



Evaluation of physical and chemical proxies used to interpret past glaciations with a focus on the late Paleozoic Ice Age

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ABSTRACT

The late Paleozoic Ice Age (LPIA) was one of Earth's most important Phanerozoic climatic events lasting for over 100 Mys. Despite its importance, its history is controversial with two hypotheses that portray glaciation differently (Fig. 1). Traditional views characterize the LPIA as a continuous glacial event that lasted from the Middle Mississippian until the Late Permian with a massive ice sheet that covered Gondwana throughout this interval. This approach often uses only one or two proxies to define the glaciation. The other emerging hypothesis suggests that numerous ice sheets occurred in Gondwana with individual glacial events lasting up to 10 Mys alternating with glacial minima/non-glacial intervals of similar duration. Both views are still prevalent. Both near- and far-field proxies are used to define the ice age. Near-field proxies include the occurrence/absence of diamictites, glaciotectionic deposits/landforms, striated clasts and clast pavements, outsized clasts (dropstones), rhythmites, cyclic diamictite-bearing successions, glendonites, grooved and striated surfaces, streamline landforms, and U-shaped paleovalleys. Detrital zircons and chemical index of alteration (CIA) studies help to delineate the occurrence, extent, and location of glaciation. Multiple complexities occur with the use of these proxies as different non-glacial processes and driving factors can produce similar features or results. Far-field proxies focus on identifying changes in eustasy. These include the occurrence of cyclic successions composed of alternating nonmarine and marine strata (cyclothem), depth of incised valleys, paleotopographic relief, phosphatic black shales, and changing oxygen isotope ratios. Like the near-field record, far-field proxies are complex indicators with varied nuances that make their application challenging. Here we discuss the limitations and use of these proxies and promote a multiproxy approach to investigating Earth's glacial intervals. We suggest that studies incorporate multiple proxies coupled with detailed environmental, paleoflow, and paleogeographic analyses to better constrain the occurrence, timing, and extent of glaciation and its influence on global systems. This approach will provide a robust view of the LPIA. We also consider the magnitude and nature of sea-level response to changing ice volumes by discussing ice-volume fluctuations, basin subsidence's modification of glacioeustasy, and sea-level's response to global isostatic adjustment (GIA). In considering these features, it becomes apparent that glacioeustasy is more complex than previously envisioned.

1. Introduction

The late Paleozoic Ice Age (LPIA; Late Devonian to Late Permian) and its transitions from greenhouse-to-icehouse-to-greenhouse was one of Earth's most important climatic events as it played an important role in driving linked oscillations in climate (e.g., Poulsen et al., 2007; Rosenau

et al., 2013; Cecil et al., 2014), ice volume (e.g., Crowley and Baum, 1991, 1992; Isbell et al., 2003, 2016), sea level (Wanless and Shepard, 1936; Heckel, 1994, 2008; Rygel et al., 2006; Horton et al., 2012), and floral and faunal restructuring (e.g., Stanley and Powell, 2003; Stephenson et al., 2007); DiMichele et al., 2009; Shi and Waterhouse, 2010; DiMichele, 2014). Despite its importance (e.g., Eyles, 1993;

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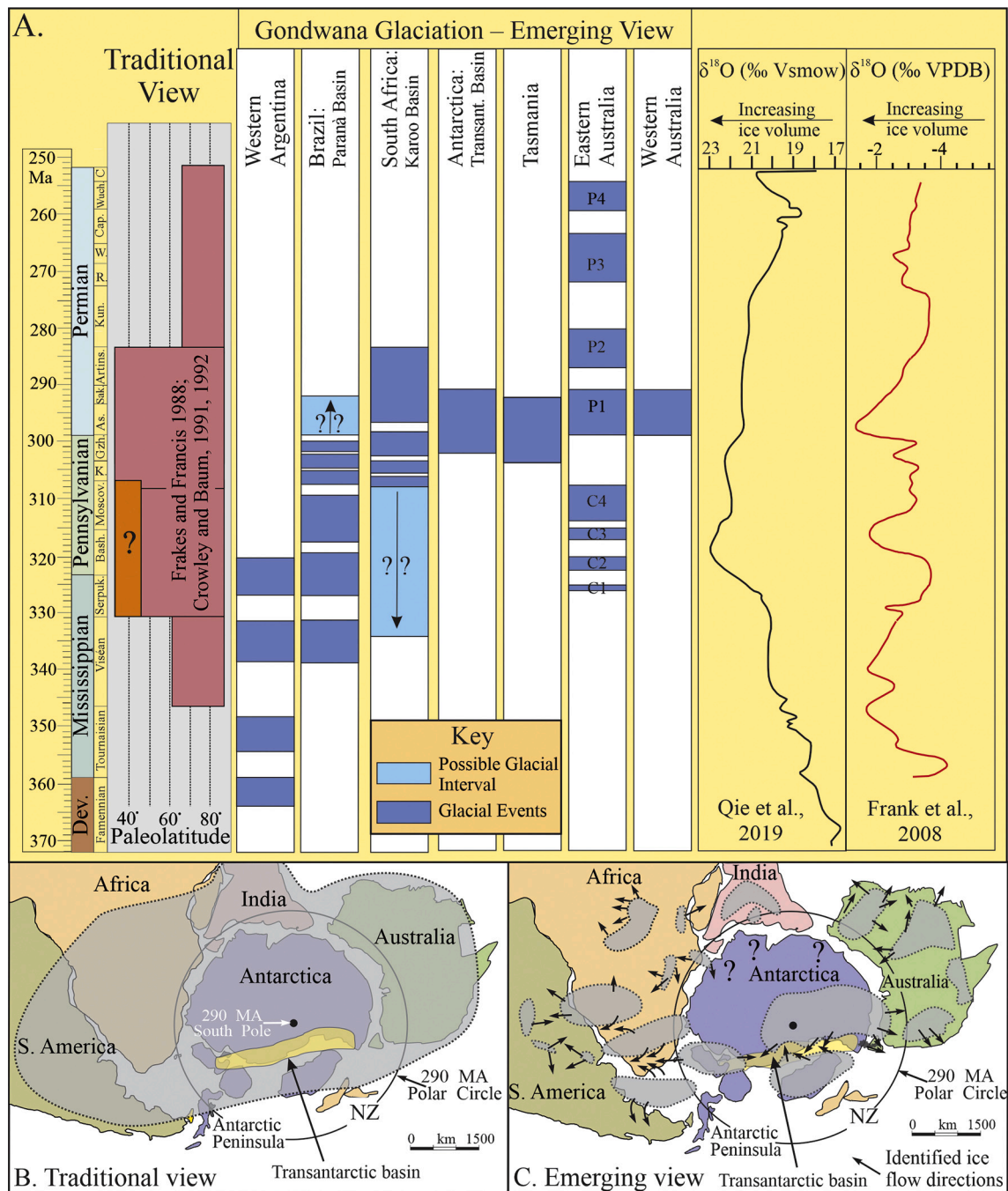


Fig. 1. Traditional and emerging views of the timing and extent of glaciation during the late Paleozoic Ice Age. A) Chronostratigraphic chart with both traditional (Frakes and Francis, 1988; Crowley and Baum, 1991, 1992) and emerging data fields for selected locations (modified from Isbell et al., 2012, 2016; López-Gamundí et al., 2020; Rosa and Isbell, 2021) and $\delta^{18}\text{O}$ curves. The $\delta^{18}\text{O}$ curve from Qie et al. (2019); (VSMOW = Vienna Standard Mean Ocean Water) is a composite curve synthesized from Buggisch et al. (2008), Joachimski et al. (2009) and Chen et al. (2016). The $\delta^{18}\text{O}$ curve by Frank et al. (2008); (VPDB = Vienna Pee Dee belemnite) is a composite from work reported by Brand (1989), Mii et al. (1999), Brand and Brenckle (2001), Veizer et al. (1999), Korte et al. (2005), and Saltzman (2005). B) Traditional reconstruction with a massive, Gondwana-wide ice sheet (modified from , Scotese and Barrett, 1990; Scotese, 1997; Isbell et al., 2012). C) Emerging view of glacial distribution during the latest Pennsylvanian–Early Permian (Cisuralian; Modified from Isbell et al., 2012).

Fielding et al., 2008a; Montañez et al., 2007; Isbell et al., 2012; Montañez and Poulsen, 2013; Chen et al., 2016;), an accurate understanding of glaciation (e.g., glacial dynamics, ice-volume/sea-level fluctuations, location of ice centers, timing of glacial/interglacial episodes, and paleoclimate) remains elusive (Vesely, 2020).

Researchers widely employ sedimentologic, stratigraphic and isotopic proxies to define glacial intervals during the Phanerozoic, and specifically the LPia. Our evolving understanding of these proxies has led to conflicting interpretations of LPia glaciations. Many LPia models,

depicted a Gondwana-wide ice sheet that lasted for over 100 Mys (Fig. 1; Du Toit, 1937; Lindsay, 1970; Veevers and Powell, 1987; Scotese and McKerrow, 1990; Ziegler et al., 1997; Blakey, 2008). These models are still used today (e.g., Starck et al., 2020), which led Craddock et al. (2019) to “ponder the possibility of a late Paleozoic snowball Earth.” Recently, the establishment of better time control on deposits and a judicious application of the various proxies used to characterize the LPia suggest that a series of smaller, widespread, and more dynamic ice centers occurred diachronously across Gondwana. These ice centers

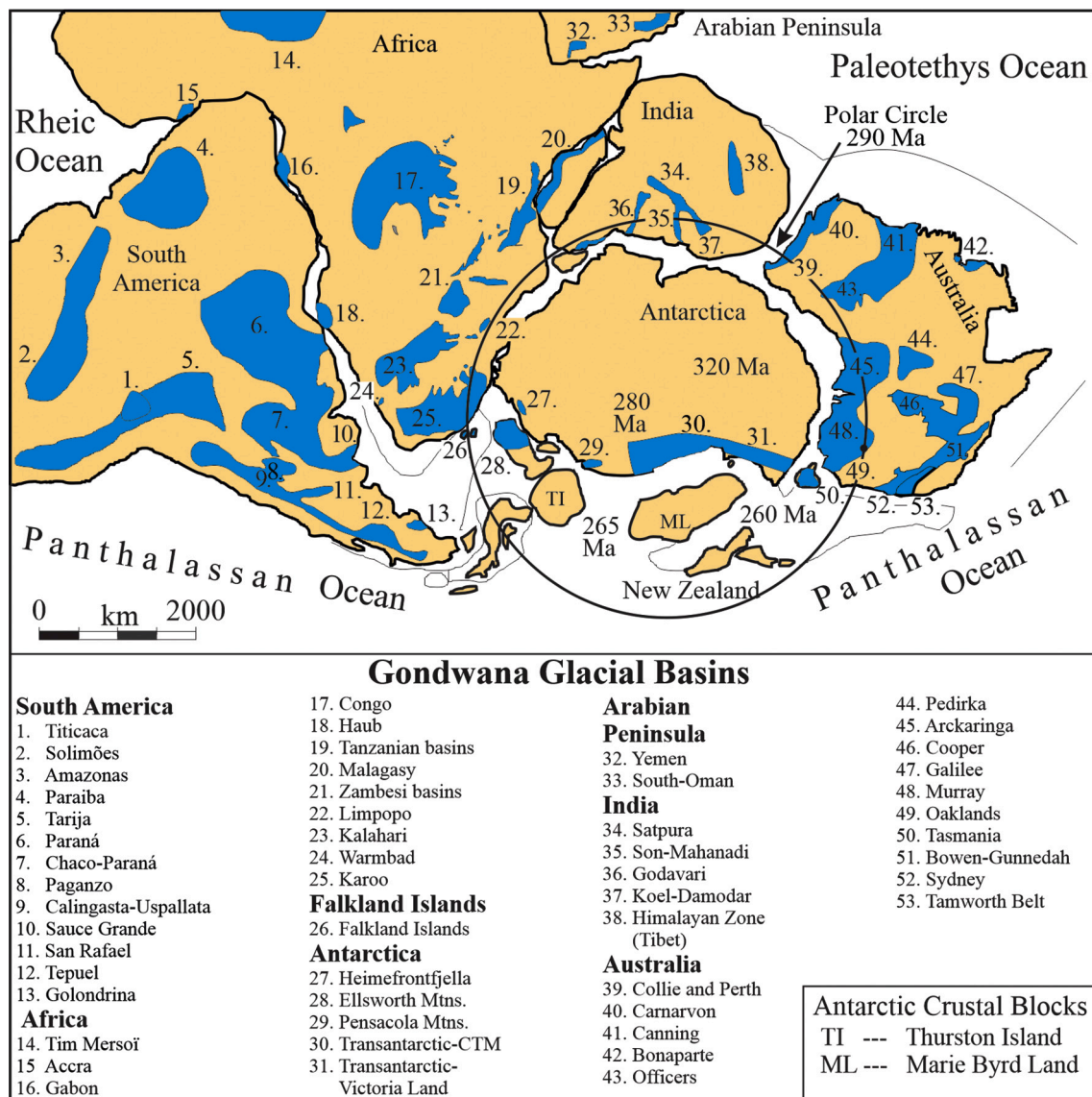


Fig. 2. Distribution of Gondwana basins containing glacial, glaciomarine, and glacially influenced marine strata. Note that glaciation was diachronous across Gondwana and that many of the basins contain glaciomarine records (See Fig. 1A; Modified from Eyles, 1993; Isbell et al., 2012; Rosa and Isbell, 2021). Glaciers nucleated in areas between the basins and shed ice and sediment into the adjacent basins (See Fig. 1C).

waxed and waned with individual glacial events lasting as long as 10 Mys alternating with ice-minimal or non-glacial intervals of equal duration (Fig. 1; cf. Visser, 1997; López-Gamundí, 1997; Isbell et al., 2003, 2012; Vesely and Assine, 2004, 2006; Fielding et al., 2008a, 2008b, 2008c; Horton and Poulsen, 2009; Montañez and Poulsen, 2013; Frank et al., 2015; Griffis et al., 2019; López-Gamundí et al., 2020; Rosa and Isbell, 2021). Because of the importance of proxies in defining Earth's climatic history, an evaluation of their use and limits is warranted as studies try to better define this and other ice ages.

Proxies used in defining individual glacial events within icehouse intervals, and their magnitude and duration, include both near- and far-field indicators. Near-field indicators include the occurrence of: diamictites; grooved and striated surfaces on bedrock and soft-sediment; soft-sediment folding and faulting; streamlined basement erosional features (e.g., whalebacks, roche moutonnée); U-shaped valleys; striated, faceted, and/or bullet-shaped clasts; limestones (interpreted as dropstones); rhythmites; and glendonites (Lindsay, 1970; Thomas and Connell, 1985; Fielding et al., 2008b, 2008c; Isbell, 2010; Vesely and Assine, 2014; Assine et al., 2018; Valdez Buso et al., 2020). However, most of these features can be produced by both glacial and non-glacial

processes (Gilbert, 1990; Posamentier and Martinsen, 2011; Moxness et al., 2018; Vesely et al., 2018). Far-field proxies include the occurrence of cyclothems, 3rd to 5th order eustatic changes, the occurrence of incised valleys within coastal plain successions, paleotopographic relief, and changing geochemical signals (e.g., $\delta^{18}\text{O}$; Heckel, 1994, 2008; Rygel et al., 2008; Eros et al., 2012; Chen et al., 2016). Nevertheless, most of these proxies represent complex responses to changing Earth processes (i.e., glacioeustasy, isostatic adjustment, aquifer eustasy, basin accommodation, climate; Jervey, 1988; Rygel et al., 2008; Whitehouse, 2018; Sames et al., 2020) that may or may not be driven by extensive glaciations.

1.1. Near-field examples of evolving views: Permian Transantarctic Basin

As an example of the application of near-field proxies and how their varied use result in disparate degrees of interpretation (one large ice sheet vs. many smaller ice sheets), we offer the following example from the Permian Transantarctic Basin (Figs. 1B, 2), a narrow (~100 km wide), elongate (~1100 km long) basin oriented sub-parallel to the modern trend of the Transantarctic Mountains, Antarctica (Collinson,

1990; Collinson et al., 1994). Early reports of Lower Permian strata in the Metschel Tillite, Darwin Tillite, Pagoda, and Whiteout Conglomerate formations, and other equivalent strata, identified diamictites as lodgement tills deposited subglacially from a single ice sheet of considerable extent (e.g., Long, 1964; Lindsay, 1970; Barrett and McKelvey, 1981; Barrett et al., 1986; Miller, 1989; Isbell et al., 1997, 2001). Under this interpretation, grooved and striated bedrock and soft-sediment surfaces resulted from subglacial abrasion (Lindsay, 1970; Miller, 1989). Lindsay (1970) and Barrett et al. (1986) concluded that the ice sheet extended out over Antarctica from an ice center located in Victoria Land. Later, researchers inferred that Gondwana was covered by a massive ice sheet extending out of Antarctica (cf. Veevers and Powell, 1987; Scotese and McKerrow, 1990; Ziegler et al., 1997; Blakey, 2008).

In contrast, more selective applications of proxies have determined that the Antarctic diamictites occurred in a wide variety of environmental settings representing erosion and deposition by subglacial (sliding, deforming, and lodgement), rockfall, glaciomarine (grounding line, morainal bank, meltwater plume, iceberg and sea/lake ice rafting; iceberg turbation), sediment gravity flow, and mass transport processes (e.g., Matsch and Ojakangas, 1991, 1992; Lenaker, 2002; Isbell et al., 2008, 2012; Isbell, 2010; Koch and Isbell, 2011; Cornamusini et al., 2017; Ives and Isbell, 2021). In this scenario, grooved and striated surfaces resulted from subglacial ploughing/sliding, turbation by icebergs, and as mass transport glide planes (Isbell et al., 2008). Such studies contributed to the identification of multiple ice sheets, that supplied sediment and ice to the basin (Figs. 1C, 2; Isbell et al., 2008; Isbell, 2010; Elliot et al., 2016; Ives and Isbell, 2021; Zurli et al., 2021). Ice transversely entered the trough-shaped basin from ice sheets located on the East Antarctic Craton and in Marie Bryd Land with subglacial and ice proximal glaciomarine sedimentation occurring along basin margins while predominantly meltwater plume, iceberg rafted, sediment gravity flow and mass transport deposition occurred along the basin axis (Lenaker, 2002; Isbell et al., 2008; Isbell, 2010; Ives and Isbell, 2021). This emerging view contributed to the idea that multiple ice sheets occurred scattered across Antarctica and Gondwana (Isbell et al., 2003, 2012; Fielding et al., 2008a; Montañez and Poulsen, 2013; Rosa and Isbell, 2021).

1.2. Far-field examples of evolving views: cyclothems

An evolving view of cyclothems and eustasy illustrates how far-field proxies influence interpretations of the LPIA. Many researchers adhering to traditional views of glaciation hypothesize that the waxing and waning of massive Gondwana glaciers served as the primary drivers of eustatic change and the deposition of cyclothems (e.g., Wanless and Shepard, 1936; Heckel, 1986, 1994, 2013; Algeo and Heckel, 2008; Rygel et al., 2008; Le Cotonnec et al., 2020). However, attributing eustatic change during the late Paleozoic entirely to glaciation ignores contributions made to global and relative sea-level change made by thermal expansion of sea water, water stored in and discharged by aquifers (aquifer eustasy), basin accommodation, and glacial isostatic adjustment (cf. Coe and Church, 2003; Catuneanu, 2006; Rygel et al., 2008; Cazenave and Llovel, 2010; Church et al., 2011; Whitehouse and Bradley, 2013; Dietrich et al., 2019; Sames et al., 2020). Thus, developing views of late Paleozoic sea level and cyclothems involve consideration of complex interactions between multiple driving mechanisms. Likewise, changing $\delta^{18}\text{O}$ values recovered from benthic organisms are used to suggest changing water volumes in the world's oceans, much of which is assumed to have cycled through glacial ice. In some studies, (e.g., Buggisch et al., 2008; Chen et al., 2013, 2016), it is assumed that $\delta^{18}\text{O}$ values used to calculate sea-level changes for the Last Glacial Maximum to Holocene interval (0.08–0.11‰ for each 10 m change in sea level; Fairbanks and Matthews, 1978; Fairbanks, 1989) are directly applicable to magnitudes of sea-level change during the LPIA. However, differing ocean basin volumes between the late Paleozoic and the Pleistocene/

Holocene, long term temperature changes over the LPIA, the space occupied by marine ice sheets or ice shelves, and the extent of flooding of the continents among other factors, likely influenced the relationship between $\delta^{18}\text{O}$ and sea level during the LPIA. Such problems will be addressed in our discussion of far-field proxies.

1.3. Objectives

Although researchers utilize physical proxies to identify glaciation in deep time, it is apparent that multiple factors often contribute to the development of these features and their various coupled relationships. This review attempts to evaluate and clarify the use of various proxies employed in delineating LPIA glaciation and other glacial intervals throughout Earth's history. It also assesses how ice volume, basin accommodation (eustasy and subsidence) and glacial isostatic adjustment influence relative sea-level changes as seen within various basins and across the world's oceans, and how those factors result in deviations from what is expected for a glacioeustatic signal. Although this review may appear to question the value of previous LPIA studies by examining the proxies used to identify icehouse conditions, our intent is to identify the limits of these proxies to aid in understanding widespread and complex icehouse events and to encourage a multiproxy approach to study the LPIA and other glacial intervals, thus making interpretations of these important climatic intervals more robust.

