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GR focus review

# Glacial paradoxes during the late Paleozoic ice age: Evaluating the equilibrium line altitude as a control on glaciation

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ABSTRACT

The late Paleozoic ice age (LPIA) consists of multiple glaciations that waxed and waned across Gondwana during the Carboniferous and Permian. Three key intervals are evaluated using the concept of the equilibrium-line altitude (ELA) as a control on glaciation to provide insight into two intervals of paradoxical ice distribution during and following glaciation. The LPIA began in the mid-latitudes during the Viséan in western Argentina with the growth of glaciers in the Protoprecordillera. Glaciation was initiated by uplift of the range above the ELA. In the Bashkirian, deglaciation occurred there while glaciation was beginning at the same latitude in uplands associated with the Paraná Basin in Brazil. Analysis suggests that deglaciation of the Protoprecordillera occurred due to extensional collapse of the range below the ELA during a westward shift in the location of plate subduction. During Late Pennsylvanian–Early Permian peak glaciation for the LPIA, extensive glacimarine deposits indicate that glaciers reached sea level, which corresponds to a major lowering of the ELA due to global cooling. Finally, during the Early to early Late transition out of the LPIA, polar Gondwana was unglaciated. However, three glacial intervals occurred at mid- to high-latitudes in eastern Australia from the Sakmarian to the Capitanian/earliest Wuchiapingian. The magnitude of global cooling during these events is debatable as evidence indicates ice-free conditions and an elevated ELA at the South Pole in Antarctica. This suggests that severe global cooling was not the cause of the final three Australian glaciations, but rather that ELA-related conditions specific to eastern Australia drove these late-phase events. Possible causes for the Australian glaciations include: 1) anomalous cold conditions produced by coastal upwelling, 2) the presence of uplands allowing nucleation of glaciers, 3) fluctuations in $pCO_2$ levels, and 4) increased precipitation due to the location of the area in the subpolar low pressure belt.
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## 1. Introduction

The late Paleozoic Ice Age (LPIA) and the ensuing Early to Late Permian icehouse to greenhouse transition were two important intervals during the Phanerozoic as they had a major impact on Earth's physical, chemical, and biological systems (Heckel, 1994, 2008; Falcon-Lang, 2004; Joachimski et al., 2006; Clapham and James, 2008; Grossman et al., 2008; Isbell et al., 2008a; Falcon-Lang and DiMichele, 2010). The LPIA, which lasted for approximately 72 Myr from the Viséan (Mississippian) to the Capitanian/earliest Wuchiapingian (Middle-earliest Late Permian), occurred on a biologically complex Earth characterized by an extensive south polar landmass; low atmospheric partial pressure of $CO_2$ ($pCO_2$); and multiple, possibly bipolar, glacial events (Isbell et al., 2003; Fielding et al., 2008a). Because of these features, which also characterize Cenozoic glaciation, the LPIA serves as the most recent and complete analog for modern environmental change associated with global climate change (Gastaldo et al., 1996; Montañez and Soreghan, 2006; Isbell et al., 2008a).

The stratigraphic, geochemical, and tectonic records of the LPIA continue to improve in resolution (e.g., papers in Fielding et al., 2008b; Gulbranson et al., 2010; papers in López-Gamundí and Buatois, 2010) revealing glacial and non-glacial intervals that occurred across Gondwana. To date, these events are roughly correlated with changes in the paleolatitude of Gondwana and to fluctuations in greenhouse gasses (Caputo and Crowell, 1985; Royer et al., 2001; Montañez et al., 2007), whereas shorter-term Earth system fluctuations (e.g. eustasy) are attributed to Milankovitch forced glacial events (Heckel, 1994; Davydov et al., 2010). However, the drivers behind the waxing and waning of each LPIA glacial event, as well as the shorter duration glacial/interglacial cycles, are likely a complex interplay of local, regional and global conditions. These conditions are difficult to quantify under the current state of knowledge, and determining why glaciers were maintained in one area while adjacent areas were ice-free or undergoing de-glaciation, remain problematic. Unresolved problems, which are addressed in this paper, include: 1) identifying the causes for the initiation of glaciation in western Argentina during the Middle Mississippian to Early Pennsylvanian, 2) determining why glaciers disappeared in western Argentina during the Pennsylvanian while glaciers were forming in adjacent areas to the east, 3) identifying causes for the LPIA glacial maximum during the latest Pennsylvanian and Early Permian, and 4) determining the controls on glaciation in eastern Australia during the late Early to early Late Permian while areas located at higher paleolatitudes were ice free. Much work is still necessary to unravel the causes of climatic perturbations and their influence on LPIA glaciation.

The role that the Equilibrium Line Altitude (ELA) played in glaciation and the insights it provides on the formation, waxing and waning, and collapse of the Gondwana glaciers has not been previously investigated. Traditionally, Gondwana glaciation is modeled as a single, massive, ice sheet centered over the paleo-South Pole located in Antarctica and extending outward into the mid-latitudes (e.g., Scotese, 1997; Ziegler et al., 1997; Hyde et al., 1999). Ice is also hypothesized to have formed in high northern latitudes on the East Asian crustal block and in low latitude uplands in North America (Raymond and Metz, 2004; Soreghan et al., 2008, 2009). The size and configuration of the hypothetical Gondwanan ice sheet appears to have been determined by encircling all known glacial deposits on paleogeographic maps. However, such practices are highly inaccurate and would be misleading if conducted on modern deposits due to the occurrence of alpine glaciers and ice caps in low latitude uplands. A single, massive, Gondwanan ice sheet is untenable (cf., Horton and Poulsen, 2009). Such a model does not take into consideration the mass balance required to sustain such an ice sheet (Isbell et al., 2003), nor the fact that the ELA varies in elevation with respect to latitude. The ELA is the theoretical altitude on a glacier that separates areas of annual net accumulation from areas of annual net ablation (Benn and Evans, 2010). Because the ELA determines where glaciers can form, consideration of factors that influence both the local position and the global distribution of the ELA provide insight on the formation and demise of glaciers in time and space (Fujita, 2008). Although it is exceedingly difficult to accurately estimate paleoelevation, consideration of the ELA concept, comparison of synchronous glaciated vs. non-glaciated areas, and the comparison with modern ELA curves can provide a tool for understanding ancient glacial successions.

