Depositional history of pre-Devonian strata and timing of Ross orogenic tectonism in the central Transantarctic Mountains, Antarctica

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ABSTRACT

A combination of field mapping, detailed sedimentology, carbon isotope chemostratigraphy, and new paleontological finds provides a significantly improved understanding of the depositional and tectonic history of uppermost Neoproterozoic and lower Paleozoic strata of the central Transantarctic Mountains. On the basis of these data, we suggest revision of the existing stratigraphy, including introduction of new formations, as follows. The oldest rocks appear to record late Neoproterozoic deposition across a narrow marine margin underlain by Precambrian basement. Siliciclastic deposits of the Neoproterozoic Beardmore Group—here restricted to the Cobham Formation and those rocks of the Goldie Formation that contain no detrital components younger than ca. 600 Ma—occupied an inboard zone to the west. They consist of shallow-marine deposits of an uncertain tectonic setting, although it was likely a rift to passive margin. Most rocks previously mapped as Goldie Formation are in fact Cambrian in age or younger, and we reassign them to the Starshot Formation of the Byrd Group; this change reduces the exposed area of the Goldie Formation to a small fraction of its previous extent. The basal unit of the Byrd Group—the predominantly carbonate ramp deposits of the Shackleton Limestone—rest with presumed unconformity on the restricted Goldie Formation. Paleontological data and carbon isotope stratigraphy indicate that the Lower Cambrian Shackleton Limestone ranges from lower Atdabanian through upper Botomian.

This study presents the first description of a depositional contact between the Shackleton Limestone and overlying clastic units of the upper Byrd Group. This carbonate-to-clastic transition is of critical importance because it records a profound shift in the tectonic and depositional history of the region, namely from relatively passive sedimentation to active uplift and erosion associated with the Ross orogeny. The uppermost Shackleton Limestone is capped by a set of archaeocyathan bioherms with up to 40 m of relief above the seafloor. A widespread phosphatic crust on the bioherms records the onset of orogenesis and drowning of the carbonate ramp. A newly defined transitional unit, the Holyoake Formation, rests above this surface. It consists of black shale followed by mixed nodular carbonate and shale that fill in between, and just barely above, the tallest of the bioherms. This formation grades upward into trilobite- and hyolithid-bearing calcareous siltstone of the Starshot Formation and alluvial-fan deposits of the Douglas Conglomerate. Trilobite fauna from the lowermost siltstone deposits of the Starshot Formation date the onset of this transition as being late Botomian.

The abrupt transition from the Shackleton Limestone to a large-scale, upward-coarsening siliciclastic succession records deepening of the outer platform and then deposition of an eastward-prograding molassic wedge. The various formations of the upper Byrd Group show general stratigraphic and age equivalence, such that coarse-grained alluvial-fan deposits of the Douglas Conglomerate are proximal equivalents of the marginal-marine to shelf deposits of the Starshot Formation. Paleocurrents and facies distributions from these units indicate consistent west (or southwest) to east (or northeast) transport of sediment. Although the exact structural geometry is unknown, development of imbricate thrust sheets in the west likely caused depression of the inner margin and rapid drowning of the Shackleton Formation carbonate ramp. This tectonic activity also caused uplift of the inboard units and their underlying basement, unroofing, and widespread deposition of a thick, coarse clastic wedge. Continued deformation in the Early Ordo-
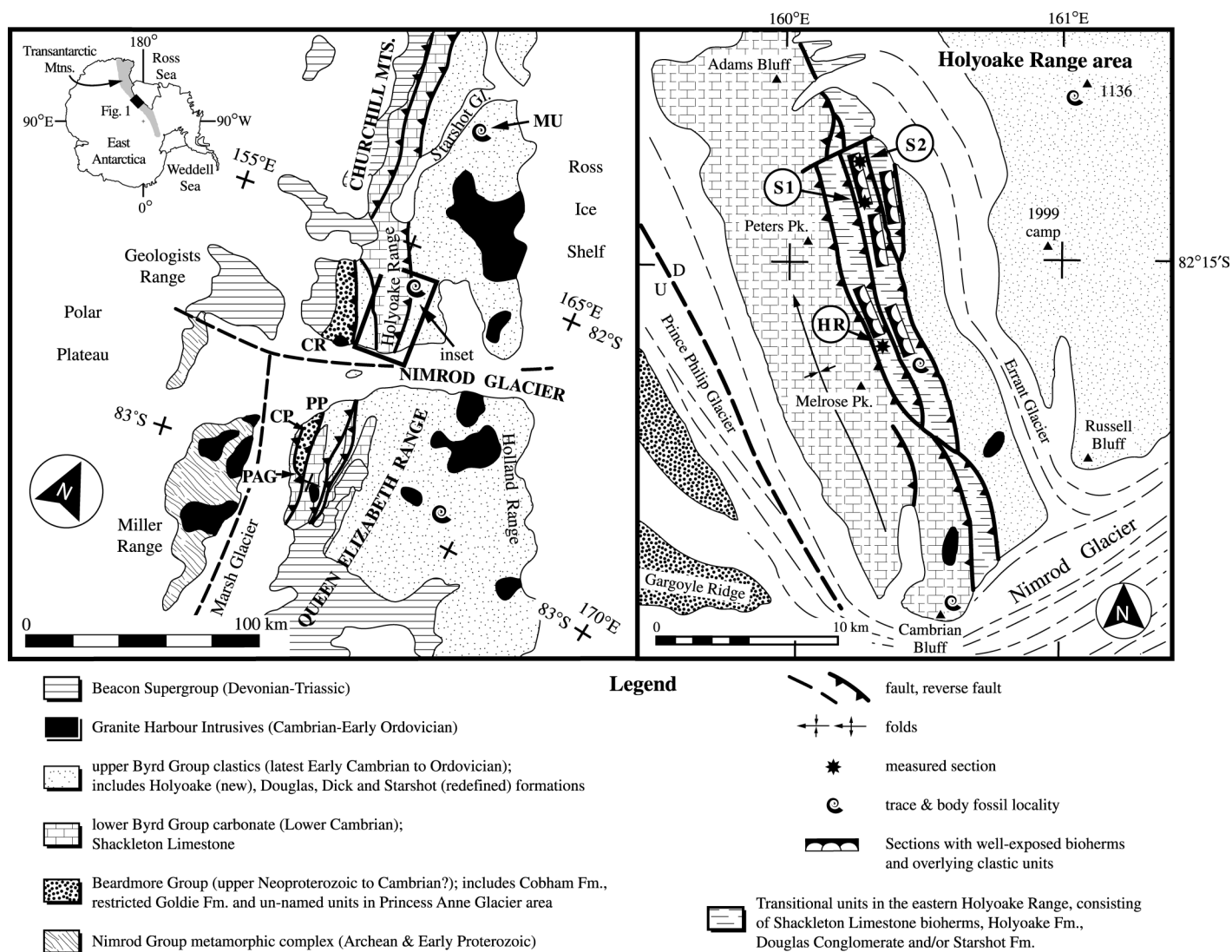
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Figure 1. Geologic map of central Transantarctic Mountains with inset (upper left) of Antarctica showing location of field area. CP—Cotton Plateau, CR—Cobham Range, MU—Mount Ubique, PAG—Princess Anne Glacier, and PP—Panorama Point. Close-up of Holyoake Range on right shows basic geologic units. Details of thrust slices of transitional units cannot be shown at this scale, but rectangular blocks indicate locations where upper Shackleton through Douglas transitions occur, each with well-developed bioherms that all dip vertically and face west. S1 (section 1), S2 (section 2), and HR (High Ridge) are described in text.

vician (younger than 480 Ma) in turn affected these synorogenic deposits, causing folding and thrust repetition of all pre-Devonian units.

Keywords: Archaeocyatha, Cambrian, Neoproterozoic, reef, Ross orogeny, Transantarctic Mountains.

INTRODUCTION

Stratigraphic relationships of pre-Devonian formations in the central Transantarctic Mountains (Fig. 1) have been the subject of much conjecture and debate (Fig. 2), in large part because of a lack of biostratigraphic constraints, detailed sedimentology, and physical stratigraphy of these units. The traditional view of stratigraphic relationships in this region, as reviewed by Stump (1995) and discussed in more detail herein, is that basement rocks of the Nimrod Group (Archean and Early Proterozoic) are directly overlain by Neoproterozoic rocks of the Beardmore Group, although this relationship has not been observed. The latter consists of unfossiliferous metamorphosed siliciclastic and carbonate rocks of the Cobham Formation and overlying less metamorphosed siliciclastic rocks (mostly sandstone with minor shale) of the Goldie

Formation. Fossil-bearing strata of the Byrd Group are generally considered to have been deposited unconformably upon the Goldie Formation. The basal formation of the Byrd Group, the Lower Cambrian Shackleton Limestone, consists of dominantly carbonate rocks with a range of facies including archaeocyathan-bearing bioherms. Clasts of the Shackleton Limestone occur within overlying synorogenic conglomerate- and sandstone-rich units of the Douglas and Starshot Formations, respectively. The stratigraphic relationships and ages of these younger units were poorly defined; possible ages range from latest Early Cambrian to Devonian (Fig. 2).
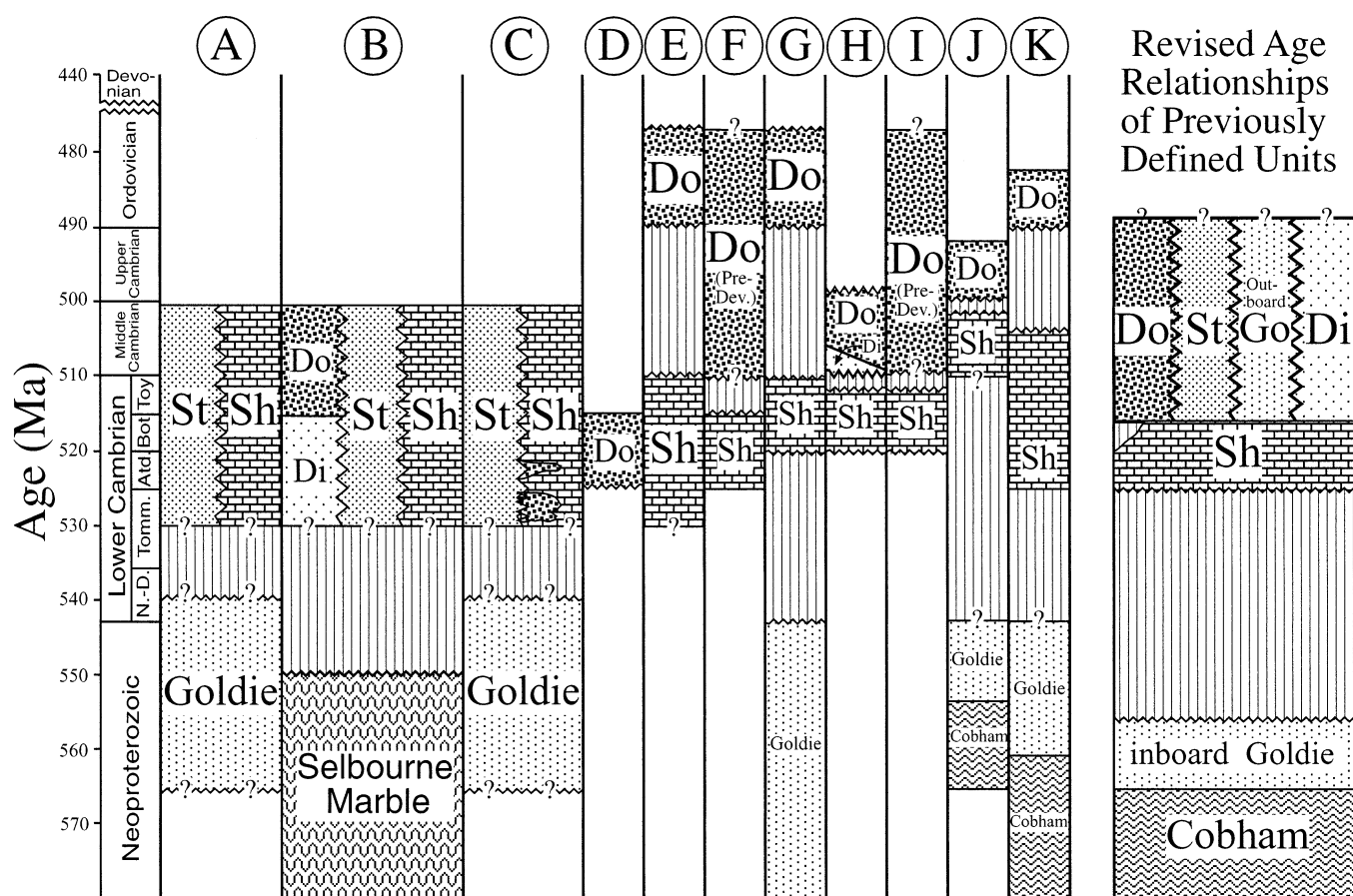
Figure 2. Interpretations of stratigraphic relationships of pre-Devonian Formations in the central Transantarctic Mountains. Right-hand column is revised view of these relationships. Formations: St—Starshot Formation, Sh—Shackleton Limestone, Do—Douglas Conglomerate, Di—Dick Formation, Go—Goldie Formation. Authors: A—Laird (1963), B—Skinner (1964), C—Laird et al. (1971), D—Laird (1981), E—Rowell et al. (1986), F—Rowell et al. (1988b), G—Rees et al. (1989), H—Rowell and Rees (1989), I—Rees and Rowell (1991), J—Goodge et al. (1993) and Goodge and Dallmeyer (1996), K—Goodge (1997). For reports simply describing "Lower Cambrian" strata, base of formation arbitrarily chosen as upper Tommotian.