2. Near-field proxies and their limitations

2.1. Depositional proxies

Studies of potential glaciogenic and glacially influenced depositional environments use an array of near-field sedimentologic, stratigraphic, and geochemical proxies to infer the occurrence of glaciers as well as their extent and duration. Although many of these proxies result from glacial activity, they may also result from non-glacial processes, and are therefore potentially ambiguous when considered outside a glacial context. Near-field depositional proxies include the occurrence of: 1) diamictites, 2) ice-marginal glaciotectionism and deposition, 3) striated and faceted clasts and clast pavements, 4) outsized and striated clasts (lonestones/dropstones), 5) rhythmites, 6) glaciogenic sequences and cycles, and 7) glendonites. Erosional proxies include: 1) grooved and striated surfaces, 2) streamline landforms, and 3) paleovalleys (cf. González Bonorino and Eyles, 1995; Fielding et al., 2008b, 2008c; Milana and Di Pasquo, 2019). Chemical proxies include: 1) detrital zircon studies and 2) the chemical index of alteration (CIA; cf. Goldberg and Humayun, 2010; Fedorchuk et al., 2019b, 2020; Griffis et al., 2019; Pauls et al., 2019). While previous authors (e.g., Eyles and Januszczak, 2004; Fielding et al., 2008b, 2008c) have noted that the identification of glaciogenic units should be based on the occurrence of multiple criteria, or where a single unmistakable example occurs, in implementation, there are numerous examples where this practice was not observed resulting in later studies suggesting that alternate non-glacial conditions occurred (cf. Socci and Smith, 1987, 1990; Passchier and Erkanure, 2010; Carto and Eyles, 2012; Socha et al., 2014; Tedesco et al., 2016; Moxness et al., 2018; Fedorchuk et al., 2019b; Pauls et al., 2020; Valdez Buso et al., 2020).

2.1.1. Diamictites

Many LPIA studies use diamictites as the primary proxy for identifying glacial events in time and space. Traditional models typically equate the occurrence of a diamictite as the result of glacial deposition and the global distribution of diamictites as the maximum extent of glaciation (e.g., Frakes and Francis, 1988). By encircling the area encompassed by such deposits, a continuous glacial cover is assumed, which then supports the idea of a single massive ice sheet covering much of Gondwana in the Carboniferous and Early Permian (cf. Veevers and Powell, 1987; Scotese and McKerrow, 1990; Ziegler et al., 1997; Blakey,

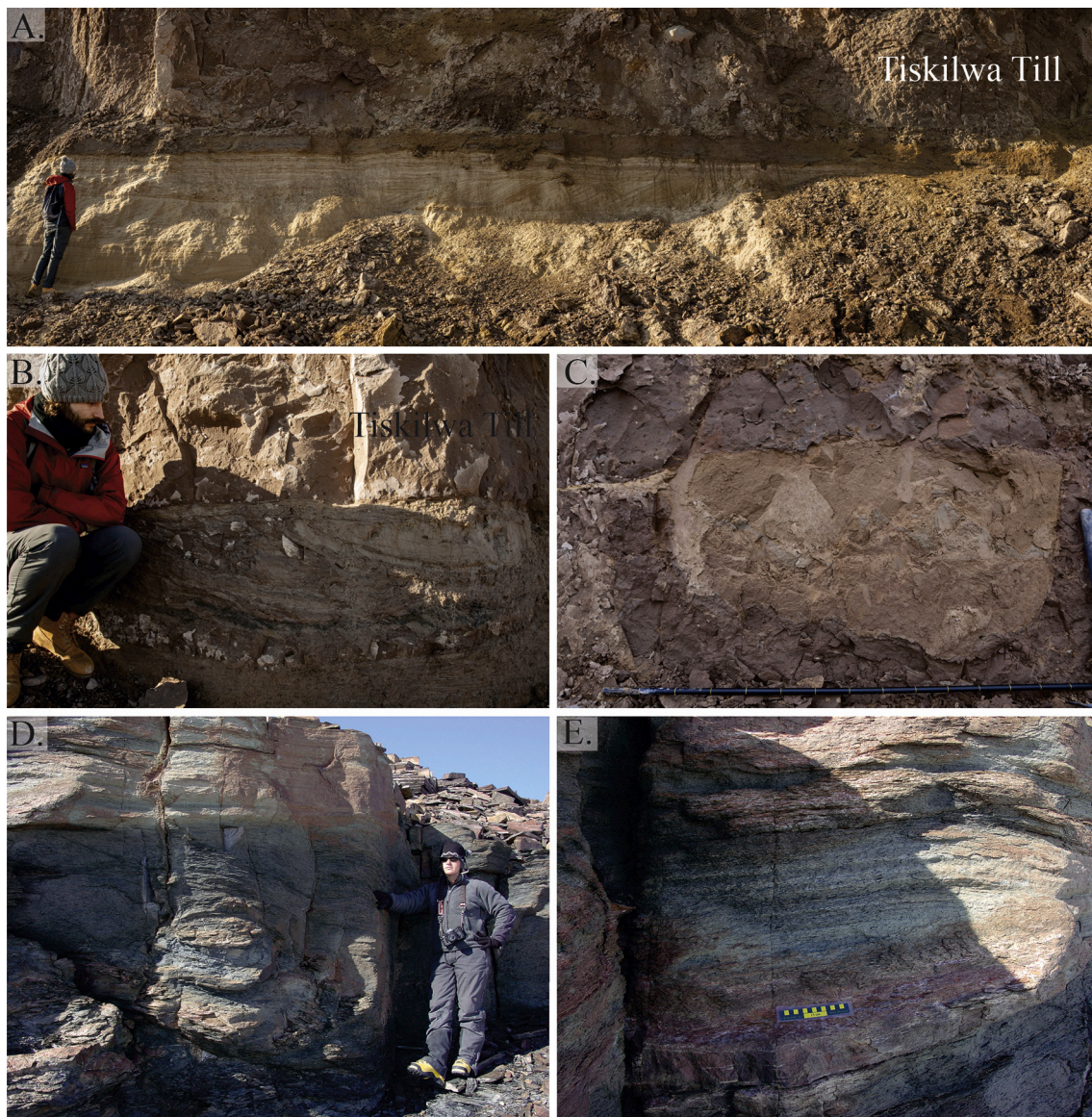


Fig. 3. Subglacial glacioteconites and traction tills from the Permian and Pleistocene. A) Pleistocene Tiskilwa Till, Lake Michigan bluffs, Cudahy, WI, USA. The massive diamict at the top is a traction till that overlies a thin glacioteconite developed in lacustrine muds and sands. Cross-stratified outwash sands occur at the base of the profile. Person for scale. B) The contact between a glacioteconite and the Tiskilwa traction till, Cudahy WI. Person for scale. C) Sheared siltstone clast within the Tiskilwa Till. Note the tail extending away from the clast to the left. Increments on the Jacob's staff are 10 cm. D) A glacioteconite in the Permian Pagoda Formation, Tillite Glacier, Antarctica. Note increasing degree of deformation upwards within the deposit. Person for scale. E) Glacioteconitic laminations at the top of deposit shown in photo 3D. Ruler is 15-cm long.

2008) and mid Carboniferous and Permian glaciation in the Russian Far East (Raymond and Metz, 2004). However, this simplistic view does not recognize the range of environments and processes that deposit, erode, and deform both glacial and non-glacial diamicts (Eyles and Lazorek, 2014; Isbell et al., 2016; Vesely et al., 2018; Dietrich et al., 2019; Rosa and Isbell, 2021). These processes and environments include subglacial, proglacial ice-contact, glaciomarine (proximal and distal), non-glacial or glacially-influenced debris flows and mass transport, and volcanic mudflow deposition.

Subglacial diamicts: Subglacial processes produce highly sheared and over compacted diamicts (glacioteconites and subglacial traction tills) that may appear homogeneous, or heterogeneous, and occur with or without readily apparent shear structures (Figs. 3, 4; e.g., Benn and Evans, 1996; Evans et al., 2006; Evans, 2018; Menzies et al., 2018). These deposits rest on unconformities with striated surfaces, striated clast pavements, décollement surfaces, or non-diamict deposits

displaying an upward increase in soft-sediment shear strain (Figs. 3B, D, 4; Van der Wateren, 2002; Evans et al., 2006; Evans, 2018; Le Heron et al., 2020). Both glacioteconites and subglacial traction tills display an upward increase in shear strain with shear strain highest at the former ice-substrate interface (Evans et al., 2006). Glacioteconites represent the deformation of pre-existing soft sediment and sedimentary rocks that retain some primary structures of the original material (Figs. 3B, D, E, 4A, B), while subglacial traction tills represent sediment deposited from the sole of a glacier and deformed by the overriding glacier (Figs. 3A-C; 4C, D; Evans, 2018). Deformation in subglacial materials can include both ductile (asymmetrical folds, foliation, tectonic laminations rotational turbates, necking structures, clast boudinage, unistrial and plasmic fabrics) and brittle features (thrust faults, hydrofracture, Riedel shears; e.g., Menzies et al., 2018; Phillips, 2018). Additional processes identified in subglacial traction tills include abrasion of large clasts, clast collision, grain breakage, and the development of micro-shears that are

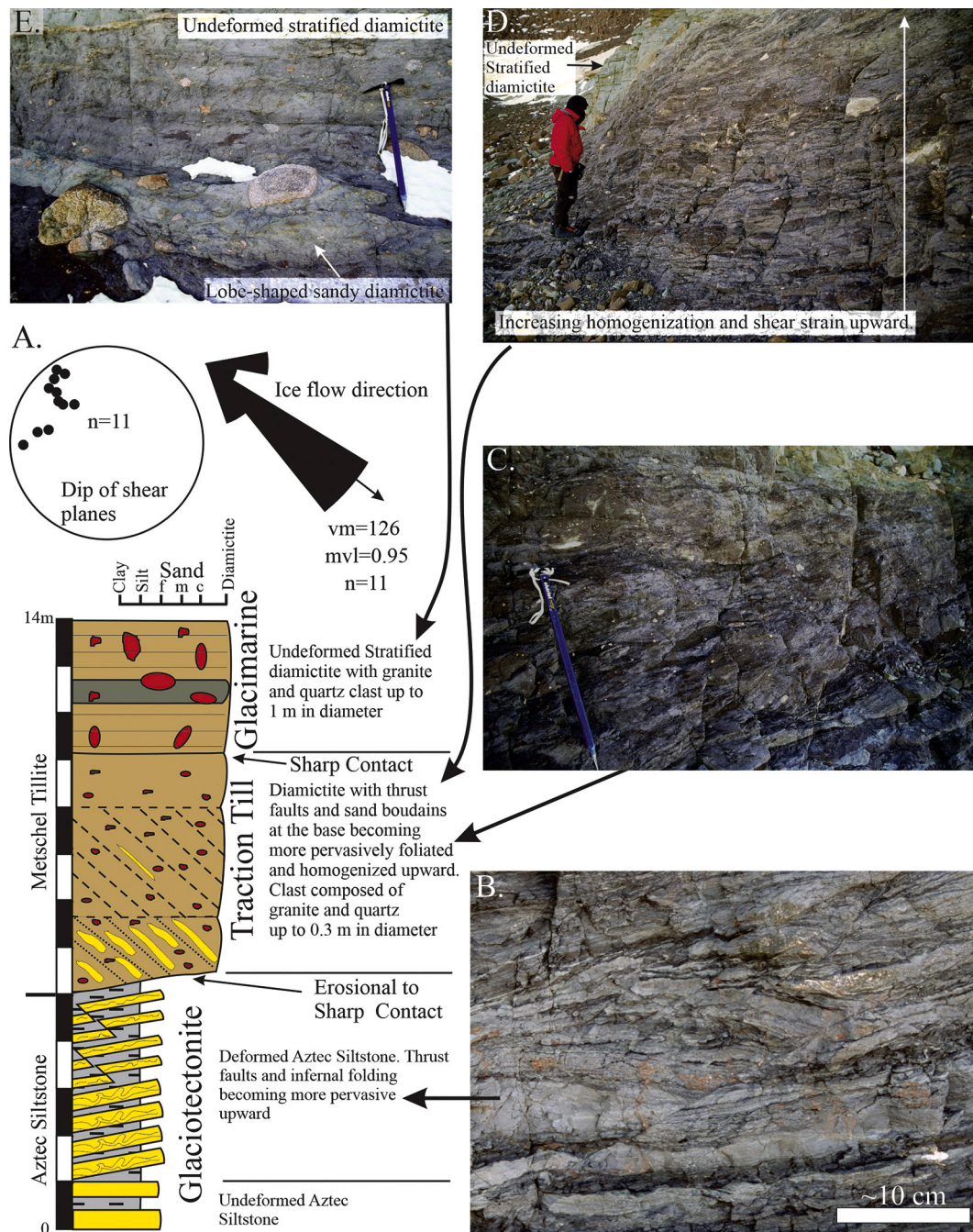


Fig. 4. A) Stratigraphic column of the glacioteconite and traction till deposits at the base of the Permian Metschel Tillite, Mount Metschel, southern Victoria Land, Antarctica. Note increasing deformation upwards. B) Thrust faulting with cm-scale displacement (glacioteconite). C) Pervasively foliated diamictite (traction till). Ice axe for scale. D) Pervasively foliated and partially homogenized diamictite (traction till). Person for scale. E) Stratified diamictite (interpreted as a glaciomarine diamict) with lobe-shaped, boulder-bearing, sandy diamictite (interpreted as a debrite) near the bottom of the photo. Ice axe for scale. Diagram is modified from Isbell (2010).

manifested as fissility in the deposits. At the ice-sediment interface, subglacial deformation, particle lodgement, clast ploughing, and basal sliding form a continuum representing different subglacial water pressure conditions allowing the glacial sole to be either coupled or decoupled from the underlying substrate (Evans et al., 2006; Evans, 2018). Deformation occurs when the glacier is coupled to its substrate (Evans et al., 2006). Ploughing occurs where clasts frozen into the base of the glacier are dragged along by the moving ice producing linear grooves (Fig. 5). A prow forms in front of the clast as it bulldozes its way through the sediment, ultimately lodging as the friction of the developing prow exceeds the force of the forward moving ice (Clark and

Hansel, 1989; Jørgensen and Piotrowski, 2003; Benn and Evans, 2010). Sliding occurs as rising water pressure allows decoupling of the glacier from the substrate and the sole of the glacier slides over the underlying sediment (Evans et al., 2006; Evans, 2018). Examples of glacioteconites, subglacial traction/deformation and clast ploughing include strata from the Neoproterozoic Chuos Formation of Namibia (Domack and Hoffman, 2011; Busfield and Le Heron, 2013), the Carboniferous Itararé Group, Brazil (Rosa et al., 2019); the Permian Pagoda Formation in the Queen Alexander Range, Antarctica (Figs. 3D, E; Isbell et al., 2008; Koch, 2010), the Permian Metschel Tillite of southern Victoria Land, Antarctica (Fig. 4; Isbell, 2010) and the Pleistocene Tiskilwa Till exposed

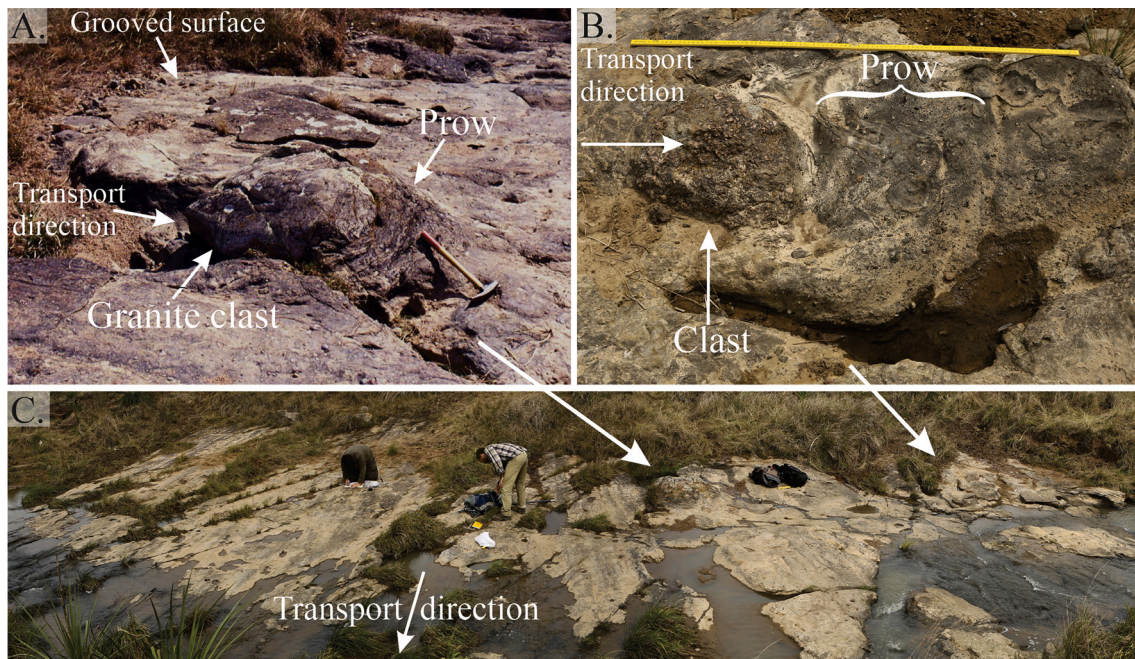


Fig. 5. Clasts lodged into a grooved and striated surface in the Taciba Formation, Itararé Group, near Cachoeira do Sul, Rio Grande do Sul, Brazil. A-B) Ploughed clasts showing prow structures. Hammer and meter stick for scale. C) Grooved surface with lodged clasts. Note the location of the clasts shown in 5A and B. People for scale.

along the Lake Michigan Bluffs in Cudahy, Wisconsin, USA. Figs. 3, 4, and 5 show examples of subglacial tills.

Glaciomarine diamicts: Subglacially deposited and deformed diamicts (glacioteconites and subglacial traction tills) occur in both terrestrial and glaciomarine settings (Evans et al., 2006; Eyles and Lazorek, 2014; Evans, 2018). Diamicts also form in pro-glacial, marine and estuarine settings when the glacial margin is grounded subaqueously (Powell, 2005; Powell and Alley, 1997). Glaciomarine diamicts form from two-component sedimentation of fines (clay, mud, and fine sand) settling from meltwater plumes and coarse material released from iceberg rafting (Fig. 5A; Smith and Andrews, 2000; Powell and Domack, 2002; Powell, 2005). Such deposits can form glaciomarine diamicts in ice proximal settings up to tens of km from grounded tidewater ice fronts (Powell and Alley, 1997). They may display sharp or gradational basal and upper contacts and occur as clast-rich or clast-poor diamicts that are internally massive or stratified (Fig. 6A) depending on sedimentation rates, meltwater discharge fluctuations, delivery of ice-rafted material, or wave and tidal activity (Powell and Domack, 2002; Eyles and Eyles, 2010). Ice-rafted debris may penetrate or disrupt stratification, if present, and have apparent a-axis orientations that are either parallel or at high angles to bedding depending on size and shape of clasts, and the water content/strength of the substrate at the time a clast impacts the sediment-water interface (cf. Thomas and Connell, 1985; Gilbert, 1990; Bronikowska et al., 2021). Rockfall off the front of glaciers (Fig. 6B) or valley walls (glacial and non-glacial valleys), iceberg turbation, bioturbation, and liquefaction also produce diamicts (Eyles et al., 1993; Dowdeswell et al., 1994; Powell, 2005; Isbell, 2010; Murray et al., 2013; Moxness et al., 2018).

Gravity-induced diamicts: Gravity-induced subaqueous processes form a continuum of deposits that display various stages of disaggregation of original sediment packages from slides to slumps to debrites to turbidites (Nemec, 1990; Posamentier and Martinsen, 2011; Rodrigues et al., 2020; Rodrigues et al., 2021). Slides represent cohesive masses with minor internal deformation. Varying degrees of internal deformation characterize slumps. Thus, slumps represent the breakup of slides as they transition to debris flows (Cossey, 2011). Together, chaotic slumps and debris flows represent chaotic and bedded diamicts (Figs. 6C, D)

that in some cases are difficult to distinguish from subglacial diamicts (Visser, 1983; Moscardelli and Wood, 2007; Eyles and Eyles, 2010; Vesely et al., 2018; Rodrigues et al., 2020). Debrisites from sediment gravity flows consist of clast-poor or clast-rich diamicts that display a range of internal organization that includes: 1) internal grading, 2) inverse grading, 3) a lower sheared layer and an upper ridged plug, or 4) appear internally massive (Fig. 6E; Mulder and Alexander, 2001). Clasts may protrude from the tops of cohesive end members (Fig. 6E), while some deposits end in coarse, clast-supported noses (Moxness et al., 2018). Individual deposits range from mm to tens of m thick (Figs. 6E, F). Basal contacts of gravity-driven sedimentary deposits are sharp where the flow was non-erosive or hydroplaning, grooved and striated where the flow eroded the substrate, intruded by flames and diapirs where flows overrode liquified substrates, or deposits may overlie sheared and deformed strata where the flow moved across unlithified materials (Phillips, 2006; Sobiesiak et al., 2018). On a microscopic scale, the diamicts contain turbate or rotational structures (Fig. 6H) indicating differential shearing and flame/diapiric structures along with detached diamictite pellets where flows moved across underlying liquified substrates (Phillips, 2006). Therefore, basal contacts of subglacial and gravity-driven flow deposits may mimic each other (cf. Van Der Wateren, 2002; Menzies et al., 2016; Evans, 2018; Sobiesiak et al., 2018). However, deformation in subglacial deposits increases upward toward the former ice sediment interface (Van Der Wateren, 2002); whereas, deformation within debris flows decreases upward away from a basal sheared horizon toward a less deformed rigid plug in the upper parts of cohesive flows (Phillips, 2006). Subaqueous debrites may result from the transformation of mass-transport deposits as slides and slumps disaggregate into debris flows, and flow transformation can result in linked turbidity currents (Fig. 6G). In turbidity currents, the incorporation of mud clasts into the flow may transform some currents into linked debris flows (Haughton et al., 2009; Talling et al., 2012). Lahars are volcanic debris flows and may be identified by their volcanic components (Figs. 6G, H; e.g., volcanic clasts, glass shards in the matrix; Sigurdsson et al., 1980; Isbell et al., 2016). Gravity-driven resedimentation in glaciomarine environments can transport clasts downslope that were originally striated and deposited by glaciers. Although