This paper summarizes the current state of knowledge on the LPIA and its main environmental drivers, and highlights problems that are unresolved concerning the distribution of glaciers and timing of glaciation across Gondwana. We then present a discussion on the ELA and how it controls glaciation in time and space, and speculate on the role that the ELA had as a driver of glacial and non-glacial intervals during the LPIA in an attempt to better understand Earth's transition from icehouse to greenhouse conditions.

## 2. Late Paleozoic Ice Age in Gondwana

Traditional models of LPIA glaciation depict a massive ice sheet(s) that waxed and waned continuously across Gondwana for up to 100 million years (Figs. 1A and 2; cf. Veevers and Powell, 1987; Frakes and Francis, 1988; Frakes et al., 1992; Ziegler et al., 1997; Hyde et al., 1999; Blakey, 2008; Buggisch et al., 2011). This concept is prevalent in geologic literature up to the present day. However, recent work has identified evidence of numerous small ice centers that waxed and waned diachronously across Gondwana through multiple glacial intervals of 1–8 million years in duration alternating with non-glacial periods of approximately equal duration (Figs. 1B and 2; Crowell and Frakes, 1970; Caputo and Crowell, 1985; Dickins, 1997; López-Gamundí, 1997; Isbell et al., 2003; Fielding et al., 2008a, 2008c, 2008d; Gulbranson et al., 2010). Although glaciation occurred in northern South America and northern Africa during the Late Devonian and early Mississippian (Caputo and Crowell, 1985; López-Gamundí, 1997; Crowell, 1999; Isbell et al., 2003; Caputo et al., 2008), most authors consider the LPIA to have begun in western South America during the Viséan (Caputo et al., 2008; Pérez Loinaze et al., 2010) and to have concluded in eastern Australia during the Middle to earliest Late Permian (Capitanian/earliest Wuchiapingian; Fielding et al., 2008a, 2008c, 2008d).

Continental drift of Gondwana across the South Pole (Fig. 3) has long been recognized as a major control for the diachronous shifting of glacial centers across the supercontinent during the LPIA (DuToit, 1921; Wegener, 1929; Crowell, 1978; Caputo and Crowell, 1985).

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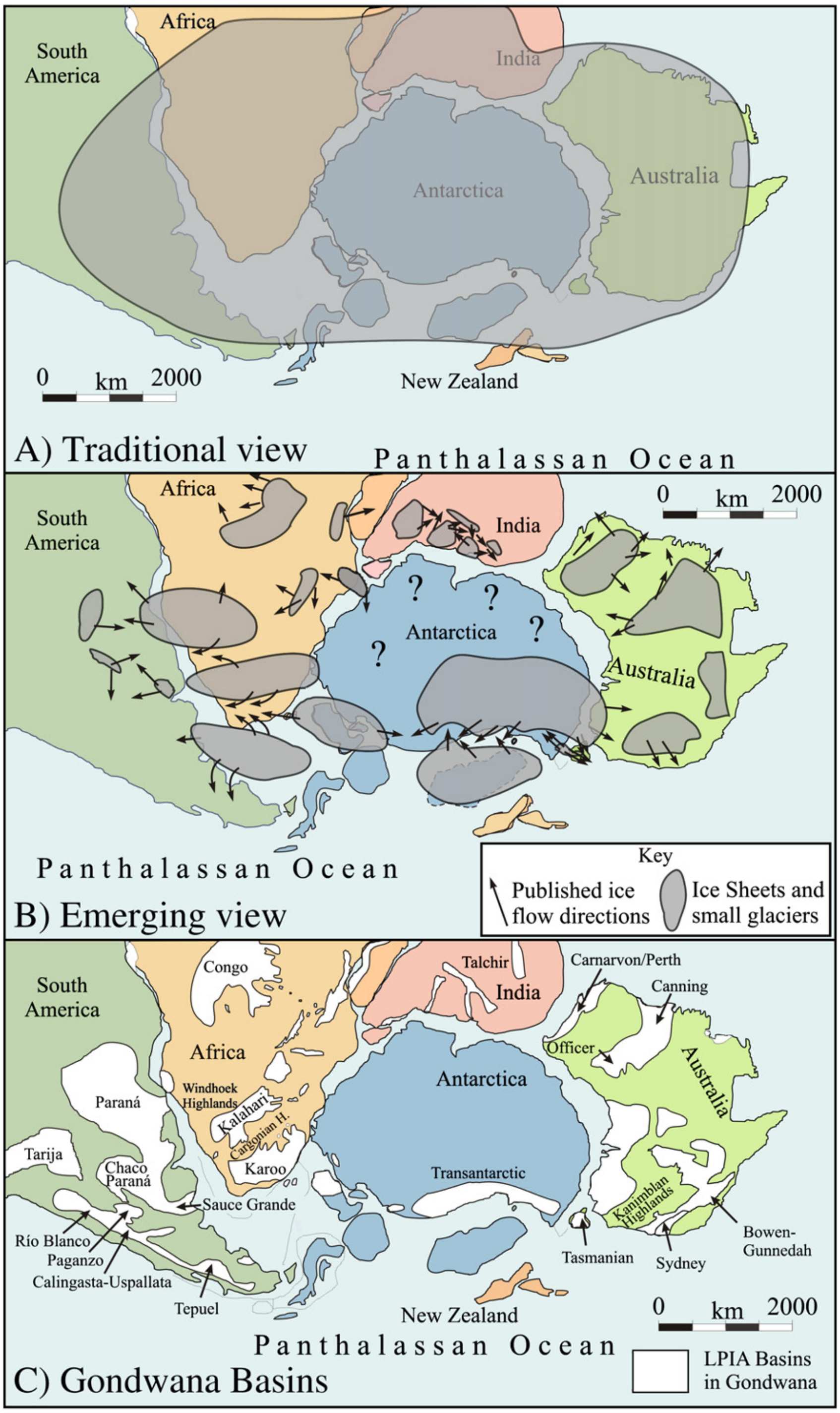