New geochronologic data recast the age and stratigraphic relationships of these units. Detrital-zircon geochronology on sedimentary units of the region indicates that the bulk of outcrops mapped as Goldie Formation are in fact younger than the Shackleton Limestone (Goodge et al., 2000, 2002). Detrital zircons from these sandstone deposits are as young as ca. 520 Ma, making their depositional age younger than the Atdabanian Stage of the Lower Cambrian (Landing et al., 1998; Bowring and Erwin, 1998; Fig. 3). These young provenance ages come from strata in an areally widespread belt of "Goldie" rocks that occupied an outboard (eastern) position along the continental margin at the time of deposition. On the basis of detrital-zircon age spectra and field relationships, we interpret the "outboard" Goldie to be a temporal and stratigraphic equivalent of the younger siliciclastic deposits of the Starshot Formation of the Byrd Group, and these "outboard" Goldie strata are now assigned to that formation. A very narrow exposure of enigmatic, generally more metamorphosed, rocks in the vicinity of Cotton Plateau and the eastern Cobham Range were also originally mapped as Goldie Formation. These have older detrital-zircon signatures (1000 Ma or older) and probably represent bona fide Neoproterozoic deposits ("inboard assemblage" of Goodge et al., 2000, 2002).

This study provides new field, fossil, and isotopic data as the basis for additional stratigraphic revisions to the pre-Devonian sedimentary succession in the region. Sedimentological, physical stratigraphic, and chemostratigraphic data set limits on the nature and timing of a Cambrian transgression onto the Antarctic craton. We also describe the first documented depositional contact between the Shackleton Limestone and younger synorogenic deposits of the upper Byrd Group from the southern Holyoake Range. Sedimentological aspects of this transition from massive, archaeocyathan bioherms to overlying siliciclastic-dominated strata allow us to reconstruct the depositional response to Ross orogenesis. In addition, a new trilobite collection in strata above the bioherms provides constraints on the ages of several formations and the timing of tectonic events. These new findings, in conjunction with the geochronological data already mentioned, have profound implications for the depositional and tectonic history of the Transantarctic Mountains. The findings also help clarify a number of unresolved first-order stratigraphic issues of correlation, and the relative ages, of nearly all of the pre-Devonian units in this region.

FIELD AREA

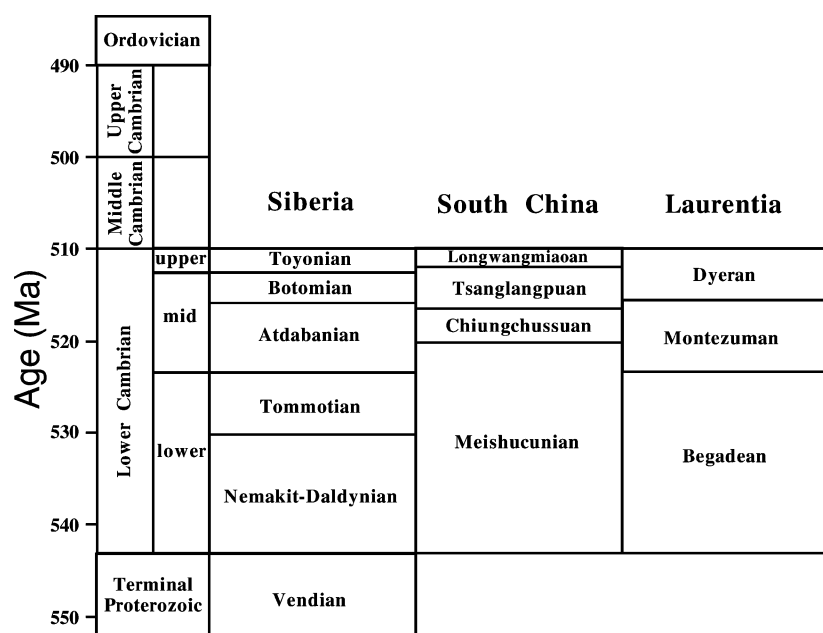Geologic mapping, sampling, and measurement of numerous stratigraphic sections were

Age (Ma)			Siberia	South China	Laurentia
490	Ordovician				
	Upper Cambrian				
500	Middle Cambrian				
510	Lower Cambrian	upper	Toyonian	Longwangmiaoan	Dyeran
		mid	Botomian	Tsanglangpuan	
			Atdabanian	Chiungchussuan	Montezuman
520					
530		lower	Tommotian	Meishucunian	Begadean
540			Nemakit-Daldynian		
550	Terminal Proterozoic		Vendian		

Figure 3. Comparison of Lower Cambrian stages of Siberia, South China, and Laurentia. Age constraints from Landing et al. (1998).

completed during the 1998–1999 and 1999–2000 austral summers. Field camps were established at Moody Nunatak and Cotton Plateau in the western Queen Elizabeth Range, in the Mount Ubique area of the Surveyors Range in the Churchill Mountains, and in the central Holyoake Range on Errant Glacier, providing access to many key localities. Detailed field observations were made possible over a geographically large area by a combination of overland travel and helicopter transit. Previous mapping and stratigraphic descriptions by several authors, cited subsequently, established the distribution of sedimentary units upon which our work is based.

BEARDMORE GROUP

The Beardmore Group, consisting of the Cobham and Goldie Formations, is a thick assemblage of graywacke, carbonate, and minor volcanic units that have been correlated with similar siliciclastic deposits along the length of the Transantarctic Mountains (Stump, 1982; Stump et al., 1986). The inferred stratigraphic position of the Beardmore Group between basement rocks and Lower Cambrian carbonate deposits led previous workers to suggest a late Precambrian age for the group (e.g., Laird et al., 1971), but the youngest constraint on it is unclear (i.e., whether it extends into the Cambrian). A Sm-Nd isochron age of 762 ± 24 Ma for basalt and gabbro (Borg et al., 1990), interpreted as interlayered with Goldie

rocks near Cotton Plateau (Fig. 1), has been widely cited as evidence for rifting at that time. A new U-Pb zircon age of 668 ± 1 Ma for the gabbroic phase (Goodge et al., 2002) suggests that these rocks and their associated sedimentary succession are significantly younger than previously thought.

The Goldie Formation is a thick succession dominated by fine-grained sandstone. On the basis of the presence of graded beds, flute marks, and abundant climbing ripples, it was considered a deep-sea turbidite succession (Laird et al., 1971; Stump, 1995), but these observations may have been made on "outboard" Goldie rocks now included in the younger Starshot Formation. There is little diagnostic paleoenvironmental evidence in much of the "inboard" Goldie, in part because of low-grade metamorphism. However, one locality ~3 km southeast of Panorama Point above Prince Edward Glacier contains a well-developed suite of sedimentary structures, including large-scale hummocky cross-stratification, which indicates that at least part of this formation was deposited above storm wave base. Coarse conglomerate beds were tentatively suggested to represent Neoproterozoic diamictites (Stump et al., 1988), but intense deformation of sections immediately below Panorama Point and at other localities, including isoclinal folding of the deposits and stretching of pebbles to aspect ratios up to 20:1, makes such a determination tenuous.

BYRD GROUP

Deposits of the Byrd Group are widespread in the central Transantarctic Mountains. The stratigraphic units include the Shackleton Limestone and overlying clastic deposits of the Starshot, Douglas, and Dick Formations.

Shackleton Limestone

The Lower Cambrian Shackleton Limestone crops out between Beardmore and Byrd Glaciers (Fig. 1); its most complete exposures are in the Holyoake Range (Rowell et al., 1988b; Rees et al., 1989; Rowell and Rees, 1989) where Laird (1963) designated a type locality. Archaeocyathan-bearing rocks from this formation were collected during Ernest Shackleton's ill-fated 1908–1909 expedition, but extensive study did not begin on this formation until much later (e.g., Laird et al., 1971; Rees et al., 1989). The archaeocyathans indicate a Botomian depositional age for at least part of the formation (Debrenne and Kruse, 1986). Trilobite fauna described by Rowell et al. (1988a) and Palmer and Rowell (1995) confirmed a Botomian age but suggested that some deposits are Atdabanian and that younger parts were possibly deposited in the Toyonian (Fig. 3). Nearly all of the trilobites have close affinities with Chinese forms of the Tsanglangpuan Stage, which largely overlaps the Botomian Stage. Carbonate units in the Queen Maud Mountains that are considered equivalents of the Shackleton Limestone also contain Botomian and possible Atdabanian trilobites and archaeocyathans (Rowell et al., 1997).

The Shackleton Limestone is a thick carbonate deposit with a lower unit of unfossiliferous interbedded quartzite and carbonate that is exposed below Panorama Point (Fig. 1) on the western slopes of Cotton Plateau, where it is in contact with the Goldie Formation. At other localities in the region, the contact was considered by Laird (1963) to be an unconformity, and at Cotton Plateau the same contact was interpreted similarly by Stump et al. (1991) and Stump (1995). However, this contact at the base of a syncline in Shackleton Limestone at Cotton Plateau shows structural evidence for fault displacement (Goodge et al., 1999), including angular discordance, a mineral-elongation lineation and stretched pebbles, asymmetric folds, asymmetric microstructures, and mineralization indicating a structural relationship. The contact is also faulted at other localities that have been rechecked (Rowell et al., 1986; Rees et al., 1989; Palmer and Rowell, 1995).

Therefore, although we disagree with previous workers about the present nature of the Cotton Plateau contact and although we have nowhere observed the stratigraphic base of the Shackleton Limestone, it is possible that it was originally deposited upon the currently underlying ''inboard'' Goldie units.

Detailed reconstruction of paleoenvironments and depositional history of the formation has been hampered by locally severe deformation and numerous fault contacts. Formation thickness is also difficult to assess, and estimates of 1000–2000 m (e.g., Burgess and Lammerink, 1979) are not well constrained. The depositional setting of the formation in our field area was previously interpreted as a simple ramp with intertidal facies passing laterally into a high-energy oolitic shoal complex that contains many individual archaeocyathan bioherms; no basinal facies were reported (Rees et al., 1989). Deposits of the oolitic shoal facies are laterally extensive and commonly include small algal-archaeocyathan bioherms (Rees et al., 1989). The bioherms were locally ecologically zoned as a function of depositional energy and in cases coalesced to form composite bioherm complexes up to 50 m thick. Burrow-mottled mudstone and skeletal or peloidal packstone were interpreted as shallow-water, low-energy deposits formed on either side of the oolite shoal complex (Rees et al., 1989). However, stratigraphic position, lithologic characteristics, and facies relationships suggest that these deposits are entirely deeper subtidal in origin and formed basinward (eastward) of the oolite shoal complex.

Douglas Conglomerate

Clastic rocks of the upper Byrd Group have been assigned to several formations including the Douglas Conglomerate, a succession of coarse clastic deposits exposed throughout the Churchill Mountains. Clasts derived from the Shackleton Limestone were recognized early on within the Douglas Conglomerate (Skinner, 1964, 1965); these clasts indicate uplift of the Shackleton and older units prior to deposition of these conglomeratic strata. Laird (1981) made the case for a sharp but conformable contact between the Shackleton and Douglas Formations. However, abundant faults in this region make this claim suspect, and locally at sections along the Starshot Glacier, the Douglas Conglomerate rests unconformably on folded Shackleton Limestone (Fig. 1; Rees et al., 1988; Rowell et al., 1988b). Its age has traditionally been bracketed as being younger than the Lower Cambrian Shackleton Limestone and older than the Devonian Beacon Su-

pergroup. The Douglas Conglomerate at its type locality is purportedly in depositional contact with fine-grained siliciclastic deposits of the Dick Formation (Skinner, 1964; Rees et al., 1988; see subsequent discussion).