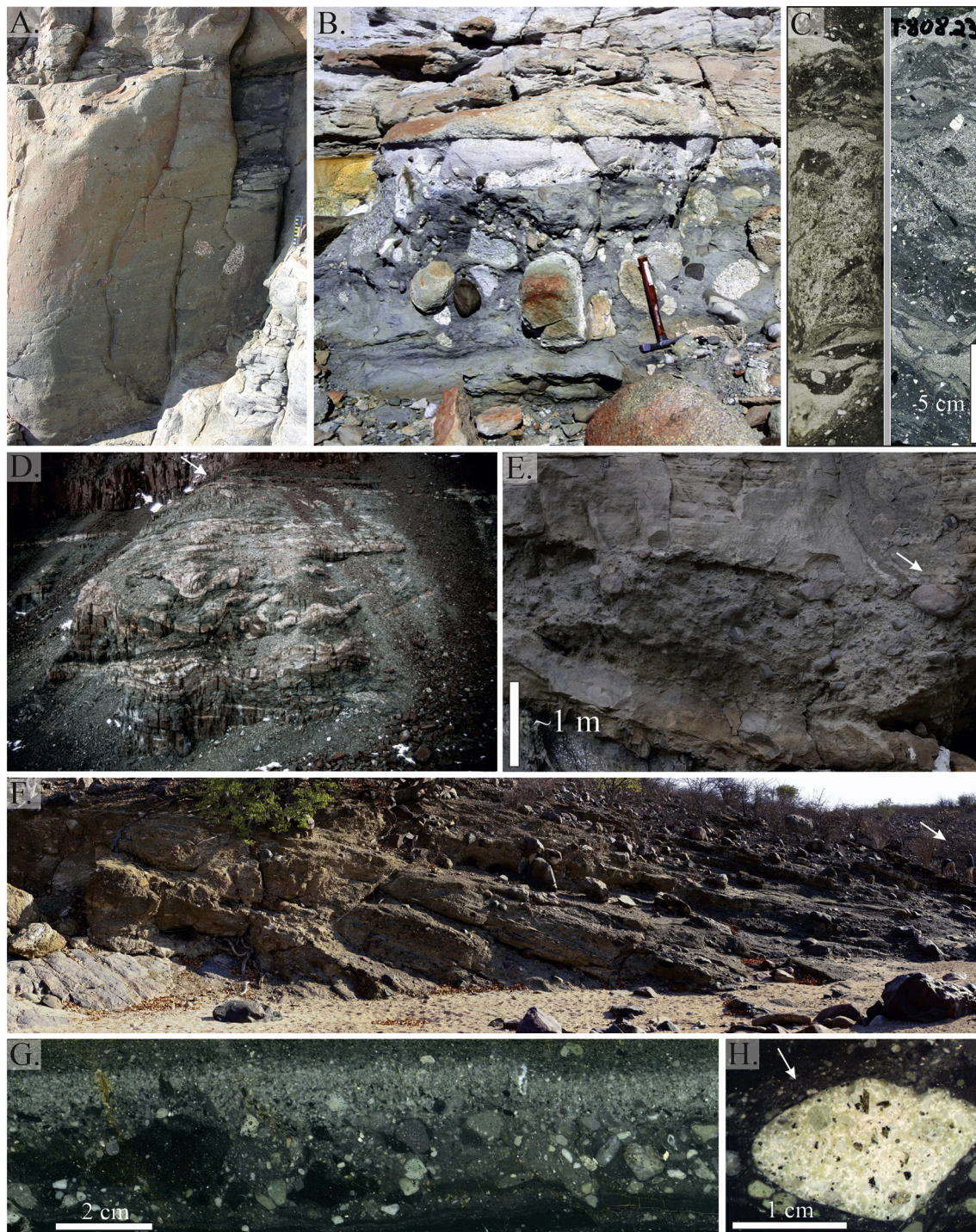


Fig. 6. Glaciomarine and sediment gravity flow deposits. A) Stratified diamictite interpreted as combined meltwater plume and iceberg rafted debris sedimentation in the Pagoda Formation, Tillite Glacier, Antarctica. 15-cm long ruler for scale. B) Massive diamictite (lower 2/3 of the photo) will abundant clasts displaying near vertical a-axis orientations at Mt. Ritchie, southern Victoria Land, Antarctica. Hammer for scale. C) Stockers Tillite core of a chaotically bedded diamictite from Tasmania, Australia. D) Folded, boudinaged, and chaotic slumps in the Pagoda Formation, Nimrod Glacier, Antarctica. People for scale (arrow). E) Massive diamictite interpreted as a debrite in the Malanzán Formation, Olta River Valley, La Rioja, Argentina. Note the boulder at the top of the unit (arrow). F) Bedded diamictites interpreted as morainal bank debrites within the Dwyka Group, Otjihaa, Namibia. People for scale (arrow). G) Cut and polished rock slab from the Atkan Formation in the Russia Far East showing a linked clast-rich debrite and a graded turbidite sandstone. H) A 2-cm clast displaying evidence for clast rotation (arrow) from the Atkan Formation in the Russia Far East. Clast is contained in the debrite layer lateral to photo 6H.

resedimentation products may contain glaciogenic clasts, they do not necessarily require the presence of a glacier during their deposition, as resedimentation may occur long after the glacier(s) has disappeared.

Diamict summary: The occurrence or absence of diamictites alone are unreliable indicators of glaciation (Crowell, 1957; Vesely, 2020).

However, coupled with other features and environmental analyses, they may provide identification of the processes that formed these sediments and sedimentary rocks. Interpretations of diamict depositional environments often requires fabric analyses (clast orientation, micromorphology, anisotropy of magnetic susceptibility (AMS), X-radiography,

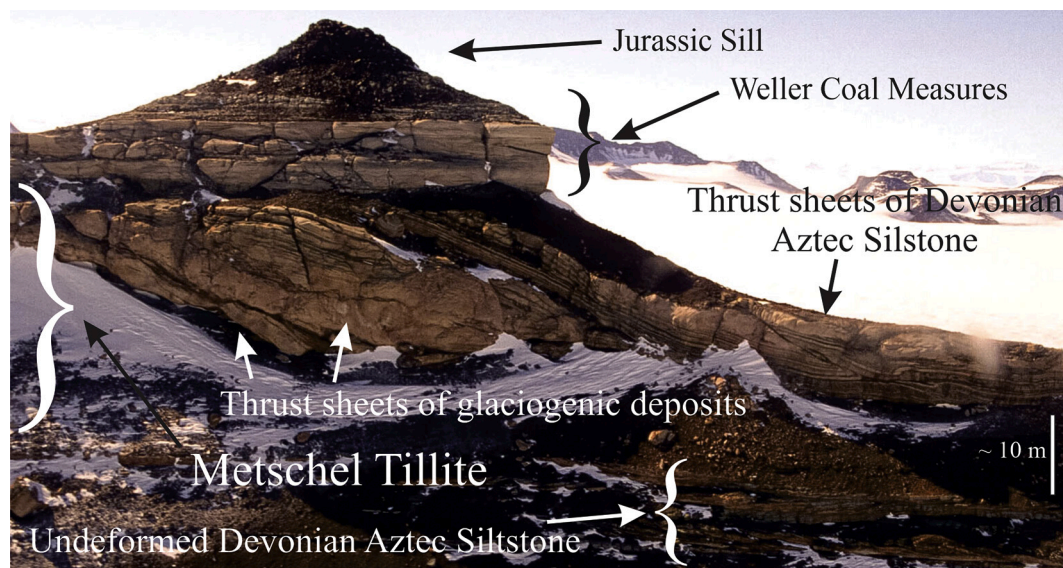


Fig. 7. Intraformational thrust sheets interpreted as proglacial glaciotectionic thrusting within the Metschel Tillite at Mt. Ritchie, southern Victoria Land, Antarctica.

and/or computed tomography (CT) scans and are typically interpreted by their relationship to adjacent facies (cf. Boulton and Deynoux, 1981; Gravenor and von Brun, 1987; Evans and Benn, 2004; Phillips et al., 2007; Hooyer et al., 2008; Menzies et al., 2010, 2016; Penkrot et al., 2018).

2.1.2. Ice-marginal glaciotectionism/deposition and mass transport deposits

Ice-marginal depositional features: Moraines are complex subaerial or subaqueous ice-contact landforms that form perpendicular to a glacier's margins, and may contain diamicts, sands, and gravels deposited as glaciotectionites, subglacial traction tills, rockfall, debris flows, ice-collapse features, and meltwater deposits. They may also be deformed by glaciotectionic (push and shove) deformation (Bennett, 2001; Benn and Evans, 2010). In glaciomarine settings, subaqueous morainal banks form at tidewater glacial fronts and are similar to subaerial moraines; however, in addition, the banks contain sediment that settled from suspension out of meltwater plumes and coarse material released from icebergs, as well as outwash in the form of grounding line fans (Powell, 1990, 2005; Powell and Alley, 1997). Grounding zone wedges are thick (15–100 m), wedge-shaped (< 15 km long) accumulations of diamicts that form at the grounding zone of marine-terminating ice sheets and are most often associated with floating ice shelves and non-temperate glacier margins (Batchelor and Dowdeswell, 2015; Batchelor et al., 2018). These deposits consist of dipping diamict beds overlain by horizontal subglacial diamicts (till). Sediment is supplied as subglacially deforming sediment to the grounding zone where it is redistributed as debris flows in front of the glacier (Powell and Domack, 2002; Demet et al., 2019). Marensi et al. (2005) and Henry et al. (2010) reported on morainal bank deposits from the mid-Carboniferous Guandacol Formation, (near Huaco, Argentina) and from the Agua de Jagüel Formation (near Uspallata, Argentina), respectively. Dietrich and Hofmann (2019) reported a LPIA grounding zone wedge deposit from the Dwyka Group in the Karoo Basin of South Africa.

Glaciotectionic landforms: Proglacial glaciotectionic landforms and deformation in the rock record result from excavation, push, and shove by an advance of the ice front, although this may represent either long term or just seasonal readvance. These features occur both subaerially and subaqueously. Depending on their occurrence and size, they are termed push moraines, thrust moraines, composite ridges, hill-hole pairs, cupola hills, rafts, mega-blocks, and De Geer moraines (Bennett, 2001; Benn and Evans, 2010). They form thrust complexes consisting of stacked thrust sheets and associated folded subglacial and proglacial

sediment, and/or pre-existing sediment/bedrock (Fig. 7; Croot, 1988; Van der Wateren, 1995; Van Der Wateren, 2002; Bennett, 2001; McCarroll and Rijdsdijk, 2003; Benn and Evans, 2010). Without an association with other glaciogenic features, glaciotectionic structures are difficult to separate from mass transport deposits. Initially, any soft-sediment deformation in the LPIA strata was interpreted as a product of glaciotectionism, but currently most modern interpretations of soft-sediment deformation are better viewed as products of mass-transport deposition (e.g., Rodrigues et al., 2020). Rare glaciotectionic complexes within the LPIA record are reported by Isbell (2010), Fedorchuk et al., 2019a, and Rosa et al. (2019) who reported glaciotectionic thrusting from the Mississippian and Pennsylvanian of the Paraná Basin (Brazil) and the Permian of the Transantarctic Basin (Antarctica). Well-preserved glaciotectionic features occur in the Metschel Tillite of southern Victoria Land, Antarctica. At Mt. Ritchie, Antarctica, proglacial deposits and Devonian bedrock occur as stacked, echelon-like, thrust wedges (Fig. 7; Isbell, 2010).

Mass transport deposits: As described and discussed in section 2.1.1, mass transport deposits form a continuum of deposits that range from slides (relatively underformed bodies)-to-slumps (varying degrees of internal deformation and disaggregation)-to-debris flows (Figs. 6C-H, 8; complete disaggregation; Cossey, 2011), and display various types of basal contacts (sharp, eroded, liquified/injected, and deformed; Sobiesiak et al., 2018). Where disaggregation of the original bodies is near completion, these deposits resemble glaciogenic diamictites (see section 2.1.1). However, where the bodies show only mild deformation (slides and early stages of slumping), they resemble proglacial glaciotectionic landforms. Depending on the degree of deformation, these bodies may display folding (Fig. 8C), echelon-like thrust sheets stacked on the backs of underlying sheets (Figs. 8A, B), or they may show boudinaged features or chaotic bedding where the slump is beginning to disaggregate (Figs. 6C, D, 8D; Nemec, 1990; Cossey, 2011; Posamentier and Martinsson, 2011; Rodrigues et al., 2020). Such features are almost ubiquitous within near-field LPIA basins (e.g., Isbell et al., 2008; Valdez Buso et al., 2015; Sobiesiak et al., 2016a, 2016b; Columbi et al., 2018; Mottin et al., 2018; Vesely et al., 2018; Rodrigues et al., 2020), but occur worldwide where mass wasting occurs. All the similarities between glaciotectionic landforms and mass transport deposits make it difficult to distinguish between the two. The presence of glacially modified clasts within the deposits could also be misleading as mass transport may re-sediment pre-existing deposits (Vesely et al., 2018).

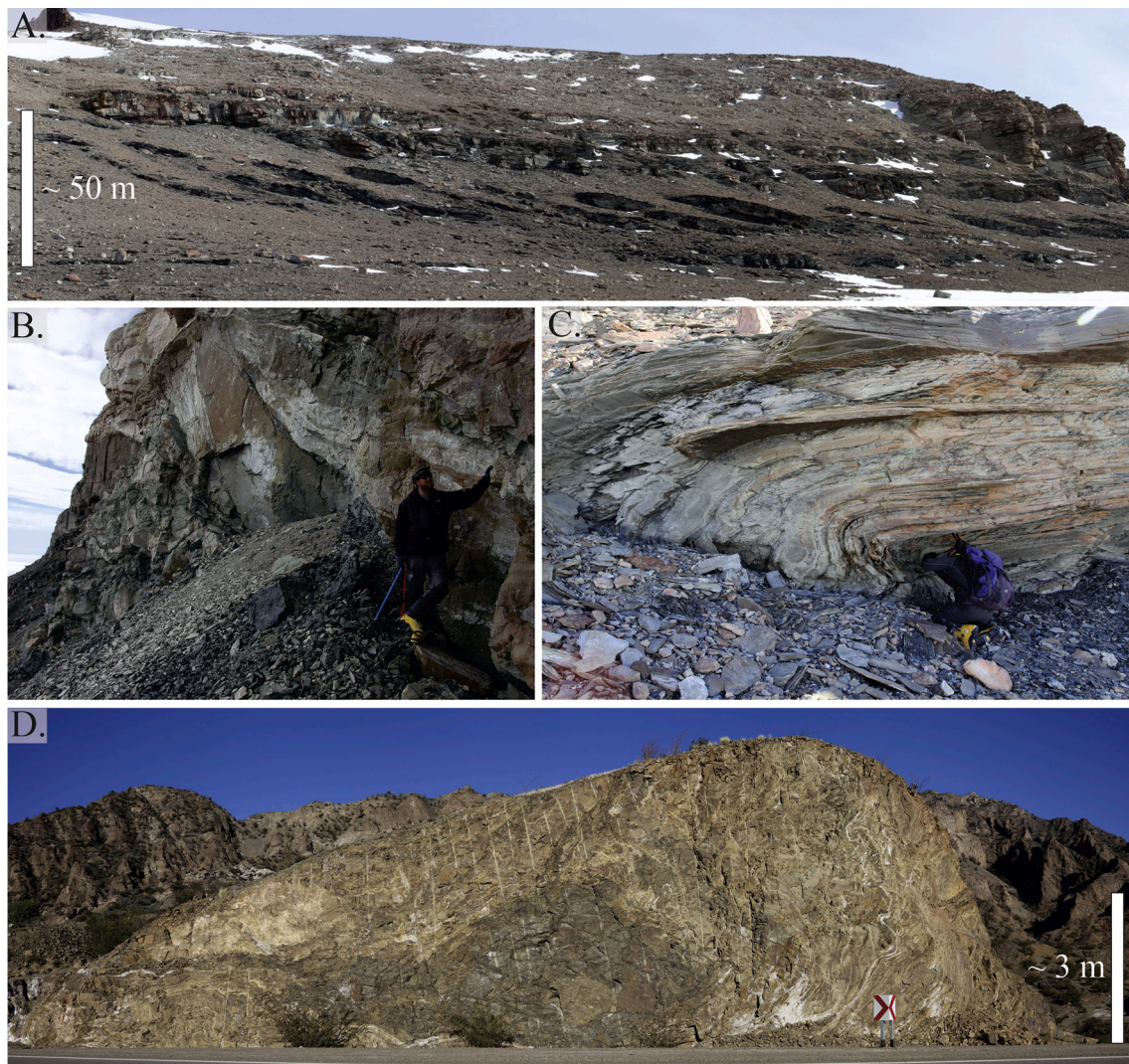


Fig. 8. Mass transport deposits. A) Echelon stacked intraformational thrust sheets in the Mackellar Formation, Robb Glacier, Antarctica. B) Stacked intraformational thrust sheets interpreted as a small, stacked mass-transport complex in the Pagoda Formation, Nimrod Glacier Antarctica. Person for scale. C) The folded front of a thrust sheet that is shown in Photo 8A from the Mackellar Formation, Robb Glacier, Antarctica. D) Chaotic slump deposit in the Tuminico Formation, along Agua de la Peña Creek, Parque Provincial Ischigualasto, San Juan, Argentina.

2.1.3. Glacially modified clasts and clast pavements

Striated and faceted clasts: Clasts modified in subglacial settings are typically striated (depending on hardness of contained mineral grains), bullet-shaped, and/or faceted depending on where in the subglacial deformation profile abrasion and clast modification occurred (Figs. 9A, C-E; Krüger, 1984; Benn and Evans, 2010). Striated clasts developed within subglacial settings tend to form “stoss and lee” forms (bullet-shaped clast) due to abrasion on the up-glacier and plucking on their down-glacier ends (Figs. 9A, C-E; Krüger, 1984; Benn and Evans, 2010). When bullet-shaped clasts are present, the orientation of striations and the long axis of the clasts are nearly always parallel. Although bullet-shaped (stoss-lee forms) clasts are diagnostic of subglacial modification (Krüger, 1984; Benn and Evans, 2010), striated clast themselves are not diagnostic as a number of processes striate clasts including: 1) subglacial abrasion, 2) abrasion by sea and lake ice, 3) abrasion by partially grounded ice shelves, 4) abrasion by grounded icebergs, 5) tectonism (Fig. 9B), 6) abrasion by high-velocity water flow, 7) wind abrasion, 8) mass movement (i.e., slides, slumps, debris flows, rock fall), and 9) volcanic activity including volcanic blasts and pyroclastic flows (Judson and Barks, 1961; Nichols, 1961; McLellan, 1971; Zamoruev, 1974; Krüger, 1984; Powell, 1984; Dionne, 1985; Eyles, 1988; Atkins, 2003; Benn and Evans, 2010). Bullet-shaped clasts can also be

transported far from their subglacial site of origin if transported by fluvial, shoreline, and mass-transport processes, or incorporated into debris carried by icebergs, sea ice, or lake ice.

Clast pavements: Clast or boulder pavements (Figs. 9C-E) may form as a lag deposit from aeolian, fluvial, or subaqueous winnowing (tides and waves); due to nearshore fast (sea/lake) ice; or due to glacial activity (Dionne, 1985; Eyles, 1988; Evans, 2018). Within glacial settings, these pavements form at the ice-till interface, between stagnant and active ice, within a deforming bed, or as pre-existing lags reworked by partially floating ice shelves or grounded icebergs (Powell, 1984; Eyles, 1988; Clark, 1991, 1992; Mickelson et al., 1992; Evans, 2018). Inter- and intraformational striated clast pavements within diamictites suggest grounded ice conditions. In such cases, the clast pavement can occur in clast-rich or clast-poor diamicts and form horizons where clast tops are faceted, polished and striated, and the striations often parallel the a-axes of clasts (Eyles, 1988; López-Gamundí et al., 2016; Evans, 2018). However, striated clast pavements where the clast a-axis is not parallel with the orientation of striations suggest that a pre-existing lag (current, wave, or meltwater winnowed) deposit was overridden by icebergs, sea/lake ice, or a partially floating ice shelf (Dionne, 1985; Eyles, 1988). Numerous examples of striated clast pavements occur, but an excellent example occurs in the Hoyada Verde Formation near Barreal, San Juan

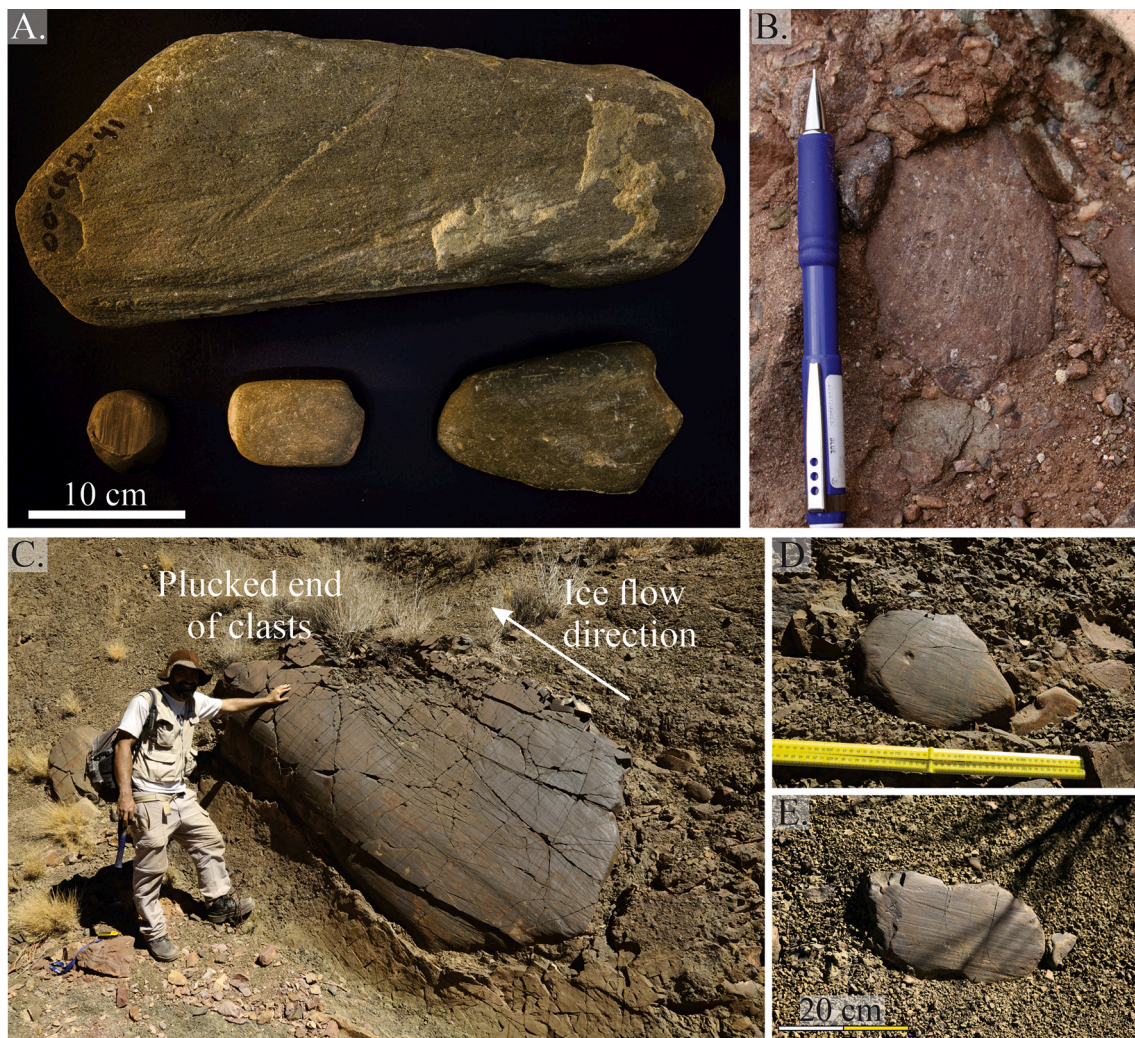


Fig. 9. A) Bullet-shaped and striated clast from the Carboniferous of Patagonia and the Permian of Antarctica. B) Tectonically striated clast in a Triassic fluvial deposit, Porterillos, Mendoza, Argentina. C-E) Striated clast pavement in the Hoyada Verde Formation, Barreal, San Juan, Argentina. Note that the bullet-shaped clasts are widely dispersed and that striations parallel the long axis of the clasts.