**Fig. 1.** Traditional and Recent reconstructions of maximum glaciation during the late Paleozoic Ice Age. A) Traditional reconstruction showing a massive ice sheet covering much of southern Gondwana (modified from Scotese, 1997, and Scotese and Barrett, 1990). B) Reconstruction of Gondwana during maximum glaciation during the Gzhelian to early Sakmarian (Pennsylvanian–Early Permian) based on recent data and ice flow directions. Ice flow directions are from Frakes et al. (1975), Hand (1993), Veevers and Tewari (1995), López-Gamundí (1997), Visser (1997a, 1997b), Visser et al. (1997), Fielding et al. (2008a), Isbell et al. (2008c), Mory et al. (2008), Rocha-Campos et al. (2008) and Isbell (2010). C) Location map for selected Gondwana basins and highlands for the Carboniferous and Permian mentioned in the text.

Glaciation began in the Precordilleran region of Argentina and in the Parnaíba and Amazon Basins of Brazil during the Middle Mississippian (Viséan) with possible ice centers also occurring in South Africa and in Patagonia (Figs. 1 and 2; López-Gamundí, 1997; Isbell et al., 2003; Limarino et al., 2006; Caputo et al., 2008; Isbell et al., 2008b; Gulbranson et al., 2010; Taboada, 2010). During the Late Mississippian–Early Pennsylvanian (Serpukhovian–Bashkirian), glaciers expanded in western and southern South America (López-Gamundí, 1997; Isbell et al., 2003; Limarino and Spalletti, 2006; Limarino et al., 2006; Henry et al., 2008; Holz et al., 2008; Rocha-Campos et al., 2008) and first appeared in eastern Australia (Fielding et al., 2008a, 2008c, 2008d). During the Late Pennsylvanian (Gzhelian) to Early Permian (Sakmarian), widespread glaciation occurred, reaching its maximum extent across Gondwana (Laskar and Mitra, 1976; Veevers and Tewari, 1995; López-Gamundí, 1997; Visser, 1997a; Isbell et al., 2003, 2008b, 2008c; Fielding et al., 2008a, 2008c, 2008d; Martin et al., 2008; Mory et al., 2008; Rocha-Campos et al., 2008; Stollhofen et al., 2008; Melvin et al., 2010). This glaciation

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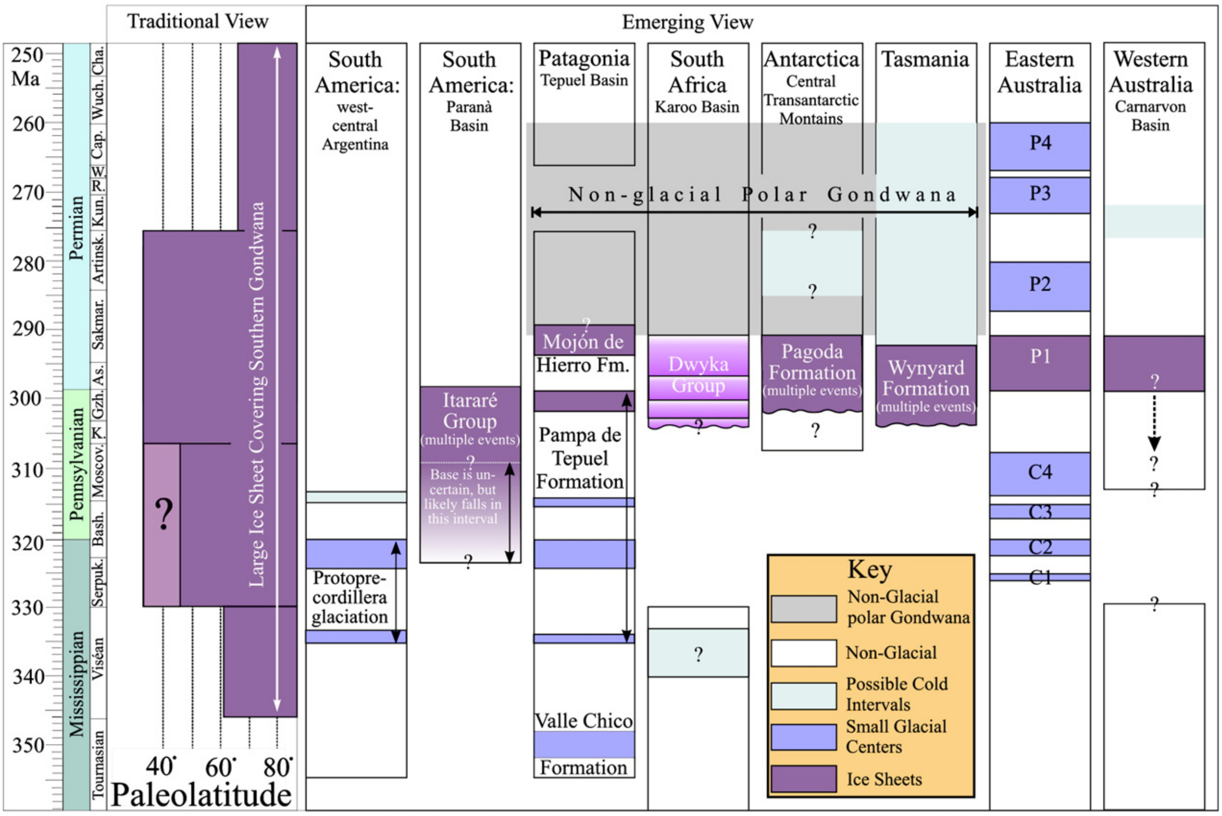


