250-km-long buildup did not form a significant barrier to circulation in the intrashelf basin and is best described as a mound-and-channel belt (see also descriptions of buildups in Cowan and James 1993). This view is further supported by the predominantly micritic composition of the boundstone. Much of the wave energy generated in the epicontinental sea was likely dissipated by the shoal-water facies described below.
the total thickness of cycles is 0.5±3 m. Abraded skeletal grains, detrital ooid packstone. The oolitic packstone intervals are 0.1±1.5 m thick, and 4D). Cycles ideally consist of (1) a basal unit of nodular argillaceous lime-
stone facies. Arrow points to “sculpted” pebble. Granite Falls section.


Shoal-Water Facies Association (Open Door Formation)
Ooid Packstone/Grainstone.—This facies occurs in noncyclic shoal de-
positions (Fig. 4C) and at the tops of upward-coarsening subtidal cycles (Fig. 4D). Cycles ideally consist of (1) a basal unit of nodular argillaceous lime-
stone, (2) a middle micritic, algal, or pebbly limestone unit, capped by (3) ooid packstone. The oolitic packstone intervals are 0.1–1.5 m thick, and the total thickness of cycles is 0.5–3 m. Abraded skeletal grains, detrital glauconite, and flasers of lime mudstone/dolomite are interstratified with the ooids (Fig. 5B). Some of the ooid packstone beds occur as planar sand sheets, whereas others are lenticular on scales of tens to hundreds of meters laterally. These packstone deposits show no internal structure. In contrast, the ooid grainstone deposits, which occur only in the Gros Ventre and Teton Ranges, exhibit low-angle cross-bedding (Fig. 4C). These grainstone units are severely recrystallized and/or dolomitized.

Lime Mudstone.—This facies occurs as thin beds within the ooid pack-
stone cycles and also as a thick (> 5 m), homogeneous unit overlying the cyclic interval. Within the cycles, the mudstone beds are 0.1–0.5 m thick. Contacts between the lime mudstone and ooid packstone are gradational to sharp and locally developed as hardgrounds exhibiting truncated intraclasts and pitted and etched surfaces (Fig. 4D). Glaucosil grains are concentrated along the hardgrounds but also occur dispersed throughout the lime mud. Minor quartz silt is also present above hardgrounds. Discrete vertical bur-
rows occur rarely within the lime mudstone of the cyclic succession. In contrast, the homogeneous lime mudstone unit overlying the cyclic suc-
cession is highly burrowed and thin skeletal pavements locally occur on bedding surfaces.

The bases of some lime mudstone beds consist of pebbly lags. The peb-
bles are derived from the lime mudstone facies and are typically rounded to subrounded and average 1 cm in diameter. In the cyclic succession, beds containing smaller (tens of millimeters in diameter), rounded to angular pebbles are common. These smaller pebbles are buff in color and exhibit a wide range of shapes that stand out sharply against the dark lime mud matrix (Fig. 5C).

Nodular Argillaceous Limestone.—This facies occurs mainly in beds at the bases of the upward-coarsening cycles and range from 0.1 to 1.5 m in thickness. It consists of lenses or nodules of micrite, similar in com-
position to the associated lime mudstone, but within a distinctive greenish blue clayey shale matrix. The contacts with the underlying oolitic facies within the cycles are typically sharp. Contacts with the overlying lime mudstone are gradational.

Coquinites.—This facies occurs as echinoderm, trilobite, or brachiopod shell concentrations above the oolitic facies. The Irvingella major and Eoorthis-Billingsella coquinites are very similar to those found in the Snowy Range Formation (Fig. 4B), although they are more variable in thickness and taxonomic composition. Echinoderms are generally in greater abundance in the Open Door Formation, including a distinctive 20-cm-thick echinoderm coquina just below the Billingsella coquina at Warm Springs Creek.