Rees and Rowell (1991) provided the only detailed sedimentological study of the Douglas Conglomerate, but noted that no complete section of the formation exists and that faulting has made stratigraphic correlations difficult. Their facies analysis indicated deposition mainly in proximal to distal alluvial-fan environments, but also in marine and lacustrine settings. Clast compositions in the conglomeratic alluvial-fan deposits are dominated by quartzite and limestone, the latter eroded from the Shackleton Limestone. However, the following clast types were also noted: argillite, sandstone, metaquartzite, diorite, two-mica granite, gneissic granite, amygdaloidal spilite, rhyolite, basalt, chert, and dolomite (Skinner, 1964, 1965; Rees et al., 1987; Rees et al., 1988). Clasts of Shackleton Limestone in the Douglas Conglomerate are locally folded and contain some calcite veins, which indicate deformation prior to Douglas deposition (Rowell et al., 1986). The Douglas itself is folded and contains cleavage, indicating that Ross deformation outlasted deposition of the unit.

Starshot Formation

The Starshot Formation was defined by Laird (1963) to include shale, sandstone, and minor conglomerate exposed along the east side of the lower Starshot Glacier (Fig. 1); its type section is at Mount Ubique. Clasts in the conglomeratic units have compositions similar to those of the Douglas Conglomerate and also contain archaeocyathan fossils, thus indicating that the Starshot Formation was deposited contemporaneously with or after the Shackleton Limestone. No depositional contacts have ever been described between the Starshot Formation and any other formation in the region. Likewise, no fossils have been recovered from the Starshot Formation, other than those inherited in clasts, so determinations of its minimum depositional age are generally unconstrained.

The depositional environments of the Starshot Formation are poorly understood, and interpretations range from deep-water turbiditic to shallow-water shelf (Laird, 1963; Laird et al., 1971). Laird et al. (1971) recognized probable shoaling of the deposits toward the north where conglomeratic facies occur. Descriptions of fine-grained parts of the Douglas Conglomerate (Rees and Rowell, 1991) indicate that in lithology and sedimentological char-

acter, they are similar to the bulk of the Starshot Formation. Recent detailed analysis indicates that the Starshot Formation was deposited in environments that ranged from shoreline and possible fluvial facies to deeper shelf (Myrow et al., 2002). Trace fossils occur in several sections, but they are not common. The vast majority of the formation was deposited in inner-shelf to shoreline environments, in part as wave-modified turbidites (Myrow et al., 2002). Abundant paleocurrent data indicate transport toward the east and northeast. The presence of wave-modified turbidites and conglomeratic beds, the inferred facies relationships with the Douglas Conglomerate, and the paleocurrent patterns all suggest development of a high-relief proximal hinterland adjacent to a storm-influenced sea (Myrow et al., 2002).

Dick Formation

The Dick Formation consists primarily of fine siliciclastic deposits (shale through fine-grained sandstone) exposed in the northern Churchill Mountains (Fig. 1). Its type locality is ~15 km east of Mount Dick, south of the outlet of Byrd Glacier at Stations P and N (Skinner, 1964). At Station P, 150 m of this unit is exposed below the Douglas Conglomerate. It is presumably conformably overlain by Douglas Conglomerate at this locality (Skinner, 1964; Rees et al., 1988), but little is known about its age or relationship to other stratigraphic units. Burgess and Lammerink (1979) claimed that it conformably overlies the Shackleton Limestone (no details provided), but this contact was considered a fault by Rees et al. (1988) and we agree. The formation contains thin polymictic conglomerate beds with carbonate clasts whose character is similar to that of clasts of the Douglas and Starshot Formations (Laird, 1963).

Rees et al. (1988) considered the Dick Formation to have been deposited after the initial episode of early Paleozoic deformation, and they viewed it as genetically related to the Douglas Conglomerate. They interpreted the unit to represent fluvial to marginal-marine deposits that were essentially the distal equivalent of the Douglas alluvial-fan facies. Rare trace fossils suggest that the deposits are at least partly marine. We did not study the Dick Formation in detail, but our examination of the unit at section N of Skinner (1965) indicates that it contains most of the sedimentological aspects of the Starshot Formation. These include abundant fine-grained and very fine grained sandstone beds, graded bedding, thick climbing-

ripple divisions, and minor carbonate-clast conglomerate beds as thick as 1 m.

In summary, the age relationships of the various formations just described are poorly known because of discontinuous regional exposure, faulted formation contacts, lack of biostratigraphic control, and a near absence of geochronologic data. Many stratigraphic schemes have been suggested, as summarized in Figure 2. We present data from several critical sections that reveal important stratigraphic and age relationships among these formations. In combination with other stratigraphic and sedimentological information, as well as newly acquired U-Pb detrital-zircon dates, a coherent stratigraphic scheme is presented for all sub-Devonian sedimentary units in the central Transantarctic Mountains.

LOWER SHACKLETON LIMESTONE

The sandstone-rich lower member of the Shackleton Limestone is exposed at Cotton Plateau beneath Panorama Point, where it consists of up to 133 m of interbedded white- to cream-weathering, vitreous, quartz sandstone and brown-weathering, white, fine-grained dolomitic grainstone (Figs. 4, 5A). These beds, in both the upright and overturned limbs of a large syncline, are in fault contact with adjacent Goldie Formation (see previous description of the contact; Goodge et al., 1999). In the less deformed Shackleton section above the lower member, the percentage of carbonate increases up section, and the sandstone component eventually disappears as the section shifts into a thick, massive boundstone unit. A detailed section of the lower Shackleton Limestone at Cotton Plateau was measured on the lower limb of the large overturned syncline (Fig. 4). This section contains meter-scale, mixed siliciclastic-carbonate cycles (Fig. 5B) that begin with 7–181-cm-thick, fine- to medium-grained sandstone beds (average = 60 cm) having sharp, commonly erosional basal contacts with underlying carbonate beds (Fig. 5C). These sandstone beds locally show evidence for shallow-water deposition including paired clay drapes diagnostic of tidal deposition (Fig. 5D) and small-scale two-dimensional wave ripples. The sandstone beds have gradational upper contacts that include zones of carbonate-rich sandstone and sandy carbonate. Carbonate caps of these cycles consist of grainstone with abundant hummocky cross-stratification (Fig. 5E). These cycles appear (atypically) to record upward-deepening patterns. The upward decrease in sandstone from cycle to cycle and eventual stratigraphic transition from these cy-

clic deposits into a carbonate platform represents a long-term transgression and establishment of a long-standing carbonate ramp.

As already mentioned, we consider the contact between the Shackleton Limestone and Goldie Formation to be a fault at all locations we examined, including at Cambrian Bluff and Cotton Plateau, where Laird et al. (1971) and Stump (1995) described it as an unconformity. The faulted nature of the contact between the lower Shackleton Limestone and ''inboard'' Goldie Formation raises two possible scenarios. First, faults may have developed at or near a possible unconformity between these units, owing to mechanical weakness coupled with a sharp rheological contrast, leaving the basal Shackleton relatively intact. Second, some amount of the lower Shackleton may be missing owing to fault truncation. Therefore, we maintain that a depositional relationship between Shackleton and inboard Goldie has yet to be demonstrated and that an unknown amount of the lowermost Shackleton section may be missing regardless of its original relationship with the underlying Goldie rocks. Given the overall upward transition from sandstone to carbonate observed at Cotton Plateau, one would predict that any older parts of the Shackleton would be dominated by quartz sandstone. A possible exposure of such a section occurs at the outlet of the nearby Princess Anne Glacier. This outcrop contains a thick section of white quartz sandstone and minor shale that Laird et al. (1971) attributed to the Shackleton Limestone. The strata contain evidence for shelf and shoreline deposition (e.g., mud cracks, paired mudstone drapes, hummocky cross-stratification). An interval of coarse, fluvial cross-bedded conglomerate, pebbly sandstone, and sandstone occurs near the top of the unit. The entire section, several hundred meters thick, is unrepresented in the Cotton Plateau section just 17 km away. Unfortunately, neither outcrop is fossiliferous, although given the facies characteristics, it is unlikely that much time is represented by these strata. An archaeocyathan fossil was collected from the Cotton Plateau section at 127.95 m (Fig. 4), but this fossil is poorly preserved and only indicates that this part of the formation is Lower Cambrian.

To better determine the depositional age of the lower Shackleton Limestone, a series of samples for chemostratigraphic analysis was taken approximately every 50–100 cm from the top through the basal limb of the syncline below Cotton Plateau (Fig. 4). Carbon isotope ratios were determined by using a Finnegan MAT Delta plus mass spectrometer housed at the University of Tennessee, Knoxville. These

carbon isotope ratios were measured against established standards (PDB—Peedee belemnite) and plotted against stratigraphic height to produce a chemostratigraphic curve (Fig. 6). This curve shows a prominent, long-term, positive isotopic excursion that begins at $\delta^{13}C$ ratios of +2‰ and rises to nearly +5‰; the curve then gradually decreases to about +2‰. The prominent positive excursion occurs over nearly 75 m of shallow-marine carbonate-rich section and therefore appears to represent a geologically significant excursion of probable global significance. A number of systematic isotopic excursions are documented in Lower Cambrian rocks, particularly for the Siberian stages, and 10 (I–X) positive excursions are known (Kirschvink et al., 1991; Brasier et al., 1994; Brasier and Sukhov 1998). Six assemblages of trilobites were noted in the Shackleton Limestone by Palmer and Rowell (1995). The oldest assemblage contains forms, including redlichiids, that overlap in age with fallotaspidid trilobites, which occur at the base of the Atdabanian Stage (Fig. 3) and are associated with carbon isotope excursion IV (Brasier et al., 1994). The positive excursion at the base of the Shackleton Limestone is considered to be excursion IV on the basis of the following. First, the redlichiid trilobites were likely recovered from very low in the formation, presumably not much above the mixed siliciclastic-carbonate deposits of the lower Shackleton. Second, the upper half of the underlying Tommotian is globally characterized by a long-standing negative excursion. Finally, no faunas of clearly Tommotian age have been recovered from the Shackleton Limestone. Thus, identification of the basal Shackleton Limestone excursion as excursion IV dates the timing of marine transgression onto the Antarctic craton in this region as being early Atdabanian.

SHACKLETON–UPPER BYRD GROUP TRANSITION

Exposures of the Shackleton and Douglas Formations along the east side of the Holyoake Range were examined by Rees and Rowell (1991). They mapped several faults in the region and concluded that no contacts between these formations were depositional. Our study, aided by helicopter support, revealed that there is large-scale repetition of section in thrust sheets carried by faults that trend subparallel to the axis of the range. A depositional transition between the Shackleton Limestone and overlying siliciclastic units of the upper Byrd Group was discovered and mapped in successive thrust sheets on numer-