**Fig. 2.** Glacial intervals of the LPIA discussed in the text showing the traditional view of glaciation (modified from Frakes and Francis 1988; and Crowley and Baum, 1991, 1992) and the emerging view based on data from Truswell (1978), Collinson et al. (1994), Isbell et al. (2003, 2008b, 2008c), Fielding et al. (2008a, 2008c), Mory et al. (2008), Rocha-Campos et al. (2008), Stollhofen et al. (2008), Taboada (2010), Gulbranson et al. (2010) and Henry et al. (in press). The Carboniferous time scale is from Davydov et al. (2010); the Permian time scale is from Gradstein et al. (2004).

was centered over Antarctica, with ice centers also occurring in eastern South America, Patagonia, Africa, the Arabian Peninsula, India, Australia, and some southern Asian crustal blocks (Figs. 1 and 2; Isbell et al., 2003, 2011a; Fielding et al., 2008a; Wopfner and Jin, 2009a, 2009b; Taboada, 2010). Finally, from the late Sakmarian to the Capitanian/earliest Wuchiapingian small ice centers and alpine glaciation were located only in Australia (Fig. 2; Fielding et al., 2008a, 2008c, 2008d).

The apparent polar wander path (APWP) proposed by Powell and Li (1994) and Li and Powell (2001) is commonly used in interpreting paleolatitude positions during the LPIA (Fig. 3; Isbell et al., 2003; Fielding et al., 2008a; 2008b). Powell and Li (1994) referenced paleomagnetic poles from eastern and central Australia for their model, in which the pole occurred in southern Argentina in the middle Devonian and shifted to central Africa in the latest Devonian to Mississippian, then migrated into Antarctica in the Pennsylvanian (Fig. 3; Powell and Li, 1994). The pole then shifted to the present-day central Transantarctic Mountains and Marie Byrd Land in Antarctica by the Middle Permian, and then moved toward and into eastern Australia by the end of the Permian. This APWP is best constrained for the interval from 320 to 280 Ma; however, from 280 to 250 Ma, including Australia's apparent drift onto the pole, the APWP is "less well constrained" (Li and Powell, 2001), thus the APWP from 280 to 250 Ma is equivocal.

More recent APWPs for Gondwana include reconstructions based on paleomagnetic data from Baltica and Laurentia (Van der Voo, 1993; Dalziel, 1997; Torsvik and Cocks, 2004; Lawver et al., 2008; Torsvik et al., 2008), and combined data from various Gondwanan crustal blocks (Fig. 3; Torsvik et al., 2008). Although the Laurentian and Gondwanan APWPs have similar shapes, there is considerable offset between the two curves (Fig. 3; Domeier et al., 2011), and between either curve and the APWP of Powell and Li (1994). Domeier et al. (2011) showed that this offset is likely the result of the incorporation of low-quality or systemically biased data "many of which are considered unreliable by modern standards." Using a rigorous filter on the South American data set, they refined the Permian to Early Triassic APWP for Gondwana (Fig. 3). This reconstruction does not have the excursion into Australia depicted on the Powell and Li (1994) curve. This paper uses APWP of Domeier et al. (2011) due to its more highly constrained data set.

Although glaciation began in western Gondwana (South America) and ended in eastern Gondwana (Australia), which supports the concept of drift of Gondwana over the South Pole during the LPIA (Fig. 2), the APWP only accounts for general trends in the distribution of glaciation through time. However, drift across the South Pole does not explain the vacillation between glacial and non-glacial conditions within the ice age, the timing and variations in the extent of the various glacial events across Gondwana or within a particular region, or the continued Middle to earliest Late-Permian glaciations in eastern Australia following the Early Permian LPIA acme (Figs. 1 and 2). For example, drift does not explain the disappearance of ice centers from the Protoprecordillera region in western Argentina during the Early Pennsylvanian, which occurred just prior to and during growth of glaciers in the highlands surrounding the Paraná Basin farther to the east in South America. In addition, during the Middle–earliest Late Permian glaciations in eastern Australia, Antarctica remained unglaciated despite being located closer to the pole (cf. Collinson et al., 1994; Isbell et al., 2008c; Henry et al., in revision). Furthermore, glaciation in eastern Australia ended in the late Capitanian/earliest Wuchiapingian (~260 Ma), at a time when it may have made its closest approach to the Permian South Pole (Figs. 2 and 3; cf. Fielding et al., 2008a, 2008c, 2008d; Domeier et al., 2011).

Although the record is not well constrained, glaciation may have also occurred in Northeastern Asia in the northern hemisphere during the Pennsylvanian and Permian (Ustritsky and Yavshits, 1971; Frakes

