A lithofacies analysis of a South Polar glaciation and the glacial-to-postglacial transition in the early Permian: Pagoda Formation and Lower Mackellar Formation, Shackleton Glacier region, Antarctica

--Manuscript Draft--

Manuscript Number: 

Article Type: Research Article 

Corresponding Author: Libby R.W. Ives, M.S. University of Wisconsin–Milwaukee Milwaukee, Wisconsin UNITED STATES 

First Author: Libby R.W. Ives, M.S. 

Order of Authors: Libby R.W. Ives, M.S.

John L. Isbell, M.S., Ph.D. 

Abstract: The evolving understanding of the extent, character, and asynchronous timing of the glacial-interglacial cycles during the Late Paleozoic Ice Age (LPIA) has created a need to better characterize the glacial histories of discrete ice centers across Gondwana. The Transantarctic Basin was located in a South Polar position near the margin of the East Antarctic Craton during the Early Permian. The basin contains a Permian – Jurassic succession. In the Shackleton Glacier region of the central Transantarctic Mountains, Early Permian glaciogenic strata (Pagoda Formation) occur at the base of the Transantarctic Basin succession and are succeeded by a basinal to a pro-deltaic shale and sandstone system (Mackellar Formation). This study uses detailed lithofacies and paleo-transport analyses of the Pagoda Formation and Lower Mackellar Formation from five localities in the Shackleton Glacier region to characterize LPIA glaciation and the glacial-to-postglacial transition in a South Polar, basin-marginal setting. These analyses show that the massive, sandy, clast-poor diamictites of the Pagoda Fm were deposited in a basin-marginal subaqueous setting through a variety of glacial and glacially-influence mechanisms in a depositional environment with depths similar to a continental shelf. Current-transported sands and stratified diamictites that occur at the top of the Pagoda Fm were deposited as part of grounding-line fan systems. Up to at least 100 m of topographic relief on the erosional surface underlying the Pagoda Fm strongly influenced the thickness and transport-directions in the Pagoda Fm. Similar to elsewhere in the Transantarctic Basin, the contact between the Pagoda Formation and the Mackellar Formation is sharp and apparently abrupt. However, glacier retreat out of the Transantarctic Basin was likely not as rapid in this region as in other areas, as evidenced by dropstones that persist 5 – 6 m above the lower contact of the Mackellar Fm. The majority of the Pagoda Fm in the Shackleton Glacier region was deposited during a single retreat phase, represented by a single glacial depositional sequence, of a temperate glacier with high meltwater discharge. This glacier was likely flowing off of an upland on the side of the Transantarctic Basin closer to the Panthalassan/Gondwanide margin. Uniform subglacial striae orientation across 100 m of paleotopographic relief suggest that the glacier was significantly thick to “overtop” the paleotopography in the Shackleton Glacier region. Though sedimentation in the Shackleton Glacier region was deposited during a single glacier retreat phase, evidence from this study does not preclude prior glacial cycles or additional cycles preserved elsewhere in the basin.
November 17, 2020

Dear Journal of Sedimentary Research Editor:

Please consider our manuscript “A lithofacies analysis of a South Polar glaciation and the glacial-to-postglacial transition in the early Permian: Pagoda Formation and Lower Mackellar Formation, Shackleton Glacier region, Antarctica” for publication in JSR.

In this paper we describe in detail the sedimentology an early Permian glacial and post-glacial sedimentary succession in the Shackleton Glacier region of the Transantarctic Basin. This succession is significant because it contributes a unique perspective (very high latitude, basin-marginal, non-marine) on glacial sedimentation during one of the maximums of the Late Paleozoic Ice Age. We conclude that the strata hold evidence for only a single glacial retreat sequence, that deposition of this sequence was strongly controlled by 100+ m paleotopography, and that glaciers persisted in the Transantarctic Basin after the dominant depositional ceased to be glaciogenic. Our observations also strengthen the hypotheses that this area was not continually covered by glacial ice during the Late Paleozoic Ice Age and that a significant ice center existed on the side of the Transantarctic Basin proximal to East Antarctic Craton’s Panthalassan margin.

We believe that JSR is the appropriate place for the publication of this manuscript because of the thorough discussion of glacial processes and depositional environments involved in our interpretation of the facies described in this paper.

John Isbell and I thank you for the consideration of this manuscript.

Sincerely,

Libby Ives
A lithofacies analysis of a South Polar glaciation and the glacial-to-postglacial transition in the early Permian: Pagoda Formation and Lower Mackellar Formation, Shackleton Glacier region, Antarctica

AUTHORS
Libby R.W. Ives* and John L. Isbell

University of Wisconsin – Milwaukee, Department of Geosciences, Lapham Hall, 3209 N. Maryland Ave.,

Milwaukee, WI, 53211, USA

*Corresponding author, woodfor5@uwm.edu
ABSTRACT

The evolving understanding of the extent, character, and asynchronous timing of the glacial-interglacial cycles during the Late Paleozoic Ice Age (LPIA) has created a need to better characterize the glacial histories of discrete ice centers across Gondwana. The Transantarctic Basin was located in a South Polar position near the margin of the East Antarctic Craton during the Early Permian. The basin contains a Permian–Jurassic succession. In the Shackleton Glacier region of the central Transantarctic Mountains, Early Permian glaciogenic strata (Pagoda Formation) occur at the base of the Transantarctic Basin succession and are succeeded by a basinal to pro-deltaic shale and sandstone system (Mackellar Formation).

This study uses detailed lithofacies and paleo-transport analyses of the Pagoda Formation and Lower Mackellar Formation from five localities in the Shackleton Glacier region to characterize LPIA glaciation and the glacial-to-postglacial transition in a South Polar, basin-marginal setting. These analyses show that the massive, sandy, clast-poor diamictites of the Pagoda Fm were deposited in a basin-marginal subaqueous setting through a variety of glacial and glacially-influence mechanisms in a depositional environment with depths similar to a continental shelf. Current-transported sands and stratified diamictites that occur at the top of the Pagoda Fm were deposited as part of grounding-line fan systems. Up to at least 100 m of topographic relief on the erosional surface underlying the Pagoda Fm strongly influenced the thickness and transport-directions in the Pagoda Fm.

Similar to elsewhere in the Transantarctic Basin, the contact between the Pagoda Formation and the Mackellar Formation is sharp and apparently abrupt. However, glacier retreat out of the Transantarctic Basin was likely not as rapid in this region as in other areas, as evidenced by dropstones that persist 5 – 6 m above the lower contact of the Mackellar Fm. The majority of the Pagoda Fm in the Shackleton Glacier region was deposited during a single retreat phase, represented by a single glacial
depositional sequence, of a temperate glacier with high meltwater discharge. This glacier was likely
flowing off of an upland on the side of the Transantarctic Basin closer to the Panthalassan/Gondwanide
margin. Uniform subglacial striae orientation across 100 m of paleotopographic relief suggest that the
glacier was significantly thick to “overtop” the paleotopography in the Shackleton Glacier region.
Though sedimentation in the Shackleton Glacier region was deposited during a single glacier retreat
phase, evidence from this study does not preclude prior glacial cycles or additional cycles preserved
elsewhere in the basin.
INTRODUCTION

Strata of the Transantarctic Basin (TAB) outcrop along the Transantarctic Mountains in the Ross Sea sector of East Antarctica (Fig. 1B). The TAB contains a complete South Polar sedimentary record of the global “icehouse” to “greenhouse” transition of the Permian Period (Collinson et al. 1994; Collinson et al. 2006; Isbell et al. 2008b). Sedimentation in the TAB was dominated by glaciogenic processes during the late Carboniferous - early Permian (Isbell et al. 2008c). This interval was part of the Late Paleozoic Ice Age (LPIA, ~374 – 256 Ma) (Fielding et al. 2008c; Montañez and Poulsen 2013). Widespread glaciation across Gondwana characterized the LPIA, as did low $pCO_2$, high $pO_2$, generally low eustatic levels with large magnitude fluctuations, low solar luminosity, and increased $\delta^{18}O$ and $\delta^{13}C$ values relative to the rest of the Phanerozoic (Gastaldo et al. 1996; Raymond and Metz 2004; Montañez and Soreghan 2006; Fielding et al. 2008d; Rygel et al. 2008; Montañez and Poulsen 2013). Glaciogenic and glacially-influenced sediments deposited during the LPIA can be used to constrain the extent and style of glaciation that occurred in a basin. This work can be used to tie near-field records of glaciation to the global and “far-field” records of climate change during the LPIA (Montañez et al. 2007; Rygel et al. 2008; Soreghan et al. 2019). Together, near-field and far-field records can be combined to approach a holistic understanding of the effects that the onset, duration, and ultimate collapse of a global icehouse influenced Earth, both geologically and biologically.

As a whole, the LPIA glaciogenic succession in the TAB represents a near-field record of South Polar glaciation and glacial-to-postglacial transition during an interval of wide-spread glaciation in Gondwana (Gzhelian – Sakmarian). In the Shackleton Glacier region, these strata differ from those in better-studied areas of the TAB because they were deposited at a higher paleolatitude, are more likely to be comprised of basin-marginal (as opposed to basinal) facies associations, and have flow directions that indicate sediments were sourced from the TAB’s Panthalassan side instead of the Antarctic craton. By
evaluating the sedimentology of LPIA deposits in the Shackleton Glacier area we aim to contribute to several significant discussions that remain unsettled.

First, this study will contribute to a better understanding of South Polar LPIA glacial environments by constraining glacial and postglacial depositional environments. Early workers interpreted Permian diamictites in the TAB as having a subglacial or terrestrial proglacial origin (Long 1964b; Lindsay 1970a; Miller 1989; Isbell et al. 2001), but more recent work has identified facies from subaqueous, pro-glacial environments (Isbell et al. 2008c; Koch and Isbell 2013; Cornamusini 2017).

Additionally, there is no consensus on whether the body of water occupying the TAB during and after glaciation was marine, lacustrine, or brackish (Collinson et al. 1994; Isbell et al. 2008c; Miller and Isbell 2010; Flaig et al. 2016; Jackson et al. 2016).

Second, researchers have not agreed upon the distribution of ice centers contributing to TAB glaciation (Fig. 2). Isbell (2010) and Isbell et al. (2008c) presented evidence for multiple glaciers entering the TAB from both the Panthalassan margin and the East Antarctic Craton, while others posit a single ice center entering the basin from the interior of the East Antarctic craton or from Victoria Land (Frakes et al. 1971; Isbell et al. 1997a; Fielding et al. 2010; Craddock et al. 2019).

Third, the basin-marginal position of TAB strata in the Shackleton Glacier region allows for the study of non-glacial components of the South Polar environments during the early Permian. The study area contains evidence of sub-aerial weathering and non-glacial deposition during the Early Permian that occurred prior to and/or during the deposition of glacial sediments (Isbell et al. 2001). By characterizing and constraining the relationship of this evidence with glaciogenic strata we can better understand the early Permian South Polar environment as a whole.

Finally, the driving mechanisms for the withdrawal of glaciers from the TAB during the early Permian remains unclear. López-Gamundi (2010) and Horan et al. (2019) interpreted the glacial to post-
glacial transition in other glaciated Gondwanan basins as a transgressive sequence (or series of sequences) resulting from a eustatic sea-level rise during a global warm interval. This idea is supported by recent geochronology studies that show millennia-scale glacial-interglacial cycles were likely synchronous, at least across western Gondwana, during the LPIA (Griffis 2019; Buso et al. 2020; López-Gamundí et al. in review). These glacial-interglacial cycles occur within a much longer-term deglaciation trend that mark the end of the LPIA in Gondwana; evidence of LPIA glaciation in western Gondwana ends prior to central and eastern Gondwana, likely due to the drift of Gondwana across the south pole (Isbell et al. 2012; Montañez and Poulsen 2013). While most LPIA basins have sedimentary records that contain several multi-million-year glacial-interglacial cycles, basin marginal areas of the TAB appear to have only a single major surface of glacier retreat (Fielding et al. 2008a; Isbell et al. 2008a; Fielding et al. 2010; Garbelli et al. 2019; Horan et al. 2019). Additionally, the limited fossil record suggests that glacial deposits in the TAB are constrained to the early Permian (Asselian) (Masood et al. 1994; Askin 1998; Babcock et al. 2002). This contrasts with the idea that higher-latitude basins should have more prolonged records of glacial sedimentation. Therefore, careful consideration of the driving mechanisms behind the glacial-to-postglacial transition in the TAB, and its driver(s), whether climatic or geologic, is warranted.

In this paper, we evaluate the physical sedimentology of four glacial (the Pagoda Fm) and five post-glacial (the Mackellar Fm and Weaver Fm) successions in the Shackleton Glacier region of the TAB (Fig. 1B).

GEOLOGICAL SETTING

Permian glaciogenic and post-glacial strata occur in outcrops along the margin of the East Antarctic Craton including Victoria Land (Isbell 2010; Cornamusini 2017), the central Transantarctic Mountains (Isbell et al. 2008c), Wisconsin and Ohio Ranges (Long 1964b; Aitchison et al. 1988), Pensacola Mountains (Schmidt and Williams 1969; Frakes et al. 1971; Nelson 1981), Ellsworth Mountains (Ojakangas
and Matsch 1981; Collinson and Miller 1991), and Dronning Maud Land (Lindström 1995b, a) (Fig. 2A).

During the Lower Permian, the TAB was a narrow (~100 – 200 km-wide), trough-shaped basin that formed parallel and proximal to the Gondwanide margin of the East Antarctica Craton (Fig. 3B) (Collinson et al. 1994; Elliot 2013; Isbell 2015; Elliot et al. 2017). In the central Transantarctic Mountains and Victoria Land, glaciogenic strata occur within four sub-basins; the Ohio Range to the Scott Glacier (Horlick Sub-basin), the Amundsen Glacier to the Darwin Glacier area (Beardmore Sub-basin), south Victoria Land (SVL), and north Victoria Land (NVL) (Fig. 1B; Fig. 3) (Frakes et al. 1966; Isbell et al. 2008c). The TAB contains Devonian – Jurassic strata (Fig. 3) (Collinson et al. 1994; Elliot et al. 2017).

The nature of the TAB during the early Permian is not well understood, and hypotheses include intracratonic and extensional settings (Collinson et al. 1994; Isbell 2015; Elliot et al. 2017). Regardless of what process drove basin formation at that time, the TAB was a relatively high accommodation, trough-shaped basin during deposition of the Pagoda and Mackellar fms, but was overfilled during the deposition of the Fairchild Fm, before developing into a foreland basin during the Middle - Late Permian (Fig. 3) (Isbell et al. 1997b). The development of the TAB into a foreland basin is associated with the onset of diachronous volcanism in the early-mid Permian along the convergent, Panthalassan-Gondwanan plate margin (Fig. 3B) (Collinson et al. 1994; Veevers et al. 1994a; Elliot et al. 2017). Though there is no evidence of orogenic activity of Late Carboniferous to Early Permian age in the central Transantarctic Mountains or adjacent Marie Byrd Land, volcanic arcs and tectonic compression were occurring elsewhere along the super-continent’s Panthalassan margin during this time; including in eastern Australia, the Ellsworth Mountains, Thurston Island, and the Antarctic Peninsula (Fielding et al. 2001; Elliot 2013). This same margin was extremely active and subject to repeated, complex accretion events throughout the Paleozoic (Veevers et al. 1994b; Domeier and Torsvik 2014). During the early Jurassic, strata in the central Transantarctic Mountains were pervasively intruded by sills associated with Ferrar Group volcanism and the break-up of Gondwana (Elliot 1992).
SEDIMENTOLOGY AND STRATIGRAPHY OF LOWER PERMIAN STRATA

The glaciogenic Pagoda Fm and the post-glacial Mackellar Fm are the lower-most strata of the Victoria Group in the Beardmore Sub-basin. In the Beardmore Sub-basin, two regional unconformities exist: the Kukri Erosional Surface, which separates Cambrian plutons and Proterozoic metasediments from the Devonian – Jurassic Beacon Supergroup, and Maya Erosional Surface which separates Devonian(?)-aged terrestrial sandstones (Castle Crags and Alexandra fms) from a near-continuous Permian – Jurassic succession of the Victoria Group (Fig. 3, Fig. 4) (Collinson et al. 1994; Isbell 1999; Elliot 2013). Significant relief of up to ~150 m occurs on the unconformity underlying the Pagoda and Mackellar fms (Fig. 5) (Isbell et al. 1997a; Isbell 1999; Isbell et al. 2008c). The relief on these unconformities led to the Early Permian succession in the central Transantarctic Mountains having a wide range of thicknesses overlying a variety of basement lithologies (Fig. 4). The Pagoda Fm and the Mackellar Fm both fill and lap onto this relief, and the Mackellar Fm often overtops it (Fig. 5).

Rare palynomorphs and conchostracans suggest that the Pagoda Fm and Mackellar Fm are early Permian in age (Masood et al. 1994; Askin 1998; Babcock et al. 2002). Analyses of palynomorphs in the Mackellar Fm suggest periglacial or tundra-like terrestrial conditions around the TAB following glacier retreat (Masood et al. 1994; Askin 1998).

Pagoda Formation

Since their discovery, the Pagoda Fm and its equivalents throughout the Transantarctic Mountains have been unanimously interpreted as glaciogenic or glacially-influenced because their predominant lithologies are massive and laminated, sandy and silty diamicritites (Long 1964a; Lindsay 1970a; Coates 1985; Barrett et al. 1986; Collinson et al. 1994; Isbell et al. 2008c). Minor lithologies of the Pagoda Fm include conglomeratic sandstones, sandstones, mudrocks, and lonestone-bearing mudrocks (Fig. 4) (Isbell et al. 2008c). Besides being poorly sorted, evidence indicating a glacial origin for the
Pagoda Fm includes striated and polished basement surfaces, the prevalence of striated and faceted clasts, and a clear relationship between local basement composition and lithologies of large clasts within the diamictites (Lindsay 1969; Coates 1985). While the Pagoda Fm often overlies striated and polished basement rocks, it is occasionally underlain by weathered bedrock (Minshew 1967; Coates 1985; Isbell et al. 1997a). Weathered bedrock and non-glacial deposits at the base of some Pagoda Fm sections have been interpreted as evidence for ice-free conditions in the TAB during the early Permian (Barrett 1965; Minshew 1967; Coates 1985; Isbell et al. 2001; Isbell et al. 2008c). Detailed interpretations of depositional environments have only been made for a few Pagoda Fm localities (Waugh 1988; Isbell et al. 2001; Lenaker 2002; Long et al. 2008-2009; Koch 2010; Koch and Isbell 2013) and its equivalents in Victoria Land (Askin et al. 1971; Barrett 1972; Barrett and McKelvey 1981; Isbell 2010; Cornamusini 2017), Horlick Mountains (Frakes et al. 1966; Aitchison et al. 1988), and Ellsworth Mountains (Ojakangas and Matsch 1981; Matsch and Ojakangas 1991). Without exception, these analyses have invoked sub-aqueous, glacial-proximal depositional settings. This is in contrast to early surveys that interpreted diamictites as subglacially-deposited “tillites” (Coates 1985; Miller 1989; Isbell et al. 1997b).