Province, Argentina (Figs. 9C-E; López-Gamundí et al., 2016).

2.1.4. Outsized clasts (lonestones/dropstones)

Outsized clasts from the LPJA record are predominantly viewed as dropstones released from icebergs (e.g., Chumakov, 1994; Crowell, 1999). However, numerous non-glacial processes deposit outsized clasts, many of which are true dropstones (e.g., Gilbert, 1990; Lisitzin, 2002; Archer, 2013; Isbell et al., 2013, 2016; Moxness, et al., 2019). Bennett et al. (1996) defined dropstones as any “clast of anomalous size, and/or lithology, indicative of vertical or oblique introduction into the host sediment” by rafting or floatation processes.

Iceberg rafted debris: Release of glacial debris from icebergs occurs due to melting or rolling of icebergs. Once released, sand- to boulder-sized material drops through the water column before impacting and deforming the substrate by bending (down folding of stratification that is conformable with the base of the clast), penetrating (stratification is bent down and terminates against the clast), rucking (lateral displacement of the sediment that is bent upward around the edges of the clasts), or rupturing (where sediment beneath the clast was liquified destroying stratification) sediment on the sea or lake floor (Fig. 10A; Thomas and Connell, 1985; Bennett et al., 1996). Such clasts occur in both ice-proximal and ice-distal glaciomarine settings with distance from the original ice front ranging from m to thousands of km (Powell and Domack, 2002; Domack and Powell, 2018). Although ice-rafted debris

occurs in fine- to coarse-grained sediment, typically, ice-rafted debris, is best identified within fine-grained sediment and consists of exotic rock clasts, diamictite pellets, or frozen sediment aggregates. Although dropstones may not indicate local glaciation, iceberg rafted debris imply the occurrence of marine- or lake-terminating glaciers somewhere within the basin (cf. Powell and Domack, 2002). Rockfall and grainfall off glaciomarine or glaciolacustrine ice fronts, proximal to the retreating terminus, also produce dropstones (Fig. 6B; Powell and Molnia, 1989). The release of an agglomeration of particles by the rolling of icebergs and surface ice (sea and lake ice) produce clast clusters and dump structures on the sea or lake floor (Thomas and Connell, 1985). Dump structures usually consist of mound-like diamict deposits (Fig. 10B).

Sea- or lake-ice rafted debris: Sediments can also be incorporated into and transported by sea/lake ice, without the influence of glaciers. Clay- to gravel-sized material can be incorporated into, or on, sea, lake, or river ice by rock fall; freezing onto and lifting of clasts in the intertidal zone; freezing and flotation of clasts by anchor ice in rivers, lakes and oceans; freezing in of suspended material in surface ice; river flow across surface ice during spring melt; aeolian processes; ‘bulldozing’ of clasts by river, lake or sea ice; as well as several other processes (Rosen, 1979; Lauriol and Gray, 1980; Dionne, 1981, 1988; Syvitski et al., 1987; Kempema et al., 1989, 2001; Gilbert, 1990; Reimnitz et al., 1998; Dethleff et al., 2000; Lisitzin, 2002). For sea-ice rafting, a 2-m-thick floe can lift and raft boulders up to 6 m or more in diameter (Fig. 10C; Drake



Fig. 10. A) Outsized clast interpreted as iceberg rafted debris in the Dwyka Group, near Sanitatis, Namibia. Major division on the ruler are 10 mm. B) Lenticular diamictite interpreted as an iceberg dump structure in the Dwyka Group, near Sanitatis, Namibia. Major division on the ruler are 10 mm. C) Tidal flat in Ungava Bay, Canada at low tide showing debris transported by sea ice including the giant boulder and gravel. Person for scale. Photo courtesy of A. Archer. D) Boulder outsized clast within the Darlington Limestone, Maria Island, Tasmania, Australia. The Darlington Limestone was a shallow water shell bank composed of the bivalve *Eurydesma*. Jacob's staff divisions are 10 cm. E) Boulder and pebbles rafted in a driftwood root mass. Agua Bay, Lake Superior, Ontario, Canada. F) A limestone boulder rafted by a driftwood root mass, from Falls of the Ohio State Park, Jeffersonville, Indiana, USA.

and McCann, 1982; Syvitski et al., 1987). Boulders up to 2 m in diameter are commonly lifted and rafted on the top of ice floes (Rosen, 1979; Archer, 2013). Such processes produce boulder strewn substrates (Fig. 10C; Archer, 2013). The boulders are lifted by the freezing of seawater at the bottom of the sea ice during tidal activity, ultimately lifting the particles off the bottom. Continued freezing at the bottom of the ice column and sublimation and ice melt at the surface ultimately transports clasts upward until they sit on top of the ice surface where they can be transported considerable distances (Rosen, 1979; Gilbert, 1990; Archer, 2013). Anchor ice can also lift and transport boulders (Kempema et al., 2001). Dayton et al. (1969) reported sediment masses of as much as 25 kg being lifted from the sea floor by this process. Diamictites are not typically associated with surface ice-rafted materials. However, clay- to gravel-sized material can be incorporated into shore ice through freeze-on, wave, stream, or aeolian transport. Several authors have suggested the occurrence of sea ice transported clasts in the LPIA record of Gondwana (Fig. 10D; e.g., Domack et al., 1993; Dickins, 1996; Jones et al., 2006; Isbell et al., 2013).

Other non-glacial rafting agents: Coarse debris shed off steep valley walls into standing bodies of water produce dropstones (Fig. 11; Moxness et al., 2018). Plants and algae also raft particles producing individual dropstones and clast clusters when material is released and

dropped through the water column (Gilbert, 1990; Bennett et al., 1996; Doublet and Garcia, 2004). Driftwood with root masses can transport clasts up to boulder-size (Figs. 10E, F). Pumice sand and gravel are also common as dropstones. Due to their density, such floating particles are deposited when they become waterlogged and settle through the water column following volcanic eruptions (cf. Bennett et al., 1996; Manville et al., 1998, 2002; Bryan et al., 2004).

Outsized clasts other than dropstones: Outsized clasts easily mistaken for “dropstones” include: 1) clasts at the base of diamicts loaded into underlying muds, which deform but do not pierce stratification; 2) clasts protruding from the tops of debrites where large clasts were carried in suspension due to the viscosity/yield strength of the flow (Fig. 6E; Bennett et al., 1996; Talling et al., 2012); 3) clasts within thin debrites that are larger than the surrounding diamictite lamina, which resulted when clasts (out-runner clasts) are stranded on soft substrates by thinning debris flows that continued on down slope (Figs. 6G, H; Carto and Eyles, 2012; Isbell et al., 2016); and 4) sand and granule clasts deposited by mudflows. Some of these outsized clasts are core stones within rotational structures indicating that they were transported and deposited by ductile, viscous flows (Fig. 6H; Phillips, 2006) rather than having been “dropped” into the succession.

Outsized clasts summary: Outsized clasts have multiple origins, which



Fig. 11. A) A boulder breccia interpreted as rockfall off the valley walls in the Malanzán Formation, Olta River Valley, near Olta, La Rioja, Argentina. Note highly deformed sandstone and mudstone squeezed between angular granite boulders. B) Boulders deforming flat topped, rippled lacustrine sandstones and siltstone in the Malanzán Formation, Olta River Valley, near Olta, La Rioja, Argentina. The large boulders are several m in diameter, and some have associated ruck structures. The boulders are interpreted as rock fall off valley walls. Yellow meter stick for scale. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

include deposition as iceberg or surface-ice rafted debris, anchor ice-rafted debris, rockfall, organic rafted debris, debris flow transported debris, deposition as outrunner clasts, and pumice deposition. The presence of bullet-shaped and or striated clasts may aid in a glacial interpretation. However, the clasts relationship to surrounding facies may provide the best clues to their origin.

2.1.5. Rhythmites

Within the high-latitude record, fine-grained laminated units or rhythmites, with or without ice-rafted debris, are typically interpreted as glacial or glaciomarine in origin (cf. [Frakes and Crowell, 1969](#); [Chumakov, 1994](#); [Isbell et al., 2008](#); [Milleson et al., 2016](#)) many of which are interpreted as varves (e.g., [Banks et al., 1955](#); [Fedorchuk et al., 2019a](#)). Although collectively termed “rhythmites,” documented late Paleozoic rhythmites units are variable, including mm-thick, graded or sharply bounded silt-clay couplets, cm-thick sandstone-mudstone alternations in which coarser divisions may have sole marks, parallel laminations and/or current ripples, and sandy heterolithics showing oscillatory-flow ripples or climbing ripples. These include rhythmite couplets likely deposited by numerous physical and biological processes over semi-diurnal, diurnal, monthly, annual, decadal, centennial, and millennial-scale cycles as well as deposition over irregular, episodic intervals within lacustrine, estuarine, fjordal, and oceanic waters (cf. [Cowan](#)

[et al., 1997, 1998, 1999](#); [Franco et al., 2011, 2012](#); [Franco and Hinnov, 2012](#); [Milleson et al., 2016](#); [Fedorchuk et al., 2019b](#); [Tedesco et al., 2020](#)). However, rhythmites also occur in numerous non-glacial environments including lake, tidal flat, estuary, delta (e.g., hyperpycnites), and deep-marine (e.g., turbidites and contourites) settings (e.g., [Kuenen and Migliorini, 1950](#); [Martino and Sanderson, 1993](#); [Greb and Archer, 1995](#); [Zolitschka et al., 2015](#); [Zavala and Arcuri, 2016](#); [Shanmugam, 2018](#)). In part, bioturbation is inhibited or greatly reduced in abundance in some of these glacial and non-glacial settings because of high sedimentation rates (cf. [Archer et al., 1995](#); [Mulder and Chapron, 2011](#); [Zavala et al., 2011](#); [Jaeger and Koppes, 2016](#)).

2.1.6. Glaciogenic sequences and cycles

The stratigraphic record of high-latitude LPIA basins commonly exhibit recurrent successions where diamictite units alternate with mudstones and sandstones. The most common understanding is that these “cycles” are the stratigraphic signature of ice-margin advance and retreat. However, ambiguity occurs concerning the nature of sequence boundaries and systems tracts and their relationship with glaciation/deglaciation. For instance, the concept of a deglaciation sequence assumes that glaciogenic sequences are bounded by glacial erosional surfaces formed during glacial maxima ([Visser, 1997](#); [Vesely and Assine, 2006](#)), whereas in other cases, lowstand incised valleys are interpreted

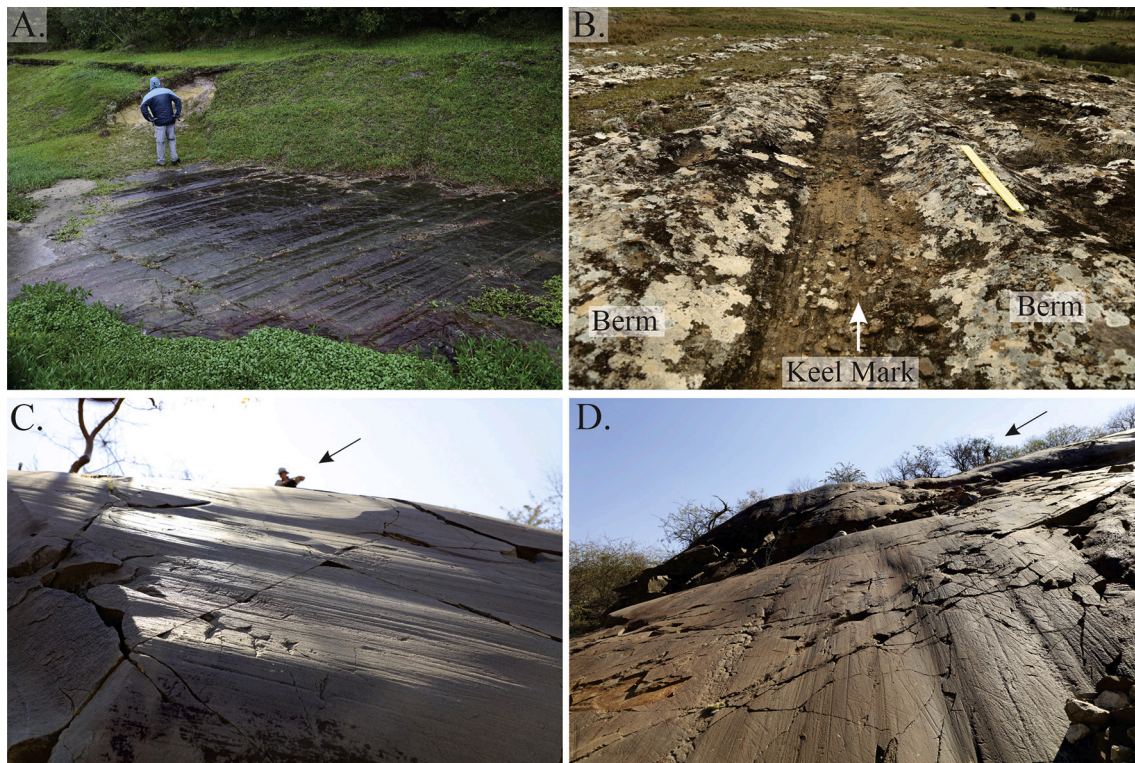


Fig. 12. A) Striated and grooved surface at the base of the Itararé Group cutting into the underlying Furnas Sandstone at Witmarsum, Paraná, Brazil. The surface is interpreted as a product of subglacial abrasion and erosion. B) Linear striated and grooved trough bounded by berms interpreted as an iceberg keel mark in the Carboniferous San Gregorio Formation, near Cerro de las Cuentas, Uruguay. The folded ruler is 0.5-m long. C–D) Subglacially polished and striated surface cut on the Nosib Quartzite near the settlement of Orotjitoambo, Namibia. People for scale (arrows).

as sequence boundaries (e.g., [Starck and del Papa, 2006](#); [Holz et al., 2008](#)). Another source of ambiguity is the assumption of the timing of diamictite units within the glacial cycle, which may range from ice maxima to ice minima (e.g., [Franca and Potter, 1988](#); [Visser, 1997](#); [Mottin et al., 2018](#); [Vesely et al., 2018](#); [Valdez Buso et al., 2019](#)) depending on whether the diamictites are interpreted as subglacial, ice marginal or non-glacial (see [section 2.1.1.](#)).

2.1.7. Glendonites

Glendonites typically form in high-latitude oceanic waters and are known from the Gondwana LPIA record ([González, 1980](#); [Domack et al., 1993](#); [Frank et al., 2008](#)). These bladed crystals are calcite pseudomorphs of the syndepositional or early authigenic mineral ikaite. Ikaite is thought to precipitate in cold or near-freezing bottom waters between -1.9 to 7 °C at or just below the sediment-water interface. These crystals typically form under elevated alkaline and phosphatic conditions and commonly occur in association with organic-rich muds ([Selleck et al., 2007](#); [Frank et al., 2008](#); [Wang et al., 2020](#)). Although they occur in glacial or high-latitude shelf settings and are used as a criteria to recognize past glacial environments (cf. [Fielding et al., 2008b, 2008c](#)), they also occur in nonglacial settings in recent deep-water sediments from the mid- to low- latitudes ([Zabel and Schulz, 2001](#); [Hensen et al., 2003](#); [Zhou et al., 2015](#)), from high mid-latitude cold-water seeps (Sea of Okhotsk; [Greinert and Derkachev, 2004](#)) and in saline lakes with tufa formation ([Benson et al., 1995](#)). Currently, Antarctic bottom waters form much of the cold (< 2 °C) bottom waters in the world's oceans. These waters form on Antarctic continental shelves by the formation of sea ice and the release of brines forming dense cold waters that sink and flow away from the continent ([Gordon, 2001](#)). In the Atlantic, these waters extend to at least the Equator, while in the Pacific they extend to 40°N ([Gordon, 2001](#)). Although the temperatures are driven by formation in polar regions, temperatures capable of forming glendonites occur

throughout the bottom waters of Earth's major oceans. Similar conditions likely occurred in the Panthalassan Ocean during the LPIA. Recently, in laboratory tests, [Tollefsen et al. \(2020\)](#) reported that ikaite can nucleate at temperatures >7 °C and as high as 35 °C from a mixture of natural seawater and sodium carbonate-rich solution, thus challenging the use of glendonites as paleotemperature indicators.

2.2. Erosional proxies

2.2.1. Grooved and striated surfaces

Grooved and striated surfaces: Glacially produced grooved and striated surfaces result from abrasion of basal debris-rich ice sliding across bedrock surfaces ([Figs. 12A, C, D](#)), grooving soft-sediment substrates, ploughing of clasts dragged across soft-sediment substrates ([Fig. 5](#)), or due to differential sliding along decollement surfaces within deforming substrates ([Deynoux and Ghienne, 2004](#); [Evans et al., 2006](#); [Benn and Evans, 2010](#)). The scale of such surfaces is dependent on the size of the glacier that produced them and on various local conditions (e.g., topography, sliding speed, glacier rheology) that occurred at the base of a glacier and within the substrate, and may range in width and length from cm to tens of km scale or larger ([Deynoux and Ghienne, 2004](#); [Dowdeswell et al., 2016](#)).

Other glacially-influenced grooved and striated surfaces: Other glacially influenced processes that produce grooved surfaces on soft-sediment include gouging of the substrate by grounded icebergs or sea/lake ice ([Thomas and Connell, 1985](#); [Gilbert, 1990](#); [Woodworth-Lynas and Dowdeswell, 1994](#); [Vesely and Assine, 2014](#); [Lynch and van der Meer, 2015](#); [Rosa et al., 2016](#); [Assine et al., 2018](#)). Iceberg keel scour marks are common in proximal glaciomarine settings, but also occur km to thousands of km away from the ice fronts wherever icebergs become grounded. These marks range from cm- to km-scale in width and cm to tens of km long depending on the size of the iceberg keel grounding

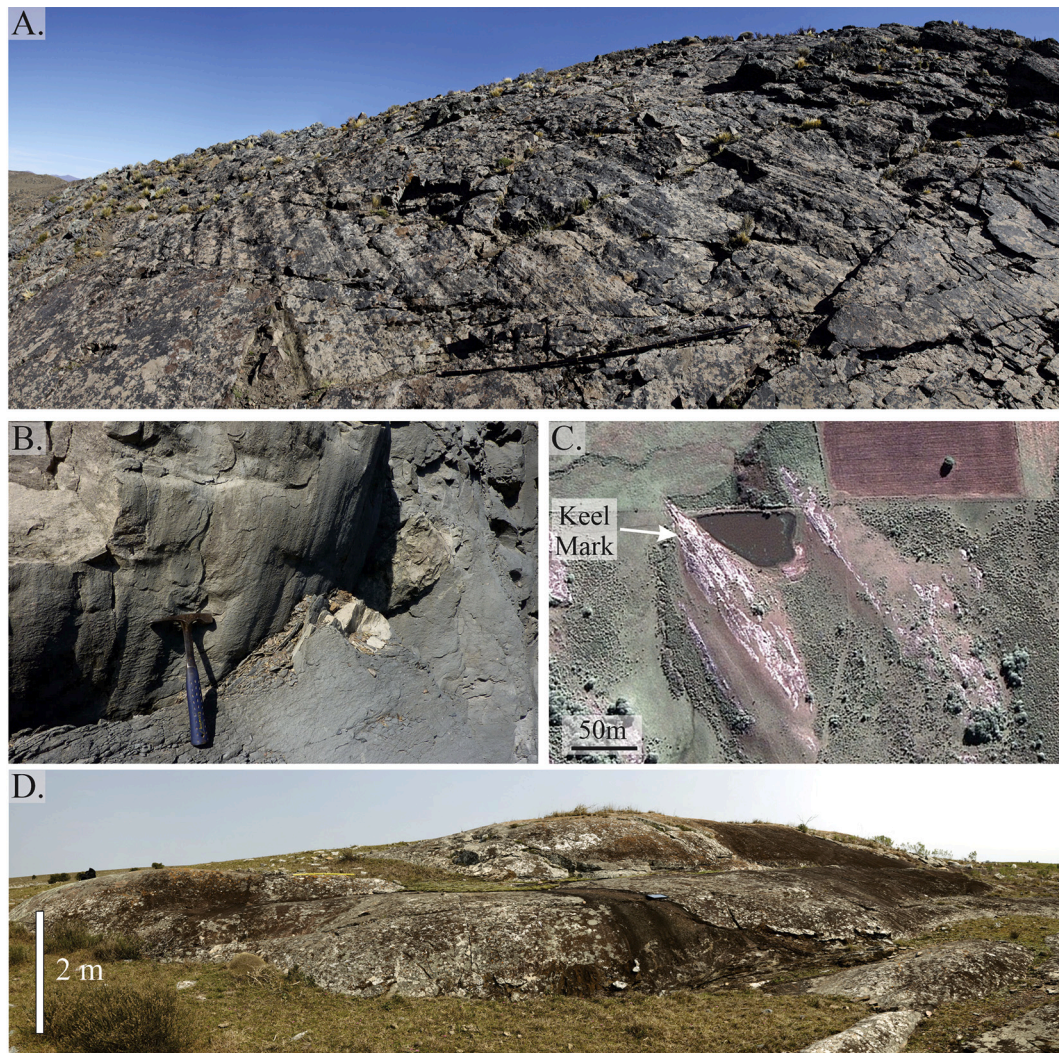


Fig. 13. A) Grooved surface in the Carboniferous Pampa de Tepuel Formation at El Molle, Chubut Province, Argentina. The grooves occur at the top of a thick succession of strata dominated by mass-transport deposits. Jacob's staff is 1.6-m long. B) Grooves at the base of a mass-transport deposit in the Paraná Basin, Brazil. Hammer for scale. C) Google Earth view of whalebacks cut into basement near Cerro de las Cuentas, Uruguay. A thin layer of the San Gregorio Formation overlies part of the whaleback. The location of the keel mark shown in Fig. 12B is marked. D) Whaleback shown in Fig. 13C.

against the substrate (Eyles et al., 2005; Vesely and Assine, 2014; Batchelor et al., 2018). They typically have one or two parallel berms that laterally bound the drag mark. Grooves and striations are restricted to the trough between berms, while the surface beyond the berms is unmodified (Fig. 12B; Woodworth-Lynas and Dowdeswell, 1994). In some cases, the keel mark may end in a prow-like structure representing material bulldozed by the ice, and the mark may end abruptly representing the site where the keel may have lifted off the bottom.