Interpretation.—The oolitic facies reflects the presence of a high-energy shoal environment during mid-late Elvinia Zone time that formed a fringing bank (Read 1985) that is transitional between a vast open-marine ramp and the intrashelf basin (Fig. 6). The migrating shoal environment is, however, recorded only by beds at the top of the oolitic interval that exhibit low-angle cross-bedding (Death Canyon section, Teton Range; Fig. 4C). Most ooids occur in sheets or lenses of packstone that show no internal structure, and the intimate associations with abraded skeletal debris, hardgrounds, intraclasts (Fig. 5B), and detrital glauconite are interpreted to indicate an allochthonous origin for these oolitic deposits. Furthermore, the ooid packstone units are associated with lime mudstone (which is not pelleted in thin section) and argillaceous limestone deposits, which reflect low-energy con-
ditions (Fig. 4D). These deposits are interpreted to reflect quiet-water de-
position in topographic depressions on the flanks of the migrating shoal. The migrating shoal is represented in the study area by cross-beded oolitic units in the Gros Ventre and Teton Ranges (Fig. 4C). Storm surges or currents likely moved ooids from the shoal into topographic depressions (Aitken 1978; Westrop 1989), and then the predominant low-energy con-
ditions allowed for the stabilization of these deposits, perhaps by algal films. A lack of agitation is also indicated by the development of glauco-
nitic horizons within the oolitic packstone deposits.

The overlying coquinites and burrowed lime mudstones deposited in the
area during *Taenicephalus* Zone time are interpreted to reflect quiet-water conditions that were affected only by severe storms. These deposits indicate a well-circulated marine setting below storm wave base rather than a restricted lagoon environment, an interpretation based on the lack of evaporites, the highly burrowed fabric, and the presence of abundant trilobite fragments, which occur as pavements on bedding surfaces. The disappearance of glauconitic horizons that were well developed in the underlying *Elvinia* Zone deposits also indicates a well-circulated water mass. The complete lack of shale or ooids in these homogeneous lime mud deposits signals the shutting off of the nonskeletal carbonate factory as well as the source of fine siliciclastics.

The lithofacies associations of the Snowy Range and Open Door Formations suggest similar water depths, typically between fair-weather and storm wave base, which resulted in alternating high- and low-energy conditions. However, the lack of shared lithofacies between the formations is striking—there are no ooids in Snowy Range strata and no thrombolite boundstone in Open Door strata. It is difficult to evaluate the nature of the separation of these environments (i.e., antecedent topography) because the Absaroka volcanics cover much of the transitional area (Fig. 1) between the Beartooth Range (Snowy Range depocenter) and the Wind River, Gros Ventre, and Teton Ranges (Open Door depocenter). It is also possible that the two environments are simple lateral equivalents, rather than occurring as parallel facies belts (Fig. 6).

**Inner Platform-Margin Facies Association**

**Quartz Sandstone.**—This facies occurs in beds 10–20 cm thick at the base of the Three Rivers Peak section in the Gallatin Range. The beds consist of very fine-grained quartz (~90%) and minor feldspar and mica. The sand grains are arranged in weakly planar laminae and are well sorted and subrounded.

**Carbonate Breccia.**—This facies (Fig. 7) occurs as a unit approximately 1 m thick consisting of randomly oriented pebbles and slabs of lime mudstone and wackestone. The pebbles are rounded to subangular and "float" in a matrix of recrystallized lime mudstone to calcisiltite. Locally, the pebbles are coated or penetrated by reddish oxide rims and occur in irregular clusters, lenses, or beds that project (weather out) from the more matrix-dominated regions of the bed. Pebby units are in sharp contact, both laterally and vertically, with dark-colored, highly recrystallized lime mudstone units (Fig. 7B) that appear as well-defined tabular bodies against the lighter-colored matrix.

Fractures within the breccia penetrate downward a few tens of centimeters (Fig. 7C) and are filled from above with pebbles that appear to have flowed into and completely filled the fractures (Fig. 7D). Evidence of solution collapse, such as calcite cement, angular fragments, or fitted, irregular surfaces, are not associated with the fractures. Rather, the boundaries of the fractures are linear and well defined by dissolution surfaces.