Isbell et al. (2008c) separated the Permian glaciogenic units in the Transantarctic Mountains into basin-margin and basinal facies associations (Figure 5). Basin-margin successions are predicted to occur near basement highs and along basin margins, are relatively thin (<100 m), contain evidence for subglacial deformation and erosion, have deformation resulting from proglacial glaciotectonism, and small (m-scale) gravity-driven deposition. Basinal successions are thicker (100 - 500 m), have little-to-no evidence for subglacial processes, and are more likely to contain laminated diamictites, mudrocks, and larger (up to 10’s of meters) mass transport deposits. Based on its paleogeographic position and Pagoda Fm thickness, the Shackleton Glacier area is here predicted to contain the basin-margin facies.
Mackellar Fm

The Mackellar Fm is the postglacial facies in the central Transantarctic Mountains, consists of multiple 15 – 50 m-thick coarsening upward successions (CUS) of silt, to inter-laminated mudrock and sandstone, to medium-grained sandstones (Seegers 1996). These CUS are interpreted as deltas, delta fronts, and associated turbidites deposits that prograded across and filled the TAB from north to south (Miller and Collinson 1994; Seegers 1996; Miller and Isbell 2010; Jackson et al. 2016). In basin-margin areas, including the Shackleton Glacier region, the Mackellar Fm tends to be sandier, includes little true mudstone (Barrett 1965), and contain ice-rafted sediments immediately above the contact with the underlying Pagoda Fm (Seegers 1996; Isbell et al. 2008c; Flaig et al. 2016).

Where the Mackellar Fm overlies the Pagoda Fm the contact between the two is most often described as sharp and abrupt. This sharp contact has been interpreted as being the result of the rapid retreat of glaciers out of the Transantarctic Basin (Isbell et al. 1997a; Isbell et al. 2008c). The mineralogy of the Mackellar Fm and the overlying, fluvial Fairchild Fm indicates that this prograding delta/fluvial system has a similar provenance to the Pagoda Fm (Barrett 1968; Lindsay 1968; Waugh 1988). There is no consensus as to whether the Mackellar Fm was deposited in a lacustrine, estuarine, or marine environment (Miller and Isbell 2010; Jackson et al. 2016). Recent work by Jackson et al. (2016) and Flaig et al. (2016) in the Beardmore Glacier region suggests that ambient depositional conditions in the TAB during Mackellar Fm were dominantly brackish with large freshwater inputs. The salinity of the Mackellar Fm’s depositional environment is difficult to resolve because of its dearth of body fossils and the metamorphism experienced by the region due to pervasive Jurassic intrusions.

Equivalent Strata in the Scott and Amundsen Glacier Area

Mt. Weaver, one of the sites described in this study, is located at the head of the Scott Glacier, in an area between the Beardmore and Horlick Sub-Basins (Fig. 3C). The Weaver Fm is considered the
equivalent to the Mackellar Fm in the central Transantarctic Mountains due to their similar age,

stratigraphy, and lithofacies (Fig. 3) (Long 1964b; Minshew 1967; Long et al. 2008-2009). Where the Weaver Fm doesn’t overlie basement rocks, it is underlain by the Scott Glacier Fm, which is considered to be a lateral equivalent to the Pagoda Fm (Long 1964; Minshew 1967; Long et al. 2008-2009). Occurrences of pyrite and the trace fossil *Paleodictyon* in the Weaver Fm indicate that it was deposited in marine conditions (Minshew 1967; Collinson et al. 1994).

**STUDY AREA AND METHODS**

The sedimentary sections described in this paper were examined as part of the US Antarctic Program’s helicopter-supported Shackleton Glacier Deep Field Camp, located at the confluence of the Shackleton and McGregor glaciers, during the 2017 – 2018 austral summer (Table 1; Fig. 1B). These sections are located on the Mt. Butters Massif (MB-17, S84° 51.029' W177° 25.216'; MBSE-17, S84° 53.003' W177° 22.354') on the west side of the Shackleton Glacier, on the east face of Reid Spur (RS-18, S84° 47.035' E178° 46.680') of the Ramsey Glacier, and Mount Munson (MM-17, S84° 45.359' E173° 41.118') at the head of Barrett Glacier. The two sections at Mt. Butters are separated by approximately 2 km. A short section was also described and detailed photographs taken of Mt. Weaver (S86° 58.354' W153° 26.801') at the head of the Scott Glacier, but contrary to reports by Minshew (1967) no glaciogenic facies were identified.

Structural and sedimentary orientations were measured using a brunton compass, with the azimuth set to 000°. Measurements were corrected for magnetic declination using the NOAA Magnetic Declination Estimated Value Calculator (NOAA 2019). Orientations were corrected for structural dip and aggregate orientations calculated using Stereonet (v.11) software (Allmendinger et al. 2011; Cardozo and Allmendinger 2013). Some measurements from Mt. Munson were collected by JLI during the 1997-1998 deep field campaign to the Shackleton Glacier region.
Sites were selected for section descriptions based on accessibility, continuity of exposure, and preliminary observations from previous expeditions to the area. At MB-17 several shorter sections were measured to capture lateral variability. Sedimentological data (texture, grain shape, sedimentary structures, current directions, etc.) were logged in each section. Measured sections were placed in context using outcrop-scale photographs taken from helicopters (Fig. 5).

**LITHOFACIES AND FACIES ASSOCIATIONS**

The following descriptions are of facies associations that comprise the Pagoda Fm and lower-most Mackellar and Weaver fms in the Shackleton Glacier region (Table 2). The characteristics of distinct sediment grain sizes are consistent throughout the Pagoda Fm. Very-fine grained sandstones and shales are black in color, similar to the overlying Mackellar and Weaver fms. Fine- to medium-grained sandstones are generally quartz-rich with some lithic and potassium feldspar grains. Cobble and boulder-sized clasts are sourced from local basement lithologies (Fig. 1B), and include predominately phaneritic granitoids, with some gneisses, quartzite, and grey, fine-grained meta-sedimentary rocks. All sand-sized and coarser-grained material in the Pagoda Fm occurs in all categories of particle roundness (angular to well rounded), although finer-grained sands grains are typically better-rounded than medium- and coarse-grained sands. Striations occur on large clasts throughout the Pagoda Fm but are not common. The lack of striations is possibly due to the hardness of individual clasts (mainly granite, quartz, and feldspar).

**Basal Fine-Grain-Dominated Facies Association (BFG)**

This facies association (FA) is part of the Pagoda Fm, occurs only at the Mt. Butters site MB-17, and laterally varies in thickness from ~0m to 5.5 m. This FA includes four lithofacies: a basal conglomerate (BFG1), horizontally-stratified fines (BFG2), horizontally-stratified fines with outsized clasts (BFG2), and pervasively sheared fines (BFG4). The weathered, granitic basement rocks (BFG0) that
underlie this FA are also discussed in this section. Fossils are rare in this FA, but those that have been found include fresh-water ostracods, conchostracans, euthycarcinoid arthropods, plant fragments, and the trace fossil *Planolites* (Isbell et al. 2001; Babcock et al. 2002; Collette et al. 2017). This unit continues ~200 m laterally before it is pinched out by the overlying diamictite. Isbell et al. (2001) suggested that this FA was protected from subglacial by being on the lee side of a bedrock knob.

**BFG Stratigraphy**

Variations in the thickness of this FA are due to relief on the underlying bedrock as well as the erosional nature of the overlying contact. This FA is underlain by a nonconformity with an underlying weathered granite profile (BFG0) and is overlain by massive diamictites (MSD) that are characteristic of the Pagoda Fm. In the lower part of the succession (0 – 1 m), the basal conglomerate facies (BFG1) laterally interfingers with fine-grained facies (BFG2 and BFG3). The conglomerate is absent in the upper part of the section (1 – 5.5 m) and appears to lap onto and pinch out against a bedrock nob. Sedimentary units alternate between facies BFG2 and BFG3 in the upper part of the section. Facies BFG3 comprises the upper-most 20 -50 cm, directly below the contact with the MSD facies association.

**BFG0: Weathered basement**

A weathered profile in granitic bedrock underlies a sharp contact with the rest of the BFG facies association (Fig. 6A). The granite basement underlying much of the Shackleton Glacier Area is part of the Cambro-Ordovician Queen Maud batholith (Borg 1983; Wade 1986). The profile’s thickness is variable, but averages ~0.75 m. The weathered bedrock is only present where the BFG facies association is present, and is absent where diamictite directly overlies basement (Fig. 1B; Isbell et al. (2001)). The weathering is most intense in the 10cm beneath that contact and gradually decreases in intensity downward. In some areas, several centimeters of mobilized and redeposited, massive grus separates the weathered mantle.
from the overlying conglomerate (Fig. 6A). The weathering profile has limited chemical alteration of the granites abundant potassium feldspars.

The sharp contact between this weathering profile and the Pagoda Fm is a non-conformity separating the Cambro-Ordovician Queen Maud granitoids and the early Permian Pagoda Fm. The primarily physical weathering observed in this weathering profile suggests sub-aerial exposure of this bedrock surface in a relatively dry environment before the deposition of the overlying Beacon Supergroup strata (Migoń and Thomas 2002).

**Facies BFG1: Basal conglomerate**

This facies is a polymictic (in terms of clast lithology), clast-supported, sandy, bedded conglomerate that includes minor amounts of thinly-bedded coarse sandstone and laminated siltstone (Fig. 6A, 9B). This unit occurs at the base of the section and its thickness varies between 5 cm and ~1 m. Conglomerate clasts are angular to sub-rounded, and range in size from granule to ~10 cm. Clast compositions include quartz, granite, siltstone, feldspar, and a medium-grained, grey, meta-sandstone. The conglomerate’s matrix is very poorly sorted; grain size ranges from fine- to very coarse-grained sand. Matrix grains are predominately angular to sub-rounded, though some coarse-grained, quartz sand grains are well-rounded. Conglomerate beds are massive and 5 – 10 cm thick, and laterally discontinuous. Bed boundaries are horizontal but irregular. This facies generally fines upward. Beds fine due to a decrease in the abundance of clasts while the matrix composition remains constant. Relatively finer-grained conglomerate beds are more likely to be overlain with a siltstone drape. Thin beds of poorly-sorted sandstones occur in association with the conglomerate and have compositions similar to the conglomerate’s matrix. Polygonal syneresis cracks (McMahon et al. 2016), rare ostracods, conchostracans, and plant fragments occur in the silt laminae.
Sandstone beds just below the contact with BFG2 are fine- to medium-grained, well-sorted, contain symmetrical to slightly asymmetrical ripples, and have rare Planolites trace fossils. The contact between these sandstone beds and the overlying fine-grained facies is sharp.

The massive, poorly-sorted nature of the conglomerate beds suggests they were deposited through repeated gravity-driven, concentrated density flows (Mulder and Alexander 2001). This facies was most likely deposited subaqueously, as indicated by symmetrical (wave) ripples, arthropod fossils, and silt laminae draped over protrusive conglomerate clasts. The ripples in this facies are most likely wave-induced. Isbell et al. (2001) observed that this facies thins laterally and interfingers with the finer grained facies in this FA, BFG2 and BFG3.

**Facies BFG2: Horizontally-stratified fines**

This facies consists of thinly-laminated to thinly-bedded siltstone to fine-grained sandstone. The lithology color is dominantly black to dark grey (Fig. 6C). Laminae and beds are internally well-sorted. Bedding contacts are predominately sharp, horizontal, and laterally continuous. Irregularly-shaped laminae and symmetrically-rippled beds are common. Shallow, horizontal Planolites trace fossils occur along some bedding planes. Rare body fossils occur in this facies, including conchostracans, ostracods, euthycarcinoid arthropods, and plant fragments. This facies was deposited subaqueously and was likely the result of settling from suspension. Symmetrical ripples and irregular laminae were created through wave action. Fossils in this facies suggest a fresh-water, lacustrine setting (Isbell et al. 2001; Babcock et al. 2002; Collette et al. 2017).

**Facies BFG3: Horizontally-stratified fines with out-sized clasts**

This facies is similar to BFG2, but is more likely to be bedded (not laminated), trends towards fine-grained sand instead of silt, and contains out-sized clasts. The out-sized clasts are coarse-sand to granule sized particles, composed of quartz, and are angular to semi-rounded. Outsized clasts are not
confined to bedding surfaces, but are randomly distributed throughout the strata. The concentration of outsized clasts varies and is sometimes high enough to be classified as a clast-poor, stratified diamictite (Moncrieff 1989; Hambrey and Glasser 2003). Ripple marks are symmetrical and most often internally composed of medium-grained sandstone with fine-grained drapes.

The depositional processes of this facies was similar to BFG2, except that the setting was likely more proximal to higher-energy dispersal processes. The outsized clasts were most likely dropped-in because they are not confined to bedding planes, are randomly distributed, and occur in variable concentrations. Clasts can be dropped into low-energy settings from either rock falls or through ice rafting - either by icebergs, sea ice, lake ice, or anchor ice (Gilbert 1990; Domack and Powell 2018). Since the clasts are well-sorted and of uniform lithology, they are not likely to be sourced from rock falls or glacial sediments. The outsized clasts in this facies were likely transported by lake ice, with sorting taking place before or during the sediment’s emplacement on or in the ice.

**Facies BFG4: Pervasively sheared fines facies**

This facies occurs in the uppermost 0.2 – 0.5 m of this facies association. The lower contact is sharp to gradational, undulatory, and is conformable with BFG2 and BFG3 facies. This facies’ upper contact is sharp and undulatory with a massive, sandy diamictite (facies MSD). Cobble- and boulder-sized clasts from the overlying diamictite penetrate through the upper contact of this facies. The grain-size composition of this facies is similar to that of BFG3, but with a higher outsized clast content.

The sediments in this facies are denser than underlying sediments composed of the same lithologies. Homogenous sediments are separated into decimeter-scale lenticular bodies by curvilinear fractures (Fig. 6M, N). All observed fracture surfaces are covered with slickensides (Fig. 6F). Stiations on the slickenside surfaces appear to have been made by the movement of outsized clasts through the surrounding fine-grained sediment. Where the lenticular bodies are not well-developed, this facies has a
highly fissile structure (Fig. 6E). Fissility is especially well-developed in the decimeter directly below this facies' upper contact.

Due to the identical nature of their lithologies, this facies was likely deposited in the manner of facies BFG3 and was later subject to soft-sediment deformation. Slickenlines along the surfaces between the curvilinear surfaces, the common fissile structure, and the massive nature of this unit suggest that the facies experienced pervasive simple shear. The fissility immediately below this facies' upper contact suggests that the strain magnitude was greatest there, and that strain magnitude generally decreased downward. This trend in strain magnitudes suggests that the deformation was caused during deposition of the overlying massive diamictite facies (Facies Association MSD). The deformation observed in this facies is consistent with subglacial deformation of a soft bed (Boulton and Hindmarsh 1987). Clasts protruding down into this facies suggest that plough or lodgment of clast occurred at some point during subglacial activity.

Strain conditions resulting in sheared horizons similar to BFG4 can also be formed beneath mass-transport bodies (Sobiesiak et al. 2018). Deformation of facies BFG4 by a mass-movement is a less likely explanation than subglacial deformation for several reasons. The contact with the overlying diamictite is generally flat, sharp, and lacks any loading structures. This suggests that the compaction and deformation of BFG4 were not due to rapid loading, as is characteristic of many types of mass-transport deposits. Loading structures would especially be likely, if mass-transported sediments were deposited over weak, water-logged mud, as was likely the case at the BFG facies association upper contact.

**BFG Depositional Environment**

The lateral interfingering of the units in this FA indicate that they were part of the same depositional environment. Therefore, the over-all fining-upward of the facies association, first within the
basal conglomerate (BFG1) and then from the conglomerate to laminated fines (BFG2 and BFG3),
suggests a deepening-upward trend.

Sediments in this FA were deposited in a lacustrine environment, with little to no direct
depositional influence from glaciers. The coarse, poorly-sorted sediments (BFG1) at the base of the section
were most likely deposited as grus/scree and by gravity-driven processes, indicating that the sediment
source is proximal relative to the overlying fine-grained facies. The presence of non-granitic grains in the
conglomerate indicates that some of the sediment grains were likely transported from outside the
immediate area. The fine-grained, laminated, occasionally wave-worked facies (BFG2 and BFG3) were
most likely deposited through settling from suspension at a depth below average normal wave base.
Alternatively, the presence of sea or lake ice would have limited any wave activity. Facies BFG4 was
deposited in the same way as BFG3 but was later consolidated and deformed as a glacier overrode these
deposits.

Cold conditions prevailed during the deposition of this FA. The climate during the time of
deposition was at least seasonally cold enough to produce lake or sea ice, as indicated by the small out-
sized clasts in BFG3 and BFG4. The depth of weathering and apparent degree of chemical weathering
underlying this FA (BFG0) is similar to pre- or inter-glacial, Quaternary grus/saprolite profiles described
in Scandinavia, Canada, and Antarctica (Olvmo et al. 2005; Sugden et al. 2005; Goodfellow 2007). These
weathering profiles can take anywhere between several thousand to a few million years to develop, and
can survive multiple glacial events, so long as the glacier was incapable of subglacially eroding bedrock
in a given location (Migoń and Thomas 2002). Therefore, placing significant temporal or climatic
constraints on the development of grus is unwise. However, these weathering profiles do indicate that

glacial cover was not constant in the Shackleton Glacier area throughout the LPIA. These inferences are
consistent with the TAB’s position at polar latitudes during the Late Carboniferous and Early Permian,
which would likely yield a generally cold climate (Lawver et al. 2011; Boucot et al. 2013; Torsvik and Cocks 2013).

In this succession, subglacially deformed sediments (BFG4) are not preceded by the deposition of sedimentary systems characteristic of temperate glaciers. The advance of a temperate glacier over the area would have been preceded by the deposition of proglacial sediments characteristic of such a system. The proglacial sediment system of a temperate glacier would most likely include glacially derived, ice-rafted debris in ice-distal fine-grained sediments (like BFG 2-3), ice-proximal to ice-medial plume sedimentation, and/or ice-proximal deposits such as high-discharge ice-contact systems like fans, sandur, or deltas (Matsch and Ojakangas 1991; Hambrey and Glasser 2012; Kurjanski 2020). The lack of evidence for temperate, proglacial deposits between this FA and the overlying massive diamictite (facies MSD) suggest that the advancing glacier was not discharging large amounts of meltwater, sediment, or debris-rich icebergs before its advance over this location (Atkins 2013). This evidence suggests that the glacier did not have a temperate thermal regime when it advanced over the area, and was therefore either polythermal or cold-based (Hodson and Ferguson 1999; Hambrey and Glasser 2012). Of course, the possibility also exists that proglacial sedimentary systems characteristic of temperate glaciers could have been deposited but were later eroded.

**Massive Sandy Diamictite Facies (MSD)**

This diamictite facies is the characteristic facies of the Pagoda Fm. Similar lithologies occur throughout the Transantarctic Basin, though they are often not as thick or interrupted as they are in the Shackleton Glacier region (Fig. 4).

The thickness of this facies ranges from 3 m at Mt. Munson (MM-17) to 73 m at Mt. Butters (MB-17C). Where exposed, the lower contact overlies either a striated and polished unconformity with the Queen Maud Batholith (MM-17 and MBSE-17) or subglacially deformed lake sediments (BFG4; MB-17C).
Upper portions of this facies are more likely to grade laterally into other facies, including Laminites Sands (LS) facies and Heterolitic Agglomeration (HA) FA. The upper contact of MSD is sharp or erosive with current-transported facies, including Cross Bedded Sandstone (CBS), Upper Fine-Grained Dominated (UFG), and Mixed Heterogeneous (HA) facies associations. Where the Mixed Heterogeneous includes stratified diamicrites, the contact between MSD and HA is gradational.

MSD Description

This facies is a massive to crudely-bedded, clast-poor to clast-rich diamicrite with minor amounts of sorted sands and gravels (Fig. 7). Clast abundances fluctuate throughout the succession. Some intervals are sufficiently clast-poor that they could be classified as muddy sandstones with dispersed clasts (<1% clasts) (Moncreiff 1989; Hambrey 1994). Clasts range from pebble-sized to 4 m in diameter. Clast compositions includes granite, vein quartz, gneiss, and fine-grained meta-sandstone. Clast shape ranges from rounded to angular. Faceted clasts are common, but bullet and striated clasts are rare. The matrix is very poorly sorted, with sizes ranging from muds through granule-sized grains. Matrix grain-size distribution remains constant within and between outcrops of this facies, though mean matrix grain size increases slightly in clast-rich sections relative to clast-poor sections.