Shackleton Syncline Section

Legend

~~~ Hummocky Cross-Stratification

⌒⌐ Intraclasts

⊤ Mud cracks

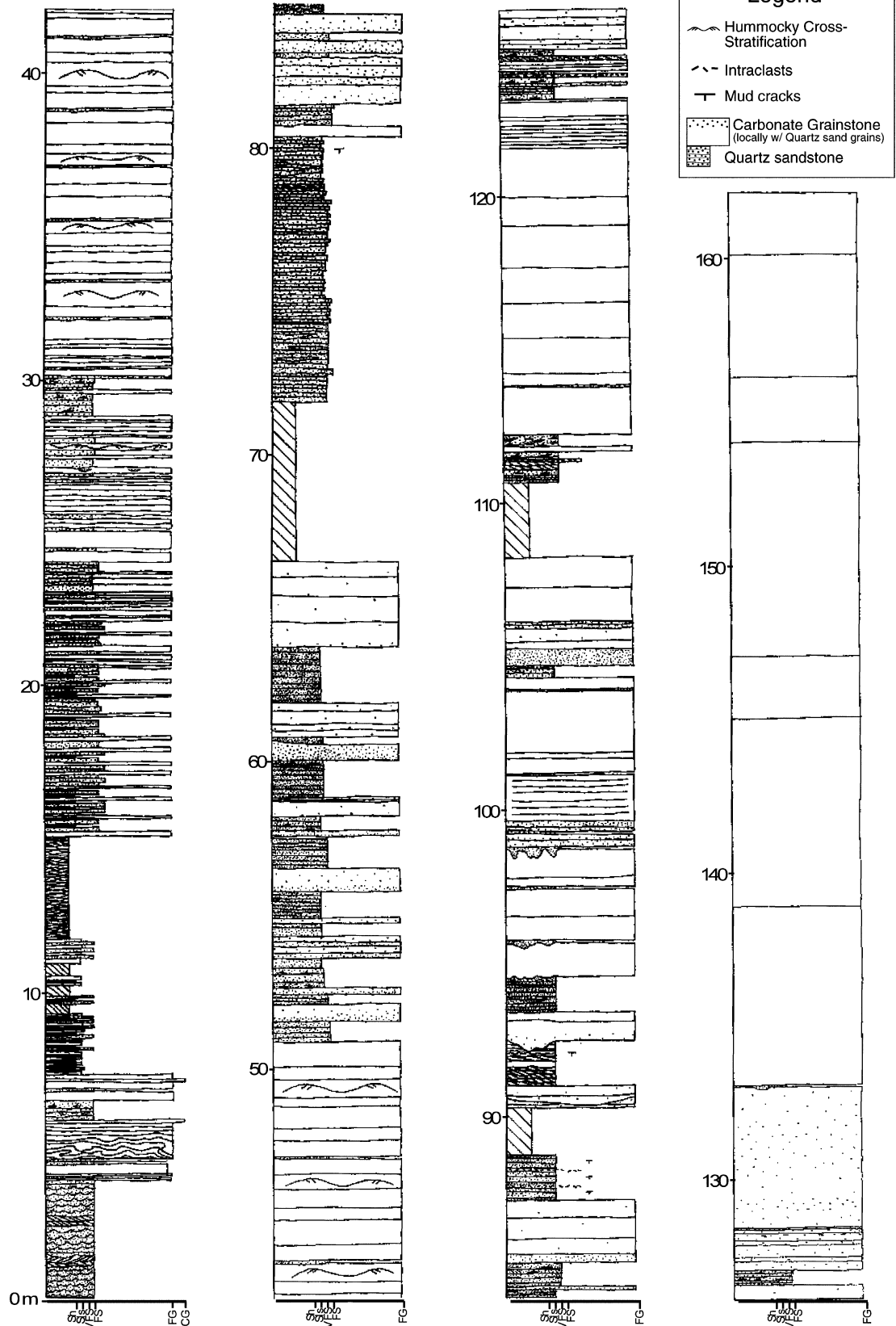Carbonate Grainstone (locally w/ Quartz sand grains)

Quartz sandstone

Figure 4. Bed-by-bed measured section of mixed siliciclastic-carbonate deposits of the lowermost Shackleton Limestone, Cotton Plateau. Abbreviations: Sh—shale; Sls—siltstone; VFS—very fine grained sandstone; FS—fine-grained sandstone. Carbonate: FG—fine-grained grainstone, CG—coarse-grained grainstone.
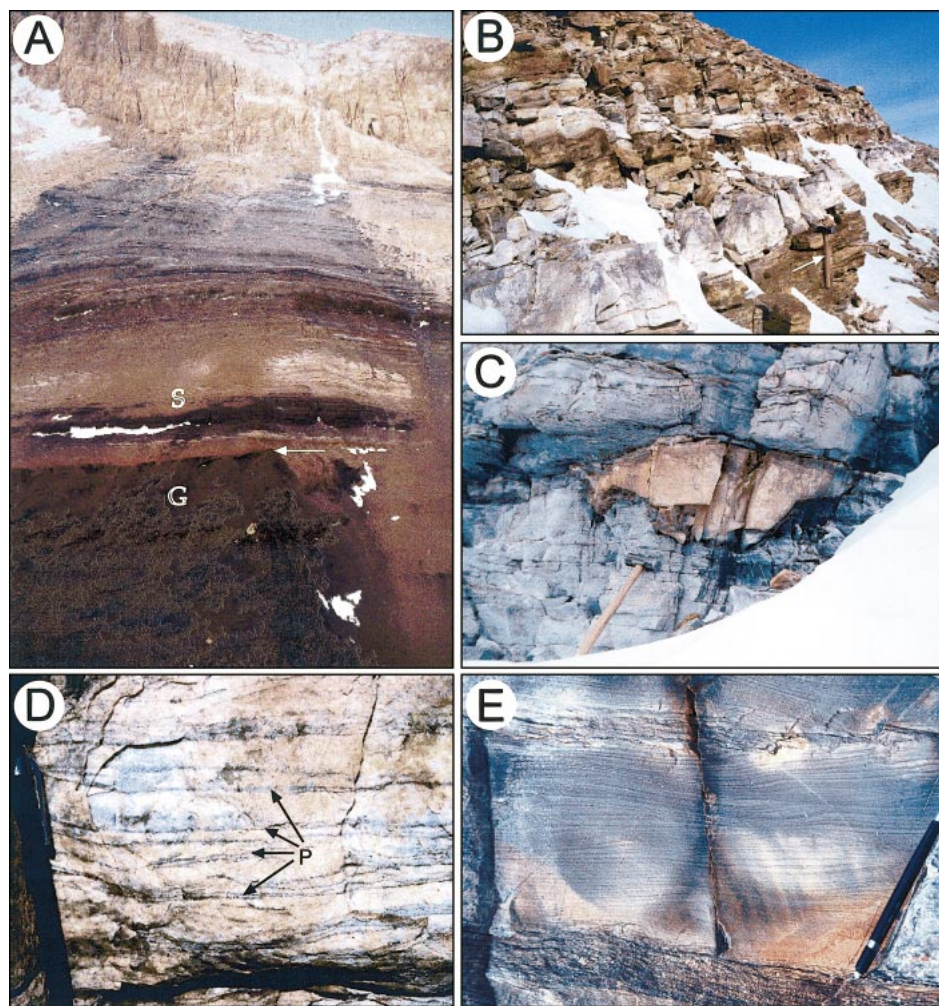
Figure 6. Chemostratigraphic curve of $\delta^{13}C$ vs. stratigraphic height for lower Shackleton Limestone at Shackleton syncline section (Fig. 4).

Figure 5. Shackleton syncline section, Cotton Plateau (Fig. 4). (A) Lower Goldie Formation (G) is overlain by interbedded quartz sandstone and dolostone of lower Shackleton Limestone (S). Arrow points to contact. Upper part of section consists of massive archaeocyathid-bearing dolostone cliffs. (B) Interbedded quartz sandstone and dolostone beds in 60–63 m zone (distances refer to measured section; Fig. 4) of lower Shackleton Limestone. Hammer for scale (arrow). (C) Deep-channel fill of quartz sandstone in cross section at 98.81 m. Hammer for scale. (D) Paired clay drapes (arrows) in quartz sandstone bed at 74.15 m. Pencil (at left) is 14 cm long. (E) Hummocky cross-stratification in carbonate grainstone at 27 m. Pencil is 14 cm long.

ous spurs along an ~30 km north-south section of the Holyoake Range.

In the Holyoake Range, the upper Shackleton Limestone consists of black to dark gray, burrow-mottled and skeletal wackestone that grades upward into large, massive, isolated archaeocyathan bioherms (Fig. 7). Both the bioherms and the interbioherm strata on which the bioherms rest are capped by an extensive, multigeneration, black, phosphatic surface, which is overlain by a unit of shale and nodular lime mudstone in a brown, calcareous shale matrix. This facies grades upward into fossiliferous siltstone and then into a complex

assemblage of carbonate-clast conglomerate and fine-grained sandstone. The transition-zone strata were carefully measured in several localities, and two measured sections are presented in Figures 8 and 9. A description and interpretation of these facies follows in the stratigraphic order in which they occur.

Nodular Carbonate (Shackleton Limestone)

The nodular carbonate facies makes up a tabular unit that is overlain by isolated archaeocyathan bioherms (described subse-
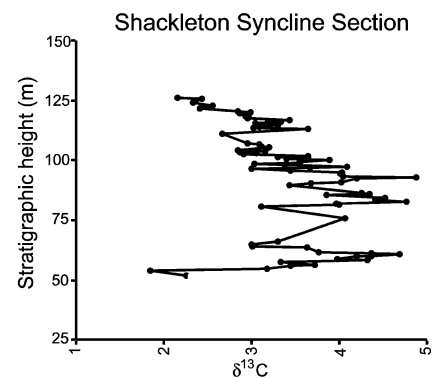
quently). It consists of gray to dark gray, nodular beds of skeletal-pelletal wackestone and lime mudstone. This facies contains rare, thin, laterally discontinuous skeletal packstone and grainstone beds with bioclasts of trilobites, archaeocyathans, and eocrinoids. A few thin intervals (commonly <2 m thick) of dark, thin-bedded lime mudstone occur within this nodular carbonate.

The dark color and nodular bedding suggest that this facies developed in a deep subtidal setting colonized by burrowing organisms. The lack of interbedded archaeocyathan bioherms, ooids, and light-gray lime mudstone (cf. Rees et al., 1989) suggests that it formed basinward of the ooid shoal facies. The thin interbeds of skeletal grainstone/packstone are interpreted to be storm deposits that infrequently affected this deeper subtidal environment.

Archaeocyathan Bioherms (Shackleton Limestone)

Isolated bioherms at the top of the Shackleton Limestone consist of massive-bedded, light gray to white algal boundstone with archaeocyathans. The bioherms range in thickness from 0 to >38 m (Figs. 7, 10A). Most of the bioherms are simple structures with only one episode of growth. However, at a few locations the bioherms are composite structures with at least two separate growth phases, each capped by an extensive phosphate-encrusted surface (Fig. 11). The bioherms form isolated bodies, and between them the phosphate-encrusted surface occurs directly on top of the nodular carbonate unit described earlier.

The upper surface of the lower growth phase of the composite bioherm at section 1

Figure 7. Large light-colored archaeocyathan reefs (white arrows), tens of meters thick, just south of section 1 of the Holyoake Range. Bedding is vertical, and stratigraphic up is to right; dashed line represents bedding orientation. Reefs are surrounded by shale, nodular carbonate and shale, and fine-grained sandstone of Starshot and Holyoake (new name) Formations.

has local small, irregular depressions with up to 1.5 m of relief (Fig. 12A). These were filled by wackestone, shale, and carbonate-clast conglomerate with boundstone fragments. A second growth phase of microbial boundstone is capped with another thick composite phosphatic surface (Fig. 12B). This composite surface is composed of multiple irregular layers of phosphatic micrite cement incorporating a thin (10–15 cm thick) skeletal packstone/grainstone with a fine-grained fauna of hyolithid, eocrinoid, and trilobite bioclasts as well as quartz silt and phosphatic fragments (Fig. 12C). Many of the phosphatic surfaces are bored with 1–5-mm-diameter, nearly circular borings.

Development of ≥40 m of relief and a lack of interbedded oolite shoals suggest that these algal bioherms developed in a deep subtidal setting basinward of the oolitic shoal complex. The phosphatic surfaces formed during early cementation of the archaeocyathan bioherms and represent a complex submarine hardground. Most of the bioherms grew during a single phase of growth, although some of them show at least two phases of growth separated by hardground development and onlap of fine carbonate and shale. The carbonate conglomerate on the side of the bioherm formed by submarine erosion of the growing bioherm, possibly by storms. The irregular upper surfaces of the bioherms were likely produced by nonuniform growth of the bioherm or possibly by submarine erosion. Both the

cessation of bioherm growth and the succeeding development of the hardground were responses to rapid drowning of the carbonate platform. The hardground records sediment starvation during flooding and replacement of carbonate by phosphate that was likely introduced by upwelling. The borings indicate the hardgrounds were hard, cemented surfaces and that they acted as surfaces of sediment bypass (Bathurst, 1975; Mullins et al., 1980). The multiple-stage hardground on top of the archaeocyathan bioherms indicates terminal drowning of the downslope bioherms, which we interpret as likely caused by tectonic subsidence related to initiation of orogenesis in this region.

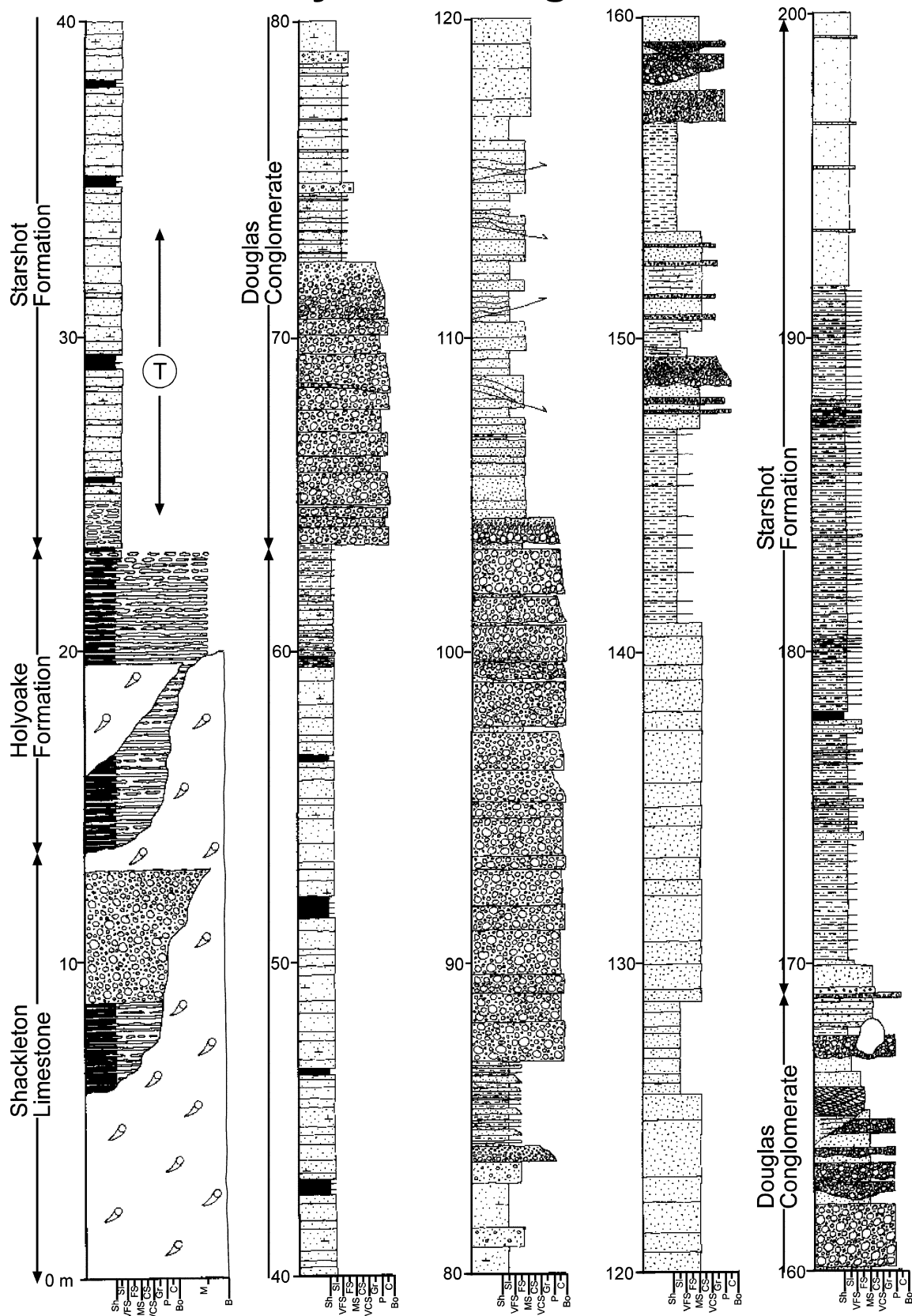## Shale and Nodular Carbonate/Shale (New Holyoake Formation)

Directly overlying the hardground surface, both between the bioherms and on their flanks, is dark gray to brown, calcareous fissile shale. This passes upward into nodular carbonate in a similar-colored calcareous shale that onlaps the uppermost parts of the taller bioherms (Fig. 10B). The nodular carbonate consists of lime mudstone nodules 2–10 cm thick and up to 40 cm wide in a brown shaly matrix. The nodules include rare trilobite and very rare hyolithid bioclasts and up to 1% disseminated pyrite. Near the top of some bioherms, in shale just below the first occurrence of the nodular carbonate facies, are rare, small (1–2 m wide, <1 m high) archaeocyathan bioherms that are capped by a thick phosphatic surface. The archaeocyathans in these bioherms are large and locally replaced by phosphate (Fig. 12D).

This shale records deep-water deposition above the phosphate-encrusted bioherms in response to drowning of the carbonate platform. The nodular carbonate and shale also constitute a deep-water facies, and although the increase in carbonate content may indicate a change in climate (Weedon, 1986), it more likely represents a response to shoaling and a stratigraphic shift into overlying calcareous siltstone (described in the next section). Shale and nodular shale within pockets in the upper part of some bioherms (between lower and upper growth stages described earlier) and the
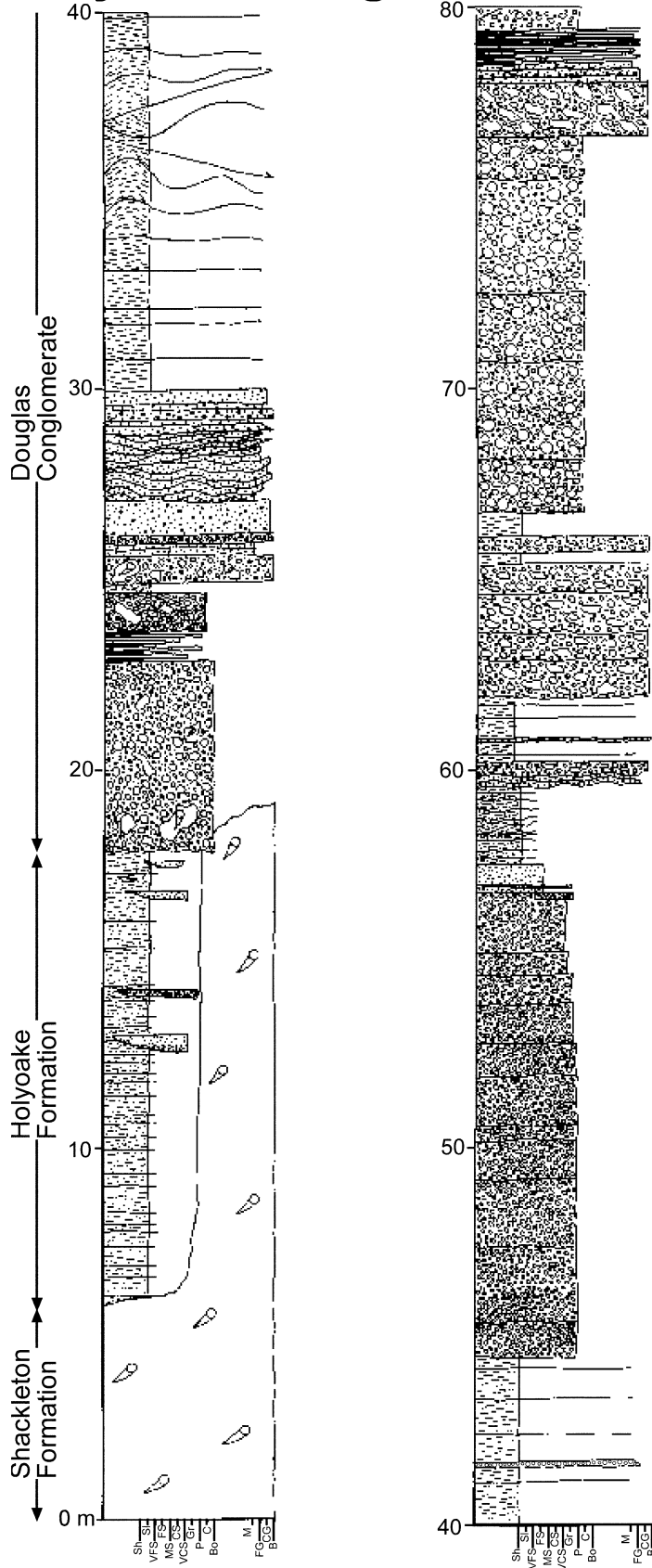
Figure 8. Stratigraphic section 1 from east side of Holyoake Range. Standard grain sizes from Sh (shale) to Bo (boulder). M—micrite, B—boundstone, T—trilobites. See legend.

Holyoake Range Section 1

Holyoake Range Section 2

Legend

small bioherm bodies within overlying shale and nodular carbonate facies reflect an interval of repeated drowning and attempted bioherm growth. Because of its distinction as a mappable unit, we define this facies of the transition zone as the "Holyoake Formation," as proposed later in this paper.