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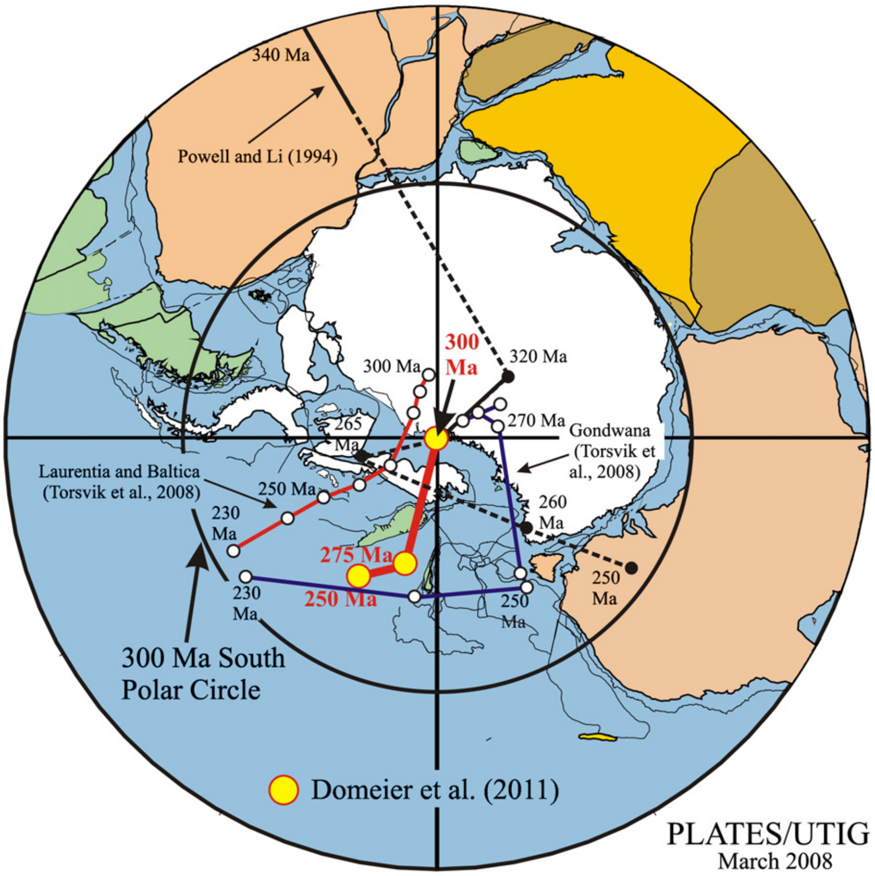


**Fig. 3.** Polar view of Gondwana at 300 Ma and the apparent polar wander path of Domeier et al. (2011), which is used in this paper. The APWP of Powell and Li (1994) and Li and Powell (2001) is plotted as a solid line where data is more reliable and as a dashed line where the data is less well constrained. The map also shows the APWPs of Torsvik et al. (2008) for Laurasia (Larentia and Baltica) and Gondwana. The reconstruction of Gondwana is from Dalziel (1997) and Lawver et al. (2008). To construct the curve for Domeier et al. (2011), their poles were attached to South America then rotated into the Gondwana configuration presented by Lawver et al. (2008) with the view centered over the 300 Ma pole of Domeier et al. (2011). The reconstruction and rotations are courtesy of the Institute of Geophysics' PLATES Project at the University of Texas-Austin.

et al., 1975; Pavlov, 1979; Epshteyn, 1981a, 1981b; Chumakov, 1985, 1994). However, the diamictites and lonestone-bearing beds of presumed glacial origin may have been deposited by shore or river ice (Frakes et al., 1975; Crowell, 1999), or as Biakov et al. (2010) suggest, the result of non-glacial marine slumping and debris flows associated with the development of the Okhotsk–Taigonos Volcanic Arc. During the late Paleozoic, the diamictites and lonestones paradoxically disappeared as Siberia drifted into and across the northern polar circle (Blakey, 2008; Lawver et al., 2008).

Correlation of $pCO_2$ and carbon and oxygen isotope fluctuations within glacial and non-glacial intervals in the Carboniferous and Permian suggests that greenhouse gasses were a major control on climate fluctuations during the LPIA (Fig. 4; Royer, 2006; Montañez et al., 2007; Frank et al., 2008; Buggisch et al., 2011). However, the relationships between some isotopic oscillations and glacial/non-glacial events are not well understood, as there are assumptions and incongruities in the geochemical record for the late Paleozoic that need to be explained. One issue is that geochemical data appear incongruous with certain glacial and non-glacial intervals. For example, during Serpukhovian glaciation, $\delta^{13}C$ and $\delta^{18}O$ values of paleo-tropical marine carbonates actually decrease, which typically occurs during warming (Frank et al., 2008). Later in the Bashkirian and the subsequent C4 glaciation in Australia that extends into the Moscovian (cf. Fielding et al., 2008c, 2008d), there is a dramatic positive excursion in $\delta^{13}C$ and $\delta^{18}O$ values, but the expected increase in delta values lags behind the deposition of glacial sediments in South America and Australia by several million years. Frank et al. (2008) put forward two different explanations for the discrepancies between the isotopic and stratigraphic records: 1) The isotopic data set is too low resolution at present to accurately represent the entirety of Late Mississippian and Early Pennsylvanian glaciation; or 2) because this glaciation consisted of multiple ice centers waxing and waning at different times in different regions, the paleo-tropics were not influenced as consistently, creating variable isotopic ratios. Nevertheless, the isotopic data and sedimentologic data are recording climatic signals, the timing of which, and global significance of, require further analysis.

Another issue is that the geochemical record remains undeveloped for Polar Regions of Gondwana. Data for $pCO_2$ levels and $\delta^{13}C$ and $\delta^{18}O$ values derived from paleo-tropical carbonates from Laurentia and Eurasia are typically recovered from whole rock analyses or brachiopod calcite (cf. Hayes et al., 1999; Veizer et al. 1999; Frank et al., 2008). The derived values are assumed to represent the chemical signature of the global ocean in areas where the isotopes are considered to have been well mixed. However, the isotopic values from restricted epicontinental seaways vary from samples taken from localities where ocean waters were better mixed (Grossman et al., 2011). The far-field data representing isotopic fluctuations in seawater are assumed to record the same climatic fluctuations that forced the ice volume changes recorded in near-field glacigenic and post-glacial deposits. This assumption is rarely checked against isotopic signatures of near-field regions (with the exception of Scheffler et al., 2003), so a level of uncertainty remains about the accuracy of using far-field chemical proxies to draw conclusions about glacial intervals experienced in the Polar Regions. More work needs to be conducted on the near-field record to better correlate these two data sets. Proxies for atmospheric $CO_2$ concentrations derived from soil-formed minerals (e.g.,