This facies is generally massive; however, crude and chaotic bedding sometimes occur. Zones of clast-poor and clast-rich diamicrites can be anywhere from 1 – 30 m thick. The transitions between these zones within the diamicrite are gradational. The thicknesses of the gradational zones are on the meter-scale and are difficult to perceive in outcrop. One could not reliably follow a zone laterally for more than a few tens of meters, especially if that zone was relatively thin (<10 m). Zones of diamicrites with similar clast concentrations cannot be correlated between outcrops; the distribution of clast-poor and clast-rich diamicrites appears to be unique to each locality. Where bedding can be discerned within the diamicrite, the beds are 2 -10 cm thick, laterally discontinuous, and internally massive.
Sandstone and/or conglomerate bodies are moderately rare within the diamicrites. These bodies are poorly- to moderately-well-sorted; the sands usually better sorted than the gravels. The grain compositions within these bodies are similar to sand- to gravel-sized grains in the diamicrite. Most sand bodies are massive but can contain layers of distinct grains sizes. These layers, however, are highly deformed. Bodies of sorted sand and/or gravel occur in two distinct sizes. Small bodies consist of sorted sediments that most often occur as irregularly-shaped, horizontally elongate lenses, “whisps”, and boudinage-like bodies that are up to 20 cm thick and 100 cm long (Fig. 7A, B, E, F). Such structures may occur alone or, more frequently, within bands and zones up to 1 m thick. Occasionally, flame-like structures made of sand and/or gravel, 1 to 2 m thick are also present. Sheath folds within the diamicites are common in the upper portion of this facies at MBSE-17. Thicker bodies of well-sorted sand or gravel range in thickness from 1 to 3 m and are laterally discontinuous. These bodies have the same grain size distributions as the smaller bodies. The large sand bodies are typically massive, but where stratification does exist in the sands and/or gravels, the beds are largely deformed through soft-sediment deformation and display water-escape structures. The large sand bodies thin laterally and have over-turned folds on their thicker ends. In one instance, the main sand body was accompanied by smaller sand bodies that trailed off away from its thick end (comet-like structure) (Fig. 7F; Sandstone with dispersed clasts). The lower contact between these bodies and the surrounding diamicite has a slope of 30° to 35° above horizontal. Occasionally such contacts overlie smaller sand lenses, sheath folds, “whisps”, and boudins.

Though the diamicite is generally massive or crudely bedded, it contains some sedimentary structures. Within the diamicite, sedimentary structures include ruck structures beneath large clasts (Fig. 7D), thinning of beds over larger clasts (Fig. 7H), and thrust faults (Fig. 7F). Boulder and cobble beds occur in the lower portion of this facies at both Mt. Butters sites (Fig. 7G). In all lag deposits, the boulders were not striated, faceted, polished, or uniformly oriented. These lags are defined by their lateral extent across the visible outcrop.
This facies is most likely glaciogenic or glacially influenced. Evidence for glacial transport of sediment in this facies includes striations and polish on basement granites where diamictites rest directly on granite (MBSE-17 and MM-17), the very poor sorting of sediment in the system, and the presence of angular to rounded grains of all sizes, faceted and striated clasts, and large boulders composed of local basement lithologies. In glacially-influenced settings, massive diamictites may be the result of subglacial till deposition (Evans et al. 2006), grounding-zone-related processes in the form of a wedge (Batchelor and Dowdeswell 2015; Demet et al. 2019; Dietrich and Hoffmann 2019), morainal bank (Eidam et al. 2020), or fan (Powell 1990; Benn 1996), settling from suspension and rain-out from icebergs and iceberg scouring (Dowdeswell et al. 1994; Lisitzin 2002), mass-transport deposits (Rodrigues 2019), or debris flows (Powell and Molnia 1989). To differentiate between these potential depositional environments, the depositional processes contributing to this facies must be carefully considered.

We can narrow down this list through the inference that this facies was more likely deposited subaqueously than subaerially. Subglacial deformation of the BFG facies likely occurred subaqueously for several reasons. There is no evidence for subaerial exposure prior to the deposition of MSD. In most of the sections measured in this study the upper contact of this facies was conformable with, and in some cases gradationally transitions into, the overlying, subaqueous facies’ (HA Facies Association). At section RSP-18, the massive diamictite is interstratified with turbidites, a subaqueous process (see LS facies interpretation). Additionally, there are no characteristics in this facies that suggest subaerial exposure (e.g. paleosols, wind-transported sediments).

The lack of clear or laterally continuous stratification, as well as a wide range of grain sizes in the matrix throughout this facies, indicate that the deposition of these sediments was not principally controlled by “sorting” processes such as currents and/or low-density flows. The massive nature of beds
may be a primary depositional feature, as is the case in subglacial deposits, supraglacial debris
(Schomacker and Benediktsson 2018), or plume sedimentation. Alternatively, this facies may also be the
result of secondary processes, such as re-deposition by high-density gravity-driven processes/flows
(Wright and Anderson 1982) or homogenization due to iceberg scour (Dowdeswell et al. 1994),
dewatering (Collinson and Mountney 2019), or bioturbation (Svendsen and Mangerud 1997; Murray et
al. 2013). Where crude stratification does occur, there are indicators that settling from suspension may be
a key depositional process. Ruck structures beneath large clasts suggest those clasts were dropped into
the surrounding diamictite from ice rafted debris (Fig. 6D) (Thomas and Connell 1985). The thinning of
diamictite beds over large clasts suggests that there was some component of settling from suspension in
their depositional process (Fig. 6H). In glacial settings, this most often occurs as plume sedimentation.
Boulder/cobble beds do not contain uniform oriented, bulleted, or striated clasts. This suggests that the
boulder beds are lags; that they formed due to winnowing (Eyles 1988). Therefore, the depositional
environment is likely not a grounding line fan or some other current-dominated system.

The subglacial deposition of this facies cannot be wholly ruled out, at least in the lowermost parts
of the unit (see BFG Interpretation). The striations and deformation underlying the lower contacts of this
facies suggest subglacial processes were active before, and possibly during, the deposition of this
diamictite. The presence of these features suggests that the basal portion of the diamictite may be
subglacial in origin; a tillite. A tillite interpretation for the MSD facies is supported by the massive nature
of the diamictites and the presence of large (up to 4 m) boulders of the local basement lithology.
However, most other sedimentary structures in this facies (e.g. ruck structures) do not support the
hypothesis that the diamictite was largely deposited subglacially. There is no strong evidence for
subglacial deformation (glacier grounding) above the base of this facies.
Small, deformed, sorted sediment bodies occur in many depositional settings alongside
diamictites, including in till (Kessler et al. 2012) and proximal glaciomarine sediments (Domack 1983;
Sheppard et al. 2000). Therefore, they cannot be used to differentiate between depositional processes.
Depending on their depositional context, small sand bodies are proposed to be sourced from ice-berg
dumps, winnowing due to de-watering, or from incorporating subglacial sediment into till through
freeze-on. The small sand and gravel bodies all indicate pervasive simple shear (whisps/stringers,
boudinage, and sheath folds) or loading (flame structures). In at least one case, shearing was associated
with an overlying, thicker sandstone body (Fig. 7F). If the diamictite experienced similar strain
conditions, representative structures would likely be impossible to observe in outcrop, due to the
homogenous natures of the facies. The large sand bodies are most likely mass-transport deposits that
underwent slumping or non-turbulent flow (Posamentier and Martinsen 2011; Rodrigues 2019).

The combination of processes that most likely contributed to the deposition of this facies
(subglacial till deposition, iceberg rain-out, plume sedimentation, iceberg scouring, and mass-transport)
is most likely to occur in a glacier-proximal to glacier-intermediate, with water depths comparable to a
continental shelf (Licht et al. 1999; Powell and Cooper 2002; Powell et al. 2009). Such a setting is the best
explanation not only because it can accommodate all the required depositional processes, but because it
can best explain the stratigraphic trends of this facies. Plumes emitted from englacial and subglacial jets,
as well as icebergs, were likely the primary sources of sediments in this system. Variations in iceberg
calving and fluctuating glacial hydraulic systems may explain the variation of matrix grain size and clast
abundance throughout the facies. Glacial hydraulic systems and icebergs capable of producing sufficient
sediment to create this facies are characteristic of temperate glaciers (Matsch and Ojakangas 1991;
Hambrey and Glasser 2012; Kurjanski 2020).
In the Cenozoic, similar depositional models have been proposed for massive to crudely-bedded diamictites Yakataga Fm, Alaska (Eyles and Lagoe 1990), cores collected from the Weddell Sea, Ross Sea, and George V regions of Antarctica (Anderson et al. 1980; McKay et al. 2009), as well as St. George’s Bay, Newfoundland (Sheppard et al. 2000). Of these examples, St. George’s Bay is likely the most analogous to the Shackleton Glacier region during the Permian, since it is not an open shelf setting, but an embayment whose topography is controlled by much-older basement rocks (Batterson and Sheppard 2000; Shaw 2016).

Plume sedimentation is not likely to occur where the glacier water is denser than the ambient water in the depositional environment; a condition associated with lacustrine conditions and resulting hypopycnal flows. Therefore the deposition of this facies most likely occurred in marine or estuarine conditions (Powell 1990). The time frame in which the deposition of this facies occurred is difficult to infer, especially without any evidence within the facies for glacier grounding above its lower contact.

Sedimentation rates in glaciomarine settings are highly variable, even for the same glacier, and strongly depend on glacier conditions and proximity to the ice front (Hallet et al. 1996). Rates of accumulation will also depend on physiography of the depositional area (e.g. fjord vs. open shelf). Accumulating ~100 m of glaciomarine diamictite could take anywhere from a few years (Cowan and Powell 1991) to a few millennia (Partin and Sadler 2016; Domack and Powell 2018).

**Laminated Sands Facies Association (LS)**

This is a sandstone facies association that occurs only at site RSP-18 and is interstratified with facies MSD. This succession is ~17 m thick and laterally continuous across the outcrop. Its lower contact is erosional above MSD and its upper contact was covered. Internally, the FA consists of fining-upward packages 3-5 m thick (Fig. 8). The packages have erosional bases, relief on erosional surfaces

---

Ives and Isbell SHG Glaciogenic Lithofacies J. of Sedimentary Research
consists of coarse to fine-grained, well-sorted sandstones. The dominant facies in this FA is fine- to medium-grained sands that occur in thin, planar beds with primary current lineations. Coarse-grained sandstones at the base of some packages are trough cross-bedded. Fine- to very-fine grained sandstones are laminated, or thinly-bedded with unidirectional cross-laminations and/or ripples. The sandstone is quartz-rich and grains are sub-angular to rounded. The larger sand grains (medium to coarse) are better-rounded than the smaller sand grains (fine to medium). Pebbles up to 8 cm in diameter occur at the base of some troughs. The uppermost portions of fining-upward packages sometimes include fine- to very-fine-grained black-colored sandstone.

**LS Interpretation**

This facies is most likely the result of a series of non-cohesive density flow events, either in the form of high-density turbidites and/or a transitional concentrated density flows (Mulder and Alexander 2001). Since these turbidites are interstratified with facies MSD, they are most likely the distal or medial portion of an ice-contact fan or delta (Lønne 1995; Dowdeswell et al. 2015).

**Heterolithic Agglomeration Facies Association (HA)**

This FA (Fig. 9) occurs at both Mt. Butters sites above a gradational contact with the massive diamictite (MSD), and below a sharp, erosional contact with the cross-bedded sands facies (CBS) at MBSE-17 and a sharp, horizontal contact with upper fine-grained facies (UFG) at MB-17 (Fig. 10). The lower part of the facies association begins as interbedded, discontinuous bodies of deformed, sorted sands and gravels (HA2) within stratified or massive diamictites (HA1). Undeformed, moderately-well-sorted, stratified sandstone bodies with a range of grain-sizes (HA3) occur in the middle of the succession and eventually become the dominant facies near the top of the succession. This FA ranges in thickness
from 1 – 15 m. Lithologies in this facies are inter-stratified. Coarse-grained lithologies range in thickness from 0.25 -1 m.

**Facies HA1: Stratified diamictite**

This facies is a clast-rich, sandy diamictite (Fig. 9D, G, I; Fig. 10). This diamictite facies is similar to the massive diamictite facies (MSD), but is more consistently stratified, is more clast-rich, and the matrix is better sorted. The matrix in this facies is moderately-well Sorted to well Sorted. Matrix grain sizes range from medium to very-fine sand. The mean matrix grain size varies between beds. Most beds have a mean matrix grain size of medium-sand, but some beds have a dominantly very-fine-grained matrix. Clasts in this facies are angular to sub-rounded. The beds in this facies are 3-7 cm thick, are horizontal, laterally discontinuous, and have sharp contacts. The distribution of clasts is random and unrelated to bedding planes. This facies has a gradational lower contact overlying the massive diamictite (MSD) facies. Vertical and lateral contacts with other facies in this FA (HA3 and HA2) are most often sharp, sometimes loaded, deformed, or erosional. This bedded diamictite is frequently interbedded with facies HA2. Strata in this facies adjacent to contacts with facies HA2 often display soft-sediment deformation.

Similar to facies MSD, this facies was likely deposited in a glacier proximal glaciomarine setting, but was dominated by plume sedimentation and ice-berg rainout (Eyles 1987; Licht et al. 1999; Powell and Domack 2002; McKay et al. 2009). The consistency of the stratification in this facies compared to the MSD facies suggests that the sediment in this facies was not as frequently subject to remobilization, either through gravity-driven transport or iceberg scour. The better sorting of the matrix in this facies relative to the massive diamictite facies indicate that sediment-sorting processes were more active than during the deposition of MSD. In a glacial-proximal setting, this likely means that turbulence kept fine-grained sediment suspended in the water column and did not allow it to settle out. The clast-rich composition
and thin nature of the beds also suggest that depositional processes were relatively constant, compared to the high variability of clast-contents in facies MSD. Loaded, deformed, and erosional contacts within this facies, and between this facies and other facies in this FA, indicate that this facies was weak and therefore water-saturated, that sedimentation was rapid, and that they were deformed by gravity-driven deposits (Fig. 9D, F; Fig. 10).

**Facies HA2: Chaotic Heterolithic**

Grain size ranges from conglomerate to very-fine-grained. Fine- to medium-grained sandstones tend to be well-sorted, while coarse-grained sandstones and conglomerates are poorly sorted. Grains and clasts are sub-angular to rounded. Medium- and coarse-grained sandstones are more likely to be less well-rounded and less well-sorted than finer lithologies. Well-sorted sandstones are quartz-rich. Grains in coarse-grained sandstones and conglomerates contain a variety of lithologies including granite, vein quartz, gneiss, and fine-grained meta-sandstone. The medium- and coarse-grained sandstones occur more frequently than finer-grained lithologies. Very fine- and fine-grained lithologies have a black or grey color. The fine- to coarse-grained sandstones are quartzitic. Bedding in this facies is often massive, and pervasive soft-sediment deformation is common (Fig. 9E, F). Primary sedimentary structures are sometimes preserved, but this is rare and only occurs in a small area of any given bed. Secondary structures in this facies include fold noses, boudinage, faulting, and other simple shear structures above and below contacts (Fig. 9D – G), and ruck structures associated with rare outsized clasts (Fig. 9I).

Beds in this facies are laterally discontinuous (Fig. 9G and Fig. 10). The thicknesses of sandstone bodies are laterally inconsistent and range in thickness from 0.5 m – 2 m. Widths of sandstone bodies range from ~1 m to outcrop scale. Sandstone beds may be interbedded with one another, but are dominantly interbedded with the surrounding either massive of stratified diamictite facies. Sandstone bodies are irregularly shaped, but generally have planar to lenticular shapes. Lower and lateral contacts
are deformed, sharp, or erosional, and often show evidence of soft-sediment deformation (Fig. 9D, F).

Upper contacts are sharp and conformable. Contacts between sandstone bodies are erosional or loaded/deformed.

The sandstone bodies in this facies are most likely the result of mass-transport, gravity-driven processes (Posamentier and Martinsen 2011; Sobiesiak et al. 2018; Rodrigues 2019). Preserved primary sediment structures in some of these bodies indicate that sediment sorting due to current transport likely occurred before the remobilization and final deposition of these sediments. Irregular lateral contacts between this facies and the two diamictites facies (HA1 and MSD) indicate that mass-transported bodies were deposited into pre-existing diamictites. The deposition of mass transport deposits (MTD) into the diamictite indicates that diamictite deposition was contemporaneous with MTD emplacement. This also suggests that the diamictite was weak (highly susceptible to deformation), likely due to high pore-water pressures/lack of consolidation.

**Facies HA3: Stratified heterolithic**

This facies is comprised of coarse- to very fine-grained sandstones (Fig. 9A-C). The medium- and coarse-grained sandstones occur more frequently than finer-grained lithologies. The sandstones are well- to very-well-sorted and sand grains are sub-angular to rounded. Medium- and coarse-grained sandstones are more likely to be less well-rounded and less well-sorted than finer lithologies. Very fine- and fine-grained lithologies have a black or grey color. The fine- to coarse-grained sandstones are quartzitic.

Medium- to coarse-grained sandstones are thickly laminated to bedded. Common sedimentary structures in medium- to coarse-grained sandstones include planar cross beds, trough cross beds, climbing ripples, 3D ripples that are asymmetric or climbing (Fig. 9A). Rare sedimentary structures include hummocky and swaley cross stratification and symmetrical ripples with bundled upbuilding (Fig. 9B and 12C). Variations in sedimentary structures within a unit of the same lithology is common.
Coarser sands are occasionally massively bedded or contain laterally discontinuous sand or gravel lenses. Trough cross-beds occasionally have pebbles at the bases of troughs. Very fine and fine sandstones are laminated or thinly bedded. Sedimentary structures in fine-grained lithologies include uni-directional cross-laminations/ripples, some flaser bedded cross-laminated units, and climbing ripple laminations. Lithologies of all grain sizes also had minor amounts of soft-sediment deformation, including dewatering structures, minor folds, and loading. At the upper contact of the Heterolitic Agglomeration FA in section MB-17 (Fig. 14), these sandstones had at least five large, shallow, east-west oriented grooves (Table 3) that occur within a massive, well-sorted sandstone (Fig. 9H). All grooves were ~1-2 m wide and 10 cm deep. Berms ~10 - 20 cm high surround the grooves on their long sides. All of the grooves gradually shallow towards the west (215˚), and two of the groves had prow-like berms at their eastern terminus. Strata in this facies are laterally continuous within the outcrop. Sandstone bodies are generally wedge-shaped and thicken in the direction of flow. Erosional surfaces are common within the facies. This facies has an erosional lower contact with both diamicrites facies (MSD and ME3), and a sharp, conformable contact with overlying UFG facies association.

The sedimentary structures in this facies indicate that current-dominated transport and deposition occurred, followed by syn- or post-depositional slumping of some deposits. The wide range of grain sizes and sedimentary structures indicate a sediment source with a wide range of grain sizes and huge variations in current velocities during deposition. Common sedimentary structures, such as planar cross beds, trough cross beds, and asymmetrical ripples suggest unidirectional, relatively high-velocity currents. Climbing ripples suggest unconfined flow. Flaser bedding, as well as abrupt changes in grain size and sedimentary structures within and between beds, indicates that current velocities were highly variable and fluctuated. Rare soft deformation suggests high pore-water-pressure during rapid deposition. Rare hummocky and swaley cross-stratification and bundled, up-building symmetrical ripples are most likely formed under oscillatory flow conditions created by surface waves, possibly
during reworking by storms (Reineck and Singh 1980; Dumas and Arnott 2006). Their rare occurrence
and inter-stratification with fine-grained sediments suggest that these wave features formed below
normal wave base. The shape of the grooves on the upper contact of this FA at site MB-17 and their
position in the upper contact of a massive (homogenized) sandstone suggest that these features are
iceberg keel marks (Dowdeswell et al. 1994; Vesely and Assine 2014).