Non-glacial grooved and striated surfaces: non-glacial grooved and striated surfaces can be produced by the sliding of mass-transport bodies along basal and internal glide planes and the ploughing of tools dragged along the bottom at the base of sediment gravity flows. Grooved and striated surfaces beneath and within mass-transport deposits range in scale from cm to tens of km in width and length (Figs. 13A, B; Posamentier and Walker, 2006; Draganits et al., 2008; Dakin et al., 2013; Sobiesiak et al., 2018). Tool marks are the result of the impact of solid objects striking the substrate. If pulled along by currents (e.g., sediment gravity flows), such objects produce elongate grooves (Middleton, 2003). In some cases, pervasively grooved and striated surfaces may result. Grooved surfaces produced by mass-transport and sediment gravity flows occur at all latitudes in both the presence and absence of glaciers. Difficulty occurs in distinguishing between glacial and non-

glacially modified surfaces. Without critical scrutiny and examination of adjacent facies the occurrence of grooved and striated surfaces may lead to erroneous paleo-environmental interpretations.

2.2.2. Streamlined landforms

Erosional features such as whalebacks and roche moutonnées are reported from several LPIA basins (Frakes and Crowell, 1970; Rocha Campos, 2002; Long et al., 2009; Bussert, 2010; Rosa et al., 2016; Assine et al., 2018). These features are streamlined features carved through subglacial erosion into bedrock. Whalebacks are narrow near symmetrical features with smoothed stoss and lee sides (Figs. 13C, D; Benn and Evans, 2010). However, Bennett and Glasser (2009) Benn and Evans (2010), and Hambrey and Alean (2017) show a whaleback with a steep stoss and gently tapering lee slopes and mention the term rock drumlin for these features, although the term is not commonly used. Assine et al. (2018) reported whalebacks 40–200 m long, 10–60 m wide and 1–15 m high from the Carboniferous Chaco-Paraná Basin in Uruguay. Roche moutonnées are asymmetrical streamline features with smooth abraded stoss sides and plucked or quarried lee faces (Benn and Evans, 2010). Such erosional features are indicative of subglacial erosion and abrasion.

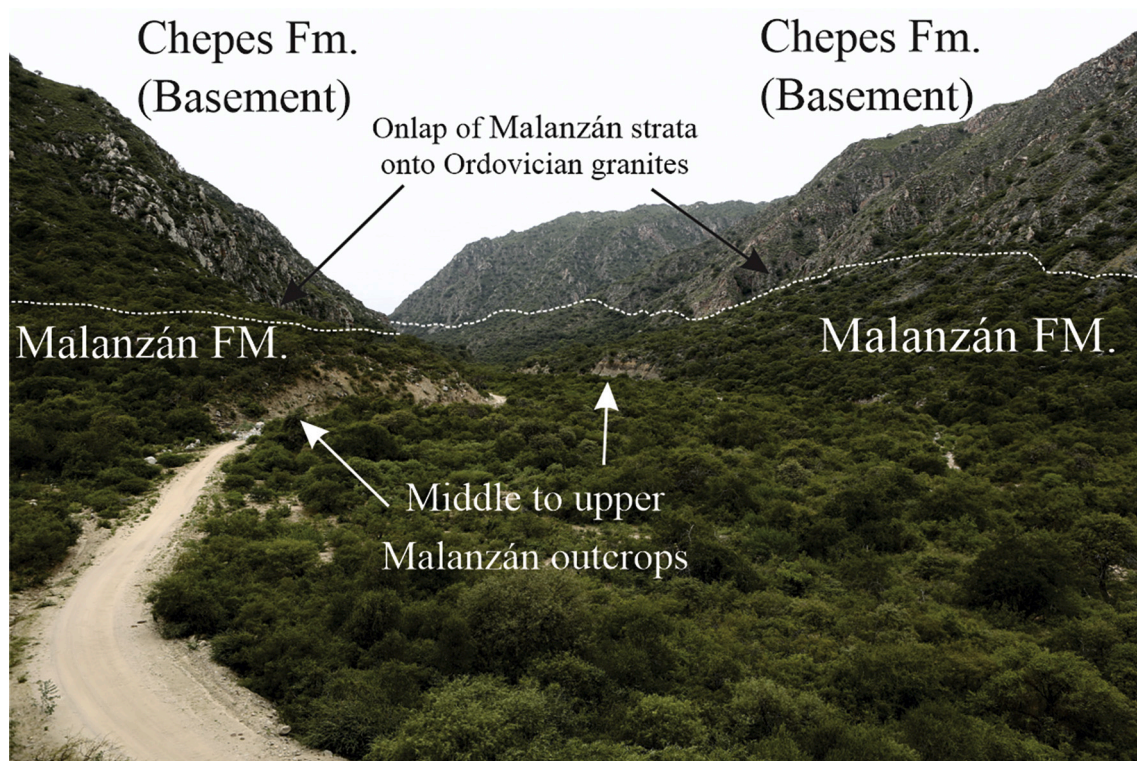


Fig. 14. A) The Olta paleovalley near Olta, La Rioja, Argentina showing a U-shaped profile. Note that the U-shaped profile is the shape of the modern valley cut on top of an unknown thickness of the Carboniferous Malanzán Formation and not the shape of the pre-Carboniferous valley cut on granites of the Ordovician Chepas Formation. Dirt road for scale.

2.2.3. Paleovalleys

Numerous paleovalleys are described from the late Paleozoic of Gondwana as well as Euramerica, many of which are interpreted as glacially carved valleys or fjords (Fig. 14). Although many show evidence of former glacial activity in the form of glacially striated or grooved floors and walls, or contain glaciogenic deposits (e.g., Martin, 1981; Visser and Kingsley, 1982; Henry et al., 2010; Aquino et al., 2014; Rosa et al., 2019; Valdez Buso et al., 2020), the origins of some paleovalleys are ambiguous and disputed (cf. Sterren and Martínez, 1996; Starck and del Papa, 2006; Enkelmann et al., 2014; Socha et al., 2014; Tedesco et al., 2016; Moxness et al., 2018; Fedorchuk et al., 2019b; Milana and Di Pasquo, 2019; Pauls et al., 2020; Valdez Buso et al., 2020). Modern fluvial processes are exhuming many of the LPIA paleovalleys, some of which display extant U-shaped profiles that researchers use to suggest a glaciogenic origin. However, it must be noted that the purported U-shape (e.g., the Olta and Malanzán paleovalleys, western Argentina) of some of these valleys is not the shape of the paleovalley cut on basement but the shape of the modern fluvial valley cutting into a late Paleozoic fill of unknown thickness. For example, Socha et al. (2014) and Valdez Buso et al. (2020) report U-shapes for the Olta and Malanzán paleovalleys (Fig. 14A) where they reconstruct the shape of the valley on top of the late Paleozoic fill. In the case of the Olta and Malanzán paleovalleys, the modern U-shape valleys are cut into strata 5–25 million years younger than the strata that rest directly on the buried basal unconformity. Basement rock at the base of the Olta paleovalley is also characterized by deep weathering profiles beneath the late Paleozoic fill (Socha et al., 2014) rather than having a striated and polished surface as expected if it were the result of subglacial erosion (Moxness et al., 2018).

U-shaped valleys also arise from fluvial activity and/or structural control. While V-shapes are typically assigned to rivers that are actively incising, incision coupled with lateral stream migration (erosion and deposition) during a base-level fall (eustasy or uplift) and lowstand produce erosional-based, U-shaped stratal bodies (stratigraphic valleys)

in cross-section (cf. Greb and Chesnut Jr., 1996; Smith and Read, 2000; Boyd et al., 2006; Strong and Paola, 2008; Gibling et al., 2011). However, it is easier for rivers to cut U-shaped valleys into sediment and sedimentary rocks than it is for lateral erosion cutting into crystalline basement rocks. Structurally generated transtensional and extensional depressions (grabens, half grabens and pull-apart basins) with steep walls also display characteristic U-shapes (Wu et al., 2009; Sagan et al., 2014; Zhu et al., 2014). Such features could easily be misidentified as glacially carved, especially if the paleovalleys are inferred by subsurface data.

Tunnel valley is a category of glaciogenic valley that forms subglacially due to erosion of poorly consolidated substrates by pressurized meltwater (O'Cofaigh, 1996; Van der Vegt et al., 2012). They are narrow and elongated depressions a few kms wide, up to tens of kms long, and tens to hundreds of meters deep (typical width/depth ratio = 10) filled by ice-contact to non-glacial sediments. In some formerly glaciated areas like the Pleistocene of the North Sea Basin, buried tunnel-valley networks seen in seismic surveys are the most obvious glacial signature (e.g., Stewart et al., 2013). Along with dimensions, the abrupt beginnings and ends and the undulating talwegs of tunnel valleys are characteristics commonly used to differentiate them from other types of glacial and non-glacial valleys (e.g., Huuse and Lykke-Andersen, 2000). However, these criteria are not straight forward, especially in exhumed valleys where only parts of their original morphology and sediment fill are preserved (e.g., Vesely et al., 2020). In these cases, alternative hypotheses should be carefully evaluated, as the recognition of tunnel valleys has strong implications for the extent of glaciation over depositional basins. Although well-known from late Ordovician glacial formations in northern Africa and the Arabian Peninsula (e.g., Ghiennie and Deynoux, 1998), tunnel valleys are still underreported in the LPIA, with few known examples in South Africa (Visser, 1988), southern Australia (Eyles and Broekert, 2001) and southern Brazil (Vesely et al., 2020).

2.3. Near-field geochemical proxies

2.3.1. Detrital zircons

Recently, U—Pb detrital zircon (DZ) geochronology is being used to identify the size, distribution, flow orientations, and age of ancient ice sheets (Canile et al., 2016; Elliot et al., 2016; Linol et al., 2016; Martin et al., 2019; Craddock et al., 2019; Ezpeleta et al., 2020; Griffis et al., 2019; Koltonik et al., 2019; Martin et al., 2019; Tedesco et al., 2019; Zieger et al., 2019, 2020; Fedorchuk et al., 2020; Malone, 2020; Pauls et al., 2020; Starck et al., 2020). In some instances, such studies hypothesize DZ fingerprints within studied sedimentary rocks that directly connect sediment within depositional basins with former ice nucleation sites located thousands of km away (e.g., Craddock et al., 2019). Such conclusions are incompatible with glacial dynamics and sediment dispersal patterns within ice sheets as viability studies suggest complexities in classic direct source to sink models when applied to glaciogenic systems. Multiple factors influence the final DZ age populations including the occurrence of: 1) recycled grains; 2) Pb-loss and complex zircons; 3) variable abundance of zircons in the parent rocks; 3) age biases from the size and sorting of zircon crystals; 4) heterogeneous patterns of subglacial erosion; 5) segregation due to varying erosional, transport, and depositional mechanisms; 6) bias due to particle size distributions; and 7) the fact that primary sources prior to reworking may reside on landmasses that are no longer connected (Thomas, 2011; Licht and Hemming, 2017; Ibañez-Mejia et al., 2018; Zieger et al., 2019, 2020). DZ studies of sediment deposited in other clastic systems have similar problem. However, using DZ analyses without knowing the constraints on erosion, transport, and depositional processes unique to glacial systems makes reconstructing the drainage patterns of ancient ice sheets suspect. In some cases, glacial flow is depicted in a system similar to dispersal patterns for fluvial drainage basins (e.g., Craddock et al., 2019), which contrasts with drainage in ice sheets which typically have radial flow away from ice domes (ice spreading centers). Maximum age of glacial deposits are constrained by the age of the youngest zircon grains within a deposit. The age of the glaciation can be no older than the age of the youngest detrital zircon ages.

Glaciers are complex source-to-sink erosional and depositional systems. The provenance of glacially derived sediment needs to be viewed in the context to the nature of glacial erosion. Glacial erosion rates and sediment yield are dependent on glacial thermal regime, substrate composition, and ice velocity. In general, sediment yield is zero or low under ice divides and increases to a maximum near the equilibrium line (point where net accumulation and loss of ice are balanced) before dropping off near the ice margins (Bennett and Glasser, 2009). For ice sheets, the equilibrium line occurs at, or relatively close, to the end of the glacier, suggesting that most of the glacial sediment is generated close to the ice margin. Once subglacially eroded, sediment concentrations typically decrease exponentially away from its source becoming so diffuse that it is difficult to identify the source on average on the scale of <50 km (Clark, 1987; Licht and Hemming, 2017). However, longer distance transport can occur where the subglacial debris are shielded from abrasion from the underlying substrate by a deforming bed. Likewise, englacial transport from a cold-based glacier may also allow longer transport distances, but low generation of sediment in such systems becomes problematic for provenance studies. Subglacial erosion and transport suggest that local sources dominate over distant sources when considering provenance for detrital zircons. However, as an ice front retreats, the zone of maximum erosion also retreats up glacier, thus, bringing sediment from sources farther up the glacial drainage to the retreating glacial terminus. At the terminus, meltwater can transport sediment hundreds to thousands of km from the ice front (e.g., the Missouri-Mississippi River drainage system for Pleistocene glaciation; Mason et al., 2017). For individual sites, Griffis et al. (2019) and Fedorchuk et al. (2020) both show that ice-proximal environments are dominated by sediments with local provenance and once the ice front has retreated from the area, more distal provenances show up higher in

the section within ice-distal sediments.

Bias in detrital zircon chronologies can also be imparted through sediment sorting. Hydraulic sorting influences the fractionation of various grain sizes in sedimentary systems including heavy minerals and detrital zircons. Using U—Pb detrital zircon studies of samples within a single channel reach of the Río Orinoco delta, Venezuela, Ibañez-Mejia et al. (2018) showed that detrital zircons varied in age populations within different fluvial sub-environments in the river. If observed in the ancient, age density plots for the various samples would suggest different source terranes, thus introducing inaccuracies in identifying provenance. However, Leary et al. (2020) suggested that size bias has little influence on DZ populations. Glaciers separate materials into different size fractions based on rock-on-rock abrasion (silt-sized particles) vs. particle crushing (gravel- and sand-sized particles; Haldorsen, 2008; Licht and Hemming, 2017). Further fractionation of grain sizes occurs due to the presence of meltwater or the release of material into glaciomarine settings where various processes segregate sediment differently based on size and sorting (e.g., meltwater streams vs. meltwater plumes vs. iceberg rafting, vs. delivery and deposition of diamict at a grounding line; i.e., Licht and Hemming, 2017).

In most cases, the path from primary source to ultimate glacial depositional site for detritus and hence detrital zircons is an indirect path temporally and spatially and often involves multiple stages of recycling through multiple sedimentary systems and basins prior to the last depositional stage (Thomas, 2011; Fedorchuk et al., 2020; Zieger et al., 2019, 2020). The durability of zircons guarantees that they will survive multiple stages of recycling thus complicating interpretations. In some cases, primary sources are detached from depositional sites due to plate tectonic activity, which make it difficult to determine the location of the primary source. For example, Laurentia served as a source, in part, for material deposited in what is now the Precordillera Range in western Argentina, and these materials underwent multiple stages of reworking following separation of the Cuyania terrane from Laurentia in the Cambrian (Thomas et al., 2015; Malone, 2020). Identifying roundness values for various detrital zircon age populations may aid in identifying recycled grains (Zieger et al., 2019, 2020), while Lu—Hf and Oxygen isotopes along with heavy mineral compositional data can discriminate between potential sources with overlapping U—Pb ages but with different crustal histories, thus, helping to alleviate some of these issues (Hawkesworth and Kemp, 2006; Griffis et al., 2019). However, to be able to accurately reconstruct the location of ice centers and the size of LPIA glaciers, requires integration of dispersal pathways (provenance, paleoflow directions) into a “stratigraphic, sedimentologic, tectonic, and paleogeographic framework” (Thomas, 2011; Pauls et al., 2020) that for glacial systems may not be applicable if classic direct source to sink models are used.

2.3.2. Chemical index of alteration (CIA)

Multiple geochemical proxies are used in the study of ancient climates. However, one of the most commonly applied to near-field glaciogenic deposits is the use of the chemical index of alteration (CIA), which attempts to determine the degree of alteration that clay minerals have undergone during the weathering process (e.g., Nesbitt and Young, 1982; Maynard, 1992; Fedo et al., 1995; Bauluz et al., 2000; Price and Velbel, 2003; Sheldon, 2006; Soreghan and Soreghan, 2007; Sheldon and Tabor, 2009; Goldberg and Humayun, 2010; Bahlburg and Dobrzinski, 2011). Hence, this proxy is used to determine the degree of physical versus chemical weathering, and hence cold/arid versus humid climatic conditions that minerals have been exposed to. In glacial-bearing successions, low CIA values suggest the occurrence of physical weathering under arid/glacial conditions while higher values suggest warmer more humid settings. This proxy is defined as the proportion of labile cations (i.e., Ca^{2+} , Na^{+} and K^{+}) that are lost due to chemical weathering processes in relation to the stable Al^{3+} content (Nesbitt and Young, 1982, 1989). CIA is calculated using the following equation:

$$\text{CIA} = (\text{Al}_2\text{O}_3 / \text{Al}_2\text{O}_3 + \text{CaO}^* + \text{Na}_2\text{O} + \text{K}_2\text{O}) \times 100$$

where CaO^* is the amount of CaO incorporated in the silicate fraction of the rock. Nesbitt and Young (1982) geochemically analysed a variety of typical clastic sedimentary rocks and sediments and calculated their average CIA values. In the case of shale and mudrock examples, CIA value averages range from 70 to 75, which corresponds to the high content of illite and chlorite that are contained in those types of rocks. However, muds that developed in tropical or highly weathered environments result in average values of 80 to 100, and generally contain higher percentages of kaolinite. Deposits from glacial and arid environments were found to have values ranging from 50 to 70 and corresponded to the presence of unweathered K-feldspar and plagioclase (Nesbitt and Young, 1982). Therefore, the mineralogical composition of the source rock (e.g., soils, paleosols, sediments or igneous or meta-sedimentary rocks) must be determined prior to using the CIA as a paleoclimate proxy as different minerals produce varying CIA values (Nesbitt et al., 1996; Price and Velbel, 2003; Garzanti and Resentini, 2016).

The CIA can be applied to interpret the weathering trends of most igneous rocks; however, caution should be applied in cases that involve certain source rock compositions and reworked or altered minerals derived from metamorphic, metasedimentary, or sedimentary rocks (Price and Velbel, 2003). For instance, CIA performs under the assumption that Al_2O_3 , CaO , Na_2O , and K_2O originate in feldspars, which works in most cases due to the abundance of feldspars in the Earth's crust (Nesbitt and Young, 1989). CIA values of fresh feldspars and biotite plot at 50, muscovite and illite plot at 75, and kaolinite plots at 100 (Nesbitt and Young, 1989). Therefore, for different compositions of igneous rocks (i.e., mafic or acidic), each mineral will weather to a different clay mineral. Considering that each igneous rock contains a different percentage of each mineral, the CIA value can vary from rock type to rock type (Nesbitt et al., 1997). In cases where the source rock contains abundant amphiboles or pyroxenes, problems can occur because the starting CIA values are already low. Amphibolites may have a starting CIA value of around 25 and a highly weathered amphibolite sample may only have a CIA value of around 50. Whereas 50 is the starting CIA value of most granites and rhyolites, which weather toward a CIA value of 100 (Nesbitt et al., 1996, 1997).

Additionally, the problem with reworked metasedimentary and/or sedimentary rocks is that they may inherit a weathering signature from a past sedimentary cycle (Krissek and Kyle, 2000b,a; Price and Velbel, 2003; Bahlburg and Dobrzinski, 2011; Pauls et al., 2019). Compositional heterogeneity in the source rock may also cause considerable differences in the CIA value of the weathered sediment (Price and Velbel, 2003). This compositional variation could be caused by differences in the sorting of sedimentary rocks, such as the winnowing or infiltration of fines, or foliations in metamorphic rocks (cf. Maynard, 1992; Price and Velbel, 2003; Soreghan and Soreghan, 2007; Goldberg and Humayun, 2010; Bahlburg and Dobrzinski, 2011). Furthermore, the diagenetic reactions associated with the introduction and interaction with groundwater (i.e., metasomatism and illitization leading to increased K^+ concentrations) can skew the results of a CIA analysis (cf. Fedo et al., 1995).