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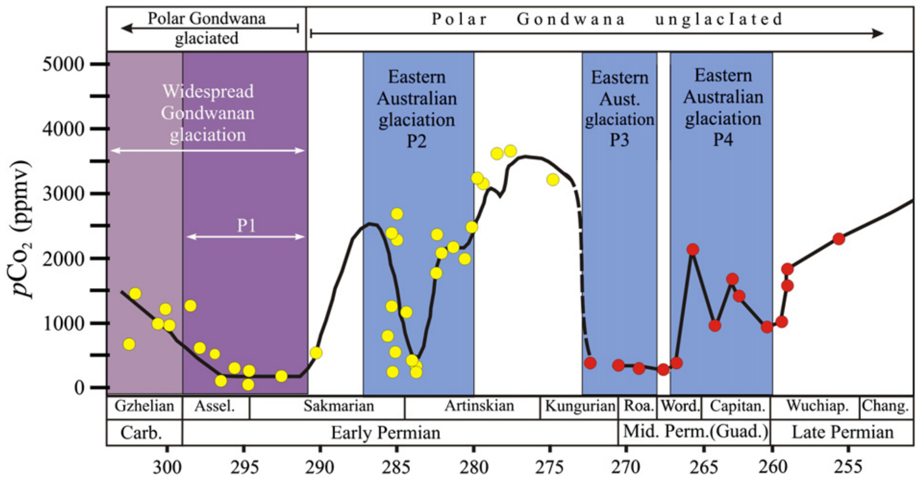


**Fig. 4.** $pCO_2$ fluctuations and glacial and non-glacial intervals in the Permian. Yellow dots are from Montañez et al. (2007), red dots represent data compiled by Royer (2006), and the dashed line links the two data sets.

calcite and goethite) circumvent these seawater mixing problems as they reflect $pCO_2$ values estimated from physical and chemical constraints in soils (cf. Ekart et al., 1999). The pedogenic calcite paleobarometer is used to estimate much of the $pCO_2$ record for the Phanerozoic (Breecker et al., 2010).

In regards to the LPIA, the $pCO_2$ record derived from paleosols has only been developed for the latest Pennsylvanian (Gzhelian) through the Permian (Fig. 4; Montañez et al., 2007), and does not exist for older LPIA events. When this gap in the dataset is filled in, it will be important to note if decreases in atmospheric $pCO_2$ occurred during the Mississippian and Pennsylvanian glacial events. The $pCO_2$ record shows strong correlation with the latest Pennsylvanian and Permian glacial events (Royer, 2006; Montañez et al., 2007; Fielding et al., 2008c). However, it should be noted that only the Late Pennsylvanian (Gzhelian) to Early Permian (early Sakmarian) glacial event was widespread. Later Permian lows in the $pCO_2$ record correspond only to regional glaciation in Australia and not to global glacial events as much of polar Gondwana was unglaciated by that time. Therefore, caution must be taken when applying the $pCO_2$ record to terminal events of the LPIA.

## 3. The equilibrium line altitude's control on glaciation

Topography is an obvious control on glaciation, as initiation and growth of glaciers take place at altitudes where net accumulation of snow and its conversion to ice occur. Nevertheless, paleotopography and its relationship with glacial mass balance is often overlooked or deemphasized as a glacial driver during the LPIA. Instead, most studies have focused on the paleolatitude of Gondwana and greenhouse gasses as the primary forcing factors. Regardless, late Paleozoic mountain ranges undoubtedly influenced atmospheric circulation, precipitation, and snow accumulation, and therefore were important in initiating glacial conditions during the LPIA (cf. Ziegler et al., 1997). Powell and Veevers (1987), Eyles (1993), and Crowell (1999) suggested that tectonic uplift along the Panthalassan margin of Gondwana, beginning with the collision of the Chilenia terrane against western South America in the late Tournasian–early Viséan (early Mississippian), initiated the LPIA. Initiation of glaciation in South America, and later initiation in Australia occurred while these regions were located at mid-latitudes (30°–60°; Powell and Veevers, 1987). These regions appear to have served as ice centers before regions closer to the South Pole were glaciated (Figs. 2 and 3). Therefore, the onset of glaciation in Gondwana appears to have coincided with orogenesis rather than with high polar latitude: in South America, convergence and accretion along the Panthalassan margin, and in Australia, uplift along the Tasman Fold Belt (Powell and Veevers 1987; Eyles, 1993). At the end of the LPIA, glaciation continued in eastern Australia for approximately 30 Myr after most of the Permian Polar Regions became ice free (Fig. 2; cf. Fielding et al., 2008a, 2008c, 2008d; Isbell et al., 2008b, 2008c; Isbell, 2010; Stollhofen et al., 2008; Henry et al., in press). At that time, most of eastern Australia was located outside of the polar circle (Fig. 3), suggesting that local factors including paleotopography may have played a role in continued glaciation. Therefore, the availability of topographic surfaces at high elevations may have been an important mechanism that facilitated nucleation and growth of LPIA glaciers (cf. Powell and Veevers, 1987; Eyles, 1993; Isbell et al., 2011b).

Isbell et al. (2011b) and Henry et al. (in press) suggested that many of the local variations and problems with the distribution and timing of Gondwana glaciations can be explained by considering the relationship between the equilibrium line altitude (ELA) and the paleo-land surface for a given region through time and space. The ELA is the elevation in a region above which ice accumulation can occur (Fig. 5A). The ELA is equivalent to the snowline, and on a glacier the ELA is the boundary separating an upper area of accumulation from a lower area of ablation (Fig. 5A; Benn and Evans, 2010). At the ELA, the rate of accumulation is equal to the rate of ablation. Therefore, glaciers cannot form if the ELA resides above the elevation of the land in a particular region. In this sense, glaciation results from lowering of the ELA relative to the elevation of the land surface, whereas, glacial retreat or cessation occurs due to a rise in the ELA relative to the elevation of the surrounding landscape. The ELA for a given location is controlled by glacial mass balance. Therefore, energy input (local to global heat flux), precipitation, topography, and latitude influence the position of the ELA (Fujita, 2008). Latitudinally, the ELA is typically close to or at sea level near the poles and rises towards the equator (Fig. 5B, C). The present ELA is just above sea level at the North Pole, rises to between 4000 and 6000 m in the tropics, and then falls in elevation southward where it intersects seal level at ~65° S latitude (Broecker and Denton, 1990). In this sense, glaciers in tropical regions only occur at high altitudes (e.g., above 4000 m in Irian Jaya, Indonesia; Allison and Peterson, 1989), but in high-latitude regions like Antarctica and Greenland, glaciers form at or near sea level (Miller et al., 1975). Variations in precipitation and/or local heat input/output can cause deviations in the overall latitudinal trend and can depress or elevate the ELA locally from its overall latitudinal tendency (Broecker and Denton, 1990). Moreover, high snowfall amounts will lower an ELA, but in locations with low snowfall,