The fractured breccia lacks true bedding and forms the lowest exposed unit above a mostly grass-covered shaly slope (Fig. 7A). The contact with the shale is planar to undulating and locally coincides with rotated blocks of club-shaped, thrombolitic limestone (10–20 cm thick). An irregular stylolitic surface separates the top of the breccia from a poorly sorted oolitic wackestone unit. The breccia and the overlying and underlying normally bedded successions can be traced for tens of meters along the outcrop, and the breccia appears as a lens-like stratigraphic unit on this scale.

**Cherty Wackestone.**—This facies occurs above the breccia as dark, burrow-mottled beds averaging 50 cm thick. Ooids and quartz silt laminae occur sporadically in the lowest beds. Phosphatic brachiopods and trilobites are present as thin (~1 cm) pavements on bedding surfaces. Chert occurs as nodules or thin stringers in the wackestone. The ledges of wackestone alternate with poorly exposed calcareous shale and siltstone.

**Interpretation.**—The inner platform margin facies is transitional between the shallow-water carbonate factory and the deep-marine ramp. This
is indicated by the presence of blocks of boundstone within the basal part of the carbonate breccia that appear to have been transported from the thrombolitic boundstone facies to the east (perhaps where this facies was laterally equivalent to the oolitic shoals). The breccia is thus interpreted as a debris sheet (Cook 1983) that accumulated during an episode of mass movement down a low-angle slope. The lateral extent of the strata and the rounded nature of the pebbles rules out a postdepositional, fault-related origin. Furthermore, the depositional environments of the lithofacies in contact with the breccia are subtidal, and thus an in situ origin for brecciation related to dissolution of unstable mineral assemblages (gypsum) is unlikely. Meteoric cements that would be associated with karstification are not present. The reddish coatings on the pebbles are randomly distributed among non-coated pebbles and are most likely a later diagenetic feature (however, a brief episode of exposure and coeval oxidation of glauconitic coatings cannot be ruled out). The fracturing of the lithified breccia (Fig. 7C, D) may have occurred in a later episode of slumping or could reflect seismic activity.

In addition to the debris sheet (breccia), evidence for a deeper-water, transitional setting is suggested by the overlying beds of dark, cherty wackestone that contain abundant, well-preserved inarticulate brachiopod valves. These beds are similar to deposits of this age that characterize the deep-ramp setting of the more rapidly subsiding regions of the passive margin in the central Great Basin (Brady and Rowell 1976). The beds of quartz sandstone near the base of the section are more difficult to explain in this context and may have been derived as wind-blown deposits from a nearby basement high, possibly in the Madison Range to the east, where Cambrian sections are relatively sandy, abbreviated, or missing. More generally, the location of these inner-platform-margin facies approximately coincides with the ancient Cordilleran hinge zone (Kay 1951; Lochman-Balk 1955), which may partly explain their complex, transitional nature (Wehr and Glover 1985; Cooper and Edwards 1991).

CARBON-ISOTOPE STRATIGRAPHY

Age control and the correlation of sections are critical to discussion of the evolution of Upper Cambrian strata in northwestern Wyoming. The focus of this section is on changes in the δ¹³C values of limestones across the Pterocephaliid±Ptychaspid biomere boundary, which potentially provide a means of correlation, independent of biostratigraphy (e.g., Pelechaty et al. 1996; Saltzman et al. 1998). Carbonate samples from six of the twenty-three measured sections of Upper Cambrian strata in northwestern Wyoming were analyzed for stable-isotope ratios. Homogeneous micrite identified in thin section was microsampled from polished slabs by using a microscope-mounted drill assembly. Care was taken to sample micrite with no visible cements or skeletal grains, although ~ 15% of the samples contain sparry calcite or skeletal material. Sample preparation followed
standard procedures presented in Saltzman et al. (1995). Analytical error for δ13C and δ18O was ± 0.04‰.