Due to the complexities described above, the CIA is mainly dependent both on the protolith and on the paleoclimate. The paleoenvironmental CIA thresholds shown by Nesbitt and Young (1982) can vary from place to place according to the composition of the protolith and several other settings in a catchment area, such as vegetation, grain sorting, and slope (Wang et al., 2020). Thus, changes in CIA values in a vertical succession provide an assessment of trends in changing climates, but not a quantitative measurement that represents strict tropical/humid or glacial/arid conditions. Establishing the mineralogy of the source rock for a drainage basin can potentially allow for the projection of weathering profiles. This is because the rock is subjected to either chemical weathering or physical weathering processes over time, and

this would allow for the use of this proxy in intrabasinal and extrabasinal comparisons. DZ analysis may also provide insight into whether sediment was derived from a primary igneous or metamorphic source or whether the materials were reworked and supplied by sedimentary sources. In addition, making sure that all samples are of similar grain size (i.e., mud: silt and clay), somewhat reduces a sorting bias due to transport processes and distances (Soreghan and Soreghan, 2007; Goldberg and Humayun, 2010; Fedorchuk et al., 2019a; Pauls et al., 2019; Pauls, 2020). Therefore, CIA is a powerful tool in potential glacial successions in determining the onset and demise of glaciation, especially when combined with other lines of evidence such as the occurrence of lithologic indicators and the use of facies analyses.

2.4. Summary and importance of near-field proxies

The LPIA was the longest and most extensive glaciations of the Phanerozoic and is one of the most studied pre-Pleistocene icehouse intervals in Earth's history. However, there is still much disagreement about how extensive glaciation was in time and space. Was it the result of a single massive, long-lived ice sheet, or was it the result of as many as 25 different ice sheets and ice caps that waxed and waned diachronously across Gondwana? Did the northern hemisphere experience glaciation, and if so, what was its relationship to Gondwana glaciation? The answers to these questions are critical in understanding icehouse climatic conditions and icehouse to greenhouse transitions and will go a long way in helping us understand the evolution of climate on Earth.

Near-field depositional and geochemical proxies have played, and will play, a major role in identifying key aspects of the LPIA and other glacial intervals. Although our discussion of near-field proxies may leave the reader wondering if there is any hope in being able to address the above questions as every proxy used to identify glaciation and the occurrence of glacial deposits can also be produced by non-glacial processes. Our goal is to encourage researchers to ask the hard questions, to critically investigate the diamictites, clasts, lonestones, surfaces, rhythmites, glendonites, detrital zircons, clays and other aspects of the deposits to examine these features with a critical eye, and to rigorously test their origins by using a multiproxy approach to conducting deep-time, glacial research. By carrying out such investigations, the richness of the dynamics of the LPIA and other glacial intervals will be realized, and a more robust history of these critical intervals will be achieved.

3. Far-field proxies and their limitations

Far-field proxies for glaciation are used to identify high-frequency and high-magnitude changes in relative sea level as a proxy for determining the onset, duration, intensity, and termination of glaciation. Other proxies may aid in characterizing the climate during icehouse intervals but identifying eustatic sea level and its implications for fluctuating ice volumes are the focus of far-field proxies in this paper. These proxies include: 1) the occurrence of cyclic sedimentation, cyclothems, produced by high-frequency and high-magnitude changes in relative sea level; 2) the occurrence of deeply-incised paleovalleys suggesting large magnitude drops in base level; 3) the occurrence of paleotopographic relief on unconformities where direct measurements of relative sea level fall are recorded; the occurrence of "deep-water" phosphatic black shales sandwiched between shallow-water limestones within cyclothems used to suggest rapid high-magnitude changes in water depth over an individual sea-level cycle; and 4) changing $\delta^{18}\text{O}$ values as a means of determining high-magnitude eustatic sea-level changes associated with ice-volume fluctuations (e.g., Heckel, 1977, 1994; Smith and Read, 2000; Chen et al., 2013, 2016). There are other indicators of climatic change (i.e., stratal onlap and offlap patterns, paleosols, carbonate-siliciclastic alternations, changing flora and fauna assemblages and other geochemical proxies; e.g., Ross and Ross, 1988; Shi, 2001; Rosenau et al., 2013; DiMichele et al., 2009; Buggisch et al., 2011;

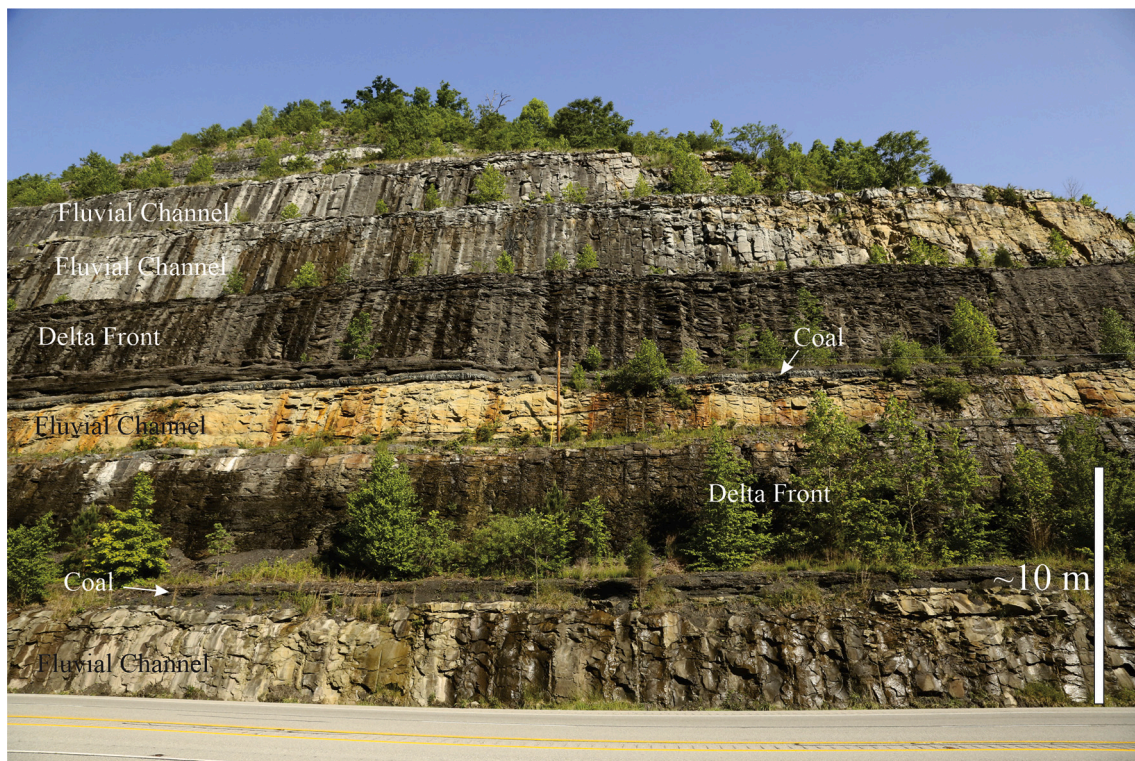


Fig. 15. Exposure of the Pennsylvanian Breathitt Group south of Pikeville, Kentucky, USA. The exposure shows alternating marine and nonmarine strata (Cyclothems) with sandstone units interpreted by [Le Cottonnet et al. \(2020\)](#) as incised valley fills.

[Zeng et al., 2012](#); [Shi and Waterhouse, 2010](#); [DiMichele, 2014](#); [Dyer and Maloof, 2015](#)); however, those listed above are commonly used at identifying glacial intervals and estimating changes in sea-level/ice-volume fluctuations during icehouse intervals. Discussions of the five proxies used here, focus on their strengths and weaknesses.

3.1. Cyclothems

[Wanless and Shepard \(1936\)](#) were the first to link sea-level fluctuation associated with the deposition of Euro-American cyclothems with ice-volume fluctuations in Gondwana. Since that time, cyclothems have been used to estimate magnitudes of glacioeustatic change and duration of climate cycles (e.g., [Heckel, 1994](#)). High-frequency Carboniferous and Permian cyclic strata (cyclothems) characterize much of North America (e.g., Midcontinent, Illinois, Appalachian, Midland, Black Warrior, Sydney basins), north-western Europe (e.g., Pennine Basin, Campine Basin) and the Russian Platform (e.g., Donets and Moscow basins) ([Ross and Ross, 1988](#); [Eros et al., 2012](#); [Heckel, 2002, 2013](#)). These strata, deposited in far-field, low-latitude settings, are characterized by transgressive-regressive depositional cycles of nonmarine sandstones mudrocks, and coal alternating with shallow marine/offshore/delta front mudrocks and limestones (Fig. 15; [Udden, 1912](#); [Weller, 1930](#); [Wanless and Weller, 1932](#); [Ross and Ross, 1988](#); [Heckel, 2013](#)). However, cycles consisting of marine mudrocks alternating with limestones or stacked shoaling upward limestone packages also occur ([Heckel, 1977, 2002](#); [Ross and Ross, 1988](#)). Other proposed causes of cyclicity include tectonism ([Weller, 1930](#); [Klein and Willard, 1989](#)), autocyclicity (i.e., fluvial avulsions or delta switching; [Wanless, 1964](#); [Beerbower, 1969](#)), and regional wet-to-dry climate cycles ([Cecil, 1990](#)). However, linkages between ice-volume fluctuations of late Paleozoic glaciers and sea-level changes that influenced the deposition of cyclothems are obvious ([Wanless and Shepard, 1936](#); [Crowell, 1978](#); [Veevers and Powell, 1987](#); [Heckel, 1994](#)), although other causes that contribute to sea-level change cannot be ruled out (see [section 4](#)). Approximately

120 minor, intermediate, and major cycles occur with estimated periodicities ranging from 20 to 480 ky (e.g., [Heckel, 1986, 1994](#); [Maynard and Leeder, 1992](#); [Weedon and Read, 1995](#); [Rasbury et al., 1998](#); [Algeo and Heckel, 2008](#); [Eros et al., 2012](#); [Hinnov, 2013](#)). These are in the range of astronomical forcing (obliquity, precession, and short- and long-term eccentricity; [Hays et al., 1976](#); [Berger et al., 1992](#); [Hinnov, 2000](#); [Strasser et al., 2006](#); [Horton et al., 2012](#)). Longer term cycles between 1.2 and 4.0 Myr also occur and may be due to longer term orbital forcing or tectonic controls (c.f. [Ross and Ross, 1988](#); [Chesnut, 1994](#); [Eros et al., 2012](#)).

In terms of the magnitude of relative changes in sea level, [Rygel et al. \(2008\)](#), studies reported therein)) synthesized data from 100 papers on Carboniferous and Permian far-field cyclothems, near-field data, and models that estimated the magnitude of eustatic change. Estimates in these studies were obtained from (1) depth of incision of paleovalleys (see [section 3.2](#)), (2) thickness of shallowing upward carbonate cycles, (3) depth of karst surfaces, (4) cycle thicknesses, (5) depth of deposition of black shales below a pycnocline (see [section 3.3](#)), (6) height of bioherms (paleotopography), (7) oxygen isotopes (see [section 3.4](#)), (8) computer modelling, (9) ice volume modelling (see [section 4.1](#)), and (10) backstripping (e.g., [Heckel, 1977, 1986, 1994](#); [Crowley and Baum, 1991](#); [Elrick and Read, 1991](#); [Maynard and Leeder, 1992](#); [Soreghan and Giles, 1999](#); [Smith and Read, 2000](#); [Isbell et al., 2003](#); [Joachimski et al., 2006](#); [Rygel et al., 2008](#)). Depending on the time interval, [Rygel et al. \(2008\)](#) suggested changes of 20–25 m in the Early Mississippian (Tournaisian), 10–50 m in the Middle Mississippian (late Viséan), 40–100 m in the latest Mississippian–earliest Pennsylvanian (latest Viséan to Late Bashkirian), <40 m in the Middle Pennsylvanian (latest Bashkirian to late Moscovian), 100–120 m in the late Pennsylvanian (Kasimovian)–earliest Permian (mid-Sakmarian), 30–70 m in the Early–Middle Permian (mid-Sakmarian through Kungurian), and 10–60 m in the Middle Permian (Roadian through Capitanian). From their study, [Rygel et al. \(2008\)](#) concluded that cyclic fluctuations of <25 m may have had a glacioeustatic cause, but other causes could not be ruled out.



Fig. 16. Pennsylvanian incised valley fills of the A) the New River Sandstone near Beckley, West Virginia, USA and B) the Pennsylvania Breathitt Group northeast of Pikeville, Kentucky, USA. Note the multiple scoured surfaces along the left edge of sandstones filling the “incised valley” in the lower rock face (channels 1–3). The scours and the sandstones above the scours extend laterally and grade into the succession to the right. Also, the sandstone above the first bench (channel 4) cuts down to the right and cuts into the sandstones in the lower rock face (channels 1–3). These features suggest that these sandstones for a composite incised valley fill representing multiple changes in base level.

Deposits that returned relative sea-level signals of 25–60 m corresponded to times of regional glaciation, while large to very large fluctuations of 60–120 m occurred during widespread continental glaciation. Rygel et al. (2008) also compared their results with widely cited onlap curves from Ross and Ross (1985, 1987, 1988) who reported eustatic changes of between 100 and 200 m for third-order cycles in the Mississippian, Pennsylvanian, and Permian. However, the dataset by Ross and Ross is confidential and is not reproducible and should therefore be ignored (Rygel et al., 2008). The link between cyclothems, sea-level fluctuations, and changes in glacial mass balance cannot be denied. However, the magnitude of sea-level change related to ice-volume fluctuations plus other factors that may have contributed to relative sea-level change across the globe also need to be considered (see section 4).

3.2. Incised valleys

River incision into shallow-marine successions is linked to adjustments of the system to changes in base level, and hence is used to estimate magnitudes of late Paleozoic sea-level change (Fig. 16; Smith and Read, 2000; Rygel et al., 2008). The depth of an incised valley cut into a coastal/delta plain or shelf succession is often considered a reliable indicator of the magnitude of sea-level fall. Conversely sea-level rise is calculated by adding the depth of incision to the estimated water depth for the marine unit marking the overlying maximum flooding interval

(Smith and Read, 2000). The depth of incision for most Carboniferous and Permian incised valleys ranges from 8 to 150 m with most studies reporting depths of incision of <60 m (Aitken and Flint, 1995; Gibling, 2006; Rygel et al., 2008; Fischbein et al., 2009; Belt et al., 2011; Wakefield and Mountney, 2013; Buller et al., 2018; Le Cottonnec et al., 2020). In a study by Gibling (2006), he noted that the thickness of Carboniferous valley fills (average of 31 m and a maximum of 200 m) where larger than Mesozoic examples (average 21.2 m and a maximum of 40 m), which he attributed to greater eustatic fluctuations during icehouse than during greenhouse climates.

Although incised valleys and incision at the base of channels occurs due to a drop in base level (e.g., sea level, tectonics), changes in energy flux due to changing climatic regime (i.e., river discharge) and/or changes in sediment supply also cause deep channel scour (Zaitlin et al., 1994; Blum and Törnqvist, 2000; Catuneanu, 2006; Gibling, 2006; Korus et al., 2008; Noorbergen et al., 2020). Channel incision also occurs due to an increase in stream discharge downstream of the confluence of two rivers/channels, scour around channel-bends, and due to increased flow through valley-constrictions (Blum and Törnqvist, 2000; Ardies et al., 2002; Gibling, 2006; Noorbergen et al., 2020). In such instances, the depth of incision may be as much as five times greater than mean channel depth (Best and Ashworth, 1997; Ardies et al., 2002; Gibling, 2006).

Deltaic distributaries tend to shallow toward their mouth (terminal



Fig. 17. Pennsylvania Phosphatic black shale (Huskpuckney Shale) overlying a rippled carbonate grainstone (Middle Creek Limestone) in Kansas City, Missouri, U.S. A. Hammer for scale.

distributaries) where shallow-water mouth bars form as flow transitions from confined to unconfined flow (Olariu and Bhattacharya, 2006). This concept and the concept of base level, “the theoretical lowest level toward which erosion of the Earth’s surface constantly progresses”... “below which a stream cannot erode its bed” (sea level) as defined by Neuendorf et al. (2005), is often used to suggest that river channels cannot scour below sea level. However, channels on coastal and delta plains do scour their bases well below sea-level even during sea-level highstands where the depth of channel scour is dependent on the size and discharge of the river system and site-specific flow dynamics (Gibling, 2006; Olariu and Bhattacharya, 2006). For example, scours on the outside of meander bends in the Mississippi River at New Orleans, USA and scours associated with confluences of anastomosed channels in the Amazon River Delta in Brazil have bases that are as much as 60 m below sea level (Meckel, 1972; Gibling, 2006; Water depth map and nautical charts, 2021). Smaller systems show lesser degrees of channel scour. For incised valleys, many measured incised valleys are in the Appalachian and Illinois basins high up fluvial valleys when the seas retreated to areas in Oklahoma during sea-level lowstands (e.g., Bristol and Howard, 1971; Beuthin, 1994; Smith and Read, 2000; Rygel et al., 2008; Le Cotonnec et al., 2020). Because coastal plain and deltaic channels scour below sea level, the use of depth of scour of an incised valley system may not directly reflection the magnitude of base-level fall as components to the total erosional relief caused by scour related to climate (i.e., discharge), channel configuration (confluence, bend, or valley constriction) or tectonism cannot be ruled out (cf. Bristol and Howard, 1971; Beuthin, 1994; Best and Ashworth, 1997; Smith and Read, 2000; Ardies et al., 2002; Gibling, 2006; Rygel et al., 2008). In some cases, composite incised valleys also occur that record multiple changes in base level or energy fluxes (Fig. 16B; cf. Zaitlin et al., 1994; Bowen and Weimer, 2003; Korus et al., 2008). If any of these factors remain unrecognized, the use of incised valleys as indicators of the magnitude of base-level change will return anomalous values of sea-

level change.

3.3. Paleotopographic relief

Topographic relief on subaerial and shallow-marine unconformities, landscapes, and drowned surfaces cut by wave action also serve as proxies for estimating relative changes in sea level especially in the Pleistocene to recent record (Kelsey, 2015). This relief serves as a measure of the magnitude of relative sea level fall. Features and surfaces measured include the heights of marine terraces, raised beaches/beach ridges, reef terraces, and relief on subaerial unconformities cut on shallow marine strata (Soreghan and Giles, 1999; Sasaki et al., 2004; Muhs et al., 2012; Kelsey, 2015). The maximum relief on subaerial unconformities cut on shallow water successions are close approximations of drops in relative sea level. In the deep time record, such surfaces and features are rare (cf. Rygel et al., 2008), difficult to identify, and require laterally extensive exposures and/or 3-D surfaces to identify. Soreghan and Giles (1999) report one such feature in the Late Pennsylvanian shallow-water carbonate rocks of New Mexico. There, in the western Orogrande Basin, an 0.5 km long exposure contains a large algal bioherm that grew into shallow water. The bioherm is cut by several unconformities in which two display a maximum relief of approximately 80 m. Because the top of the sequences displaying the greatest relief are not preserved, they estimated that the potential magnitude of sea level fall was approximately 100 m (Soreghan and Giles, 1999). This exposure does not appear to have suffered from compaction (few stylites in the limestones) and because of the limited extent, they ruled out differential tectonic subsidence for a cause of the relief. Although this exposure provides excellent insight into relative sea level change for a snapshot view of the late Pennsylvanian by using this technique, it is only a single point for the LPJA and therefore, provides limited temporal and spatial information of changes in relative sea level.

In general, relative sea level changes for the Last Glacial Maximum

(26,000 years ago) to recent recorded by marine terraces, beach ridges, reef terraces, and shallow marine unconformities are the result of multiple processes that include eustatic sea level change; tectonic uplift; and in high latitude sites, glacioisostatic rebound following deglaciation (Kelsey, 2015).

3.4. Phosphatic black shales

Within the midcontinent of North American, the occurrence of widespread phosphatic black shales sandwiched between thick regressive limestones are used to estimate water depth fluctuations within cyclothems and hence serve as a proxy for establishing late Paleozoic eustatic fluctuations (Heckel, 1977, 1986, 1994; Algeo and Heckel, 2008). In the Mid-Continental Shelf/Basin of Kansas, Nebraska, Oklahoma, Missouri, and Iowa, U.S.A., 3–10-m thick cyclothems consist mainly of marine shales and limestones. In some cases, major cyclothems are capped by paleosols deposited during sea-level lowstands (Heckel, 1977). The deepest water facies of these deposits are m-thick, organic-rich, phosphatic, gray to black shales (Fig. 17) that are bracketed above and below by shallow-water limestones. These shales accumulated in subpynoclinal, anoxic, bottom-waters (Heckel, 1977, 1994, 2008, 2013; Algeo and Heckel, 2008; Algeo and Heckel, 2008). Although water depths for black-shale deposition are difficult to ascertain, Heckel (1977, 1994) reasoned that deposition occurred beneath a pynoclinal/thermocline in approximately 100 m of water leading him to estimate a similar value for sea-level fluctuations for these mainly marine cyclothems. In discussing a super-estuarine circulation model for marine anoxia in the Late Pennsylvanian Mid-Continent Sea, Algeo et al. (2008) and Algeo and Heckel (2008) suggested glacioeustatic changes of between 60 and 150 m. Their values were based on Heckel's estimate and estimates based on conodont $\delta^{18}\text{O}$ variations within these shales (Joachimski et al., 2006). However, sea-level fluctuations of up to 150 m are not required for black shale deposition in the Mid-Continental Shelf, which had a deep-water (> 150 m), tortuous connection to the Panthalassan Ocean through the Delaware, Midland, Palo Duro, and Anadarko basins (Algeo and Heckel, 2008). Algeo and Heckel (2008) and Algeo and Heckel (2008) hypothesized a shallow pynocline (~15–30 m), and an average water depth of 50 m for the Late Pennsylvanian Mid-Continent Sea, which contrasts with their suggested eustatic change. Ahern and Fielding (2019) suggested similar relative sea-level oscillations of 18–36 m for deposition of Serpukhovian organic-rich, black shale-bearing succession from the Big Snowy Trough in Montana, U.S.A. However, the Big Snowy Trough may have had water circulation restricted by a sill within the trough, whereas the Mid-Continental Shelf developed anoxic bottom waters due to super-estuarine circulation. Regardless, eustatic changes of no more than 15 to 50 m were needed to explain the formation of these phosphatic black shales.