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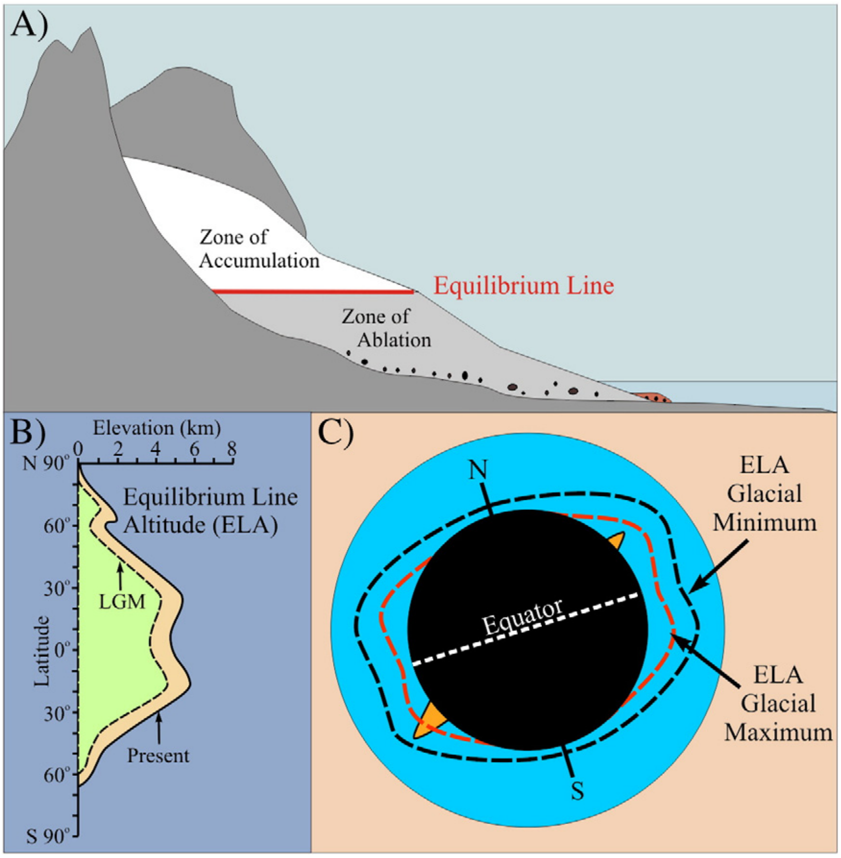


**Fig. 5.** Equilibrium line altitude (ELA). A) Diagram of the equilibrium line of a glacier, which separates the zone of accumulation above from the zone of ablation below. B) Present and Last Glacial Maxima (LGM) distribution of the ELA with latitude (modified from Broecker and Denton, 1990; Benn and Evans, 2010). C) Global change in the ELA during Glacial Minimum and Glacial Maximum.

such as arid regions, glaciers will have higher ELAs. Global warming during climate cycles will also result in a latitude-dependent rise in the position of the ELA, whereas global cooling will lower the ELA globally. The elevation of the ELA is particularly sensitive to summer precipitation and temperature, which are major controls on annual ablation rates. Further, tectonic movement in a region can influence the formation or destruction of glaciers: uplift of a region above the ELA makes it possible for glaciers to nucleate, while subsidence of a region below the ELA causes preexisting glaciers to ablate. The time scales required for significant uplift or subsidence may extend over multiple climate cycles and may produce a net effect over time rather than an instantaneous glacial/deglaciation event. Thus, it is apparent that glaciers can form during both icehouse and greenhouse conditions as long as the land surface resides above the local ELA. Therefore, the latitudinal position of the ELA is the manifestation of all controls on the formation and later demise of glaciers in time and space. Because the ELA changes due to climatic fluctuations and tectonic events, the ELA was likely a controlling factor on glacier distribution throughout the many phases of the LPIA (Fig. 5).

Precise estimates of paleotopography and the exact elevation of the ELA may be impossible to reconstruct. However, consideration of the relative position of the ELA is still a powerful tool to identify the controls on glaciation/deglaciation events. Moreover, if glaciation is occurring, then it is a requirement that the local ELA resides below the land surface on which the glacier nucleated. Likewise, the local ELA resides above the land surface during deglaciation. Use of the ELA concept, comparison of synchronous glaciated vs. non-glaciated areas, and comparison with predicted modern ELA curves can provide insight on what factors governed initiation, expansion, contraction, and collapse of glacial events during the LPIA. The interplay between the ELA and paleotopography for important intervals during the LPIA are explored in the following sections.

## 4. Carboniferous glaciation of the Protoprecordillera, western Argentina

### 4.1. Background

The Protoprecordillera (Fig. 6) was an ancient mountain belt that formed in what is now west-central Argentina due to the subduction and collision of the disputed Chilenia terrane with western South America (Cuyania terrane) during the Late Devonian and Mississippian Chañic orogeny (Fig. 7; Ramos, 1988; López-Gamundí et al., 1994; Limarino et al., 2002, 2006). Initially, the Protoprecordillera formed as an obducted accretionary prism, and then developed into a fold–thrust belt as the Chilenia terrain accreted to western Gondwana (Fig. 7). During and following collision in the Mississippian, the Protoprecordillera was a substantial mountain belt that separated the Río Blanco and Calingasta–Uspallata Basins to the west from the Proto-Paganzo Basin to the east (Figs. 6 and 7). During the Pennsylvanian and Permian, the active margin along this portion of Gondwana shifted to the west with initiation of subduction beneath Chilenia. Such subduction, with associated back-arc extension, is hypothesized to have resulted in the transformation of the Proto-Paganzo basin into a more extensive extensional back arc basin, the Paganzo Basin, which consisted of sub-basins filled with Pennsylvanian and Permian strata separated by uplifted granitic basement blocks. Collapse of the Protoprecordillera was a byproduct of this back-arc extension (Limarino et al., 2002, 2006).