HA Depositional Environment

The three facies in this FA represent a complex depositional environment that is characteristic of
subaqueous, glacier-intermediate to proximal settings in front of the terminus of temperate glaciers.
Evidence for the glaciogenic origin of this FA includes ruck structures (representing ice-rafted debris),
 iceberg keel marks (facies HA3), the very poor sorting of sediment in the system, and the presence of
angular to rounded grains of all sizes, bulleted and striated clasts, and large boulders composed of local
basement lithologies.

The bedded diamictite (facies HA1) was deposited primarily through plume sedimentation, and
is the dominant, or “background”, sedimentation type in this FA. The gradational contact between facies
MSD (massive diamictite) and facies HA1 suggests a gradual shift in depositional environments between
the two. Sediment composition is consistent between the two diamictite facies, suggesting that the
sediment source did not change, but that upward, the depositional environment shifted. The deposition
of both diamictites was likely controlled by the same processes (i.e. plume sedimentation, iceberg rain-
out, iceberg scouring, and mass-transport) but to different degrees. The variability in clast abundance and
consistent muddy-sand matrix suggests that the subaqueous portion of facies MSD was relatively distal
to the glacier margin. The bedded diamictite of the FA, facies HA1, is likely to represent a more glacier-
proximal setting of the same depositional system. More abundant clasts (more consistent iceberg
sedimentation) and better-sorted matrix (grain-size fractionation caused by plume turbidity), suggest that this facies was more glacier-proximal.

The sandstone facies in this FA (facies HA3 and HA2) most likely represent the medial portion of a grounding-line fan(s) system (Powell 1990; Lønne 1995; Dowdeswell et al. 2015). In facies HA3, the high-velocity, unidirectional current transport combined with abrupt changes in grain size (i.e. current velocity), unconfined flow, and interstratification with the bedded diamictite (facies HA1) are characteristic of grounding-line fans. The gravity-driven transport of facies HA2 sandstone bodies were likely derived from deposits similar to (or the same as) facies HA3. “Shedding” of sediments is also characteristic of the rapid sedimentation in grounding-line fan systems (Benn 1996; Powell and Alley 1997; Lønne et al. 2001). Intense, ductile deformation and loading along contacts throughout this FA indicate that all facies were water-saturated, unconsolidated, and generally had the consistency of soup, suggesting rapid deposition and therefore prone to resedimentation (Fig. 10B).

The wave reworking of some sandstone beds, as indicated by hummocky and swaley cross-stratification and bundled, up-building symmetrical ripples, suggest that this depositional environment was occasionally subjected to surface-wave activity (below normal wave base). These features indicate that there was not perennial ice cover during the deposition of this FA. This evidence for wave reworking may also suggest a similar origin for the boulder lags in facies MSD.

Where this FA is well-developed in outcrop at Mt. Butters (sites MB-17 and MBSE-17), the succession has a general coarsening and increase in sorting upward. This trend association indicates the increase in the proximity of the energy and sediment source, either through progradation of the grounding-line system or advance of the glacial front, and/or a decrease in accommodation through time.
Cross bedded Sandstone Facies (CBS)

CBS Description

This facies occurs at Mt. Butters section MBSE-17, and consists of an erosion-based, laterally extensive channel-form sandstone body 10-30 m thick and several hundred m wide that cut into a laterally continuous thick sandstone sheet at the top of a coarsening upward succession of the HA (1-3) facies association (Fig. 10; Fig. 11). The sandstone body is laterally continuous across outcrop MBSE-17 but is not present at section MB-17, which is ~2 km north (Fig. 1B). The basal CBS erosional surface has a relief of up to 10 m. This facies’ upper contact with the overlying UFG facies association of the Mackellar Fm is sharp and horizontal.

This facies occurs within a multi-storied, multi-lateral sand-filled channel-form bodies displaying non-sequential, lateral compensational stacking patterns (Fig. 11). Individual channels are m-scale thick and 10s of m wide, trough-shaped in cross section, and are filled by either vertical or downstream accretion dipping in an easterly direction. Channels are truncated by the base of overlying channels bodies. Channel stacking is non-sequential and disorganized, with some aggradation. This facies is comprised of well-sorted, medium- to very coarse-grained quartz arenite sandstone, with minor occurrences of conglomerate lenses and beds (Fig. 11). Mudrocks were not observed within the sandstone body. Very rare pebble- and small-cobble-sized clasts occur throughout the sandstones. Those clasts have similar lithologies to clasts observed in both diamicite facies (HA1 and MSD). Sedimentary structures are almost exclusively 0.15 to 1.5 m thick sets of low-angle stratification and trough cross-beds (Fig. 11A).

Thin beds with asymmetrical ripples also occur but are rare.

Adjacent to the described section (MBSE-17), the edge of the channel-form body appears to extend across the top of the underlying strata as a wing-like extension (Fig. 10A). Channel-form sandstone bodies occur within the wing. These bodies appear to transition laterally into other thick
sandstones with channelized bases. Most notably, the contact between the channelized sandstones in the wing and the underlying HA facies appears to be sharp.

**CBS Interpretation**

The CBS facies in the Pagoda Fm was deposited by strong, tractive, confined flow as indicated by the occurrence of the basal erosion surface and the internal channel bodies filled by down-stream accreting bar forms and cross-stratification. The trough-shape of the internal sandstone-filled channels and their multistoried and multilateral characteristics suggest that the channels were stationary during flow within the channels and that they did not migrate until channel switching occurred and new channels formed as older channels filled and were abandoned (cf. Friend 1983). The occurrence of low angle stratification and m-scale trough cross beds organized into down-stream accreting bodies with massive bedded to lenticular gravels suggest high flow velocities. Such features are characteristic of highly dynamic systems where aggradation within channels likely forced channel switching to adjacent areas on the depositional surface. The occurrence of this facies association on top of unconfined coarsening-upward HA (1-3) facies association and the occurrence of wings that appear as a continuation of the HA (1-3) coarsening upward succession suggest that the CBS facies formed as part of a HA-CBS larger-scale dispersal system. The presence of wings also suggests that parts of the CBS system were unconfined. Together, these patterns are most characteristic of an unconfined, distributive setting (Funk et al. 2012).

Whether this facies was deposited subaerially or subaqueously is unclear. However, the CBS sandstone body does not contain evidence for shallow water wave reworking, pedogenesis, or subaerial exposure; whereas, facies both below and above this facies represent subaqueous deposition below normal wave base, and likely below storm wave base. Therefore, a subaqueous setting seems likely.
Possible depositional environments include: 1) channelization on a delta top (ice-contact, glacio-fluvial, or non-glacial delta distributaries), 2) channelization related to turbidites, or 3) the proximal to medial portion of a subaqueous fan. The absence of delta foreset beds rules out delta progradation, and the abundance of high discharge bodies and absence of graded beds, sole marks, and reactivation structures suggest near continuous flow during deposition rather than episodic flow by turbidites. This unit is similar to some grounding-line fan systems that authors have called sub-aqueous outwash fans (Visser et al. 1987; Thomas and Chiverrell 2006; Rose 2018). Additionally, this sand body is dissimilar to those previously described in both the Pagoda Fm (Lindsay 1969; Waugh 1988; Koch and Isbell 2013) and the Mackellar Fm (Miller and Frisch 1986; Flaig et al. 2016), with the exception of a sand sheet interpreted to be a grounding-line fan in the Pagoda Fm-equivalent Metschel Fm of Southern Victoria Land (Isbell 2010) (Fig. 3).

_upper fine-grain-dominated facies association (uFG)_

uFG stratigraphy

The facies described here is the base of the Mackellar Fm in the Shackleton Glacier region and the Weaver Fm at Mt. Weaver. While the Mackellar Fm is not considered to be glaciogenic, its lower-most strata contains rare glacial signatures in the form of out-sized clasts (Grindley 1963; Seegers 1996; Flaig et al. 2016). Outsized clasts have not previously been described in the Weaver Fm. (Minshew 1967; Long et al. 2008-2009). At Mt. Butters, the Mackellar Fm onlaps directly onto bedrock highs (Isbell et al. 1997a).

The lower contact of this facies is sharp and horizontal with both the CBS facies and HA facies association. The Mackellar Fm and its equivalents are laterally extensive in outcrop throughout the TAB.

uFG description

This facies is laminated to thinly-bedded, moderately-well-sorted to well-sorted, very-fine- and fine-grained sandstone. At Mt. Munson, this facies is on average slightly coarser, and contains some fine-
to medium-grained thin beds. Laminae and beds are either massive or cross-laminated with asymmetricripples on bedding surfaces. Ripples have uni-directional flow patterns. Some cross-laminations have climbing and/or flaser-like characteristics. Laminae and beds are laterally continuous, and have sharp, horizontal, and planar contacts. Very-fine-grained sandstones are black to grey in color, and fine- and medium-grained sandstone are yellow to grey. At all locations, the lowermost 2 – 5 m of this facies contain rare outsized clasts that range in size from very-coarse sand grains to small cobbles. However, clasts are locally common (Fig. 12B and D). The mean grain size of outsized clasts is coarse-grained sand to granule. These clasts are of similar lithologies to those in the diamictite facies and include striated and faceted clasts (Fig. 12B).

At Mt. Munson (site MM-17) this facies includes a ~10-m-thick chaotically deformed and normally faulted zone overlying glide planes. The deformed area is several tens of meters across. Some primary bedding was preserved in the deformation zone, showing that the deformed sediments had the same sedimentary structures and grain-size distribution of the undeformed sediments in this facies.

Horizontal, shallow trace fossils are common on bedding planes throughout this facies, though are less abundant where out-sized clasts are present. Vertical bioturbation is negligible, with some isolate burrows. Trace fossils identified in this facies include U-shaped vertical burrows (?Arenicolites isp.), non-specialized grazing trails (Helminthopsis tenuis, Helminthoidichnites tenuis, and ?Gordia marina), simple feeding traces (?Treptichnus isp., ?T. bifurcus), and arthropod trackways (Fig. 13). Transverse wrinkle structures were also found (Fig. 13A).

**UFG Interpretation**

Deposition of this facies likely results from unconfined, unidirectional current-driven sediment transport and settling from suspension. The facies at Mt. Munson (MM-17) are coarser than those observed at all other sites, possibly reflecting higher-energy/shallower conditions there. Striations on
some clast suggest that they were glacially derived. The clasts in the lower portion of this facies association were most likely transported by ice. However, the transport mechanism could have been either icebergs or sea/lake ice. The similarities between the composition of outsized clasts in this facies to other glaciogenic facies and facies association in this succession, as well as the presence of iceberg keels on top of the ME facies at MB-17, which was deposited at or near the top of the Pagoda Fm, suggests that icebergs were responsible for at least a portion of these clasts. The deformed interval at Mt. Munson is characteristic of a slump.

The depositional environment of the Mackellar Fm is typically interpreted as pro-delta and subaqueous channel levee settings (Seegers 1996; Miller and Isbell 2010; Flaig et al. 2016). The intraformational slump in this facies at site MM-17 (Mt. Munson) may have been caused by any number of drivers including rapid deposition and/or the presence of a slope. Glaciogenic ice-rafted debris could have been transported by either icebergs or sea/lake ice transporting glaciogenic material. Not all ice-rafted debris in the Mackellar Fm is necessarily glacially-derived, especially granule-sized clasts (see facies association BFG). This study is the first to describe ice-rafted debris at the base of the Weaver Fm (Long 1964b; Minshew 1967; Katz and Waterhouse 1970; Long et al. 2008-2009).

The trace fossils observed in the Lower Mackellar Fm in this study are dominantly non-specialized grazing trails and simple feeding traces, with some U-shaped burrows and arthropod trackways. The trace fossils assemblage, ichnodiversity and other characteristics of the trace fossils observed in this facies are very similar to those described in other TAB locations (Miller and Collinson 1994; Miller and Isbell 2010; Jackson et al. 2016). Because it is dominated by horizontal, non-specialized grazing trails and simple feeding traces, this assemblage closely resembles the Mermia ichnofacies described in other, post-glacial LPIA successions (Buatois et al. 2010; Alonso-Muruaga et al. 2012; Netto et al. 2012). The Mermia ichnofacies is endemic to lacustrine, fjordal, or other freshwater-dominated settings.
These interpretations suggest that the Lower Mackellar Fm in Shackleton Glacier region was largely freshwater-dominated with sporadic brackish conditions. Such conditions could be created in an environment with large freshwater inputs that displaced the more saline waters that were likely the dominant conditions within the TAB. This interpretation is consistent with other ichnology studies of the area, specifically, the work by Jackson et al. (2016) and Flaig et al. (2016) of the Mackellar Fm at Turnabout Ridge and Buckley Island in the Beardmore Glacier area. These studies interpreted the trace fossil assemblages in the Mackellar to have been made by short-lived, marine communities of small-bodied, benthic organisms that were stressed by high sedimentation rates and large freshwater inputs into otherwise brackish or marine waters.

The transverse wrinkle structures observed in this study area may be biotic, abiotic, or some combination of physical and biological processes (Fig. 13A) (Davies et al. 2016). No additional inferences can be made about these features, because the photographed sample was the only one observed in the study area and was not *in situ*.

**SITE DESCRIPTIONS AND PALEO-TRANSPORT DIRECTIONS**

*Mt. Weaver*

At Mt. Weaver (MM-17) Beacon strata rest nonconformably on up to 2 m of weathered granite (facies BFG0). Relief on the unconformity is locally up to 20 m. The Scott Glacier Fm at this location is a matrix-supported, coarse sand conglomerate with granite pebbles cobbles and boulders, and does not appear to be glaciogenic in origin. The conglomerate appeared to be composed entirely of granite-derived sediments. In the measured section, the conglomerate is ~0.75 m thick, but the conglomerate appeared to fill up to 5 m of topographic relief on the underlying granite basement. There is no evidence of glacial abrasion on the basement or on conglomerate clasts. The Scott Glacier Fm has a sharp upper contact with the Weaver Fm, which contained dropstones up to 5.5 m above the contact. Flow directions from...
asymmetrical ripples and cross-laminations in the Weaver Fm were toward 322°, 312°, and 002° (Table 3, Fig. 15).

*Mt. Munson*

The Pagoda Fm at Mt. Munson is ~5 m thick, and is an example of (Isbell et al. 2008c)’s Basin Margin FA. This observation contradicts those of Coates (1985), who described a 30 m-deep, steep-sided channel filled with Pagoda Fm sediments. A thorough, on-the-ground investigation of this site found that the “channel” described by Coates was more likely a significant debris-shoot/talus slope with boulders of diamictite up to 5m-wide. None of the boulders of diamictite or other sedimentary lithologies within the talus slope were unequivocally *in situ* and could not be traced laterally. A near-vertical fault that runs through Mt. Munson (Miller et al. 2010) may have created conditions that could account for Coates’ observations of channel walls.

At Mt. Munson, the massive diamictite of the Pagoda Fm overlies striated and polished Queen Maud granitoids and has a sharp upper contact with the Mackellar Fm. Striations on the underlying bedrock are oriented both north-south and east-west. The vergence of folds within a slump within the Mackellar Fm at this location include measurements of 109°, 104°, and 114° (Table 3). Current-transport directions from asymmetrical ripples in the Lower Mackellar Fm indicated flow towards 157° ± 28.5° just above the contact with the Pagoda Fm and 109° ± 20.0° for up to 48 m above the Mackellar Fm’s lower contact.

*Mt. Butters*

Five sedimentary sections were measured on the Mt. Butters Massif, four at site MB-17 and one at site MBSE-17 (Table 4; Fig. 14). MB-17, and MBSE-17 are located ~2 km apart, MB-17 is located north of MBSE-17. Massive diamictites are the dominant facies in the Pagoda Fm at both sites, but the upper Pagoda Fm is comprised of different current-transported facies at each location. At each location, the
Pagoda Fm is underlain by Queen Maud granitoids (Borg 1983) and has a sharp upper contact with the Mackellar Fm. At both sites, the lowest 5 m of the Mackellar Fm contain out-sized clasts of similar composition to those within the Pagoda Fm diamictites. All sites described at Mt. Butters are examples of Isbell et al. (2004)’s Basin Margin FA. In the Shackleton Glacier Area, including locations on the Mt. Butter’s massif, the Pagoda Fm is absent and Mackellar Fm laps directly on to basement rocks (La Prade 1970; Isbell et al. 1997a; Isbell 1999). Based on the thickness of the Pagoda Fm sections described at the Mt. Butters Massif both in this paper, La Prade (1970), Coates (1985), and Isbell (1999) the relief on the Maya Erosional Surface within the Massif was on the order of 100 – 200m.

**MB-17A and MB-17B**

These sections describe the basal 5–7 m of the Pagoda Fm at MB-17 consisted of the Basal Fine-Grain-Dominated FA (Fig. 6). The surface of the basement at site MB-17 underlying these facies dips uniformly towards 262° (strike 352°) with a dip angle of 11° (Table 3). Symmetrical ripple crests in these units are oriented at 346° ± 6.0° indicating that the orientation of waves energy in this lake environment locally parallel the coast (Table 3, Fig. 16). The internal slickenlines in the over-compacted portion of the BFG4 facies are oriented 006° - 186°.

**MB-17C and D**

Section MB-17C was measured through the bulk of the Pagoda Fm at this site (Fig. 14). This section contains ~73 m of Massive Diamictite FA, overlain by ~10 m of Heterolitic Agglomeration Facies Association. The vergence of soft-sediment deformation features within the massive diamictite indicate transport directions within the diamictite. Near the base of section MB-17C (12 – 20 m), shear planes of sand “pods” indicate shear directions toward ~220° - 256° (Table 3). Shear planes beneath a thick, mass-transport deposit toward the top of the section (~58 m) indicate shear directions range from 191° – 243° (Table 3). Within the Heterolitic Agglomeration FA current transport orientations are scattered. The
transport directions of asymmetrical ripples are scattered around north, ranging from 055° - 236°, with a
Fisher mean orientation toward 325° (Table 3). Cross beds in this facies are more tightly clustered toward
the west, ranging from 221° - 356°. Two wide grooves interpreted as ice berg keel marks (Fig. 14) at the
top of the Mixed Heterolithic FA in this section have a Fisher mean lineation of 253° - 073° (Table 3). All
grooves (N = 5) shallow toward the west.

MBSE-17

Section MBSE-17 was measured at a separate local than the previous four Mt. Butter’s sites. The
Pagoda Fm here is ~75 m thick. The Pagoda Fm overlies striated and polished granitoids and has a sharp
contact with the overly Mackellar Fm. The massive diamictite FA at the base of the Pagoda Fm is ~55 m
thick here. The Pagoda Fm at this site also includes ~10 m inter-bedded stratified diamictite and
discontinuous, sorted material that has been extensively deformed, which underlies an erosional contact
with cross-bedded, channelized sands (CBS). Striations on the underlying granite are oriented ~314° -
134°. Large (~5 m long) thrust faults within the massive diamictite have vergence towards 263°, parallel to
the paleoslope. Current transport directions in the CBS facies are towards 208°± 29°.