3.5. Oxygen isotope ratios

Oxygen Isotopes ratios ($\delta^{18}\text{O}$ values) obtained from marine brachiopods, bivalves, foraminifera, corals, and conodonts serve as proxies for determining Paleozoic-recent sea surface temperatures, changes in eustatic sea level, and linked changes in ice volume (cf. Shackleton, 1987; Fairbanks, 1989; Grossman et al., 2008; Chen et al., 2016; Qie et al., 2019). Holocene/Pleistocene examples have shown that separating changes in temperature and eustasy/ice volume are challenging. The shells of near-surface (< 300 m water depth) planktonic and benthic organism reflect sea surface temperatures and regional difference in evaporation, precipitation, and freshwater influx (Shackleton, 1967; Miller et al., 2005, 2011; Brand et al., 2009; Chen et al., 2013). However, deep-sea bottom (> 1000 m) waters are not influenced by changing surface conditions as the temperature of these waters are assumed to be less variable as they are controlled by polar bottom waters. Therefore, $\delta^{18}\text{O}$ values recovered from benthic organism that lived at depths of >1000 m more closely reflect changes in ice volume and long-term

climate driven changes in temperature (Shackleton, 1967; Miller et al., 2005; Raymo et al., 2018). Typical water depths for deep-water samples during the Cenozoic are taken from depths of 2300 to 4200 m, as greater depths are below the carbonate compensation depth (Shackleton, 1987; Ahmad et al., 2008; Raza and Ahmad, 2013). Although ice-volume fluctuations control most of the observed variability in the $\delta^{18}\text{O}$ record in deep oceanic waters over tens to hundreds of thousand years, long-term temperature changes over millions to tens of millions of years can be up to triple the values attributable to ice-volume changes (Miller et al., 2005). Therefore, scaling of long-term temperature changes must be accounted for. Also, it must be acknowledged that thermohaline circulation could have been different in the Paleozoic (Beauchamp and Baud, 2002; Jones et al., 2006) and its effect on the $\delta^{18}\text{O}$ record is unknown.

Because seawater $\delta^{18}\text{O}$ values increase as ice accumulates, and fall during deglaciation, changes in $\delta^{18}\text{O}$ values derived from deep-sea benthic fossils are used to determine coupled changes in ice volume and eustasy. During the Pleistocene, $\delta^{18}\text{O}$ values of sea water from coral (aragonite) were calibrated against a number of parameters, including: 1) known changes in sea level from the Last Glacial Maximum (LGM) to recent (Fairbanks and Matthews, 1978; Fairbanks, 1989), 2) estimates of ice volume, 3) ice $\delta^{18}\text{O}$ of the LGM ice sheets (Dansgaard and Tauber, 1969), and 4) temperature-corrected changes in $\delta^{18}\text{O}$ from benthic foraminifera between the LGM and Holocene (Shackleton and Opdyke, 1973; Winnick and Caves, 2015). For the LGM to recent, $\delta^{18}\text{O}$ was estimated to vary by 0.08–0.11‰ for each 10 m change in eustatic sea level (Fairbanks and Matthews, 1978; Fairbanks, 1989; Winnick and Caves, 2015). However, caution is required for applying this calculation to deep-time (> 0.5 Ma) oxygen isotope studies due to long-term deep-water temperature changes, changes in salinity, and changing $\delta^{18}\text{O}$ values of glaciers through time as ice sheets grow (cf. Mix and Ruddiman, 1984; Miller et al., 2005; Rohling et al., 2014; Winnick and Caves, 2015; Gasson et al., 2016; Raymo et al., 2018). There is also a decoupling between $\delta^{18}\text{O}_{\text{seawater}}$ and sea-level changes due to loss of ice for marine ice sheets (ice sheet with its base below sea level). $\delta^{18}\text{O}_{\text{seawater}}$ is a total ice-mass signal not necessarily a sea-level signal and a disconnect occurs when ice is lost from subglacial basins of marine ice sheets when sea water fills the space formerly occupied by the ice mass (Gasson et al., 2016). Likewise, the growth and decay of an ice shelf (floating ice mass) does not affect eustasy as the ice is floating and its mass is already compensated by displacement of an equivalent mass of water. However, the growth of an ice shelf or its loss influences $\delta^{18}\text{O}$ values in the world's oceans. Much of the Gondwana glacial record is from glaciomarine strata (Fig. 2) suggesting that ice entering the basins and displaced sea water, thus growth or loss of these glaciers did not have a 1-to-1 relationship between fluctuations in ice mass, sea-level change, and $\delta^{18}\text{O}$ values.

Comparing late Paleozoic changes in $\delta^{18}\text{O}$ values with Fairbanks and Mathews' (1978) LGM-Recent model suggested a 160 m drop in sea level on the Russian Platform (brachiopod calcite) from late Viséan/early Serpukhovian to early Bashkirian (Mii et al., 2001) and eustatic fluctuations of >120 m for strata containing conodont apatite deposited in the Midcontinent Seaway of North America during the Virgillian (Gzhelian) (Joachimski et al., 2006). Buggisch et al. (2008) using composite values $\delta^{18}\text{O}$ from the Late Devonian to Late Mississippian suggested a progressive increase in ice volume over 38 Myr to correspond to a progressive lowering of sea level by approximately 200 m. In using oxygen isotopes to reconstruct eustatic change, $\delta^{18}\text{O}$ values are assumed to represent chemical signatures within the world ocean. However, these LPIA oxygen-isotope records come from epicontinental seas or basins that displayed restricted circulation and large salinity fluctuations (cf. Algeo and Heckel, 2008; Grossman et al., 2008; Joachimski and Lambert, 2015; Roark et al., 2017; Montañez et al., 2018; Qie et al., 2019), which make data interpretations and absolute eustatic values ambiguous. Fossil $\delta^{18}\text{O}$ values are affected by local salinity differences (e.g., evaporation-precipitation, freshwater influx), ice-volume

fluctuations, water temperatures relative to depth, surface-air temperatures above ice sheets, migration of organisms to different water depths, climate change due to emission of volcanic greenhouse gasses, and diagenesis (Emeis et al., 2000; Miller et al., 2005, 2011; Chen et al., 2013; Winnick and Caves, 2015; Montañez et al., 2018). Brand et al. (2009) and Chen et al. (2013) suggested that oxygen isotope values from late Paleozoic open-ocean sources are decoupled from those obtained from epeiric sea samples. Consequently, this epeiric sea problem may limit reconstructions of absolute changes in sea level due to ice-volume fluctuations in deep time (Miller et al., 2005, 2011). However, Frank et al. (2008) and Montañez et al. (2018) suggested that data sets from basins with restricted circulation still provide valid paleoclimatic and eustatic trends. Therefore, if true, $\delta^{18}\text{O}$ values from these basins signal the onset, demise, and perhaps the relative intensity of glaciation.

During early stages of the LPIA, peaks in $\delta^{18}\text{O}$ values (Fig. 1) occurred during the Famennian (Late Devonian; (Brand et al., 2004; Isaacson et al., 2008), Tournaisian and Viséan (Early to Middle Mississippian; Mii et al., 1999; Saltzman, 2002; Buggisch et al., 2008; Frank et al., 2008; Grossman et al., 2008; Isaacson et al., 2008; Wallace and Elrick, 2014). These peaks suggest that either long-term cooling or multiple glacial events occurred during the early part of the LPIA. However, these oxygen isotope values are primarily from the North American Midcontinental and Russian Platform epeiric seaways, while some are composite records constructed from widely scattered basins (Fig. 1).

Middle Mississippian (late Viséan) to Middle Permian (Guadalupian) strata from the Naqing section of South China contains a near continuous succession of limestones deposited in the Luodian Basin, a narrow, shallow-water (50 to ~200 m deep), tropical basin surrounded by an extensive shallow-water (< 50 m deep) platform (Chen et al., 2013, 2016). Due to its location between the Paleo-Tethys and the Panthalassan oceans, South China is regarded as having well-mixed oceanic waters. However, the extent and shallowness of the basin and surround platform, raises questions as to the vigor of circulation between the Luodian Basin and the open ocean and whether or not $\delta^{18}\text{O}$ values are free from surface temperature and salinity influences. Conodont apatite from this basin show high $\delta^{18}\text{O}$ (‰VSWOM) values in the late Viséan and highest values in the Serpukhovian to Bashkirian suggesting that the LPIA glacial maxima occurred across the Mississippian-Pennsylvanian boundary (Fig. 1; Chen et al., 2016; Qie et al., 2019; Wang et al., 2019). High values representing major glaciation also occurred in the Late Pennsylvanian and Early Permian with values falling off in the Kungurian to Capitanian suggesting that the main phase of deglaciation occurred at this time (Chen et al., 2011, 2013, 2016; Shen et al., 2018; Qie et al., 2019). This contrasts with composite $\delta^{18}\text{O}$ records by Korte et al. (2005, 2008) and Frank et al. (2008) with many of their sites including samples deposited in epeiric seaways. Frank et al. (2008) noted a major peak in the Bashkirian followed by a drop in values throughout the rest of the Pennsylvanian (Fig. 1). Both Frank et al. (2008) and Korte et al., (2005, 2008) noted a major $\delta^{18}\text{O}$ (‰VSWOM) peak in the late Asselian/early Sakmarian followed by a long-term decline in values into the Kungurian. Both show additional smaller subsidiary peaks in either the Kungurian and Wordian (Korte et al., 2005, 2008) or Roadian and Capitanian (Fig. 1; Frank et al., 2008). Korte et al., (2005, 2008) noted that these peaks could be the result of evaporative enrichment of seawater in $\delta^{18}\text{O}$ in the Delaware Basin (USA) while Frank et al. (2008) noted the peaks occurred during P3 and P4 glaciation episodes in eastern Australia.

Regardless, no truly deep-water, $\delta^{18}\text{O}$ record exists for the late Paleozoic Ice Age. Therefore, all $\delta^{18}\text{O}$ records are suspect in terms of actual quantitative sea-level or ice-volume changes. However, trends are still useful to signal changing climatic conditions and can be used to indicate times of glaciation/deglaciation and their relative intensities (Frank et al., 2008; Montañez et al., 2018).

3.6. Summary of far-field proxies

Far-field proxies have been used extensively to identify the onset, duration, and intensity of glaciation ever since the seminal work of Wanless and Shepard (1936) who linked cyclothems to Gondwana Glaciation. Although multiple causes drive eustasy, it is undeniable that fluctuations in ice volume operating on orbital time scales are the primary cause of Mississippian, Pennsylvanian, and Permian high-magnitude, high-frequency changes in eustasy. The eustatic frequency recorded by cyclothems is well established (e.g., Heckel, 1994, 2002, 2008, 2013; Eros et al., 2012). However, important questions that need further exploration include: 1) what was the magnitude of sea level change during the various cycles? 2) How robust is that record? 3) How did the magnitude of the eustatic change vary through time? 4) How much ice was needed to drive the observed changes? 5) Can individual glacial events be linked directly to far-field cycles? And 6) did other cause of eustasy contribute to the magnitude of individual cycles, and if so, how much did they contribute (See section 4.)? Researchers have insisted that the use of far-field proxies is better for estimating eustatic change than near-field proxies as it is difficult to pick out short-term cycles within the glacial deposits, and it is always assumed that the glacial record is incomplete due to subglacial erosion. All proxies are based on assumptions and that is also true for the proxies discussed above. In this section we took a critical look at the strengths and weakness of the main far-field proxies used to estimate the magnitude of eustatic change. Because any one of these proxies, when used alone, gives an incomplete view of eustasy, we again advocate for a multiproxy approach where results are compared and correlated to each other like that of the synthesis provided by Rygel et al. (2008). Through such work, a more complete picture of the LPIA can be obtained.

4. Ice volume and sea-level responses during glacial and post-glacial intervals

4.1. Relationship between ice volume and eustasy

The relationship between ice volume and eustasy is important in determining the link between upper Paleozoic cyclothems and Gondwana glaciation. The waxing and waning of LPIA glaciers resulted in orbital-driven high-frequency changes in global sea level (Heckel, 1994). However, the question remains as to the magnitude of eustatic fluctuations and how much ice volume was required to drive those changes.

Glaciers behave as Bingham plastics. Once ice reaches a critical thickness of ~40 m, the weight of the overlying ice column exceeds a critical yield stress, and the basal layers begin to deform. Due to its rheology, ice cannot be stacked in a single narrow thick column but flows outward to support the thickening column. To grow upward, ice expands outward. Therefore, a given thickness or volume (V) of ice requires a given areal footprint (S), which is calculated using the following equation (Paterson, 1994):

$$\log(V) = 1.23[\log(S) - 1]$$

The water equivalent (WE) is identified by converting the ice volume (V) into an equivalent volume of water using the average density (0.917) of glacial ice (Shumskiy, 1961):

$$\text{WE} = 0.917(V)$$

This volume distributed across the estimated area of the late Paleozoic world ocean ($374.5 \times 10^6 \text{ km}^2$) provides a sea-level equivalent (SLE; Crowley and Baum, 1991):

$$\text{SLE} = (\text{WE})/374.5 \times 10^6 \text{ km}^2$$

Then, once corrected for isostatic adjustment (SLEI), a eustatic change can be calculated for the change in ice volume in question

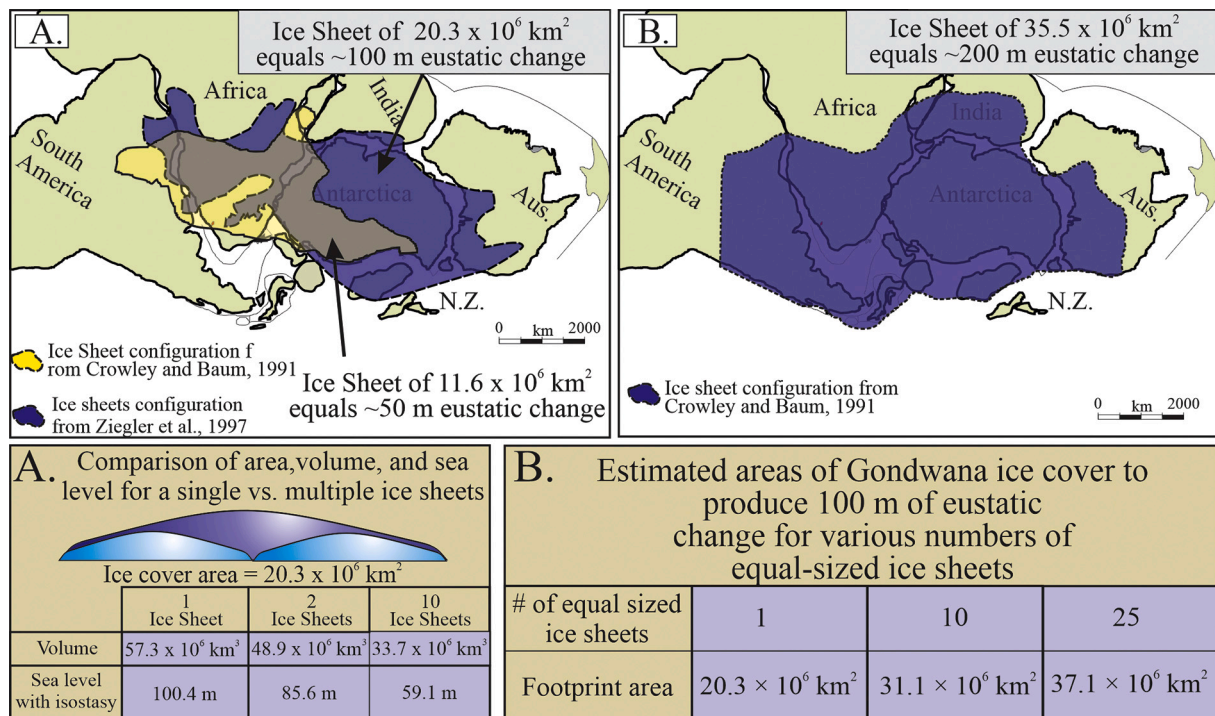


Fig. 18. Relationships between areal extent of ice cover, ice volume, and glacioeustasy (Modified from Isbell et al., 2003). A-B Visual comparison of the size of single ice sheets needed to produce a 50, 100, and 200 m change in eustacy due to complete waxing and waning of the ice sheet. C) Comparison of the area, volume and eustatic effect of ice covering an area of $20.3 \times 10^6 \text{ km}^2$ (size of a single sheet capable of producing 100 m of eustatic change) depending on the number of equal ice sheets with a total footprint of that size. Note that multiple ice sheets covering the same area contain less ice than a single ice sheet with the same footprint. D) Relationship between the number of ice sheets of equal size and the total volume of ice needed to produce 100 m of eustatic change.

(Crowley and Baum, 1991):

$$\text{SLEI} = (\text{SLE}) - 0.284(\text{SLE})$$

Fig. 18 shows the eustatic response for the complete waxing and waning of single ice sheets of various sizes (Isbell et al., 2003). During the late Paleozoic, growth and loss of a single ice sheet covering an area of between 11.6 and $20.3 \times 10^6 \text{ km}^2$ would have resulted in first a drop and then a rise of global sea level of between 50 and 100 m assuming water was distributed equally across the world's oceans (Crowley and Baum, 1991; Isbell et al., 2003). Small-scale glacial advance or retreat cycles would have produced smaller eustatic responses. Because a single ice sheet contains more ice by volume than multiple ice sheets covering an equivalent area, the potential change in sea level for melting of a given area of ice cover decreases as the number of ice sheets increases (Figs. 18C, D; Isbell et al., 2003). Throughout the LPIA, there are between 1 and 10, and possibly as many as 25 ice-spreading centers (ice domes, ice caps, and ice sheets) occurred across Gondwana at various times (cf. Crowell and Frakes, 1970; Veevers and Powell, 1987; Ziegler et al., 1997; Crowell, 1999; Isbell et al., 2003, 2012; Fielding et al., 2008a; Rosa and Isbell, 2021). Therefore, depending on the volume of ice in each ice sheet, a change of 100 m in sea level required growth and loss of an area of ice cover of between $20.3 \times 10^6 \text{ km}^2$ (single ice sheet), $31.1 \times 10^6 \text{ km}^2$ (10 equally sized ice sheets), and $37.1 \times 10^6 \text{ km}^2$ (25 equally sized ice sheets; Fig. 18D). Large ice sheets are incredibly stable and display only small, mass-balance fluctuations due to changes in insolation. High-magnitude and high-frequency eustatic fluctuations require a large number of highly sensitive, smaller ice sheets that respond quickly to climatic perturbations (cf. DeConto and Pollard, 2003; Horton and Poulsen, 2009). The numbers above represent nucleation, maximum advance, retreat, and complete disappearance of ice for the various glaciers to produce eustatic fluctuations of 100 m. However, for a eustatic fluctuation of that size to be produced by incomplete advance and retreat cycles, an even larger total volume of ice

and its areal extent are required. While small changes ($< 50 \text{ m}$) in global sea level are easily explained, large, high-frequency changes in eustacy require a tremendous area of Gondwana to be covered by ice and followed by complete deglaciation. It should also be noted that observable eustatic fluctuations were the result of the growth and decay of ice sheets ($> 50,000 \text{ km}^2$). Ice caps ($< 50,000 \text{ km}^2$) and alpine glaciers had little impact on eustacy (Isbell et al., 2003) as these ice masses contain little ice mass when compared to ice sheets.

Ice volume fluctuations are not the only cause of eustatic fluctuations. Processes that proceed on a similar time-scale ($\sim 10 \text{ m/ky}$) include: desiccation and inundation of marginal seas; thermal expansion and contraction of sea water; and fluctuations in water storage in groundwater, rivers and lakes (Miller et al., 2005). Changes in temperature can account for 1–2 m changes in sea level for each 1°C change in water temperature. Storage in groundwater and land surface reservoirs are more controversial. Miller et al. (2005) suggested only modest storage of water in these reservoirs (1–2 m of sea level equivalence); whereas, other authors suggest the storage and release of water from aquifers and lakes (aquifer eustacy) may account for greater than 50 m of sea-level change over 0.4 to 3 Myr time frames (Plint et al., 1992; Sames et al., 2016, 2020; Wendler et al., 2016; Wendler and Wendler, 2016). Therefore, all sources of climatically driven storage and release of water need to be reconciled when considering eustatic change during the LPIA.

4.2. Eustacy vs relative sea level

The far-field techniques discussed in section 3 (except for $\delta^{18}\text{O}$ studies) above return data from individual sites within separate depositional basins. Therefore, these data represent estimates of relative sea level rather than eustatic change. Because relative sea level, or accommodation, equals the sum of eustasy and subsidence (i.e., tectonic, sediment loading/unloading, compaction), these near-field proxy data

Table 1
Subsidence within selected basins over a single eustatic cycle.

Tectonic Setting	Subsidence rates mm/yr	Subsidence		
		40 k.y. Cycle	100 k.y. Cycle	413 k.y. Cycle
Cratonic Setting	0.03 – 0.05	1.2–2	3–5	12.39–20.65
Rift Basin	0.05–0.2	2–8	5–20	20.65–82.6
Foreland Basin	0.2–0.5	8–20	20–50	82.6–206.5
Forest City Basin (Klein and Kupperman, 1992)	0.028	1.12 m	2.8 m	11.56 m
Salinas Basin (Klein and Kupperman, 1992)	0.013–0.115	0.52–4.6 m	1.3–11.5 m	5.4–47.5 m
Illinois Basin (Klein and Kupperman, 1992)	0.01–0.03	0.4–1.2 m	1–3 m	4.13–12.39 m
Black Warrior Basin (Whiting and Thomas, 1994)	0.0053–0.22	2.2–8.8 m	5.3–22 m	21.9–91 m
Appalachian Basin (Klein and Kupperman, 1992)	0.016–0.1775	0.64–7.2 m	1.6–18 m	6.6–74.34 m
Donets Basin (Eros et al., 2012)	0.11–0.53	4.4–21.2 m	11–53 m	45.47–218.9 m (480 k.y. = 52.8–254.4 m)

cannot be related directly to eustacy or the ice-volume record for the whole of the LPIA without first being corrected for subsidence (Jervey, 1988; Emery and Myers, 1996; Catuneanu, 2006). Many studies neglect this consideration and simply assume that relative sea level for a given site directly equals to eustatic values. For a given study site, the degree to which relative sea level/accommodation differs from eustatic change is

dependent on the following variables: 1) the magnitude of eustatic change, 2) subsidence rates, and 3) the duration of an individual eustatic cycle (Jervey, 1988; Catuneanu, 2006). Table 1 displays subsidence rates for various cyclothem-bearing basins, while Table 2 shows the degree to which different eustatic signals, modelled as a sine curve, are modified by various combinations of these parameters. Slow basin subsidence rates and short-duration cycles have the least impact on the eustatic signal. However, high subsidence rates acting over long-duration cycles produce either a continuous rise in relative sea level for small magnitude changes in eustacy or a relative sea-level rise that is more than double that of eustacy for a high magnitude eustatic rise. Table 2 and Fig. 19 also show that the timing of a relative sea-level high lags that of a eustatic high, and depending on the combinations of the various parameters, could be unrecognizable (continuous rise) or lag the eustatic high by tens of thousands of years. Considering that all basins subside differentially, the magnitude and timing of a relative sea-level high varies across an individual basin. Basin margins track the magnitude and timing of eustatic highs/lows (little to no subsidence) while rapidly subsiding basin locations could display more than double the magnitude of eustacy with relative sea-level highs lagging eustatic highs by tens of thousands of years.