Calcareous Siltstone/Very Fine Sandstone (Starshot Formation)

This facies consists of brown-weathering, calcareous, siltstone to very fine grained sandstone. A wide range of carbonate content exists and locally exceeds 50%; thus the lithology is in cases sandy calcisiltite. The facies includes a few very widely spaced (several meters) decimeter-scale shaly intervals. The sandstone is blocky weathering and generally featureless in outcrop. This facies occurs above shale with nodular carbonate, as already described, and for less than 1 m of this facies transition, the siltstone contains thin (<4 cm) nodules of gray micritic limestone (Fig. 13B). The lowermost 5–10 m of section above this nodular-facies transition contains abundant fossils of trilobites and hyolithids. This facies contains locally developed black phosphatic material that occurs both as disseminated grains in siltstone beds up to 1.1 m thick and as nodular beds up to 8 cm thick. The latter contain abundant <2-mm-diameter borings (Fig. 13C).

Beds of pebbly fine- to medium-grained sandstone from 30 to 70 cm thick occur within this facies. These are nongraded and matrix

Figure 9. Stratigraphic section 2 from east side of Holyoake Range. For explanation, see Figure 8.
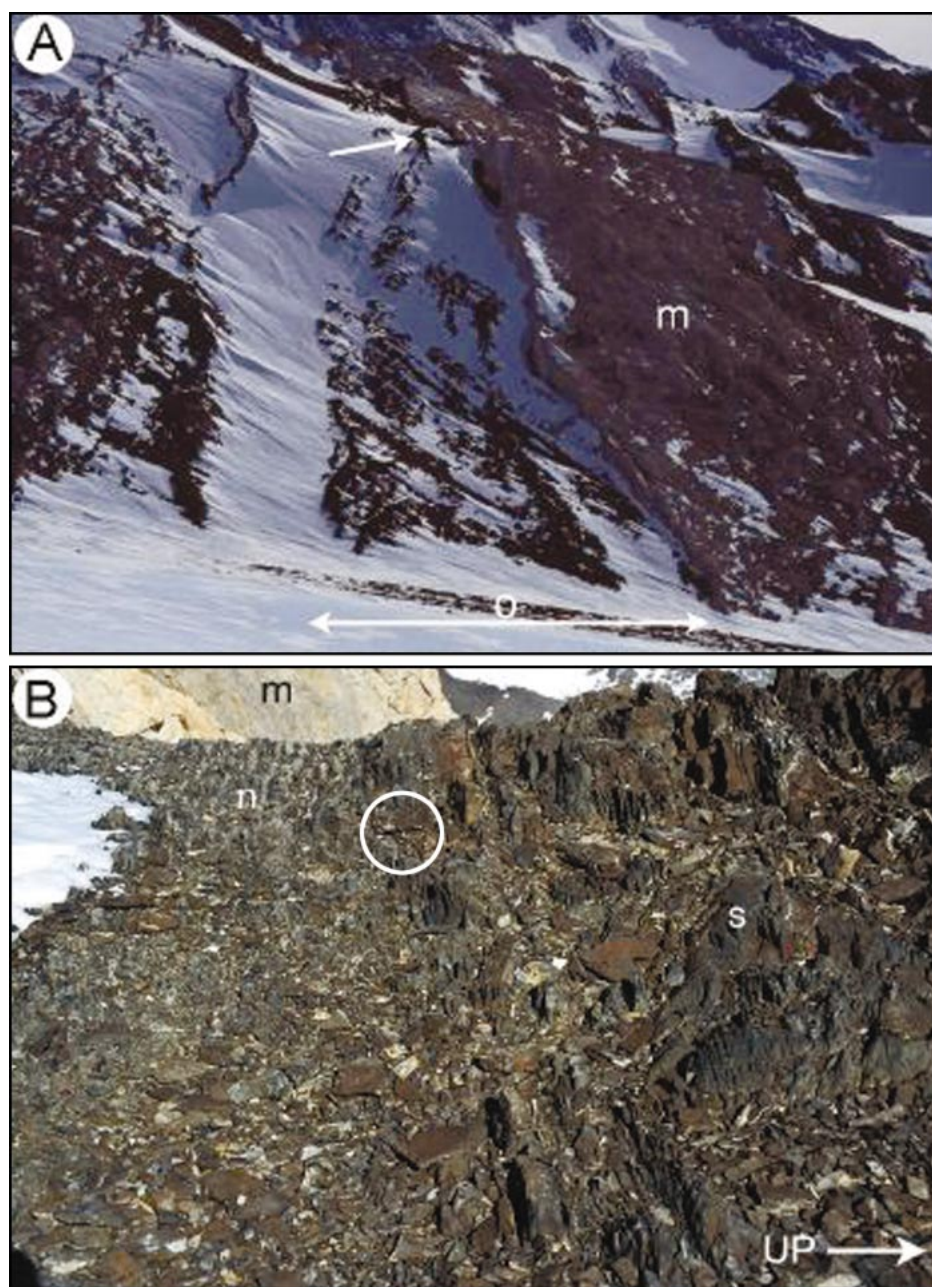
Figure 10. (A) Onlap of shale (o) against flank of large archaeocyathan reef (m). Small arrow indicates position where onlapping bed in foreground contacts the reef. The reef is ~40 m thick. (B) Transition from uppermost Shackleton Limestone archaeocyathid mound (m) into Starshot Formation at section 1 of the Holyoake Range. Nodular limestone and shale (n) directly overlie mound and are succeeded by fossiliferous calcareous siltstone and very fine sandstone (s). Pencil (circled) is 14 cm long.

glomerate and sandstone facies (described subsequently) indicates that this siltstone/very fine sandstone was deposited in nearshore to shoreline environments. The pebbly sandstone beds have many of the characteristics of debris flow deposits such as matrix support and lack of grading. These beds likely formed as shallow-marine debris flows, possibly representing storm-generated deposits.

Trilobite Fauna

The fauna from the calcareous siltstone/very fine sandstone units is of Early Cambrian age. Intercontinental correlation of Early Cambrian trilobites is very difficult, and considerable uncertainties exist (Palmer and Rowell, 1995; Palmer, 1998). However, the fauna reported here has closest affinities to faunas from South Australia and south China. Although the trilobites are slightly deformed and indifferently preserved, there appear to be specimens of *Redlichia*, *Megapaleolenus*, and *Hsuaspis* cf. *bilobata* Pocock. The latter is probably conspecific with a trilobite described by Palmer and Rowell (1995) from a collection (C-83–3) made directly below the Douglas Conglomerate on the east side of the Holyoake Range. Given that their samples came from a locality very close to ours, it is possible that the very same interval was sampled. *Hsuaspis* occurs in China in rocks of late Chiungchussuan to medial Tsanglangpuan age (Zhang et al., 1995) and in Australia in beds considered to be Botomian equivalents (Jell in Bengtson et al., 1990). *Megapaleolenus* characterizes a zone at the top of the Tsanglangpuian (Chang, 1988). The late(?) Tsanglangpuian/Botomian age of this fauna suggests a need to reevaluate the age of assemblage 5 of Palmer and Rowell (1995) from the underlying Shackleton Limestone, determinations of which have been poorly constrained but thought to be a slightly younger Early Cambrian age.

Mixed Conglomerate, Sandstone, and Silty Shale (Douglas Conglomerate)

This particularly complex facies consists of interbedded conglomerate; sandstone beds with a wide range of grain sizes, bed thicknesses, and sedimentary structures; and black silty shale. Conglomerate beds range from sandy pebble conglomerate to poorly sorted cobble/boulder conglomerate. Individual beds are generally <2 m thick, but amalgamated units of conglomerate range up to 17 m in thickness. Beds range from massive, unorganized and poorly sorted to well graded (Fig. 14A) with well-organized transitions, e.g.,

supported with widely dispersed, well-rounded, pebbles that make up <20% of the lithology. The pebbles consist mostly of quartzite, white limestone, and intraclasts of fine sandstone generally <7 cm in diameter, although one bed contains a sandstone intraclast that is 45 × 5 cm in cross section. At section 1, this facies occurs in the 10 m of strata directly

below the first stratigraphic occurrence of conglomerate beds.

Few diagnostic sedimentary structures occur in this facies, but the presence of abundant trilobite and hyolithid fossils as well as bored phosphate nodules indicates a marine depositional setting. The relative paucity of shaly beds and a stratigraphic transition into con-
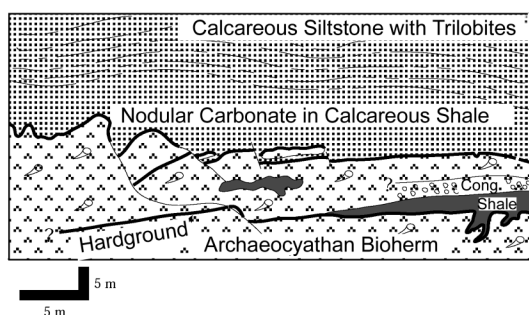
Figure 11. Drawing of upper part of composite reef from section 1, Holyoake Range. View is to the west. Two distinctive phosphate-encrusted hardground layers near the top of the reef (heavy black lines) indicate that drowning occurred in at least two phases. The initial drowning event was followed by shale deposition and renewed growth of the archaeocyathan reef. The discontinuous carbonate conglomerate above the shale likely indicates erosion of the reef, possibly by submarine currents (e.g., Schlager, 1998, 1999). Lateral expansion of the reef over the carbonate conglomerate was followed by the final drowning event and formation of the laterally persistent phosphatic hardground at the top of the reef. Nodular carbonate and shale then buried the reef following drowning. It is unclear whether the small archaeocyathan bioherm in the center of the drawing is a third phase of reef development or is a three-dimensional extension of the second phase of reef growth. Small faults offsetting the hardground at the top of the reef cannot be traced into the overlying shale or siltstone.

clast-supported cobble conglomerate to pebbly sandstone to cross-bedded sandstone. Beds are commonly lenticular with concave-up deeply scoured bases. Conglomerate beds have highly variable clast compositions from siliciclastic dominated to carbonate dominated. Large, outsized intraclast blocks of sandstone range up to 1 × 0.4 m in cross section and are particularly abundant in areas of intimate interbedding of conglomerate and sandstone (Fig. 14B). Clasts clearly derived from the underlying Shackleton Limestone are abundant and locally outsized, particularly in section 2 (Figs. 9, 15, A and B), in which conglomerate beds rest directly on the upper surface of a bioherm. Large white biohermal limestone clasts range up to 2 × 0.8 m across and indicate local erosion and transport of bioherm blocks. Large outsized blocks also locally occur stratigraphically well above the bioherms. One outsized block of fenestral mudstone, 1.5 × 1.6 m in cross section, occurs in section 1, 147 m above the high point of the underlying bioherm (Fig. 14C).

Green light-red-weathering sandstone beds (Fig. 14B) range from fine grained to coarse grained with floating small pebbles. On average, these beds are medium to coarse grained and moderately to poorly sorted. Bed thicknesses range from 4 cm to >1 m and average 30–40 cm. These beds make up amalgamated units up to 12 m thick. Fine-grained sandstone beds show parallel lamination, trough cross-bedding (Fig. 14D), and locally developed current ripple lamination.

Black shaly siltstone and silty shale units are widely dispersed in this facies. These are generally a few tens of centimeters thick, but range up to 14 m thick. In some cases, the fine-grained units contain 1–3-cm-thick very fine grained to fine-grained sandstone beds that locally reach up to 15 cm in thickness.

This diverse facies was almost certainly deposited in a variety of subenvironments. It occurs in a stratigraphic position above the previously described facies such that it records continued shoaling and increased proximity of source regions. The bed thicknesses and coarse grain size of the conglomerate indicate proximal deposition, particularly given the large outsized meter-scale blocks. These blocks, locally derived from the Shackleton Limestone, indicate local tectonic uplift and erosion, particularly as strata from lower in the formation are also represented as clasts. The conglomerate beds first appear in section 1 (Fig. 8) in close association with phosphate-rich siltstone and farther up section with organic-rich silty shale. Thus, at least some of these beds were either deposited in, or in proximity to, a subaqueous setting. Although the phosphatic beds are likely marine in origin, the black silty shale beds could be either marine or marginal marine. Thicker units of mixed conglomerate and medium- to coarse-grained sandstone most likely represent fluvial