The δ13C stratigraphic profiles for the Granite Falls and Warm Springs Creek sections were presented in Saltzman et al. (1995), along with profiles from two sections in the Great Basin. The Great Basin profiles reveal a positive excursion in δ13C (up to 2‰) across the Pterocephaliid–Ptychaspis biomere boundary, while the Granite Falls and Warm Springs Creek sections show a more gradual positive trend. The lack of a more pronounced excursion presumably reflects the much thinner, less complete stratigraphic record in the Wyoming region, although the possibility that this reflects variations in local water masses cannot be ruled out (Patzkowsky et al. 1997). Figure 8 shows that the δ13C stratigraphic profiles for the six sections (Fig. 1) that record δ13C stratigraphic profiles record an overall positive change upsection, with the highest δ13C values recorded in the Taenicephalus Zone, although again, no major excursions are seen. The δ18O values do not show any stratigraphic trends or covariation with δ13C ratios (Fig. 9) that would be expected if diagenetic alteration capable of massive resetting of carbon-isotope ratios had occurred.

The Pterocephaliid–Ptychaspis biomere boundary is well constrained by fossil collections of *Irvingella* and *Billingsella* coquina at Warm Springs Creek (Table 1) and collections containing *Comanchia* and *Billingsella* coquina at Granite Falls. In these two sections, the biomere boundary coincides approximately with the rise in δ13C ratios above 0.5‰. The biomere boundary is not as well constrained in the other four sections, and thus an attempt was made to locate the transition to δ13C ratios above 0.5‰ in the carbonates to approximate the biomere boundary.

At Death Canyon in the Teton Range, the *Eoorthis* shell bed is preserved and places an upper limit on the biomere boundary. *Elvinia* Zone trilobites reported lower down in the section by Miller (1936) are not abundant. The δ13C transition to ratios above 0.5‰ lies 2 m below the *Eoorthis* bed and is the preferred position for the biomere boundary. At Green River Lakes in the Wind River Range, the δ13C change again lies 2 m below the *Eoorthis* bed and 9 m above reported *Elvinia* Zone trilobites (*Camaraspis*; Loichman and Hu 1960). At Three Rivers Peak in the Gallatin Range, the transition to δ13C ratios above 0.5‰ lies ~50 cm above *Elvinia* Zone trilobites that occur at the top of the breccia (*Pterocephalia*; A.R. Palmer, oral communication). This change in δ13C ratios is ~2 m below known *Taenicephalus* Zone trilobites and brachiopods (A.R. Palmer, in Ruppel 1972).

At Clark Fork in the Beartooth Range, the highest δ13C ratios occur in the *Taenicephalus* Zone. However, the signal in this section is very noisy and does not allow for much confidence in placement of the biomere boundary. The biomere boundary is instead placed a meter above the thrombolite reef facies, where it occurs in most other Snowy Range sections (Grant 1965). It may be that the noisy δ13C signal at Clark Fork is a result of relatively high volumes of diagenetic fluids moving through the dominantly shaly deposits. Reworking of limestone clasts during storms may also be a problem, and similar scatter has been found in preliminary analysis of other Snowy Range sections.

The positive change in carbon isotopes spanning the Pterocephaliid–Ptychaspis biomere boundary in northwestern Wyoming is in agreement with previous data for this interval (Saltzman et al. 1995). This study thus provides additional chemostratigraphic evidence consistent with the notion that paleoceanographic changes occurred across the biomere boundary. These changes may be related to increasing rates of burial of organic matter and the development of low-oxygen conditions on the shelf. It is interesting that the base of the Pterocephaliid biomere also corresponds to the onset of a large positive trend in δ13C of nearly 4‰ that is worldwide (Saltzman et al. 1998). Although this event is larger in magnitude and duration than the Pterocephaliid–Ptychaspis event, the correspondence between rising δ13C values and extinction is striking. The evidence for a drowning event across the Pterocephaliid–Ptychaspis boundary in northwestern Wyoming discussed below is consistent with the notion that sea-level changes may have been the trigger for the apparently synchronous perturbations of the carbon cycle and the marine biota.
DEPOSITIONAL HISTORY