Reid Spur

The section measured and described at Reid Spur (RSP-18) contains ~60 m of Pagoda Fm. The
base of this section was not exposed in outcrop, but isopach maps of the region suggest that the thickness
of the Pagoda Fm here is ~100 m (Fig. 4 – inset) (Isbell et al. 1997b). The basement lithologies are likely
metasandstones and schists of the Beardmore Group (Fig. 1B). The upper contact of the Pagoda Fm at
Reid Spur is sharp with the overlying Mackellar Fm. The Massive Diamictite FA in this section does not
contain any sand “pods” or other indicators of soft-sediment deformation. This is the only succession to
contain the Laminated Sands FA. The current direction within the turbidites of the Laminated Sands FA
is towards $176^\circ \pm 29^\circ$ (Table 3). Current transport directions in the overlying Mackellar include $261^\circ$, $241^\circ$, $211^\circ$, and $281^\circ$ (Table 3).

**DISCUSSION OF DEPOSITIONAL TRENDS AND FLOW DIRECTIONS**

*Basin Margin vs. Basinal Facies Associations*

Isbell et al. (2008c) described two generalized facies association of the Pagoda Fm; one that occurs in basinal settings and another that occurs along the basin margins. These facies associations are a good tool for relating the observations from this study to glaciogenic deposits throughout the TAB. The Basin Margin FA is thinner (<100 m thick) than the Basinal FA and also generally has more evidence for subglacial to glacial-proximal sedimentation in the form of striated surfaces, subglacial shearing, grounding-line fans, plume sedimentation, and m-scale mass-transport deposits. The sites described in this study from Mt. Munson and Mt. Butters in the Pagoda Fm are characteristic of the Basin Margin FA. The Mt. Munson section (MM-17) contains the thinnest glaciogenic interval with evidence for both glacier-proximal sedimentation (facies MSD) and some grounding-line fan sedimentation (facies HA1) (Fig. 14). The thinness of the Pagoda Fm at Mt. Munson likely indicates less accommodation, likely due to a higher topographic position, than other sites examined in this study. The Pagoda Fm at both Mt. Butters sites (MB-17 and MBSE-17) is characteristic of the Basin Margin FA, as it includes evidence for subglacial erosion, abrasion and deformation; a proximal grounding-line fan system; plume; and mass-transport sedimentation. The site at Reid Spur (RSP-18) is also most characteristic of the Basin Margin FA, because diamictite facies there are thick, unstratified, and has a poorly-sorted matrix, suggesting plume sedimentation and gravity-driven redeposition. However, this section also likely represents a transition between the two facies. This is indicated by the LS facies at this site, which is attributable to the mid- to distal portion of a grounding-line fan, and by the inferred thickness of the Pagoda Fm at Reid Spur (~100 m), which is the same as the transition thickness between Isbell et al. (2008c)’s two FAs.
The transport directions within these successions strongly suggest that paleotopography (relief on the Maya Erosional Surface) played a significant role in the deposition of the Pagoda Fm and lower Mackellar Fm in the Shackleton Glacier region. All transport direction measurements and calculations are listed in Table 4 and Table 3, and displayed in Figure 2 and Figure 18. The influence of topography is particularly evident in the section MB-17 (Fig. 16). The surface of the basement at MB-17 dips toward the west at 11˚ (Fig. 16B; Table 3). Wave ripple crests in facies BFG are parallel to the strike of the basement surface at MB-17 (Fig. 16C), suggesting that the paleotopography created by this surface was sufficient to affect and orient wave action. The transport directions of slumping and other gravity-driven processes within facies MSD at both sites MB-17 and MBSE-17 are also generally towards the west (Fig. 16D), following the same slope. Flow directions in the grounding-line fan facies associations (HA and CBS) have a wide-spread ranging from the south-west toward the east. However, within the HA facies association at MBSE-17, gravity-driven transport is still towards the west (Fig. 16E). The south-south-east flow directions in the turbidite facies (LS) at site RSP-18 do not align with the flow direction at Mt. Butters, suggesting that those facies have a separate origin than the Mt. Butter’s grounding line fan(s) (Fig. 16D).

Flow directions of asymmetrical ripples in the lower Mackellar Fm also flow to the west at the Mt. Butters sites and at Reid Spur (Fig. 16E, F). Conversely, transport directions in the slump in the lower Mackellar Fm at Mt. Munson (MM-18) are toward the east-south-east (Fig. 16F). Since MM-17 was likely low accommodation and therefore located on a topographic high above the Mt. Butters sites, a slump that indicates and opposing slope direction suggests that MM-17 was located on the opposite side of that high. Though the change in slope is clear between Mt. Butters and Mt. Munson, a similar inference cannot be made between Mt. Butters and Reid Spur. Due to its basin-marginal characteristics, the Pagoda Fm at RSP-18 was likely located in deeper water.
Paleotopographic control on the deposition of Pagoda Fm and Mackellar Fm has been noted by authors throughout the TAB (Lindsay 1970b; Barrett 1972; Isbell et al. 1997a; Isbell et al. 2008c; Cornamusini 2017). In previous studies, ice-flow directions (usually striae on bedrock or clast pavements) have often been combined with other transport directions in the Pagoda Fm to infer a generalized transport direction. However, recent work in modern, high-relief, glaciated landscapes has shown that the relationship between glacier flow directions, other transport directions, and topography can be used to infer the thickness of the glacier relative to the magnitude of relief on the landscape (Landvik et al. 2014). In other words, whether or not a glacier “follows” the underlying topography is a function of the glacier’s thickness. Therefore, indicators of glacier flow should be considered separately from other transport directions.

The ice-flow directions below the Pagoda Fm in the Shackleton Glacier region are oriented generally north-west to south-east at Mt. Butters and Mt. Munson (Fig. 16D). Though none of the striae observed during this study had uni-directional indicators, previous workers in this area have found glacially-carved features in the basement underlying the Pagoda Fm that show glacier flow was basinward (toward the south) or along the basin axis (toward the south-east) (Fig. 15). These uniform flow directions on both a paleo-topographic high (MM-17) and paleo-topographic low (MB-17 and MBSE-17) suggest that the glacier, when it created these striae, was sufficiently thick to “overtop” the pre-existing topography in the Shackleton Glacier Area. Based on the difference in Pagoda Fm thickness, the local relief between site MB-17 (Mt. Butters) and MM-17 (Mt. Munson) was at least 85 m, and the onlapping of the Mackellar Fm onto basement in this area suggest that localized relief may have exceeded 100 m (Seegers 1996; Isbell et al. 1997a; Seegers-Szablewski and Isbell 1998). This scale of relief is on the scale of large hills. The topographic prominence of subglacial features on the scale of 100 m is considered negligible in studies of modern ice sheet margins (e.g. Lindbäck and Pettersson (2015); Cooper et al. (2019)), but would likely perturb or redirect the flow of relatively thin glaciers.
This discussion is all to say that the thickness of the glacier that created these striae more likely than not greatly exceeded the thickness of local topographic relief (~100 m), and that flow was most likely toward the center of the TAB.

DEPOSITIONAL MODEL

The weathered bedrock below the Pagoda Fm at Mt. Butters and the Scott Glacier Fm at Mt. Weaver is indicative of widespread sub-aerial exposure of the Maya Erosional Surface both before and during glaciation of the Transantarctic Basin. Note that weathered basement rocks underlying the Beacon Supergroup are common beneath basin marginal facies overlying granitic basement, including MB-17, Mt. Weaver, Cape Surprise, Mt. Fridtjof Nansen, and in the Geologists Range (Fig. 4F) (Barrett 1965; Isbell et al. 2008c).

The Pagoda Fm and lower Mackellar Fm in the Shackleton Glacier region was most likely deposited in a succession of environments that were part of an early Permian glaciation in a basin-marginal setting in the South Polar region of the Transantarctic Basin (Fig. 14). The majority of the Pagoda Fm in the Shackleton Glacier Area is composed of massive diamictites (facies MSD) that likely formed in glacier-proximal environments, at depths similar to those that occur on continental shelves, through a variety of glaciogenic and glacially-influenced processes. These massive diamictites are also conformably overlain by, and interstratified with, grounding line fan deposits (facies associations LS, HA, and CBS). These glacially derived lithologies are conformably succeed by pro-deltaic, fine-grained facies of the Mackellar Fm. The Mackellar Fm has some initial glacial influence in the form of ice-rafted debris, which is absent at ~5 m above the basal Mackellar Fm contact in all localities.

Deposition of the Basal Fine-Grained FA at the base of the MB-17 succession represents a pre-glacial lacustrine environment below normal wave base with periodic ice cover that was not glaciogenic but rather represented annual freeze over of surface waters (Fig. 16C). The weathered profile beneath this
FA suggests subaerial exposure prior to its deposition. Well-sorted, sand-sized, out-sized grains suggest periodic lake ice cover. The orientation of wave-ripple crests in this FA parallels the strike of the underlying bedrock structure, showing that the basement slope affected the deposition of Pagoda Fm sediments (Fig. 16B,C). The overall fining upward of this FA suggests an increase in accommodation, possibly a deepening of lake waters, prior to the onset of glaciation. The fossils present in this FA show that this lacustrine environment supported invertebrate communities, and that the surrounding landscape was vegetated. This FA has only been observed at site MB-17 (Mt. Butters). Deposition of similar sediments was either not wide-spread, not well-preserved, or both.

Evidence for the grounded advance of a glacier(s) in the Shackleton Glacier region is present at base of the MSD facies at both Mt. Butters’ sites (MB-17 and MBSE-17) and at Mt. Munson (MM-17). The lower contact of the MSD facies is not exposed at Reid Spur, so similar inferences cannot be made for that locality. No evidence for glacial erosion was observed at Mt. Weaver (MW-18). No conclusive evidence for subglacial deformation or erosion was observed higher in the Pagoda Fm at any site examined, though a subglacial origin for the MSD facies cannot be wholly ruled out. This advance was likely made by a glacier whose thickness exceed 100 m and flowed from north to south across the Shackleton Glacier region; from the direction of the present Ross Sea crossing basins margins perpendicular to the elongate trend of TAB (See discussion in prior section; Fig. 16D). This along with other data from the TAB strongly suggest that there were at least two ice centers in Antarctica during the Permian, one located on the East Antarctica craton and one in present day West Antarctica.

When the glacier margin retreated from the Mt. Butters, Mt. Munson, and Reid Spur sites, the deposition of glacier-proximal deposits was initiated. The massive diamictite (MSD) facies that dominated the Pagoda Fm at Mt. Butters and Reid Spur (RSP-18) was likely deposited through a combination of glaciogenic depositional processes characteristic of marine shelves (or their lacustrine
equivalents); a combination of settling from suspension of neritic sediments, plume sedimentation from subglacial and englacial jets, as well as iceberg sedimentation and mixing. The Shackleton Glacier region during the early Permian was likely not an open shelf, but a near-coastal setting with ample topographic relief, and water depths below normal wave base. Ultimately, many of these sediments were likely remobilized by gravity-driven slides, slumps, and flows. This is especially true at the Mt. Butters site, where the vergence of soft-sediment deformation features in the MSD facies follow the local paleotopographic slope (Fig. 16B, E). Soft sediment deformation caused by mass-transport deposits (facies HA) within bedded diamictites at Mt. Butters shows that the diamictites were weak, and likely water-saturated, suggesting relatively rapid deposition.

At Mt. Butters and Reid Spur, massive diamictites are inter-stratified with and overlain by grounding line fan deposits in the form of the LS, HA, and CBS facies associations. The stacked density flows (facies LS) at Reid Spur represent the medial to distal portion of a fan, likely in relatively deep water. Flow direction in these sediments are towards the basin axis (Fig. 16D), suggesting the basin geometry controls topography in this area. On the other hand, the fan deposits at Mt. Butters are more proximal to the glacier margin. The laminated diamictites, reactivation surfaces, and storm-wave deposits in the fan at the top of the Mt. Butters section (MB-17 and MBSE-17) suggest this fan was built gradually along a relatively stable margin and not during a single, catastrophic drainage event (Dowdeswell et al. 2017). The successions at Mt. Butters from the HA facies association through the CBS facies association represents either the progradation of one of these fans, a minor readvance associated with the deposition of the fan, or a combination of the two (Fig. 16E). The dispersed flow directions within the HA facies are characteristic of a fan, though their orientation broadly toward the north is away from the basin axis and opposed to glacier flow directions. Flow directions in the CBS facies are well clustered toward the southwest (Table 3, Fig. 16E). Influence of the paleotopographic slope at Mt. Butters and the radial nature of fan geometries, are the most likely cause from these seemingly antagonistic flow directions (Fig. 16E).
At all localities, the glacial signature in the Shackleton Glacier region persists several meters into the Mackellar Fm (or Upper Fine-Grained FA; Fig. 14; Fig. 17) in the form of glacially-transported, out-sized clasts. This indicates that a glacial margin capable of producing icebergs did not retreat out of the TAB as abruptly as previously suggested based on evidence from more northern parts of the TAB (Isbell et al. 2008c) and were still active as a pro-deltaic system became the dominant depositional regime in the central TAB. The Mackellar Fm in the Shackleton Glacier Area has a coarser average grain size (fine to very-fine-grained sandstone) than is typical for the formation in deeper parts of the basin (shale). This is likely due to its position along the margin of the TAB. Evidence for iceberg scour at the conformable contact between the Pagoda Fm and Mackellar Fm in the MB-17 section is additional evidence for this.

**GLACIAL STRATIGRAPHY**

In this study, the sequence stratigraphy of the Pagoda Fm in the Shackleton Glacier region can only be considered at Mt. Butters, because that is the only location where the authors were able to measure complete sections. Since the Mt. Butters locations are basin-marginal successions (Fig. 16), this analysis is glacial sequence stratigraphy should not be considered applicable to basinal successions of the Pagoda Fm. The Pagoda Fm at Mt. Butters is unique within the TAB because it contains only a single glacial sequence as defined by Powell and Cooper (2002) and Rosenblume and Powell (2019) (Fig. 17). The sequence described in this paper is bounded at its base by a surface of glacial erosion (defined by striae on bedrock and the deformation of facies BFG4), and at its top by an iceberg termination surface (defined by the final out-sized clast in the section). Since the Pagoda Fm in this location is overlain by the non-glacial Mackellar Fm, there is no true maximum retreat surface beyond the iceberg termination surface. Most of the succession likely represents a glacial retreat systems tract, though there is likely some fraction of the massive diamictite facies above the erosional surface that is more likely to have been subglacially deposited and would therefore represent a glacial maximum systems tract. The transition between those two systems tracts would be defined by a grounding-line retreat surface.
This sequence is most consistent with Rosenblume and Powell (2019)’s Type I “idealized glacial sequence”, which is a model developed to reflect the a sedimentation sequence deposited during the retreat of a subpolar glacier with grounding-line fan development and a glacial erosion surface as a sequence boundary. This sequence model was developed based on Upper Miocene sediments from the Ross Sea (Rosenblume and Powell 2019), whose climatic and geologic were likely reasonably similar to the TAB during the Permian. The Type I idealized sequence is interpreted to represent a dynamic climatic glacial system with very high erosional rates and debris fluxes, which is consistent with the depositional model presented for the Pagoda Fm in this paper.

LPIA GLACIATION IN THE TRANSANTARCTIC BASIN

The descriptions and interpretations of the Pagoda Fm in this paper are most meaningful when considered within the larger context of the LPIA in the Transantarctic Basin. Understanding the type, timing, and extent of LPIA glaciations within the TAB on a regional scale allows better comparisons with global records on climate change. At the beginning of this paper, four questions were proposed regarding what the sediments described in this study might tell us about the LPIA in the Transantarctic Basin:

1. What were the depositional environment and depositional mechanisms for the glaciogenic Permian sediments in the Shackleton Glacier region, and how does this compare to elsewhere in the TAB?
2. What can the LPIA strata in the Shackleton Glacier region study area tell us about the distribution of ice centers contributing to sedimentation in the TAB?
3. How does the early Permian, non-glacial record in the Shackleton Glacier region deepen our understanding of high-latitude LPIA environments?
4. What are the characteristics of the Permian glacial-to-postglacial transition in the Shackleton Glacier region and what can these characteristics tell us about South Polar glaciation during one of the apexes of the LPIA?

**Environment and Mechanisms for Deposition of Glacial Sediments**

**Depositional Environment**

Throughout the TAB, Early Permian glaciogenic depositional environments were locally and regionally variable in part due to the inherent complexity of glacial processes and preservation potential, but also due to heterogeneity of the pre-existing landscape. Most studies of glaciogenic rocks in the TAB note up to several hundred meters of topographic relief on the underlying basement that is not wholly “filled” by glacial sediments and continued to influence post-glacial sedimentary deposition (Isbell et al. 1997a). Cornamusini (2017) observed how such complexities can result in seemingly contradictory flow directions. This study is another example. Transport directions at all of the sites in this study appear contradictory, unless paleotopography and processes that created each feature are considered (Fig. 16).

For example, at Mt. Butters the flow directions of Permian glaciers in the TAB appears to be from north to south, while the transport directions of gravity-driven deposits and flow directions within the grounding line fan are perpendicular to that (Fig. 16D, E). If averaged together, these flow directions would imply general transport toward the south-west, when the most likely scenario was that the mass- and current-transported deposits followed a paleotopographic slope and the glacier did not.

**Glacier characteristics: thermal regime, thickness, and extent**

One of the ultimate goals of the study of LPIA glaciogenic sediments is to infer the “type” and distribution of glaciation experienced in any given basin. Glacier “type” typically refers to the glacier’s thermal regime and its size (i.e. ice sheet, ice cap, or alpine). Such characteristics of glaciers are controlled by many factors, but generally tie back into climate and geologic setting. Glaciogenic sedimentary
deposits are often used to infer the thermal regime of their parent glacier (Dowdeswell et al. 2017; Kurjanski 2020). Recent studies all agree that glaciogenic deposits in the TAB are most likely the result of transport and deposition by temperate glaciers (Isbell et al. 2008c; Isbell 2010; Koch and Isbell 2013; Cornamusini 2017). Temperate glaciers are far-and-away the more prolific producers and transporters of sediments and that the sedimentary record is therefore biased towards them. The presence of deposits from temperate ice does not mean the thermal regime of the glaciers were never cold-based or that there was not lateral variation in glacier thermal regime.

This study finds that the majority of sediments in the Pagoda Fm in the Shackleton Glacier region were deposited during the retreat of a temperate glacier. The key indicators for this retreat include the presence of grounding-line fan systems, ample evidence for plume sedimentation and rapid deposition, as well as abundant glacially-transported clasts with a wide size range made of local basement lithologies. The presence of sub-aqueous fan deposits interspersed within the massive diamictite facies suggests that the glacier had an active, channelized subglacial hydrologic system, and that during its retreat the glacial margin was stationary for at least a few years at a time to create grounding-line fans (Cowan and Powell 1991; Dowdeswell et al. 2015). Based on flow directions, the two grounding-line fans observed in the Shackleton Glacier Area (Mt. Butters and Reid Spur) are likely derived from two separate fans. These fans do not represent advances of the glacial margin by their presence alone, since the hydrologic systems of outlets of glacier are not necessarily fixed. However, prograding fan like the one built at Mt. Butters likely took years to build, and suggest that the glacier had a persistent, organized hydrologic system (Hunter et al. 1996; Dowdeswell et al. 2017).

**Timing and Duration of Glaciation**

Limited fossil evidence indicates that glaciogenic sediments in the TAB were most likely deposited during the early Permian (Askins, 1998; Isbell et al., 2001; Babcock et al., 2002; Colette et al.,
Palynomorphs in the TAB correlate to the Australian *P. confluens* zone (Barrett and Kyle, 1975; Kyle, 1977; Kyle and Schopf, 1982; Lindström, 1995; Askin, 1998), which has recently been constrained with U-Pb zircon ages in east Australian basins to the Asselian (Price, 1997; Smith et al., 2017).