to marginal-marine deposits. Meter-scale graded beds (Fig. 14A) are likely channel deposits that were either interbedded with fluvial sheet-flood or shoreline sandstone beds. Similar facies were described as part of a wider range of coarse clastic facies of the Douglas Conglomerate by Rees and Rowell (1991). They interpreted the wider suite of facies as ranging from proximal to distal alluvial-fan deposits.

The strata described from the Shackleton–upper Byrd Group transition zone likely represent a shift from clearly shallow-marine (trilobite-bearing) siltstone through marginal-marine and then alluvial-fan deposits. Much of the shallow-marine deposits were deposited in fan-deltas. The stratigraphic framework and implications of these deposits are discussed next.

STRATIGRAPHIC REVISION

The depositional ages and stratigraphic relationships of the Neoproterozoic to lower Paleozoic formations in the central Transantarctic Mountains have been difficult to discern because of structural complexity, abundant ice cover, and a general lack of fossils. Detailed stratigraphic and sedimentological data presented in this paper, in addition to ages of detrital zircons from these units (Goodge et al., 2002), allow for a significant stratigraphic revision (Fig. 16). The depositional transition described herein between the uppermost Shackleton Limestone and the siliciclastic deposits of the upper Byrd Group is critical in this regard.

A multitude of stratigraphic schemes and inferred depositional ages of the various formations have been proposed (Fig. 2). In many cases, previous workers had limited access to particular formations and did not examine others, which hindered stratigraphic synthesis. For instance, Rees, Rowell, and various colleagues (Fig. 2, columns E through I) examined the Shackleton and Douglas Formations, but did not study the Starshot Formation. Conversely, all previous workers uniformly placed the Cobham and Goldie Formations in the Beardmore Group and generally considered them to be Neoproterozoic in age because of their apparent great thickness, lack of fossils, and inferred position beneath the Lower Cambrian Shackleton Limestone. However, the bulk of rocks previously mapped as Goldie Formation are now known to be late Early Cambrian in age or younger. Thus, we suggest that the Beardmore Group should be redefined to include only those rocks that are known, or reasonably inferred to be older than the Shackleton Limestone, namely, the Cobham Formation and "inboard" Goldie assemblage

of Goodge et al. (2002). The ''inboard'' Goldie is best defined as that part of the Beardmore Group lacking detrital minerals having ages younger than 600 Ma. This criterion restricts the mapped extent of the group to the eastern Cobham Range and the vicinity of Cotton Plateau (Fig. 1). The available detrital-zircon ages only indicate that this restricted Beardmore Group is younger than ca. 1000 Ma, although a new zircon U-Pb age of 668 Ma from gabbro associated with pillow basalts at Cotton Plateau (Goodge et al., 2002) indicates that deposition is at least in part late Neoproterozoic. Although the same geochronological constraints allow the ''inboard'' Goldie to be as young as the lowermost Shackleton Limestone, it is unlikely to be Cambrian. Sandstone samples with a young detrital signature (younger than 520 Ma) from the Lowery and Beardmore Glacier areas, as well as the northern Churchill Mountains, suggest that the outer ranges contain younger ''outboard'' Goldie deposits with a different provenance (Goodge et al., 2002). These should hereafter be mapped as part of the Starshot Formation (Figs. 1, 16).

Early workers recognized that the presence of Shackleton Limestone clasts in both the Douglas and Starshot Formations indicated that these units were equivalent to and/or younger than that formation. Most subsequent workers considered the Douglas to be younger than, and locally in unconformable contact with, the underlying Shackleton Limestone (Fig. 2), and we concur. Several lines of evidence indicate age equivalence between the Starshot and Douglas Formations (Fig. 16). First, ''Douglas''-like conglomeratic units occur in rocks mapped as Starshot Formation in the Mount Ubique area. Second, these conglomeratic parts of the Starshot occur spatially between more inboard regions mapped as Douglas Conglomerate that are dominated by coarse conglomerate and more outboard regions of the Starshot Formation that are almost exclusively sandstone and minor shale. Third, paleocurrent patterns in these sandy units are uniformly oriented from inboard areas toward more outboard regions (i.e., flow from west to east and southwest to northeast) (Myrow et al., 2002). Finally, the age distributions of detrital zircons from samples of these formations are similar (J. Goodge, unpubl. data).

Rees et al. (1988) and Rowell et al. (1988b) clearly demonstrated an unconformity between the Shackleton and Douglas Formations at one locality, but noted fault contacts between these units at other locations. The depositional transition from the Shackleton Limestone into the younger siliciclastic units of the upper Byrd Group described in this paper indicates that, in this area, the latter entirely postdate Shackleton carbonate deposition (i.e., none is even partly coeval). The siltstone/sandstone facies within this upward-coarsening transition (Fig. 8, 23.8–63.5 m) is similar in grain size, texture, color, and weathering to the bulk of the Starshot Formation and is thus included in that formation. Overlying coarse-grained units are assigned to the Douglas Conglomerate. It is apparent from spatial distributions of conglomeratic facies and paleocurrents that the coarse-grained units of the Douglas Conglomerate make up a thick sedimentary wedge of limited lateral extent. This is a logical geometry because this unit represents the deposits of an alluvial-fan complex (Rees and Rowell, 1991). The distal fingers of this wedge make up the conglomerate beds of the Starshot in the Mount Ubique area. The result of progradation of this alluvial wedge into a marine setting is that sandstone facies both underlie and are laterally equivalent to this conglomeratic body (Fig. 16). In fact, this wedge is also overlain by a thick section of sandstone (Fig. 16), as seen at High Ridge (Fig. 1). Here, an upward-coarsening succession identical to that in sections 1 and 2 is followed by hundreds of meters of sandstone and minor shale facies typical of the Starshot Formation.

The presence of sandstone below, adjacent to, and above a conglomeratic wedge helps explain a number of other stratigraphic complexities, including the relationship of these units with the Dick Formation. As described earlier, our brief study of the Dick Formation at co-type section N of Skinner (1964, 1965) revealed a remarkably similar suite of bedding characteristics and internal sedimentary structures to those of the Starshot Formation. In addition, detrital-zircon ages from the Dick Formation at section N include populations at 495, 510, 555, 1020, and 1130 Ma and a range of older Proterozoic and Archean ages; these ages are very similar to zircon-age distributions from the Starshot Formation elsewhere (J. Goodge, unpubl. data). These data and other geologic relationships lead us to consider these units as equivalents. According to Skinner (1964), Burgess and Lammerink (1979), and Rowell and Rees (1989), the Dick Formation is overlain by, and in conformable contact with, the Douglas Conglomerate at co-type section P. This is therefore analogous to the transition from the Starshot Formation (with Botomian fossils, ca. 515 Ma) to the Douglas Conglomerate in the Holyoke sections described previously herein. The presence of young 500–490 Ma detrital zircons in the Dick Formation at co-type section N indicates that the strata at this locality are potentially younger by ~20 m.y. than at its other co-type section P, including part or all of the Douglas Conglomerate. If the Dick Formation is at different locations both older than, and younger than, a (presumably) large part of the Douglas Conglomerate, then the latter likely forms a wedge within the sandy Dick Formation, as described for the Starshot Formation. The base of the Dick Formation is not exposed, but we would predict that it overlies a shaly unit that itself directly overlies the Shackleton Limestone, as described for the Holyoake Range in this paper.