4.3. Glacial isostatic adjustment (GIA)

Glacioeustatic change represents the assumed loss or addition of a uniform thickness of water to or from the world's oceans due to changes in the volume of ice sheets as they wax and wane. However, the magnitude of relative sea-level change over a glacial cycle in sites distal to the ice front can deviate by as much as 40% or more from that of the eustatic signal (Mitrovica et al., 2009; Creveling et al., 2018; Horton et al., 2018). In contrast, relative sea level in ice-proximal settings, displays an opposite, out of phase response to that of far-field sites that may be many times that of the supposed eustatic response (Fig. 20; Mitrovica et al., 2009; Hay et al., 2014; Creveling et al., 2018). It is now known that fluctuations in ice-sheet volume (and therefore mass) not only drive changes in the volume of water in the Earth's oceans, but cause gravitational and rotational perturbations to the planet, shift

Table 2
Subsidence, eustacy and relative sea level changes over a single eustatic cycle.

Subsidence rates (mm/yr)	Cycle Length (k.y.)	Relative Sea Level Rise (m)					
		For a 10 m eustatic rise/fall	For a 25 m eustatic rise/fall	For a 40 m eustatic rise/fall	For a 60 m eustatic rise/fall	For a 80 m eustatic rise/fall	For a 100 m eustatic rise/fall
0.05	41 k.y.	11/9 ~1 k lag	26/23.9 <1 k lag	41/38.9 <1 k lag	61/58.9 <1 k lag	81/78.9 <1 k lag	100.68/98.6 <1 k lag
	100 k.y.	12.6/7.6 3 k lag	27.55/22.6 1 k lag	42.52/37.5 1 k lag	62.5/57.5 <1 k lag	82.5/77.5 <1 k lag	102.5/97.5 <1 k lag
	413 k.y.	22.6/1.9 47 k lag	36.2/15.6 18 k lag	50.9/30.2 11 k lag	70.7/50 7 k lag	90.5/69.9 6 k lag	110.5/89.9 5 k lag
0.1	41 k.y.	12.1/8 1 k lag	27/22.9 1 k lag	42/37.9 <1 k lag	61.9/57.8 <1 k lag	81.9/77.8 <1 k lag	101.92/97.8 <1 k lag
	100 k.y.	15.52/5.5 5 k lag	30.2/20.2 2 k lag	45.12/35.1 ~1 k lag	65.08/55.1 ~1 k lag	85.04/75.0 ~1 k lag	105/95 ~1 k lag
	413 k.y.	Continuous Rise	49.18/7.9 37 k lag	62.9/21.5 22 k lag	82.1/37.45 15 k lag	101.6/60.31 11 k lag	121.4/80.2 9 k lag
0.2	41 k.y.	14.4/6.2 2 k lag	29.1/20.9 1 k lag	44.1/35.9 1 k lag	64/55.8 1 k lag	83.9/75.7 <1 k lag	103.9/95.7 <1 k lag
	100 k.y.	22.1/2.1 11 k lag	35/15.8 4 k lag	50.5/30.5 3 k lag	70.3/50.3 2 k lag	90.2/70.2 ~1 k lag	110.2/90.2 ~1 k lag
	413 k.y.	Continuous Rise	Continuous Rise	90.3/7.7 47 k lag	107.2/24.6 30 k lag	125.6/43.06 22 k lag	144.8/62.2 18 k lag
0.5	41 k.y.	22.5/2 5 k lag	36.1/15.6 2 k lag	50.7/30.2 1 k lag	70.6/50.1 1 k lag	90.5/70 1 k lag	110.3/89.8 1 k lag
	100 k.y.	Continuous Rise	55.27/5.27 11 k lag	68.2/18.2 7 k lag	87.1/37.1 4 k lag	106.6/56.6 3 k lag	126.2/76.2 3 k lag
	413 k.y.	Continuous Rise	Continuous Rise	Continuous Rise	Continuous Rise	212.2/5.7 64 k lag	225.7/19.25 47 k lag
Ice volume equivalent (km ³)		5.74 × 10 ⁶	1.42 × 10 ⁷	2.3 × 10 ⁷	3.4 × 10 ⁷	4.6 × 10 ⁷	5.7 × 10 ⁷

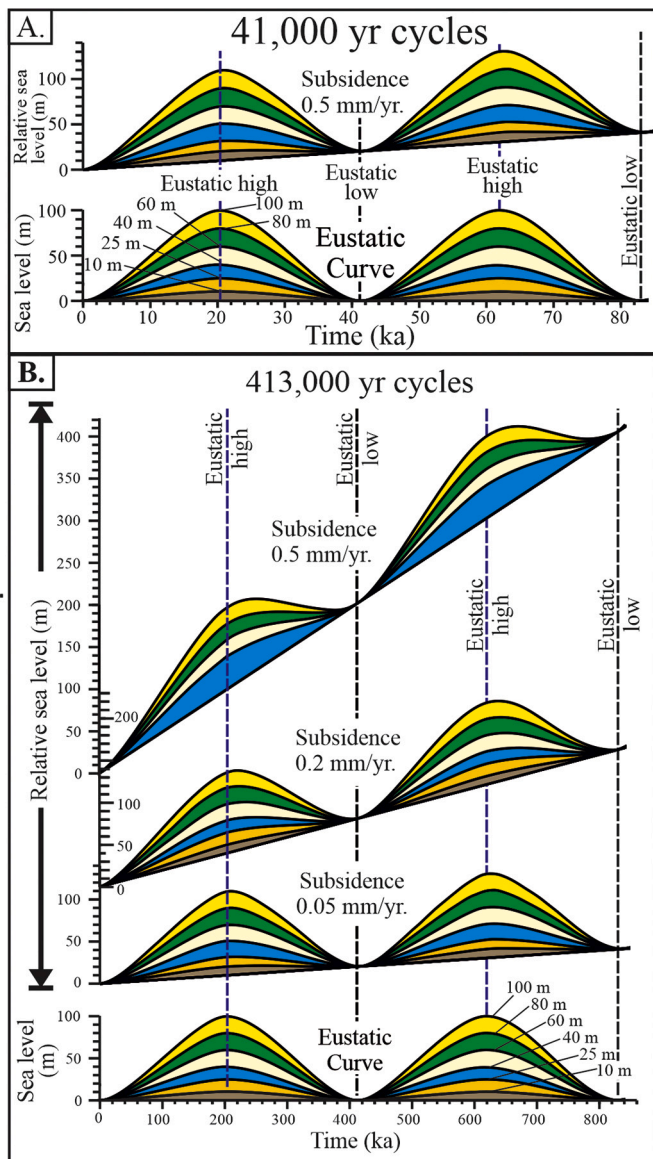


Fig. 19. Relative sea-level curves showing how orbitally driven changes in eustasy are modified by basin subsidence to produce accommodation as originally defined by [Jervey \(1988\)](#). Accommodation is identified by adding changes in eustatic sea level to basin subsidence to determine an accommodation curve or relative sea-level changes within a basin ([Jervey, 1988](#)). Space is created where the accommodation curve is positive thus promoting deposition and space is destroyed when the curve is negative, which favours erosion and the development of unconformities within basins. Accommodation for A) 41,000-year obliquity- and B) 413,000-year eccentricity-driven symmetrical (modelled as a sine curve), eustatic cycles the various colors represent different eustatic magnitudes. Relative sea-level changes along basin margins (no subsidence) are driven by eustatic cycles, whereas relative sea-level changes are accentuated due to higher subsidence rates in basin centers. Also note that as subsidence rates and cycle lengths change, the timing of relative sea-level highs and lows shift away from eustatic highs and lows.

ocean dynamics, and cause deformation to the solid Earth (elastic flexure, viscoelastic response, and viscous mantle flow; [Mitrovica and Milne, 2003](#); [Kendall et al., 2005](#); [Milne and Mitrovica, 2008](#); [Whitehouse, 2009, 2018](#); [Whitehouse and Bradley, 2013](#); [Mitrovica et al., 2015](#); [Horton et al., 2018](#); [Pohl and Austermann, 2018](#)). Together these changes result in rapid, glacially driven, highly variable changes in relative sea level across the globe in response to changes in a glacier's ice volume ([Farrell and Clark, 1976](#); [Whitehouse and Bradley, 2013](#)).

Glacial isostatic adjustment (GIA) is the composite response of the redistribution of water in the world's oceans during glacial cycles due to all these responses.

Studies of Pliocene to recent strata reveal that glacially driven changes in sea level record spatial and temporal changes that are site specific and that are not a direct record of eustasy, that is a uniform distribution of water across all of the world's oceans ([Fig. 20](#); e.g., [Raymo et al., 2011](#); [Whitehouse and Bradley, 2013](#); [Horton et al., 2018](#)). Instead, the elevation of sea level at a given point on Earth is related to the gravitational attraction at that site; elevation of ocean water by the gravitational attraction of an ice sheet's mass; redistribution of water and ice loads during a glacial cycle; redistribution of those loads across the geoid; which in turn influence, and is influenced by, oceanic and atmospheric circulation and Earth's rotation ([Fig. 20C](#); [Mitrovica and Milne, 2003](#); [Kendall et al., 2005](#); [Milne and Mitrovica, 2008](#); [Whitehouse, 2009, 2018](#); [Whitehouse and Bradley, 2013](#); [Mitrovica et al., 2015](#)). The geoid is the shape that the ocean's surface takes under the influence of gravity and rotation of the Earth. Because of unequal distribution of mass on Earth, the surface elevation of the world's oceans varies by almost 200 m. For example, the surface of the northern Indian Ocean resides at 106 m below the average surface elevation while the surface of the northern Atlantic Ocean is approximately 85 m higher than the average elevation ([Fig. 20C](#); cf. [Ihnen and Whitcomb, 1983](#); [Lemoine et al., 1997](#)). The geoid is a gravitationally equipotential surface that changes due to changes in the distribution of mass across the Earth's surface and the redistribution of Earth's internal mass due to elastic, viscoelastic, and viscous response of the crust/lithosphere and upper mantle due to loading ([Whitehouse, 2018](#)). Because ice sheets are ephemeral and may wax and wane at different times, growth and collapse of each individual ice sheet and their redistribution of mass on the surface of Earth will result in a unique, spatially variable, sea-level geometry, or "fingerprint," of sea-level change across the world's oceans ([Fig. 20D, E](#); [Hay et al., 2014](#); [Horton et al., 2018](#)). Today, melting of the Greenland ice sheet results in a fingerprint that is different than that of melting and re-distribution of waters from the west Antarctic or the East Antarctic ice sheets ([Fig. 20D](#); [Horton et al., 2018](#)), and each will result in different highs and lows at various sites across the globe ([Fig. 20D](#); [Horton et al., 2018](#)). Because the crust, lithosphere, and upper mantle would have responded to such redistribution of water and ice loads in the past as they do today, it is safe to assume that GIA operated in a similar fashion during Earth's previous ice ages.

The concept of GIA follows. As an ice sheet grows, it loads and depresses the lithosphere causing a redistribution of mass internally within the asthenosphere ([Fig. 20A](#)). The gravitational attraction of the growing mass also attracts sea water resulting in an increase in sea level (super elevation) proximal to the ice sheet ([Fig. 20A](#); [Whitehouse, 2009, 2018](#); [Gomez et al., 2010](#); [Creveling et al., 2018](#); [Dietrich et al., 2019](#)). This results in a drop in global water levels in far-field sites due to the gravitational water bulge and due to removal of water from the ocean to be locked up as ice mass/volume. Loading and unloading of the lithosphere by changing ice and water masses further results in redistribution of Earth's internal and external masses (elastic flexure, viscoelastic response, and viscous mantle flow; [Mitrovica and Milne, 2003](#); [Tamisiea and Mitrovica, 2011](#); [Whitehouse, 2009, 2018](#)). Flexural uplift occurs in ice-proximal areas during deglaciation due to unloading ([Fig. 20B](#); [Whitehouse, 2009, 2018](#)). This coupled with the loss of the gravitational attraction of sea water as the ice sheet shrinks results in a relative drop in sea level within a few thousand km of the ice front ([Fig. 20](#)). This redistribution of surface masses continues to drive a redistribution of Earth's internal masses. The water returned to the world's ocean during deglaciation is then distributed across the globe based on Earth's gravitational field and rotation at the time of deglaciation resulting in an uneven rise in sea level in far-field locations ([Fig. 20](#)). The dynamics of these processes associated with glacial isostatic adjustment indicate that the rise and fall of sea level is out of phase in ice-proximal and ice-distal locations ([Fig. 20](#); [Mitrovica et al., 2009](#); [Horton et al., 2018](#); [Pohl and](#)

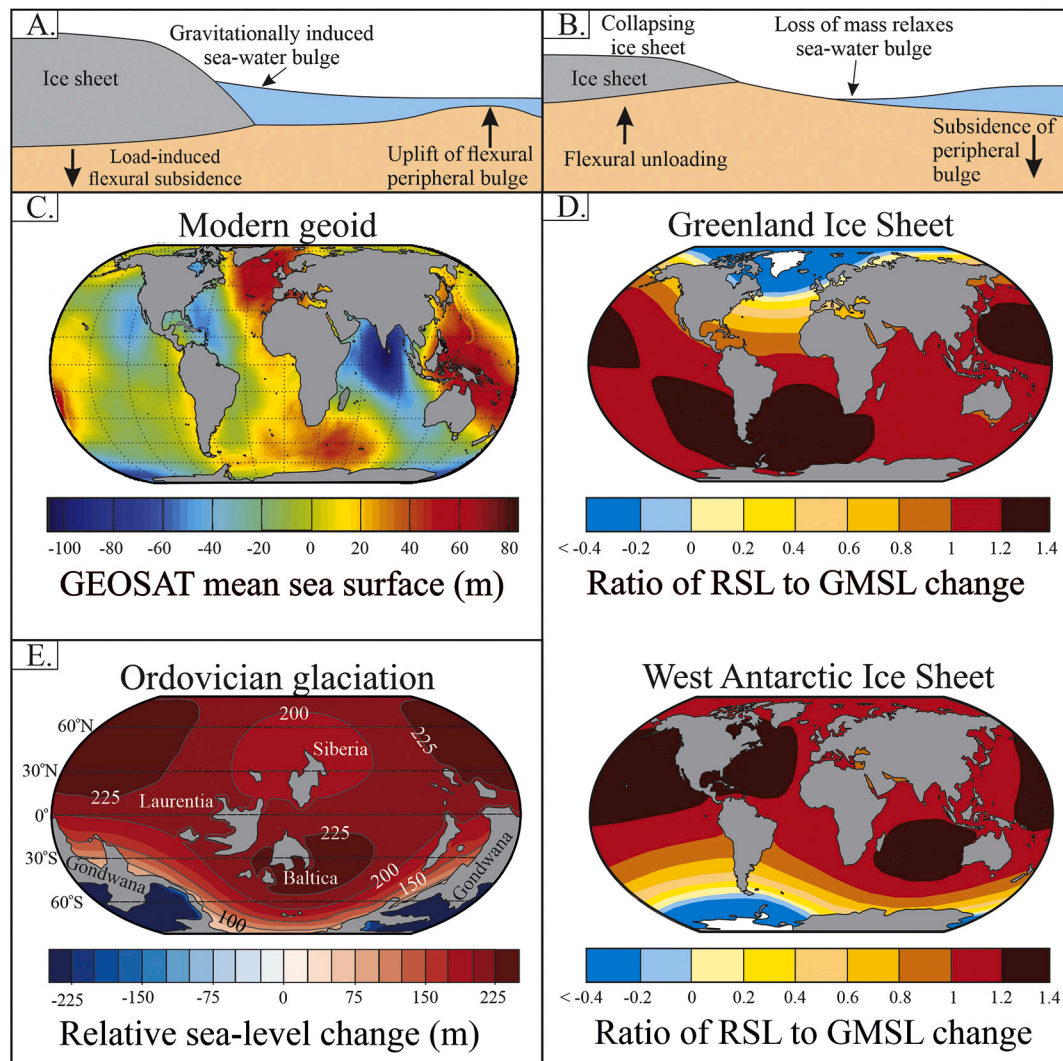


Fig. 20. Glacial isostatic adjustment. A) Development of an ice sheet with associated flexural subsidence and the uplift of a peripheral bulge. Note that gravitational attraction of the ice mass produces a sea-water bulge (super elevation) of the sea surface adjacent to the ice sheet. B) During deglaciation, flexural unloading results in lithospheric rebound beneath the ice sheet and subsidence of the peripheral bulge. Sea level falls proximal to the ice sheet due to the loss of the gravitational attraction of sea water by the disappearing ice mass. Diagrams A and B were constructed from data and diagrams presented by Whitehouse (2009, 2018). C) Mean sea-surface elevation defines the shape of the modern geoid (modified from Lemoine et al., 1997). D) Sea-level fingerprints for ice lost from the Greenland and the West Antarctic ice sheets. The ice sheet in question for each diagram is shown in white. Note that the change in sea level is unique to each ice sheet and that the change in sea level is out of phase in near-field (ice proximal) versus far-field (ice distal) sites (modified from Horton et al., 2018). The change in sea level following deglaciation is given as a ratio of site-specific, relative sea-level (RSL) change compared to global mean sea-level (GMSL; eustatic) change. E) GIA showing modelled relative sea-level change 5 k.y. following the Ordovician (Hirnantian) deglaciation (modified from Pohl and Austermann, 2018).

Austermann, 2018). Residual glacial isostatic adjustments from previous glacial cycles may also influence relative sea level (Raymo et al., 2011). Due to the loading and unloading of an ice sheet, its attraction of water masses, loading and unloading due to changing water masses in the oceans, and their influence on rotation (location of the mass), each individual ice sheet has/had a different sea-level fingerprint (Mitrovica et al., 2009; Hay et al., 2014; Horton et al., 2018).

To date, no GIA fingerprinting of the LPIA ice sheets has been conducted. However, GIA has been model for Pliocene to recent changes in sea level and are shown to explain unequal rises and falls in world sea-level with respect to the loss of ice in Greenland and Antarctica (Fig. 20D; Milne and Mitrovica, 2008; Mitrovica et al., 2009; Raymo et al., 2011; Hay et al., 2014; McHugh et al., 2017; Horton et al., 2018). They have also been used to model out-of-phase changes in sea level during the Ordovician (Hirnantian) glaciation (Fig. 20E; Creveling et al., 2018; Pohl and Austermann, 2018; Dietrich et al., 2019). Realizing that GIA responses happen during icehouse intervals, ice-volume and sea-

level relationships during the LPIA must have been extremely dynamic with the waxing and waning of its numerous ice sheets.

4.4. Reconciling ice volume, accommodation, eustasy and GIA responses

The various drivers of sea-level change discussed in section 4 do not operate independently but form a combined influence on relative sea-level change that results in a complex response to the waxing and waning of icehouse glaciers (cf. Jervey, 1988; Catuneanu, 2006; Milne and Mitrovica, 2008; Wendler et al., 2016; Wendler and Wendler, 2016; McHugh et al., 2017; Creveling et al., 2018; Whitehouse, 2018; Dietrich et al., 2019). With GIA, basin accommodation, ice volume, thermal expansion, and aquifer eustasy models in mind and what they mean to relative sea-level changes, it is apparent that not all million year or lesser relative sea-level changes are the result of glacioeustasy, or at least that ice volume fluctuations and relative sea-level changes are complex. For that matter, GIA studies suggest that glacioeustasy (a loss or addition of

a uniform thickness of water to/from the world's oceans due to changes in ice volume) may be an outdated concept as glacial and post-glacial sea-level changes are not equal across the world's oceans nor are they synchronous (high-and low-latitude response of sea-level rise are out of phase with one another) around the globe as the concept of eustasy demands. Although much is known about sea-level changes during icehouse intervals, it is apparent that the link between ice-volume fluctuations and sea-level change is a dynamic, highly complicated relationship.

5. Concluding remarks on multiproxy approach for investigating deep-time glaciations

The above discussions detail the strengths and weaknesses of many near- and far-field proxies used for identifying the occurrence, location, extent of glaciations, and/or the magnitude of glacioeustatic fluctuations through Earth's history, especially in reference to the LPIA. Although these proxies provide important information in defining the LPIA as a major climatic event, the occurrence of any individual proxy alone is not diagnostic, except for perhaps the occurrence of sculpted and striated bedrock features. Many of these proxies have multiple origins or are controlled by multiple factors. Therefore, a multiproxy approach in conjunction with detailed paleoenvironmental, paleoflow, and paleogeographic studies provides the most robust approach for investigating past glacial intervals (Eyles et al., 1983; Eyles, 1993; Fielding et al., 2008b, 2008c).

Although traditional views of glaciation persist, much progress has been made in defining the LPIA and other icehouse intervals. However, it is a work in progress as much remains to be investigated and new models require rigorous testing before we truly understand ancient glacial intervals. Rosa and Isbell (2021) and López-Gamundí et al. (2020) provide the most recent summaries of the LPIA, which represent progress reports of current views of the LPIA. Such views are snapshots and are subject to change as multiproxy-based studies and the development of new tools allow for the evolution of our understanding of this icehouse interval.

Studies of ice-volume fluctuations, accommodation, and GIA indicate a complex relationship between ice sheets and sea-level change. Such studies, especially GIA analysis, are challenging our concepts of glacioeustasy. Although much has been accomplished in developing an understanding of the LPIA, much remains to be studied. Further testing of current concepts and the applications of new methods to late Paleozoic studies, like GIA analysis, promise to dramatically challenge our views of icehouse intervals.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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