Palynological studies of the Mackellar Fm indicate that it was deposited during the same biozone as the Pagoda Fm (Masood et al., 1994; Askin, 1998). Therefore, glaciation in the TAB had likely concluded before the end of the Asselian.

The Pagoda Fm in the Shackleton Glacier region likely represents a single glacial-interglacial cycle, stratigraphically represented by a single glacial sequence. In this context, the phrase “glacial-interglacial cycle” refers to the advance of a glacier into the basin and its whole retreat out of the basin, which is stratigraphically defined by a surface of glacial retreat. This is not to say that the position of the glacier margin did not fluctuate during that cycle, but that only one grounded erosional surface is present in the study area and only one surface of glacier retreat (Powell and Cooper 2002; Rosenblume and Powell 2019). No instances of a grounded re-advance over any of the sections examined in this study were observed beyond the basal erosional surfaces. As previously discussed, the processes and environments responsible for the extensive deposition of the massive, glaciogenic diamictite facies were likely diverse and largely subaqueous. The preservation of such sediments is most probable if deposited during the final retreat phase with no subsequent advances over the area (Kurjanski 2020). This is especially true in a basin-marginal position in a basin like the TAB that was trough-shaped and not exposed to open marine conditions.

Evidence for glacier readvance above the basal erosional surface of the Pagoda Fm does exist at other Pagoda Fm outcrops in the TAB, including in the Beardmore Sub-basin (Lindsay 1970a; Miller 1989; Koch and Isbell 2013). This evidence typically occurs in thicker, basinal facies association which are likely areas with higher accommodation than the basin margins.
The Salinity Question

There continues to be a paucity of convincing evidence one way or another that water in the TAB was fresh (Cornamusini 2017), fresh to brackish (Miller and Collinson 1994; Miller and Isbell 2010), brackish marine (Flaig et al. 2016; Jackson et al. 2016), or marine (Isbell 2010; Koch and Isbell 2013). The lacustrine nature of the sediments at the base of the site MB-17 section was determined based on the paleontology in previous studies (Isbell et al. 2001; Babcock et al. 2002). Previous studies of the Pagoda Fm in the Beardmore Glacier area and its equivalents in South Victoria Land have used grounding-line fan deposits with significant “rain-out” material (plume sedimentation) as evidence for an overflow-dominated fan; suggesting marine conditions (Powell 1990; Isbell 2010; Koch and Isbell 2013). The prevalence of stratified diamictite facies (plume sedimentation) associated with grounding-line fan deposits in the Pagoda Fm of the Shackleton Glacier region also support a non-lacustrine interpretation. However, this reasoning cannot rule out brackish ambient water conditions. For example, Lunkka and Alhonen (1996) describe a Quaternary grounding line fan system in Finland with ample indicators of plume sedimentation that was deposited in the Baltic Sea during a time that the Baltic was alternating between marine and brackish conditions. Trace fossils from the Mackellar Fm in the Shackleton Glacier area also suggest dominantly freshwater conditions.

Distribution of Ice Centers

The flow-directions and facies analysis from this study strongly support the hypothesis that an ice center was positioned inboard of the Panthalassan margin of Antarctica during the early Permian (Fig. 2B, ice center “q”). The presence of such an ice sheet is a relatively new hypothesis that was first proposed by Isbell (2010) based on transport directions in South Victoria Land. In recent publications, this proposed ice center on the Panthalassan side of the TAB has been inconsistently included (Fielding et al. 2008c; Isbell et al. 2012; Montañez and Poulsen 2013) and excluded (Fielding et al. 2010; Craddock et al. 2019) from LPIA ice center reconstructions. Evidence from this study and Isbell (2010) shows that an ice
center should be included on the non-cratonic edge of the TAB in reconstructions including the Early Permian, an interval also referred to as “Glacial III” (Isbell et al. 2003), “Event 5” (López-Gamundí et al. in review), and Australian “P1” (Fielding et al. 2008b). Ice-flow directions elsewhere in the TAB clearly indicate that glaciers also flowed into the TAB off of the East Antarctic Craton and along the TAB’s basin axis toward the Wisconsin and Ohio Ranges (Fig. 2B, ice center “r’’). The multiple ice centers contributing to sedimentation in the TAB may not have been synchronous in their advances and retreats throughout the LPIA.

High Latitude Paleoenvironments

Weathered granitic surfaces on TAB basin margins and the non-glacially influenced lacustrine facies at Mt. Butters are all the result of non-glacial environments at southern polar latitudes during the LPIA. The presence of weathered granite profiles on the margins of the TAB suggests that those areas experienced subaerial weathering prior during the LPIA. Such weathering profiles may give us some insight into the timing of glaciation if compared to similar Quaternary examples. The depth of weathering and apparent degree of chemical weathering is similar to some pre- or inter-glacial, Quaternary grus/saprolite profiles described in recently deglaciated, high-latitude landscapes (Bouchard et al. 1995; Migoń and Thomas 2002; Olvmo et al. 2005; Sugden et al. 2005; Phillips et al. 2006; Goodfellow 2007; Dampier et al. 2011). In these Quaternary examples, this preservation could have been the result of the advance of glaciers with cold-based thermal regimes with non- or minimally-erosive subglacial conditions, a high landscape position above the glacier (as nunataks), or shielding from subsequent glacier advances due to landscape position or cover (Goodfellow 2007). However, placing any temporal constraints on the development of soils in granitic bedrock, especially if the climatic conditions are not known, is problematic (Migoń and Thomas 2002). Chemical weathering in these environments requires only some precipitation, deep groundwater (consistent percolation), and time (Ollier 1988; Goodfellow 2007).
Additionally, weathering profiles in modern granites have formed during intervals ranging between several thousand and 1 million years (Phillips et al. 2006; Darmody et al. 2008; Ebert et al. 2012).

Based on these constraints, the weathering surfaces on the margins of the TAB (on the Maya Erosional Surface) illuminate the potential complexities of the glacial and interglacial history of the TAB.

First, the presence of these weathering profiles suggest that the TAB and its drainage area were not glaciated throughout the entirety of the LPIA; a hypothesis that is implied by models of LPIA glaciation that suggest a single, large ice sheet emanating from Antarctica (e.g. Scotese and Barrett 1990). Instead, these surfaces suggest that the TAB experienced interglacial intervals during the LPIA. This interpretation is consistent with the modern understanding of how the geometries and extents of even large ice sheets change in response to millennia-scale interglacial intervals (e.g. Batchelor et al. 2019) or “brief” warm intervals within glacial-interglacial cycles (e.g. Yokoyama et al. 2016). Second, the preservation of these weathering profiles does not preclude any subsequent glacier advance(s) over the TAB basin margins. In section MB-17, the preservation of this surface is likely due to its protection from erosion by lacustrine sediments (BFG facies association). Because these weathering profiles could form during a single interglacial interval, their presence also does not preclude prior glacial advances over the Shackleton Glacier. At site like Mt. Weaver, where there is no evidence for glaciogenic deposition or erosion, the presence of weathered basement granites at Mt. Weaver most likely indicates the lack of subsequent glaciations. Though the presence of a weathered surface does not preclude an advance of a non-erosive, cold-based glacier over that location following the formation of the weathering profile, there is no evidence specifically for that scenario.

**Glacial to Post-Glacial Transition**

The transition from glaciogenic to post-glacial lithofacies in the Transantarctic Basin has generally been described as abrupt (Isbell et al. 2008c). The nature of the contact between the Pagoda Fm and
Mackellar Fm in the Shackleton Glacier region paints a similar picture; glaciogenic rocks have a sharp upper contact with the Mackellar Fm. Glacial influence on sedimentation in the form of dropstones (iceberg termination surface) is present for up to 6 m above this contact at Reid Spur, Mt. Butters, Mt. Munson, and Mt. Weaver. This pattern indicates that iceberg-producing glaciers likely persisted in the Shackleton Glacier and Mt. Weaver areas for a time after the dominant deposition regime of the basin switched over to pro-deltaic systems.

This abrupt glacial-to-post-glacial transition is often interpreted to represent a rapid retreat of glaciers from the Transantarctic Basin due to a similarly rapid amelioration of regional climate (Isbell et al. 2008c; Cornamusini 2017). However, that is not what is indicated by the glacier retreat sequence in the Shackleton Glacier region. The succession indicates retreat of a temperate glacier, which in turn indicates a warming of regional climate, but there is no evidence in this area that the rate at which the glacier left the basin or the regional climate warmed was especially abrupt or rapid. The iceberg keel marks at the top of the Mt. Butters Pagoda section and the ice-rafted debris in the lower Mackellar Fm are evidence of this, as they indicate that the ice front was still discharging icebergs into basinal waters during the early stages of Mackellar deposition. The Pagoda Fm in the Shackleton Glacier region represents a single glacier retreat sequence from that area of the basin. There is little evidence constraining how protracted or fast the withdrawal of the glacier margin from the basin was, but it was likely not especially abrupt.

CONCLUSIONS

- The Pagoda Fm in the Shackleton Glacier region is glaciogenic and was deposited in a basin-marginal subaqueous setting. The dominant lithology in the Pagoda Fm here is massive, sandy, clast-poor diamictite. The depositional processes governing these diamictites were subaqueous glacial processes; likely a combination of mass-transport, iceberg-rainout, iceberg scouring, plume sedimentation, and subglacial till deposition. Current-transported sands and
stratified diamictites within the Pagoda Fm were deposited as part of grounding-line fan systems.

- The transport directions and thicknesses of strata along the TAB margin were strongly controlled by topographic relief on the underlying erosional surface.
- In the Shackleton Glacier region, all glaciogenic sediments in the Pagoda Fm were likely deposited during the retreat phase of a single glacial sequence.
- Glacier flow directions (towards the south and south-east) and trends in Pagoda Fm thicknesses in the Shackleton Glacier Area support the hypothesis that an ice center was present toward the Panthalassan/Gondwanan margin of East Antarctica during the LPIA.
- Weathering profiles in granites at basin-marginal positions in the TAB could have formed during subaerial exposure lasting between tens-of-thousands and several million years. Results from surface exposure age dating from similar Quaternary sites suggest that the presence of such weathering profiles does not preclude the re-advance of non-erosive, cold-based glaciers over some areas, nor more extensive glaciation during glacial intervals prior to the formation of the weathering profiles. Weathering profiles could have formed during interglacial periods or prior to the onset of glaciation in the basin.
- The transition from Pagoda Fm to Mackellar Fm depositional environments in the Shackleton Glacier region is abrupt, which is similar to how the transition has been described in the rest of the Beardmore sub-basin. If basin-marginal Pagoda Fm successions are considered to be a single glacier retreat sequence, this abrupt transition does not mark a sudden change in glacier character or climate. Instead, this abrupt transition may simply mark the end of retreat from the area of a temperate glacier margin that was a prolific sediment-producer. In all sections examined in this study, ice rafted debris persisted in the Mackellar Fm 5-6 m
above its lower contact, suggesting that glaciers were present in the basin in this area after
the depositional regime was no longer dominantly glaciogenic.

- Though the Pagoda Fm in the Shackleton Glacier region likely only represents a single glacial
  sequence, the evidence outlined in this study does not preclude prior glaciations from
  occurring in the area. Weathering profiles in granite basement rocks underlying the Pagoda
  Fm could have formed during an interglacial interval. The basin marginal position of these
  strata makes the preservation of older sediments unlikely.
ACKNOWLEDGEMENTS

This work would have been impossible without the hard work of all the people who made the 2017 - 2018 Shackleton Deep Field Camp such a success, including the talented people of the National Science Foundation, Antarctic Support Contract, Ken Borek Air, Petroleum Helicopters, Inc., New York Air National Guard, and the U.S. Air Force. Special thanks are owed to Edith Taylor and Rudolf Serbet for aiding in the field planning, Danny Uhlmann and Ted Grosgebauer for keeping us from tumbling off of cliffs, and to Patty Ryberg, Rudolf Serbert, Brian Atkinson, and Erik Gulbranson for their companionship and cooking in the deep field. Thanks also the Kate Pauls and Eduardo da Rosa for their feedback on the manuscript, and to Pablo Alonso-Muruaga for his assistance identifying the trace fossils in this succession, and discussions about their meaning. Funding for this research came from National Science Foundation OPP-1443557, EAR-1729219, and OISE-1559231 grants, the University of Wisconsin - Milwaukee Graduate Fellowships programs, the P.E.O. Scholar Awards Program, The American Federation of Mineralogical Societies, The Wisconsin Geological Society, University of Wisconsin - Milwaukee (RGI grant), and the University of Wisconsin-Milwaukee Department of Geosciences.
FIGURE CAPTIONS

Figure 1. Generalized geologic maps of study area. Maps are South Polar projections. A. Geologic map of the Central Transantarctic Mountains and Victoria Land, with relevant outlet glaciers and mountain ranges labeled. Modified after Elliot (2013), Goodge (2016), and Estrada et al. (2016). Box on inset map indicates extent of this geologic map. Red box on geologic map indicates the extent of “map B”. B. Regional geologic map of the Shackleton Glacier area, noting the locations of sections described in this study, except Mt. Weaver. MM-17 is Mt. Munson, MB-17 is Mt. Butters 1, MBSE-17 is Mt. Butters 2, and RS-18 is Reid Spur. Geology adapted from McGregor and Wade (1969); Mirsky (1969), aerial photos from LIMA Landsat imagery (Bindschadler 2008).

Figure 2. Paleogeographic reconstructions of Gondwana near the Carboniferous-Permian Boundary. All maps are south-polar projections. Star indicated the approximate location of the Shackleton Glacier area. Continent distributions after Lawver et al. (2011) and copied from (Isbell et al. 2012). Note that there are differences in the positioning of some crustal blocks (e.g. Patagonia and New Zealand) between this reconstruction and Figure 3, which is modified after Elliot (2013). A. Yellow regions indicate the modern extent of sedimentary basins containing late Paleozoic Ice Age strata. Abbreviations include: Falkland Islands/Malvinas (FI), Ellsworth Mountain block (EM), Antarctic Peninsula (AP), Thurston Island (TI), Marie Byrd Land (MBL), and the Challenger Plateau/western New Zealand (ChP). Basins adapted from Isbell et al. (2012). B. Proposed positions of glacial centers during the Early Permian based on flow directions and position of “highlands”. Illustrated ice centers are not meant to represent the whole possible extent of each proposed glacier, but where proposed glaciers were likely to be nucleated. Confidence is based on abundance of available lithologic data, and both relative and absolute ages. Ice centers are as follows: a. Uruguay (Crowell and Frakes 1975; Assine et al. 2018; Fedorchuk 2019), b. Asunción (Frakes and Crowell 1969; Franca and Potter 1988; Limarino et al. 2014), c. Windhoek/ Koakoveld Highlands (Martin 1981; Visser 1987; Franca et al. 1996; da Rosa et al. 2016;

**Figure 3.** Regional geologic context and tectonic setting of the Transantarctic Basin during the early Permian. **A.** Stratigraphy of the Beacon Supergroup across different regions of the Transantarctic Basin. Adapted from Elliot (2013); Cornamusini (2017); Elliot et al. (2017). **B.** Tectonic setting of southern Gondwana during the Permian, adapted from Elliot (2013). Regions of sedimentary deposition are shaded yellow. Note that there are differences in the positioning of some crustal blocks (e.g. Patagonia and New Zealand) between this reconstruction and Figure 2. **C.** Modern extent of the Pagoda-ages Horlick Sub-Basin and Beardmore Sub-basin in the Transantarctic Mountains. Grey areas are outcrops/nunatuks. Lines show isopachs of the Pagoda Fm (Beardmore Sub-basin) and Buckeye Fm (Horlick Sub-basin) from Isbell et al. (2008c).
Figure 4. Representative sections of different thicknesses of the Pagoda Fm in the southern portion of the Transantarctic Basin's Amundsen-Byrd Sub-basin. This is not a cross-section. Sections adapted from Barrett (1965); Lindsay (1969); La Prade (1970); Coates (1972, 1985); Miller (1989); Koch and Isbell (2013). The inset map shows the location of each section and isopachs of the Pagoda Fm in the Beardmore Sub-basin from Isbell et al. (2008c).

Figure 5. Photographs of Mt. Butters outcrops showing the stratigraphy of the Pagoda and Mackellar fms and well as the relief of the Maya Erosional surface in this area. Formations are labeled and the Maya Erosional Surface is marked by a green, dashed line. A. Is a photograph of section MB-17 taken from a helicopter, view to the south. The saddle in the background of this photo is a location where the Mackellar rests directly on bedrock, which it appears to onlap. This saddle is approximately halfway between section MB-17 and MBSE-17. B. This is an outcrop that is part of the Mt. Butters Massif, and occurs on a spur south-east of section MB-17. View is toward the south-east from MB-17. This site has never been visited on foot, so the scale is not certain. This exposure is the same as Figure 4 in (Isbell et al. 2001). C. Photograph of section MBSE-17, view toward the SW. Purple oval shows to helicopter's shadow.

Figure 6. Photographs of the Basal-Fine-Grain-Dominated (BFG) Facies Association at the base of Mt. Butters section MB-17. Rulers are 50cm long with markings in cm. A. Granitic weathering profile and associated grus (BFG0) overlain by a bedded conglomerate (BFG1). B. A fining upward succession within the basal conglomerate (BFG1). C. A characteristic section of both horizontally stratified fine-grained facies' (BFG2 and BFG3). D. Top-down view of symmetrical ripples in BFG3. Note the out-sized clasts exposed in ripple troughs. E. Sharp contact between pervasively sheared fines (BFG4) and the overlying massive diamicrite (BFG4). F. Slickensides on the lower surface of a lenticular body in BFG4.

Figure 7. Photographs of the Massive Sandy Diamictite (MSD) facies association. Rulers are 50 cm long when folded in half and 1 m long when unfolded. Marks on rulers are in cm. The weathered
color of the diamictite is grey to beige and tend to be redder when the matrix has a higher percentage of sand in the matrix. Small sand bodies are highlighted by blue where present and orange line indicate important bedding planes. A. A characteristic clast poor section of the MSD facies with a sand “whisp” (MB-17C). B. A section of this facies’ diamictite that very clast poor, yet contains several large boulders (MB-17C). C. A clast rich section of the MSD diamictite (MB-17C). D. A ruck structure within the diamictite made by a boulder (MM-17). E. Examples of small sand bodies within the MSD diamictite that had experienced simple shear strain, result in sheath fold, “stringers”, boudinage, and small reverse faulting, (MBSE-17). F. Example of small sand body bands within an otherwise massive, clast-poor diamictite. (MB-17C). G. A boulder and cobble bed in an otherwise massive diamictite (MBSE-17). H. Diamictite beds onlapping on to a large, 4m boulder within the diamictite. Inset picture shows the whole, ~4 m diameter boulder in outcrop, with person for scale.

Figure 8. Photograph of the Laminated Sands (LS) facies at site RS-18 (Reid Spur). Triangles show fining-upward packages separated by erosional surfaces. Reddish lithologies are medium sandstone to coarse-grained sandstone, and have planar laminar and low-angle cross-stratification. Black lithologies are fine-grained sandstones.

Figure 9. Photographs of the Heterolithic Agglomeration (HA) facies association at Mt. Butters in section MB-17 and MBSE-17. Marks on all rulers are in cm. Rulers are 50 cm when folded in half, and 1 m long when unfolded. White dashed lines have been used to highlight contacts and important bedding surfaces. A. Medium-grained sandstone layer in facies HA1. Sandstone body composed mostly of climbing ripples and capped by asymmetrical ripple. Note sharp contact with black-colored, bedded diamictite (facies HA3) in lower part of image (MB-17) B. Fine- to – medium grained sandstone in facies HA1 with up-building symmetrical ripples (MB-17). C. Swaley and hummocky cross-stratification in sandstone layer of facies HA1 (MB-17). D. Loaded contact between a massive sandstone (HA2) and
bedded diamictite (HA3) (MB-17). E. Sandstone with a soft-sediment fold nose in facies HA2. (MB-17). E.