We see no benefit in continuing to use the name ''Dick Formation'' because this unit appears to be identical to the Starshot Formation in lithologic and sedimentologic character, detrital-zircon geochronology, and stratigraphic relationship with the Douglas Conglomerate. We thus suggest that the name ''Dick Formation'' be abandoned and that exposures previously mapped as Dick Formation be placed in the Starshot Formation (Fig. 16). The name ''Starshot'' is preferred given its earlier introduction (Laird, 1963). In summary, we suggest that the Starshot Formation should include all rocks originally mapped within this formation plus rocks of the ''outboard'' Goldie Formation and all of those rocks previously mapped as Dick Formation. We suggest continued use of the term ''Douglas Formation'' for those rocks dominated by conglomerate.

Holyoake Formation

The depositional transition between the Shackleton Limestone and overlying clastic deposits of the upper Byrd Group marks a major lithologic break that reflects profound changes in depositional systems, depositional history, and tectonics. This transition occurs at the contact between the phosphatized upper surface of the Shackleton Limestone and the overlying fine-grained shale and mixed nodular carbonate and shale deposits. The shaly deposits form an extensive stratigraphic unit that can be mapped throughout the Holyoake Range and may be present within adjacent ranges as well. Previous workers did not recognize this unit as a separate mappable lithosome and therefore would have likely included these deposits as a facies within the Douglas Conglomerate or Starshot Formation. This lithologically distinct unit, which consists of fine-grained dominantly siliciclastic facies sandwiched between carbonate-
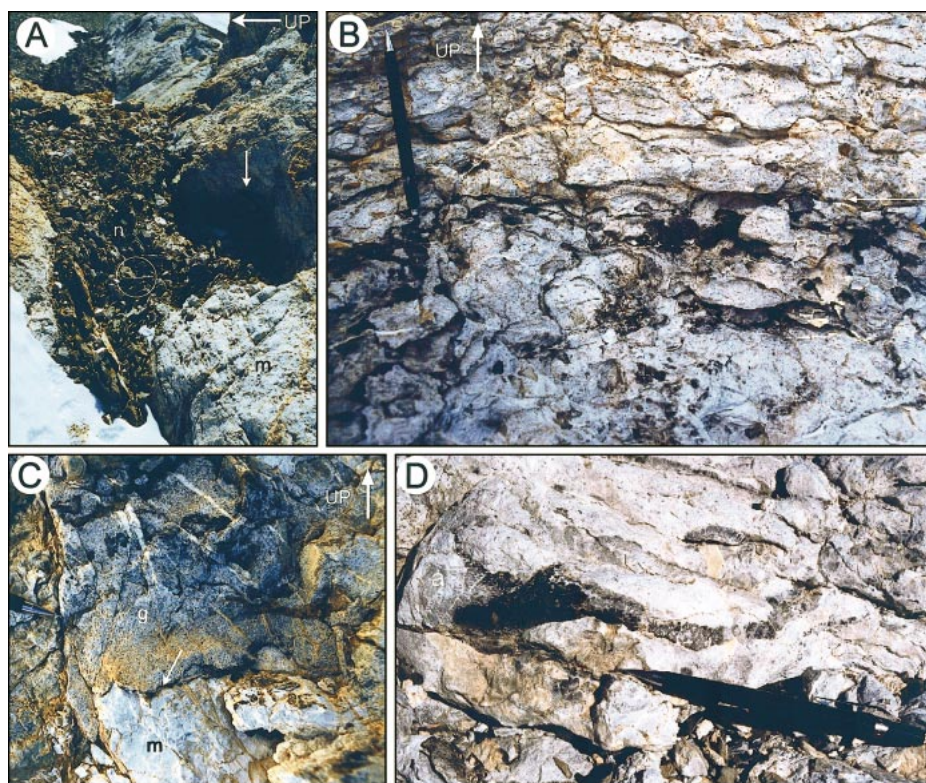
Figure 12. Section 1 of the Holyoake Range. (A) Irregular relief on top of lowermost archaeocyathan reef (m); bedding is vertical. White arrow points to composite phosphatic hardground surface that caps the reef. Shale and nodular limestone (n) overlie and onlap the hardground surface. Pen is 14 cm long (circled). (B) Phosphatic hardground surface (thin arrow on right edge) near top of composite archaeocyathan reef. Bedding is vertical. Pencil is 14 cm long. (C) Close-up of phosphatic grainstone fill (g) and multiple hardground surfaces (white arrow) on top of composite archaeocyathan mound (m). Tip of pencil is 2 cm in length. (D) Large phosphatized archaeocyathid (a) in small reef within shale and nodular carbonate facies. Bedding-plane view. Pencil is 14 cm long.

platform deposits (Shackleton Formation) and coarse, synorogenic molasse (Douglas and Starshot Formations), is herein assigned to a separate formation. Because this unit occurs within the Holyoake Range, we suggest that it be designated the "Holyoake Formation." This formation is exposed at numerous sections along the eastern Holyoake Range, in-

cluding sections 1 and 2 (Figs. 8 and 9), but is not present where uplift and erosion of the Shackleton Limestone caused angular unconformity with overlying conglomerate units of the Douglas Conglomerate (Rees et al., 1988; Rowell et al., 1988b). The base of this formation is defined as the lowermost shale that rests on the phosphatized surface at the top of

the Shackleton Limestone. The top of the formation is defined as the top of the extensive nodular carbonate and shale unit that underlies calcareous siltstone to very fine grained sandstone of the basal Starshot Formation. The formation reaches a maximum thickness of ~45 m and thins appreciably over the bioherms of the upper Shackleton Limestone. Locally, in section 2 and possibly elsewhere, conglomeratic channels of the overlying Douglas Conglomerate cut out both the Starshot Formation and the thinned Holyoake Formation at the high point of a bioherm (Fig. 9). The type section of the Holyoake Formation is designated at the section 1 locality (Fig. 1; GPS [Global Positioning System] location 82°13.185′S, 160°16.122′E).

The Holyoake Formation probably also exists in more outboard regions, between the Shackleton and Starshot Formations, but this stratigraphic relationship has not been observed. Such a transition would be similar to those at inboard localities except that the overlying clastic deposits would be mostly sand-
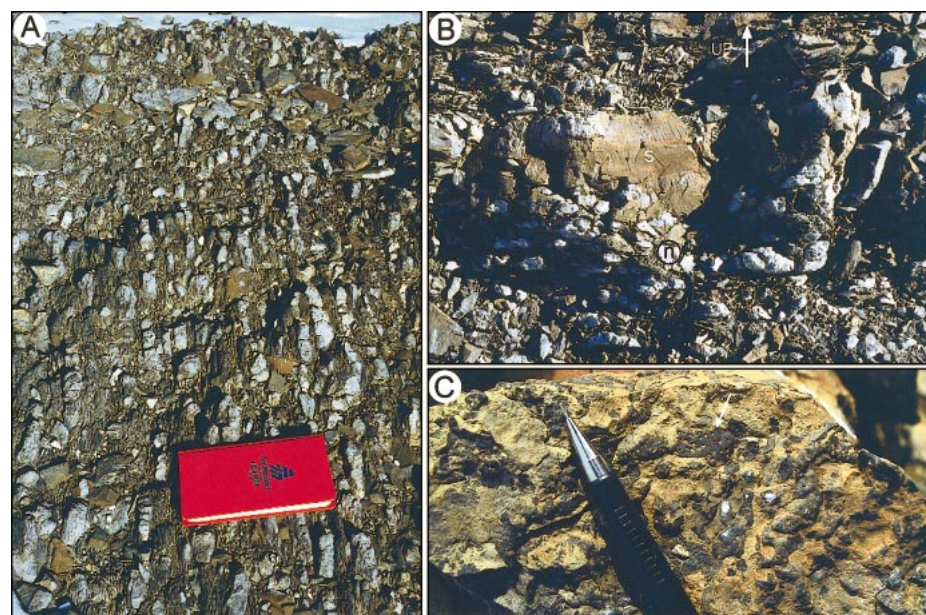

Figure 13. Section 1 of the Holyoake Range. (A) Shale and nodular carbonate of Holyoake Formation. (B) Transition from interbedded nodular limestone (n; circled) and shale of upper Holyoake Formation to very fine calcareous sandstone (s) w/limestone nodules of basal Starshot Formation. Pencil (vertical; lower center) is 14 cm long. (C) Borings in nodular, black phosphate carbonate in Starshot Formation at 59.17 m (Fig. 9). Arrow points to small borings on bedding plane. Pencil tip is 2 cm long.
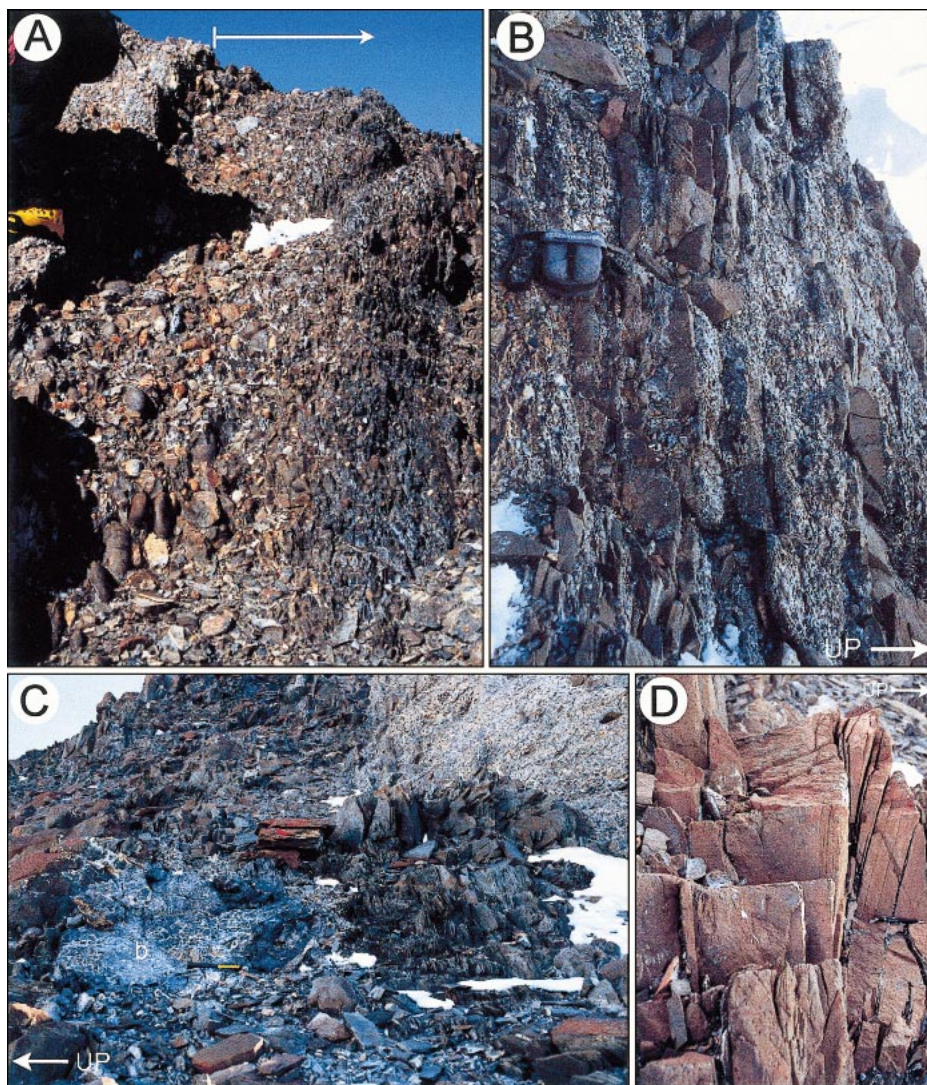
Figure 14. Douglas Conglomerate at section 1 of Holyoake Range. (A) Graded conglomerate bed 1.8 m thick at 70.2 m (Fig. 9). Yellow, size 9 men's boot on left for scale (~30 cm). Arrow points in direction of stratigraphic top. (B) Interbedded conglomerate and sandstone. Camera case for scale. (C) Large block of black fenestral mudstone with calcite veins (b) within conglomeratic bed at 166 m (Fig. 9). Block is 1.5 × 1.6 m in cross section. Interbedded sandstone and shale on right. Hammer for scale. (D) Trough cross-bedded sandstone and pebbly sandstone near the top of section. Pencil is 14 cm long.
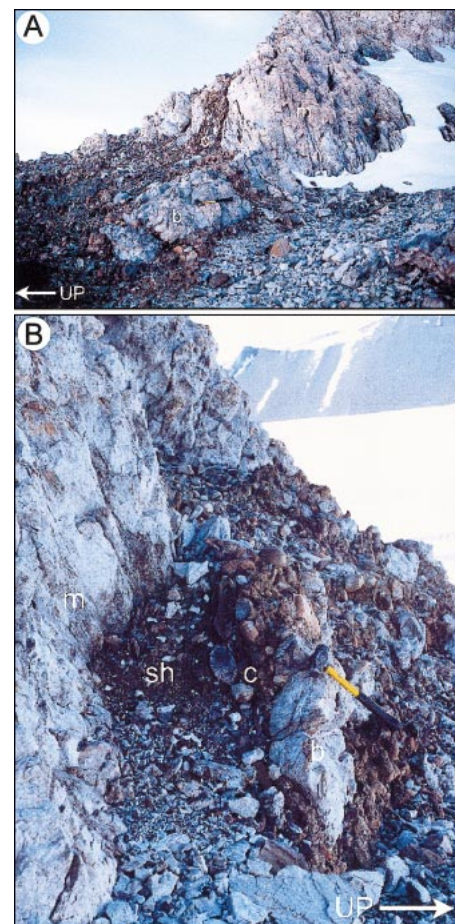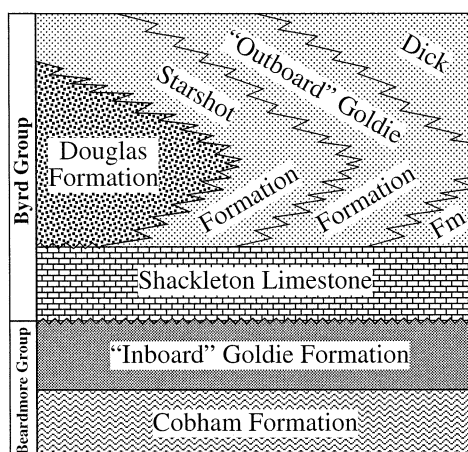

Figure 15. Section 2 of the Holyoake Range. (A) Conglomerate (c) resting directly on eroded upper surface of large archaeocyathan mound (m). Hammer (lower center) for scale. Large blocks of archaeocyathan bioherm (b) occur in basal conglomerate beds in Douglas Conglomerate. (B) Top of large archaeocyathan mound at section 2 of the Holyoake Range. Both the mound (m) and the flanking shale/nodular limestone fill (sh) are truncated by boulder conglomerate (c). Large block (b) of archaeocyathan bioherm in conglomerate. Hammer for scale.

stone of the Starshot Formation with minor, if any, conglomerate (i.e., no Douglas Conglomerate). In fact, the High Ridge section (Fig. 1) records such a transition. Here, the Holyoake Formation is overlain by deposits with only one 95-cm-thick conglomerate bed, and the rest is almost entirely sandstone and minor shale of the Starshot Formation. The Holyoake Formation is therefore considered an extensive unit of the middle Byrd Group that everywhere overlies Shackleton Limestone in the study area and underlies sandy and conglomeratic deposits of the Starshot and Douglas Formations.

DISCUSSION

New data outlined in this paper on Byrd Group strata provide multiple constraints on the timing of deposition of various units and the onset of Ross orogenesis. Previously published trilobite and archaeocyathan biostratigraphic data from the Shackleton Limestone generally constrained determinations of the age of this unit, but the unfossiliferous basal unit of the formation was never dated, and the stratigraphic top of the formation was never recognized, let alone dated. Chemostratigraphic data presented in this paper strongly suggest
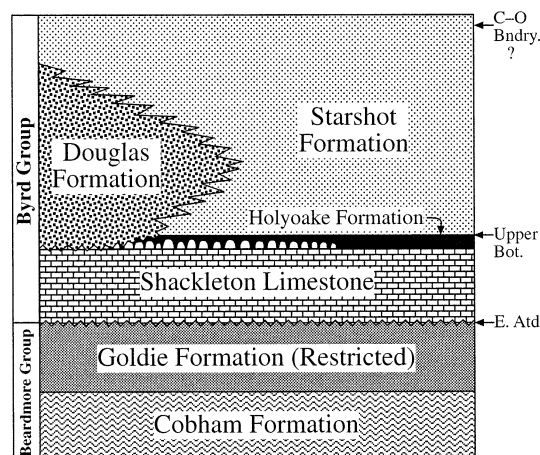
the lower sandstone-rich part of the formation to be early Atdabanian in age (isotopic excursion IV of Brasier et al., 1994; Brasier and Sukhov, 1998). Although this section, which we interpret to be in fault contact with the Goldie Formation at Cotton Plateau, may not be the absolute base of the formation, lower parts may not be much older. This quartz-rich sandstone of the basal Shackleton has analogues with relatively thin Cambrian basal transgressive sandstone units across Laurentia and elsewhere. Thus, Cambrian seas appar-

Figure 16. Inferred age relationships of previously defined units in the central Transantarctic Mountains. "Inboard" and "outboard" components of the Goldie Formation were distinguished by Goodge et al. (2002). Revised stratigraphy includes newly defined Holyoake Formation and Goldie Formation that is restricted to deposits older than the Shackleton Limestone (Beardmore Group).