In this section, the stratigraphic analyses presented above are integrated by discussing the depositional history of the Elvinia and Taenicephalus Zones in a sequence stratigraphic framework. Component systems tracts from two depositional sequences are recognized (Fig. 10). Many authors have discussed the limitations in applying sequence stratigraphic concepts to outcrop studies of Late Cambrian mixed carbonate–siliciclastic deposi- tional systems (Chow and James 1989; Mount et al. 1991; Cowan and James 1993; Montañez and Osleger 1993; Srinivasan and Walker 1993; Osleger and Moñañez 1996). The goal of this discussion is simply to identify stratal surfaces that may have chronostratigraphic significance and to tie them to important biostratigraphic and chemostatigraphic horizons.

Stage 1: Shallowing and Exposure of the Platform.—Digitate stromatolites at the top of the upward-shallowing Pilgrim Limestone succession (Lochman-Balk 1971; Thomas 1993) are truncated and overlain by beds of quartz siltstone or shale in the basal Dry Creek Shale (Shaw and DeLund 1955; Lochman-Balk 1971). This abrupt carbonate–siliciclastic transition marks the base of Sequence 1 of Figure 10. The influx of siliciclastics follows a major sedimentary hiatus on the craton during the Pterocephaliid Zone, the base of Sequence 1 of Figure 10. The influx of siliciclastics of quartz siltstone or shale in the basal Dry Creek Shale (Shaw and DeLund 1955; Montañez and Osleger 1993; Srinivasan and Walker 1993; Osleger and Moñañez 1996). The goal of this discussion is simply to identify stratal surfaces that may have chronostratigraphic significance and to tie them to important biostratigraphic and chemostatigraphic horizons.

Stage 2: Flooding and Establishment of the Carbonate Factory.—Following the Sauk II–Sauk III lowstand, sea level began to rise and gradually flooded the previously exposed siliciclastic source areas of the Dry Creek Shale. As siliciclastic input decreased, the carbonate factory was reestablished (“start-up” phase) in northwestern Wyoming during mid-late Elvinia Zone time. The approximately 2 m of wackestone and pebbly dolomite, upper Aphelaspis Zone time, is not recorded in this region.
limestone that overlie the Dry Creek Shale signal this reestablishment and constitute the upper part of the transgressive systems tract (as distinct from the overlying aggradational to progradational thrombolitic and oolitic facies of the highstand systems tract). Hardground formation within the thin transgressive package indicates reduced (or stalled) accumulation rates (Kidwell 1997), which may have resulted from a combination of low initial carbonate productivity and gradual reduction of siliciclastic input from the craton. This mid-late Elvinia Zone interval of sea-level rise and gradual flooding of cratonal interiors is recognized across Laurentia (James et al. 1989; Bond et al. 1989; Westrop 1992; Osleger and Read 1993; Osleger 1995; Runkel et al. 1998) and appears to have influenced carbon cycling in the Cambrian seas. The general trend of rising $\delta^{13}C$ values during mid-late Elvinia Zone time (Fig. 8) may in part have tracked the increased area of shallow, productive seas created during transgression.

**Stage 3: Aggradation and Progradation.**—The shoal-water and thrombolite facies that overlie the maximum flooding interval (boundary between TST and HST in Figure 10) are interpreted to reflect aggradation to progradation of a healthy shallow-water carbonate platform, perhaps during the slowing of the rate of sea-level rise. The shoal-water and thrombolite facies are thus interpreted to constitute the highstand systems tract. Ooid packstone beds within subtidal cycles (parasequences) increase in thickness up section and are capped ultimately by beds of low-angle cross-beded grainstone. This upward shoaling pattern provides evidence for the progradational geometry of these deposits, although current biostratigraphic and outcrop limitations preclude discussion of the relative synchronity and direction (seaward or lagoonward) in which these facies prograded. Clastic deposition was gradually terminated during deposition of the HST and the trend toward heavier $\delta^{13}C$ values continued.