Deformed contact between facies HA2 and HA3. Jacob’s staff for scale – marking every 10 cm. (MB-17) G.

Outcrop showing inter-fingering between facies in this facies association. Black-colored sediments are

bedded diamictites (facies HA3), other lithologies show sand and gravel of facies HA2. (MB-17) H.

Groove structures on top of facies HA1. View approximately toward north (down the Shackleton

Glacier). I. Granitic outsized clast punctuating inter-laminated sandstone beds within the bedded

diamictite facies HA3. (MB-17).

**Figure 10.** Stitched photographs (A, B, and C) and interpretive sketches (B’ and C’) of the upper

portion of section MBSE-17 at Mt. Butters. View is to the north, and the ridge runs roughly east-west.

Figure A shows lateral varations in the architecture of the Cross bedded sandstone (CBS) facies

association. Figures B and B’ show an outcrop and highlight the stratigraphic relationships between

Massive Diamictite (MSD) facies, Heterolitic Agglomeration (HA) facies association, and the CBS facies

association. Black lines in CBS denote channel erosional surfaces. Black lines in HA3 indicate soft-

sediment deformation. The part of this outcrop highlighted by C and C’ is located within the red box on

B. This portion of the outcrop is characteristic of the HA facies association, and was selected to illustrate

the pervasive nature of soft-sediment deformation within this facies association.

**Figure 11.** Photographs of the Cross Bedded Sandstone (CBS) facies at section MBSE-17 on Mt.

Butters. A. An example of low angle and trough cross-bed sets in this facies. Measuring stick in 1m long.

B. Photograph of facies in outcrop noting occurrences of minor lithologies.

**Figure 12.** Photographs of the Upper Fine Grain Dominated (UFG) facies association. Marks on

all rulers are in cm, and are 50 cm long when folded in half and 1 m long when unfolded. A. Outsized

clasts in the lower Mackellar Fm (MB-17); B. A striated and faceted out-sized clast from in the lower

Mackellar Fm (RS-18); Photographs C. (MB-17), D. (MBSE-17) and E. (RS-18) are examples of
characteristic ripples and planar lamination in the facies; F. Outcrop view of the Weaver Fm at Mt. Weaver shows the lateral continuity of beds in the facies. View to the south along the north face of Mt. Weaver. Note light-colored thin bed toward the top of the photograph. Red colored debris is dolerite is from a sill. Person in the background is sitting on granite bedrock underlying the Permian sediments at this location.

**Figure 13.** Field photographs of trace fossil in the Upper Fine-grain Dominated facies (UFG), which is part of the lower Mackellar Fm. Units on measuring sticks in cm. A. Sample collected from float at site MB-17 (Mt. Butters). Inset photo shows the whole sample, large photo shows a zoom-in version of what appear to be transverse wrinkle marks. B. Sample collected from MB-17 (Mt. Butters). Trace fossils include *?Arenicolites* isp. and *?Treptichnus* isp. C. Sample collected from MB-17 (Mt. Butters). Trace fossils include *Helminthopsis tenuis*, *Helminthoidichnites tenuis*, and *?Gordia marina*, D. Sample collected from MBSE-17 (Mt. Butters). Trace fossils include *?T. bifurcus* and *Arenicolites* isp., and E. Sample collected from RS-18 (Reid Spur). Trace fossils include and arthropod trackway.

**Figure 14.** Sedimentary logs and paleotransport directions from sites described in this study. Paleotransport directions are plotted to align with the maps of modern Transantarctic Mountains presented in this study, so that north is toward the bottom of the page and varies by geographic position (see inset map). Details of paleotransport measurements are available in Table 3. **MW-18** is Mt. Weaver, **MM-17** is Mt. Munson, **MB-17** is Mt. Butters 1, **MBSE-17** is Mt. Butters 2, and **RS-18** is Reid Spur. Lithologies are grouped by their interpreted facies or facies association. The inset map shows the location of each section and isopachs of the Pagoda Fm in the Beardmore Sub-basin from Isbell et al. (2008c). The datum for these columns were chosen using the last evidence for glaciogenic sediments, either the uppermost outsized clast or diamicrite.
Figure 15. Glacial transport directions from the literature and this study in the Pagoda Fm and equivalents in South Victoria Land, Beardmore Sub-basin, and Horlick Sub-basin. Directions are described in Table 4. Directions are from Long (1964b); Frakes et al. (1966); Minshew (1967); Lindsay (1970a); Coates (1985); Isbell et al. (1997b); Lenaker (2002); Long et al. (2008-2009); Isbell (2010), and this study. Isopachs are from Isbell et al. (2008c). Note how the orientation of north varies throughout the map area.

Figure 16. Box diagrams showing progressive phases of the depositional model for the Pagoda Fm and lower-most, glacially-influenced Mackellar Fm in the Shackleton Glacier region, alongside maps showing the modern locations of the sites described in this study with transport orientations related to each part of the depositional model. See Table 3 for flow directions. A. Map of the modern central Transantarctic Mountains. Grey areas are approximate locations of nunatuk regions. Solid black lines are isopachs of the Pagoda Fm, copied from Isbell et al. (2008c). Red square indicates map area in parts B – F. Note that the smaller map areas are rotated relative to the larger “map A”. Yellow stars show site locations described in this study. B. Paleotopography of the Shackleton Glacier Area prior to deposition of the Pagoda Fm. Map shows strike and dip of granite surface underlying the Pagoda Fm at Mt. Butters site MB-17, corrected for modern structural conditions. This map also includes the Isbell et al. (2008c) isopach lines and derived basin axis orientation for reference C. Proposed depositional conditions for the Basal Fine Grained (BFG) Facies Association at the base of the Pagoda Fm at site MB-17. The green, double-sided arrow shown on the map shows the orientation of symmetrical (wave) ripple crests, which parallels the strike of the underlying basement. D. Proposed depositional conditions for Massive Sandy Diamictite (MSD) facies and Laminated Sands (LS) facies during retreat of the glacier out of the Shackleton Glacier area. On the map, purple wedge indicates range of flow direction in the LS facies, green wedges indicate down-slope transport direction in the MSD facies. Blue double-headed arrows show glacier flow directions measured in this study. E. Proposed conditions during the deposition of the grounding-line fan represented by facies associations MSD, Heterolithic Amalgamation (HA), Cross-bedded Sands (CBS), and Upper Fine Grained
(UFG). The red wedge shows range of flow directions in the UFG facies, blue and purple show the range of flow direction for the CBS and HA flow directions, respectively and the blue arrow indicates the mean flow direction of the CBS facies at site MBSE-17. Proposed conditions during the deposition of the lower Mackellar Fm (UFG facies association).

**Figure 17.** Glacial sequence stratigraphy of the Mt. Butters sections, after Powell and Cooper (2002) and Roseblume and Powell (2019). The depositional systems are defined as N = non-glacial, D = glacier distal, P = glacier proximal, and I = ice-contact.
TABLE HEADINGS

Table 1. Names and locations of sedimentary sections described in this paper.

Table 2. Facies and Facies Association Descriptions.

Table 3. Summary of Paleo-transport measurements from this study (mean = Fisher mean ± a99).

Note that due to the proximity of the study locations to the south pole the orientation of compass directions are not the same across the basin. T = Trend, P = Plunge, D = Dip angle, DD = Dip Direction, S = strike, * = measured during 1997 field season

Table 4. Glacial transport directions from the literature and this study in the Pagoda Fm and equivalents in South Victoria Land, Beardmore Sub-basin, and Horlick Sub-basin. Directions are plotted in Figure 13. The orientation of magnetic/geographic north varies considerably with proximity to pole. Therefore, making direct comparisons between directional measurements is not advised. Transport direction of bi-directional features (striae) typically inferred from flow directions in adjacent or overlying strata. Data at Mt. Munson from Isbell (unpublished) are previously unpublished measurements that were collected during the 1995 field season.

* = measured from figure, not reported in text
REFERENCES


transgressive deposits in Gondwana; reconstructing salinity conditions in coastal ecosystems affected by

Buso, V.V., Milana, J.P., Di Pasquo, M., Paim, P.S.G., Phillip, R.P., Aquino, C.D., Cagliari, J.,
ages and correlations from Paganzo and Paraná basins: Palaeogeography, Palaeoclimatology,
Palaeoecology, v. 544.

Cardozo, N., and Allmendinger, R.W., 2013, Spherical projections with OSXStereonet: Computers
& Geosciences, v. 51, p. 193-205.

Coates, D.A., 1972, Pagoda Formation: Evidence of Permian glaciation in the central
Transantarctic Mountains, in Adie, R.J., ed., Antarctic Geology and Geophysics: Oslo,
Universitetsforlaget, p. 359-364.

Coates, D.A., 1985, Late Paleozoic glacial patterns in the central Transantarctic Mountains,
Antarctica, in Turner, M.D., and Splettstoesser, J.F., eds., Geology of the central Transantarctic Mountains:

Collette, J.H., Isbell, J.L., and Miller, M.F., 2017, A unique winged euthycarcinoid from the

Collinson, J.D., and Mountney, N., 2019, Sedimentary Structures: Edinburgh, Dunedin Academic

the central Transantarctic Mountains, Antarctica: Geological Society of America Bulletin, v. 118, p. 747-
763.

Transantarctic basin, in Vevers, J.J., and Powell, C.M., eds., Permian-Triassic Pangean basins and
foldbelts along the Panthalassic Margin of Gondwanaland: Boulder, Colorado, Geological Society of
America Memoir 184, p. 173-222.

Collinson, J.W., and Miller, M.F., 1991, Comparison of Lower Permian post-glacial black shale
sequences in the Ellsworth and Transantarctic Mountains, Antarctica, in Ulbrich, H., and Rocha-Campos,
A.C., eds., Gondwana Seven Proceedings: papers presented at the Seventh International Gondwana
Symposium: São Paulo, Instituto de Geociências, Universidade de São Paulo, p. 217-231.

Cooper, M.A., Jordan, T.M., Schroeder, D.M., Siegert, M.J., Williams, C.N., and Bamber, J.L., 2019,
Subglacial roughness of the Greenland Ice Sheet: relationship with contemporary ice velocity and geology:
The Cryosphere, v. 13, p. 3093-3115.


Dietrich, P., and Hoffmann, A., 2019, Ice-margin fluctuation sequences and grounding zone wedges: The record of the Late Palaeozoic Ice Age in the eastern Karoo Basin (Dwyka Group, South Africa): The Deposition Record.


Elliot, D.H., 1992, Jurassic magmatism and tectonism associated with Gondwanaland break-up: an Antarctic perspective, in Storey, B.C., Alabaster, T., and Pankhurst, R.J., eds., Magmatism and the...


Ives and Isbell  SHG Glaciogenic Lithofacies  J. of Sedimentary Research


Isbell, J.L., Koch, Z.J., Lenaker, P., and Askin, R.A., 2004, Gondwana glaciation in Antarctica; are the strata glacial-terrestrial or glacial-basinal in origin and were they deposited by ice sheets?, p. 961-962.


Lindsay, J.F., 1968, Stratigraphy and sedimentation of the lower Beacon rocks of the Queen Alexandra, Queen Elizabeth, and Holland Ranges, Antarctica, with emphasis on Paleozoic glaciation [unpublished Dissertation thesis]: The Ohio State University, 300 p.

Lindsay, J.F., 1969, Stratigraphy and sedimentation of Lower Beacon rocks in the central Transantarctic Mountains, Antarctica, v. 33: Columbus, Ohio, The Ohio State University Research Foundation, 58 p.


Minshew, V.H., 1967, Geology of the Scott Glacier and Wisconsin Range areas, central Transantarctic Mountains, Antarctica [unpublished Ph.D. thesis]: The Ohio State University, Columbus, Ohio, 268 p.


Noaa, 2019, NCEI Geomagnetic Calculators, National Geophysical Data Center, National Oceanic and Atmospheric Administration (NOAA), p. NOAA's National Centers for Environmental Information (NCEI), formerly the National Geophysical Data Center, and the collocated World Data Service for Geophysics, Boulder, operated by NOAA/NESDIS/NCEI, archive and make available geomagnetic data and information relating to Earth’s magnetic field and Earth-Sun environment, including current declination, geomagnetic field models and magnetic indices, geomagnetic observatory data, and geomagnetic surveys.


Pauls, K.N., 2014, Sedimentology and paleoecology of fossil-bearing, high-latitude marine and glacially influenced deposits in the Tepuel Basin, Patagonia, Argentina: University of Wisconsin - Milwaukee, Milwaukee, WI.


1960 Powell, R.D., 1990, Glacimarine processes at grounding-line fans and their growth to ice contact
deltas, in Dowdeswell, J.A., and Scourse, J.D., eds., Glacimarine Environments: processes and sediments:

and the stratigraphic record, in Barker, P.F., and Cooper, A.C., eds., Geology and seismic stratigraphy of

1966 Powell, R.D., and Cooper, J.M., 2002, A glacial sequence stratigraphic model for temperate,
glaciated continental shelves: Geological Society Special Publication, p. 215-244.


1970 Powell, R.D., Naish, T., and Anonymous, 2009, Glacial sequence stratigraphy of the ANDRILL
AND-1B core, Antarctica, p. 72-72.

1972 Raymond, A., and Metz, C., 2004, Ice and Its Consequences: Glaciation in the Late Ordovician,

Verlag, 551 p.


outcrop perspective into the evolution of deformation within mass-transport deposits: Marine and
Petroleum Geology.

1981 Rose, P.B., Grant; Vaughan, O.; Cater, J.;Rea, B.R.; Spagnolo, M.; Archer, S., 2018, Aviat: a Lower
Pleistocene shallow gas hazard developed as a fuel gas supply for the Forties Field, in Bowman, M.L., B.,
ed., Petroleum Geology of NW Europe: 50 Years of Learning – Proceedings of the 8th Petroleum Geology

reveals adynamic subpolar Antarctic Ice Sheet in Ross Sea during the late Miocene: Sedimentology, v. 66,
p. 2072–2097.

1988 Rust, I.C., 1975, Tectonic and sedimentary framework of Gondwana basins in southern Africa, in


1993 Ives and Isbell
SHG Glaciogenic Lithofacies
J. of Sedimentary Research

1994 Schomacker, A., and Benediktsson, Í.Ö., 2018, Supraglacial Environments, in Menzies, J., and van

1996 Scotese, C.R., and Barrett, S.F., 1990, Gondwana's movement over the South Pole during the
Palaeozoic: evidence from lithological indicators of climate, in McKerrow, W.S., and Scotese, C.R., eds.,

1999 Seegers-Szablewski, G., and Isbell, J.L., 1998, Stratigraphy and depositional environments of
Lower Permian post-glacial rocks exposed between the Byrd and Nimrod Glaciers, Antarctica: Journal of

2001 Seegers, G.M., 1996, Sedimentology of the Permian Mackellar Formation, Central Transantarctic

2004 Shaw, J., 2016, Paraglacial landscapes in St George's Bay, Newfoundland, Canada, in
Society, p. 97-98.

2008 Sheppard, K., Bell, T., and Liverman, D.G.E., 2000, Late Wisconsinan stratigraphy and
chronology at Highlands, southern St. George's Bay, southwest Newfoundland: Quaternary International,
v. 68-71, p. 275-283.


2013 Soreghan, G.S., Soreghan, M.J., and Heavens, N.G., 2019, Explosive volcanism as a key driver of
the late Paleozoic ice age: Geology, v. 47, p. 600-604.

erosion and weathering zones in the coastal mountains of Marie Byrd Land, Antarctica: Geomorphology,
v. 67, p. 317-334.

2018 Survis, S.R., 2015, Sedimentology and stratigraphy of high-latitude, glacigenic deposits from the
Late Paleozoic Ice Age in the Tepuel-Genoa Basin, Patagonia, Argentina: University of Wisconsin -
Milwaukee, Milwaukee, Wisconsin.

2021 Svendsen, J.I., and Mangerud, J., 1997, Holocene glacial and climatic variations on Spitsbergen,


Table 1. Names and locations of sedimentary sections described in this paper.

<table>
<thead>
<tr>
<th>Location Name</th>
<th>Section Name</th>
<th>Geographic Coordinates</th>
<th>Thickness of Pagoda Fm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mt. Butters 1</td>
<td>MB-17</td>
<td>S84° 51.029’ W177° 25.216’</td>
<td>90 m</td>
</tr>
<tr>
<td>Mt. Butters 2</td>
<td>MBSE-17</td>
<td>S84° 53.003’ W177° 22.354’</td>
<td>77 m</td>
</tr>
<tr>
<td>Reid Spur</td>
<td>RS-18</td>
<td>S84° 47.035’ E178° 46.680’</td>
<td>&gt; 62 m</td>
</tr>
<tr>
<td>Mt. Munson</td>
<td>MM-17</td>
<td>S84° 45.359’ E173° 41.118’</td>
<td>5 m</td>
</tr>
<tr>
<td>Mt. Weaver</td>
<td>MW-18</td>
<td>S86°58.354’ W153° 26.801’</td>
<td>0 m</td>
</tr>
</tbody>
</table>
Table 2. Facies and Facies Association Descriptions.