ently transgressed this part of the Antarctic margin during the early Atdabanian.

The ensuing carbonate platform existed throughout the rest of the Atdabanian and subsequent Botomian stages. The first documented depositional contact at the top of the Shackleton Limestone is described herein and records a major depositional shift in the geologic history of the Transantarctic Mountains. Archaeocyathan bioherms developed during the last stage of deposition of the formation are encrusted with multiple thick, phosphatic hardgrounds. The preserved growth morphology of these bioherms and lack of evidence for subaerial exposure indicates the bioherms were killed as a result of rapid drowning of the carbonate ramp. Multiple hardgrounds near the apex of the bioherms suggest that demise of these structures was a multiphase event. The phosphatic crusts capping the bioherms developed at, or near, the sediment-water interface in a deep-marine setting (Schlager, 1998), indicating that they formed over a considerable amount of time during which there was very little or no other sedimentation. Rapid initial drowning led to deposition of the regionally extensive phosphatic crust and subsequent deposition of the Holyoake Formation, a unit of black shale followed by mixed nodular carbonate and shale facies. The latter grades upward into trilobite-bearing calcareous siltstone of the Starshot Formation and then into mixed conglomerate and sandstone facies of the Douglas Conglomerate. Onlapping relationships between the overlying deeper-water siliciclastic rocks and the phosphate-encrusted bioherm further indicate the carbonates were drowned prior to deposition of the siliciclastic sediments. The siliciclastic facies transitions record progressive shoaling

from initial deep-water deposition (shale and nodular carbonate encased in shale) through shallow-marine to subaerial deposits (conglomerate facies). This large-scale upward-coarsening succession records a regionally significant shift from passive carbonate-ramp deposition to a major influx of terrigenous sediment. This was triggered by a major tectonic event, the structurally active phase of the Ross orogeny, which resulted in faulting, uplift, progressive unroofing, and widespread deposition of an enormous clastic wedge. The trilobite fauna from the lowermost siltstone deposits of this transition zone (basal beds of Starshot Formation) date the onset of this event as being within the Botomian Stage (ca. 515–510 Ma). This age is consistent with ca. 490 Ma and older ages of detrital zircons in the overlying Starshot and Douglas Formations (J. Goodge, unpublished data). It is measurably younger, however, than the onset of basement deformation in this area, dated to an Early Cambrian interval between ca. 540 and 520 Ma (Goodge et al., 1993).

Inception of Ross orogenesis within the supracrustal successions of the central Transantarctic Mountains is therefore recorded by the drowning and final stratigraphic occurrence of archaeocyathan reefs. It is also coincident with a dramatic change in composition, provenance, and depositional setting of marginal-basin siliciclastic deposits. Detrital-zircon ages from the Starshot and Douglas Formations show that they were deposited well into the Late Cambrian and probably into the Early Ordovician. Given that the relatively thin Holyoake Formation is Botomian in age, the overlying clastic units of the Byrd Group span at least the rest of the Cambrian (>20 m.y.). These relationships provide important new

timing constraints on clastic units of poorly known age. The stratigraphic relationships discussed here not only provide precise constraints on the timing of supracrustal deformation, but they also help to clarify the regional tectonic patterns (discussed next).

Assigning the onset of supracrustal deformation to within the Botomian Stage is consistent with stratigraphic relationships in other areas. Rowell et al. (1992) concluded that the earliest Ross deformation was likely to be Middle Cambrian in age, on the basis of (1) the presence of folded rocks beneath the Middle Cambrian Nelson Limestone in the Neptune Range of the Pensacola Mountains, and (2) the Douglas unconformity above the Lower Cambrian Shackleton Limestone (defined at that time to be Toyonian and older). The new fossil control on the basal Starshot Formation described herein indicates that deformation commenced somewhat earlier in the late Early Cambrian. New evidence shows that the Pensacola Mountains underwent an initial major phase of late Early to early Middle Cambrian deformation and subsequent episodes of late Middle Cambrian to possible Ordovician activity (Rowell et al., 2001). In the Queen Maud Mountains, deformation of Middle Cambrian limestone of the Taylor Formation (Rowell et al., 1997; Encarnación et al., 1999) is inferred to have occurred in the Middle Cambrian to Ordovician, well after the initial phase of tectonism described here. Together, these results support the idea of protracted and/or temporally punctuated deformation during the Ross orogenic cycle (Rowell et al., 1992, 2001; Goodge et al., 1993; Goodge, 1997), which can be inferred on the basis of radiometric age control to span the period from 540 to 480 Ma.

TECTONIC IMPLICATIONS

Although the transition zone between the Shackleton and upper Byrd Group clastic deposits is apparently recorded only within a relatively small outcrop area, this interval provides keys to understanding the nature and scale of crustal deformation. Our revised stratigraphic framework suggests that low-angle faults within the Shackleton Limestone and between the older Shackleton Limestone and the upper Byrd Group clastic units are of reverse geometry. Kinematic displacement may vary from one area to another, but recognition that "outboard" Goldie Formation rocks are in fact part of the Starshot Formation, and therefore younger than the Shackleton Limestone, refutes earlier assumptions that for all Shackleton-"Goldie" contacts, the Shackleton rests either unconformably or structurally upon older strata. Those exposures with Shackleton above newly recognized Starshot Formation rocks ("outboard" Goldie) are now viewed as an older-over-younger structural juxtaposition. Intraformational faults, which are well exposed at Cambrian Bluff north of Nimrod Glacier (Laird et al., 1971), probably represent initial structural repetition of the Shackleton carbonate ramp.

The consistent west to east and southwest to northeast (outboard) paleocurrent data from the Starshot Formation (Myrow et al., 2002) agrees with the spatial distribution of conglomerate within the Byrd Group. Rapid drowning of the outboard carbonate platform was likely caused by flexural loading of crust underlying the inner margin to the west (near the present-day axial zone of the orogen). We speculate that loading was initially caused by the development of imbricate east-vergent thrust sheets farther to the west, but further subsidence may have resulted from the negative buoyancy of subducting oceanic lithosphere (Dickinson, 1995). Nonetheless, the timing of the stratigraphic record relative to known deformation events in the region indicates tectonic rather than global eustatic control. As deformation progressed, the carbonate platform was uplifted and eroded, so that in some places it was entirely removed and older Neoproterozoic rocks were exposed and eroded to become clasts in the coarser molasse facies. Locally, incomplete removal of the carbonate led to an unconformity of folded Shackleton Limestone overlain by Douglas Conglomerate. Differences in the composition of the Douglas Conglomerate—relative abundance of carbonate versus siliciclastic clasts—resulted from spatial and temporal changes in source rocks during progressive deformation and unroofing. Farther outboard, rapid subsidence occurred as a result of thrust loading of the inboard crust. The stratigraphic response to thrusting was drowning of the bioherms and development of a phosphatic hardground, followed by deposition of shale and then progradation of a thick clastic wedge. Although the specific geometry of thrusting and flexural loading remains to be established, the stratigraphic transitions reported here offer a broad framework in which to understand the interplay of tectonism and sedimentation. In addition, the implied vertical and lateral stratigraphic changes reflect the shift from preorogenic passive carbonate sedimentation to synorogenic deposition of molassic strata spanning proximal alluvial-fan to shelf environments.

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