**Stage 4: Drowning of the Carbonate Factory.**—The shallow-water carbonate platform established during mid-late Elvinia Zone time was drowned again near the Pterocephalid–Ptychaspid biomere boundary (Sequence 1–2 boundary in Figures 10 and 11) [a precise correlation with the extinction event is precluded in Wyoming by the limited distribution of the I. major faunizone]. This drowning unconformity (sequence boundary of Schlager 1991) is not marked by intervening lowstand deposits or evidence of subaerial exposure, and is thus similar in some respects to the base of Sequence 1 (Pilgrim–Dry Creek surface) described above. However, unlike the base of Sequence 1, the drowning event at the top of Sequence 1 is not marked by an abrupt change from carbonates to siliciclastics and is not thought to represent a major hiatal surface. The existence of a minor hiatus is suggested by carbon-isotope evidence from the thicker, more complete sections in the Great Basin, where an isotopic excursion of nearly 2% recognized by Saltzman et al. (1995) is not recorded in Wyoming (Fig. 8). Thus, the possibility that a brief sea-level fall preceded drowning (resulting in erosion or nondeposition of sediments with the $\delta^{13}C$-enriched carbon-isotope values found in the Great Basin; Saltzman et al. 1998) cannot be ruled out at this time.

The termination of the thrombolite boundstone facies at the top of Sequence 1 in the intrashelf basin is striking, but it does not change the background pattern of sedimentation of shale, flat-pebble conglomerate, and bioturbated wackestone. In the shoal-water environment, the nonskeletal carbonate factory was terminated at the top of Sequence 1. These facies
changes may have resulted in part from a decrease in the saturation state of calcium carbonate in seawater related to inundation of the carbonate platform during sea-level rise (Cowan and James 1993). Deposition of the debris flow (breccia) in the Three Rivers Peak section may also record a highstand systems tract of Sequence 1. The reestablishment of the carbonate platform in the region occurred during the late part of the TST and maximum flooding of cratonal interiors. The overlying highstand systems tract consists of a thrombolite boundstone facies in the northern half of the study area (Snowy Range Formation) and an oolitic-shoal facies in the south (Open Door Formation). A major change in the pattern of carbonate sedimentation during I. major time (across the Elvinia-Taenicephalus boundary) corresponds to the termination of both widespread thrombolite growth and the production of ooids by the nonskeletal carbonate factory. This sequence-bounding surface might be related to changes in sea level and/or the chemistry of seawater. Thus, the Osleger and Read (1993) sea-level curve may need to be modified to reflect a sequence boundary at the transition between the Elvinia and Taenicephalus Zones. Apparently, previous studies of this interval in sections across Laurentia—including those with a greater history (thickness)—do not reflect a sea-level event because of the lack of facies belts sensitive to modest changes in sea level.

Carbon-isotope data from six sections in Wyoming reveal a change toward more positive values that spans the Pteroccephaliid-Ptychaspis biomere boundary. This study thus provides additional chemostратigraphic analyses that are consistent with the notion of a significant paleoceanographic event across the biomere boundary, perhaps triggered by a rise in sea level (Saltzman et al. 1995). Similar parallel changes in sea level, carbon isotopes, and faunal turnover have also recently been described from the late Middle Ordovician in relation to orogenic uplift (Patzkowsky et al. 1997). Further work aimed at sorting out the roles of tectonics, changes in seawater chemistry, and eustatic change in the relative sea level history observed in northwestern Wyoming will undoubtedly provide a better understanding of their connections with environmental change.

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