<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Facies Name</th>
<th>Thickness</th>
<th>Lithology</th>
<th>Structures</th>
<th>Formative Process</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basal Fine-Grain-</td>
<td>BFG1: Basal conglomerate</td>
<td>5 cm – 1 m</td>
<td>Polymictic, clast-supported, very poorly sorted, sandy, conglomerate that includes minor amounts of coarse, poorly-sorted sandstone and siltstone</td>
<td>Conglomerate sandstone beds are massive and 5 – 10cm thick, and laterally discontinuous, siltstones occur as drapes with rare polygonal syneresis cracks</td>
<td>Subaqueous, gravity-driven cohesive flows</td>
<td>Non-glacial lacustrine setting with at least seasonal lake ice coverage</td>
</tr>
<tr>
<td>Dominated (BFG)</td>
<td>BFG2: Horizontally-stratified fines</td>
<td>1 - 5 m</td>
<td>Siltstone to fine sandstone</td>
<td>Horizontal and laterally continuous laminae and thin beds with sharp contact are dominant, irregularly-shaped laminae and symmetrically-rippled beds are common</td>
<td>Settling from suspension of silt and fine-to-medium sands with some wave-ripening</td>
<td>Non-glacial lacustrine setting with at least seasonal lake ice coverage</td>
</tr>
<tr>
<td></td>
<td>BFG3: Horizontally-stratified fines with outsized clasts</td>
<td>1 - 5 m</td>
<td>Medium sandstone to siltstone with outsized clasts, clasts are angular to sub-rounded and coarse sand to granule size</td>
<td>Horizontal and laterally continuous laminae and thin beds with sharp contact are dominant, irregularly-shaped laminae and symmetrically-rippled beds are common; outsized-clasts are randomly distributed</td>
<td>Settling from suspension of silt and fine-to-medium sands with some wave-ripening; out-sized clasts are lake-ice transported and “dropped” in (i.e. ice-rafted-debris).</td>
<td>Non-glacial lacustrine setting with at least seasonal lake ice coverage</td>
</tr>
<tr>
<td></td>
<td>BFG4: Pervasively sheared fines</td>
<td>20 – 50 cm</td>
<td>Medium sandstone to siltstone with outsized clasts, clasts are angular to sub-rounded and coarse sand to granule size</td>
<td>Massive within decimeter-scale lenticular bodies by separated by curvilinear fractures, slickensides cover fracture surfaces; highly fissile laminae/structure</td>
<td>Subglacial soft-sediment deformation of facies BFG3</td>
<td>Non-glacial lacustrine setting with at least seasonal lake ice coverage</td>
</tr>
<tr>
<td></td>
<td>Massive Sandy Diamictite (MSD)</td>
<td>5 – 75 m (at least)</td>
<td>Clast-poor to clast-rich, sandy diamictite; muddy sand matrix; minor amounts of discontinuous sand and gravel bodies</td>
<td>Diamictites are massive to crudely bedded and ungraded, sands are generally massive, and sometime laminated, but always highly deformed</td>
<td>Glaciogenic, subaqueous processes; likely a combination of mass-transport, iceberg-rainout, iceberg scouring, plume sedimentation, and subglacial till deposition</td>
<td>Glacier-proximal to glacier-intermediate, continental shelf</td>
</tr>
<tr>
<td>Heterolithic Agglomeration (HA)</td>
<td>Laminated Sands (LS)</td>
<td>(~17) m</td>
<td>Coarse to fine-grained, well-sorted sandstones; quartz-rich; sub-angular to rounded</td>
<td>Fine- to medium-grained sands that occur in thin, planar beds with primary current lineations; coarse-grained sandstones are trough cross-bedded; Fine- to very-fine grained sandstones are laminated, or thinly-bedded with unidirectional ripples</td>
<td>High-density turbidites and/or a transitional concentrated density flow (Mulder and Alexander 2001)</td>
<td>Distal or medial portion of an ice-contact fan or delta</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>HA1: Bedded diamicite</td>
<td>1 – 10 m</td>
<td>Clast-rich, sandy diamicite; matrix is moderately well-sorted</td>
<td>3 – 7 cm beds that are massive with sharp, horizontal, and laterally discontinuous contacts; soft sediment deformation associated with facies HA2</td>
<td>Glaciogenic, subaqueous processes; likely a combination of iceberg-rainout and plume sedimentation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>HA2: Chaotic heterolithic</td>
<td>0 – 15 m</td>
<td>Conglomerates to very fine-grained sandstones</td>
<td>Bedding is usually massive, soft-sediment deformation is pervasive; few primary sedimentary structures preserved; secondary structures include fold noses, boudinage, faulting, shear structures above and below contacts, and ruck structures</td>
<td>Mass-transport, gravity-driven processes in the form of slides, slumps, and/or mass transport deposits.</td>
<td>Subaqueous glacier-proximal and grounding-line fan system</td>
</tr>
<tr>
<td></td>
<td>HA3: Stratified Heterolithic</td>
<td>0 – 15 m</td>
<td>Coarse- to very fine-grained sandstones</td>
<td>Medium- to coarse-grained sandstones are thickly laminated to bedded with planar cross beds, trough cross beds, climbing ripples, 3D ripples that are asymmetrical or climbing, hummocky and swaley cross stratification, and symmetrical ripples with bundled upbuilding; Very-fine and fine sandstones are laminated or thinly bedded include uni-directional ripples, some flaser ripples and climbing ripples; occasional, small-scale</td>
<td>Current- dominated transport and deposition, with some slumping; Unconfined flow; Poorly-sorted sediment source; large variations in current velocities; occasional wave re-working</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cross bedded Sandstone (CBS)</td>
<td>10 – 30 m</td>
<td>Well-sorted, medium-to very coarse-grained quartz arenite sandstone; rare pebbles, cobbles, and conglomerates</td>
<td>Low-angle and trough crossbeds; Crossbed sets range in thicknesses from ~15 cm to ~1.5 m; rare thin beds with asymmetrical ripples; Occur within amalgamated channels in multi-storied, multi-lateral sand sheet &lt;2 km wide</td>
<td>Strong tractive flow confined to a series of amalgamated channels; subaqueous</td>
<td>Unconfined, distributive flow; likely glacial-proximal subaqueous fan/grounding-line fan</td>
</tr>
<tr>
<td></td>
<td>Upper Fine-Grain-Dominated (UFG)</td>
<td>126 m + (lower 5 – 20 m included in this study)</td>
<td>Moderately-well-sorted to well-sorted, very-fine- and fine-grained sandstone; Rare outsized clasts (very-coarse sand to cobble) in lower 2 - 5 m</td>
<td>Thinly bedded to laminated; massive or asymmetrically rippled; ripples are unidirectional and often have climbing or flaser-like characteristics; Laterally continuous, and have sharp, horizontal, and planar contacts; Chaotically deformed and normally faulted zone overlying glide planes</td>
<td>Unconfined, unidirectional current-driven sediment transport and settling from suspension; slumps; ice-rafted debris</td>
<td>Mackellar Fm; Pro-delta and subaqueous channel levee with glacier influence</td>
</tr>
</tbody>
</table>
Table 3. Summary of Paleo-transport measurements from this study (mean = Fisher mean ± a99). Note that due to the proximity of the study locations to the south pole that the orientation of compass directions are not the same across the basin. T = Trend, P = Plunge, D = Dip angle, DD = Dip Direction, S = strike, * = measured during 1997 field season

<table>
<thead>
<tr>
<th>Section (height)</th>
<th>Facies</th>
<th>Feature</th>
<th>Measurement</th>
<th>Orientation</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Site MW-18</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10 m above basement</td>
<td>UFG</td>
<td>Asymmetrical ripples</td>
<td>Transport</td>
<td>T: 322°, 312°, 002°</td>
<td>3</td>
</tr>
<tr>
<td><strong>Site MM-17</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0 m</td>
<td>-</td>
<td>Striae on basement</td>
<td>Lineation</td>
<td>T: 180°</td>
<td>1</td>
</tr>
<tr>
<td>*0 m</td>
<td>-</td>
<td>Striae on basement</td>
<td>Lineation</td>
<td>T: 095° ± 3.7°</td>
<td>8</td>
</tr>
<tr>
<td>*4 m above base of Mackellar</td>
<td>UFG</td>
<td>Asymmetrical ripples</td>
<td>Transport</td>
<td>T: 157° ± 28.5°</td>
<td>7</td>
</tr>
<tr>
<td>13-17.5 m</td>
<td>UFG</td>
<td>Highly-deformed slump features</td>
<td>Vergece</td>
<td>T: 109°, 104°, 114°</td>
<td>3</td>
</tr>
<tr>
<td>*4 - 48 m of Mackellar Fm</td>
<td>UFG</td>
<td>Asymmetrical ripples</td>
<td>Transport</td>
<td>T: 109° ± 20.0°</td>
<td>10</td>
</tr>
<tr>
<td><strong>Site MB-17</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0 m</td>
<td>-</td>
<td>Dip and dip direction of basement</td>
<td>Mean pole to plane</td>
<td>T: 262°</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Plane from mean pole</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>S: 352°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>D: 11°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>DD: 262°</td>
<td></td>
</tr>
<tr>
<td>MB-17A/B</td>
<td>BFG</td>
<td>Symmetrical Ripple Crest Axes</td>
<td>Mean lineation</td>
<td>T: 346° ± 6.0°</td>
<td>17</td>
</tr>
<tr>
<td>MB-17B</td>
<td>BFG4</td>
<td>Slickenslide lineation</td>
<td>Mean</td>
<td>T: 006°</td>
<td>1</td>
</tr>
<tr>
<td>MB-17C (12 m)</td>
<td>MSD</td>
<td>Fold axes and small thrust faults</td>
<td>Mean Vergence</td>
<td>T: 220° ± 31°</td>
<td>7</td>
</tr>
<tr>
<td>MB-17C (20 m)</td>
<td>MSD</td>
<td>Plane of thrust faults</td>
<td>Vergence</td>
<td>T: 256°, 251°</td>
<td>2</td>
</tr>
<tr>
<td>MB-17C (58 m)</td>
<td>MSD</td>
<td>Slide/slump plane</td>
<td>Planes</td>
<td>DD: 243°, 191°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>S: 117°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>D: 23°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>DD: 206°</td>
<td></td>
</tr>
<tr>
<td>MB-17C/D (73 – 81m)</td>
<td>HA</td>
<td>Asymmetrical ripples</td>
<td>Mean transport</td>
<td>T: 325°, 236° - 055°</td>
<td>30</td>
</tr>
<tr>
<td>MB-17C/D</td>
<td>HA</td>
<td>Cross beds</td>
<td>Transport</td>
<td>T: 346°, 356°, 226°, 221°</td>
<td>5</td>
</tr>
<tr>
<td>MB-17D</td>
<td>HA</td>
<td>Climbing Ripples</td>
<td>Transport</td>
<td>T: 221°, 256°</td>
<td>2</td>
</tr>
<tr>
<td>MB-17C (84 m)</td>
<td>HA</td>
<td>Grooves (iceberg keel marks?)</td>
<td>Direction of shallowng</td>
<td>T: 253° ± 24°</td>
<td>5</td>
</tr>
<tr>
<td>MB-17C (18 m above Pagoda)</td>
<td>UFG</td>
<td>Asymmetrical ripples</td>
<td>Mean Transport</td>
<td>T: 219° ± 39°</td>
<td>7</td>
</tr>
<tr>
<td><strong>Site MBSE-17</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0 m</td>
<td>-</td>
<td>Striae on Basement</td>
<td>Lineation</td>
<td>T: 311°, 316°</td>
<td>2</td>
</tr>
<tr>
<td>MBSE-17 (29 m)</td>
<td>MSD</td>
<td>Sheafold Hing</td>
<td>Orientation Vergence</td>
<td>T: 016°, P: 20°</td>
<td>286˚</td>
</tr>
<tr>
<td>MBSE-17 (31 – 32 m)</td>
<td>MSD</td>
<td>Thrust faults</td>
<td>Mean pole to plane</td>
<td>T: 263°</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Plane from mean pole</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>S: 353°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>D: 21°</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>DD: 083°</td>
<td></td>
</tr>
<tr>
<td>MBSE-17 (54 m)</td>
<td>HA</td>
<td>Crenulations under slump/slide</td>
<td>Lineation</td>
<td>T: 241°, 251°</td>
<td>2</td>
</tr>
<tr>
<td>MBSE-17 (61 – 68 m)</td>
<td>CBS</td>
<td>Cross beds and asymm. ripples</td>
<td>Mean transport</td>
<td>T: 208° ± 29°</td>
<td>7</td>
</tr>
<tr>
<td>MBSE-17 (21 – 26 m above Pagoda)</td>
<td>UFG</td>
<td>Asymmetrical Ripples</td>
<td>Mean transport</td>
<td>T: 298° ± 6°</td>
<td>23</td>
</tr>
<tr>
<td><strong>Site RSP-18</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>24 – 48 m</td>
<td>LS</td>
<td>Cross beds, asymm. ripples, &amp; PCL</td>
<td>Mean transport</td>
<td>T: 176° ± 29°</td>
<td>18</td>
</tr>
<tr>
<td>68 – 70 m</td>
<td>UFG</td>
<td>Asymmetrical Ripples</td>
<td>Transport</td>
<td>T: 261°, 241°, 211°</td>
<td>281°</td>
</tr>
</tbody>
</table>
Table 4. Glacial transport directions from the literature and this study in the Pagoda Fm and equivalents in South Victoria Land, Beardmore Sub-basin, and Horlick Sub-basin. Directions are plotted in Figure 13. The orientation of magnetic/geographic north varies considerably with proximity to pole. Therefore, making direct comparisons between directional measurements is not advised. Transport direction of bi-directional features (striae) typically inferred from flow directions in adjacent or overlying strata. Data at Mt. Munson from Isbell (unpublished) are previously unpublished measurements that were collected during the 1995 field season.

* = measured from figure, not reported in text

<table>
<thead>
<tr>
<th>Point</th>
<th>Location</th>
<th>Study</th>
<th>Flow Orientation</th>
<th>Sedimentary Structure</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>Kennar Valley</td>
<td>Isbell (2010)</td>
<td>252° (toward), N=8</td>
<td>Large-scale thrust sheets</td>
</tr>
<tr>
<td>b</td>
<td>Mt. Metschel</td>
<td>Isbell (2010)</td>
<td>126° (toward), N=11</td>
<td>Deforming bed transport</td>
</tr>
<tr>
<td>c</td>
<td>Mt. Ritchie</td>
<td>Isbell (2010)</td>
<td>063° (toward), N=4, 307° (toward), N=5</td>
<td>Deforming bed transport, Large-scale thrust sheets</td>
</tr>
<tr>
<td>d</td>
<td>Colosseum Ridge CR2</td>
<td>Lenaker (2002)</td>
<td>294°</td>
<td>Striated clast pavement</td>
</tr>
<tr>
<td>e</td>
<td>Colosseum Ridge CR4</td>
<td>Lenaker (2002)</td>
<td>206°</td>
<td>Striated clast pavement</td>
</tr>
<tr>
<td>f</td>
<td>Haven Ridge</td>
<td>Lenaker (2002)</td>
<td>218°</td>
<td>Striated surface</td>
</tr>
<tr>
<td>g</td>
<td>Wallabies Nun.</td>
<td>Isbell (1997)</td>
<td>145°</td>
<td>Striated clast pavement marks</td>
</tr>
<tr>
<td>h</td>
<td>Geologist’s Range Mt Csejtey</td>
<td>Isbell (1997)</td>
<td>074°</td>
<td>Striae on basement</td>
</tr>
<tr>
<td></td>
<td>Sullivan Nunatak Arrowhead Nun.</td>
<td></td>
<td>068°</td>
<td>Striae on basement</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>129°</td>
<td>Grooves on mass transport</td>
</tr>
<tr>
<td>i</td>
<td>Moore Mtns Section P</td>
<td>Lindsay (1970)</td>
<td>114°</td>
<td>Soft-sediment grooves</td>
</tr>
<tr>
<td>j</td>
<td>Section V</td>
<td>Lindsay (1970)</td>
<td>184° ± 2°, N=3</td>
<td>Striated, Polished, Grooved surface</td>
</tr>
<tr>
<td>k</td>
<td>Section N</td>
<td>Lindsay (1970)</td>
<td>*156°</td>
<td>Grooved surface</td>
</tr>
<tr>
<td>l</td>
<td>Mt. Miller</td>
<td>Lindsay (1970)</td>
<td>182° ± 5° (oldest)</td>
<td>Grooves in basement</td>
</tr>
<tr>
<td></td>
<td>Section K</td>
<td></td>
<td>148° ± 4° (younger)</td>
<td>Striae</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>123° ± 6° (younger)</td>
<td>Striae</td>
</tr>
<tr>
<td>m</td>
<td>Mt. Mackellar Section I</td>
<td>Lindsay (1970)</td>
<td>*132°</td>
<td>Grooved surface</td>
</tr>
<tr>
<td>n</td>
<td>Mt. Mackellar Sections ABCD</td>
<td>Lindsay (1970)</td>
<td>188°, 123°</td>
<td>Pebble fabrics and boulder pavements</td>
</tr>
<tr>
<td>o</td>
<td>Buckley Island</td>
<td>Coates (1985)</td>
<td>144°</td>
<td>Grooves and striae on Alexandra Fm</td>
</tr>
<tr>
<td>p</td>
<td>Reid Spur</td>
<td>this study</td>
<td>176°</td>
<td>Mean current direction in grounding-line fan</td>
</tr>
<tr>
<td>q</td>
<td>Mt. Butters locations combined on map</td>
<td>Coates (1985)</td>
<td>180°</td>
<td>Elongate hump on granite surface</td>
</tr>
<tr>
<td></td>
<td></td>
<td>this study</td>
<td>186°</td>
<td>Slickenslide lineation in facies BFG4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>215°, 205°</td>
<td>iceberg keel gouges</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>131°, 136°</td>
<td>Striae on granitic basement</td>
</tr>
<tr>
<td></td>
<td>Location</td>
<td>Reference</td>
<td>Angles</td>
<td>Observations</td>
</tr>
<tr>
<td>---</td>
<td>-------------------</td>
<td>------------------</td>
<td>-----------------</td>
<td>-------------------------------------------------------------------------------</td>
</tr>
<tr>
<td>r</td>
<td>Mt. Munson</td>
<td>this study</td>
<td>180°</td>
<td>Striae on polished granite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>this study (1997)</td>
<td>095°</td>
<td>Striae on granite (N= 8)</td>
</tr>
<tr>
<td>s</td>
<td>Mt. Smithson</td>
<td>Coates (1985)</td>
<td><em>262</em>, <em>251</em></td>
<td>Grooves in tillite, craig and tail structures</td>
</tr>
<tr>
<td>t</td>
<td>Crown Mountain</td>
<td>Coates (1985)</td>
<td>*180°</td>
<td>Striae on metavolcanic bedrock</td>
</tr>
<tr>
<td></td>
<td>Section 16</td>
<td>Long et al. (2008)</td>
<td>130°, 090°</td>
<td>Fabrics of elongate pebbles and boulders at base</td>
</tr>
<tr>
<td>u</td>
<td>Roaring Valley</td>
<td>Long et al. (2008)</td>
<td>115°</td>
<td>Grooves on bedrock</td>
</tr>
<tr>
<td></td>
<td>Kutschin Peak</td>
<td>Coates (1985)</td>
<td>*155°</td>
<td>Striae on basement with beveling upstream</td>
</tr>
<tr>
<td>v</td>
<td>Crack Bluff</td>
<td>Long et al. (2008)</td>
<td>110°, 150°, 165°</td>
<td>Striae, grooves, crescent marks, &amp; r. moutonnée</td>
</tr>
<tr>
<td>w</td>
<td>Mt. Blackburn</td>
<td>Minshew (1967)</td>
<td>075°</td>
<td>roche moutonnée, nail and chatter marks on bedrock</td>
</tr>
<tr>
<td>x</td>
<td>Watson</td>
<td>Coates (1985)</td>
<td>160°</td>
<td>Orientation if roche moutonnée</td>
</tr>
<tr>
<td></td>
<td>Escarpment</td>
<td>Minshew (1967)</td>
<td>156°, 189°</td>
<td>Striae on basement</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>075°, 070°</td>
<td>roche moutonnée, nail and chatter marks on bedrock</td>
</tr>
<tr>
<td>y</td>
<td>Tillite Spur</td>
<td>Minshew (1967)</td>
<td>089°</td>
<td>Striae on bedrock and boulder pavements</td>
</tr>
<tr>
<td></td>
<td>Wisconsin Range</td>
<td>Frakes et al. (1966)</td>
<td>*288°</td>
<td></td>
</tr>
<tr>
<td>z</td>
<td>Mt. LeSchack</td>
<td>Minshew (1967)</td>
<td>086° ± 35</td>
<td>Plucked surfaces, nail markings, clast fabrics</td>
</tr>
<tr>
<td>α</td>
<td>Discovery Ridge</td>
<td>Frakes et al. (1996)</td>
<td>*255°</td>
<td>Straited &amp; plucked boulders, bedrock gouges &amp; striae</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>*227°</td>
<td>Paleocurrent directions (cross beds)</td>
</tr>
<tr>
<td></td>
<td>Discovery Ridge</td>
<td>Long (1964)</td>
<td>075°</td>
<td>Grooves and striae from “many levels”</td>
</tr>
<tr>
<td></td>
<td>&amp; other locals</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>β</td>
<td>Mt. Weaver</td>
<td>this study</td>
<td>109°</td>
<td>Asymmetrical ripples, Weaver Fm</td>
</tr>
<tr>
<td>r</td>
<td>Mt. Munson</td>
<td>this study</td>
<td>157°</td>
<td>Slump vergence</td>
</tr>
<tr>
<td></td>
<td></td>
<td>this study (1997)</td>
<td>109°</td>
<td>Current ripples 0 – 4 m above Mackellar Base</td>
</tr>
<tr>
<td></td>
<td></td>
<td>this study (1997)</td>
<td></td>
<td>Current ripple 4- 48 m above Mackellar Base</td>
</tr>
</tbody>
</table>

--- Non glacial flow directions plotted on Fig. 13 Map ---

β 

r