Glaciation in the Protoprecordillera began during the Middle Mississippian (Viséan; Limarino et al., 2006; Gulbranson et al., 2010;

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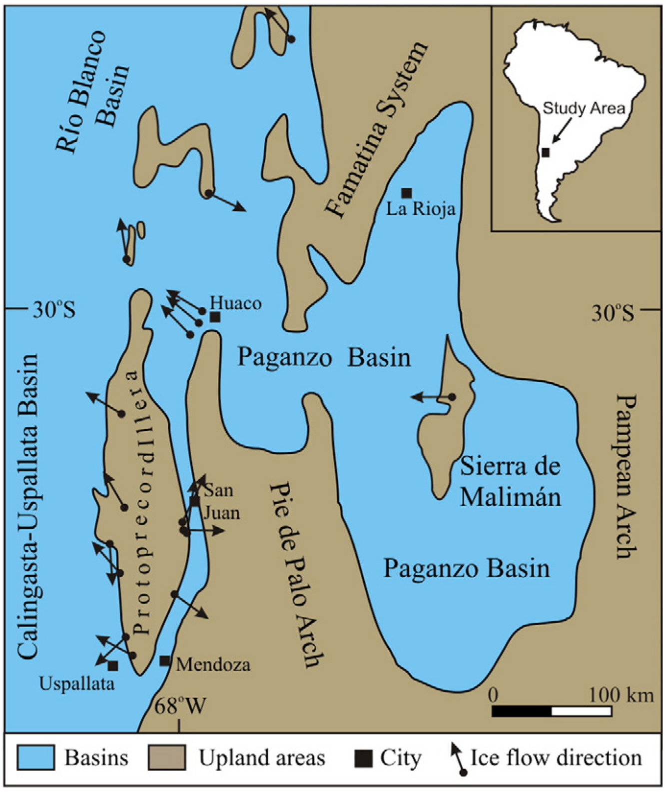


**Fig. 6.** The Protoprecordillera in west central Argentina housed glaciers in the Serpukhovian–Bashkirian (late Mississippian–early Pennsylvanian) and acted as a divide between the Río Blanco and Calingasta–Uspallata Basins to the west, and the Paganzo Basin to the east, until the collapse of the fold–thrust belt in the Pennsylvanian. Modified from Henry et al. (2010)

Pérez Loinaze et al., 2010), with a second, more extensive glacial event occurring during the Late Mississippian (Serpukhovian) to Early Pennsylvanian (Fig. 2; early Bashkirian; Limarino et al., 2002, 2006; Henry et al., 2008, 2010; Gulbranson et al., 2010).

Glacimarine deposition during the Serpukhovian to early Bashkirian (Latest Mississippian to Earliest Pennsylvanian) occurred in deeply incised paleo-valleys in the Protoprecordillera, and in the adjacent Calingasta–Uspallata and Río Blanco basins (Fig. 8; López-Gamundí, 1997; Limarino et al., 2002, 2006; Kneller et al., 2004; Dykstra et al., 2006; Henry et al., 2008, 2010; Gulbranson et al., 2010; Césari et al., 2011). The sedimentary record includes: 1) glacimarine and glacially influenced marine deposits in the form of striated boulder pavements resulting from grounded ice advance, 2) thick massive and stratified diamictites deposited from a combination of settling from meltwater plumes and iceberg rafting, 3) conglomerates and wedge-shaped sandstone bodies deposited as grounding-line fans, 4) thin bedded diamictites, massive sandstone beds, and thin-bedded graded sandstone bodies deposited from debris flows and turbidity currents, and 5) lonestone-bearing mudrocks deposited in distal marine environments (Fig. 8; López-Gamundí, 1987; Henry et al., 2008, 2010; Gulbranson et al., 2010). Such deposits are well recognized in the Calingasta–Uspallata Basin in the Agua de Jagüel, Tramojo, El Paso, and Hoyada Verde Formations. Glacial deposits in the Paganzo Basin are limited primarily to paleo-valleys cut into underlying basement rocks (Fig. 8C). Glacial strata in the Paganzo Basin consist of grooved and striated surfaces, and thin glacial and glacilacustrine rocks (Buatois and Mángano, 1995, López-Gamundí and Martínez, 2000; Pazos, 2002; Buatois et al., 2006; Dykstra et al., 2006). The deposits in the paleofjords and in the Calingasta–Uspallata, Río Blanco, and Paganzo Basins suggest that temperate tidewater glaciers reached the sea (or inland lake in the Paganzo Basin) and drained radially away from the Protoprecordillera through an extensive system of fjords (López-Gamundí and Martínez, 2000; Kneller et al., 2004; Dykstra et al., 2006; Henry et al., 2008, 2010). In the upper portions of the glacial sequence and in the base of the post-glacial sequence at many sites, including sites containing the Agua de Jagüel, Río del Peñón, and Jejenes Formations, abundant mass movement deposits occur (Fig. 8E). These deposits include meters to tens of meters thick and meters to hundreds of meters long and wide slide and slump blocks (Kneller et al., 2004; Dykstra et al., 2006; Henry et al., 2008, 2010).

Sandstone detrital modes and clast composition of conglomerates in the basins surrounding the Protoprecordillera indicate that the mountain belt served as a barrier to sediment dispersal, and that the basins were segregated with respect to composition during early glacial time

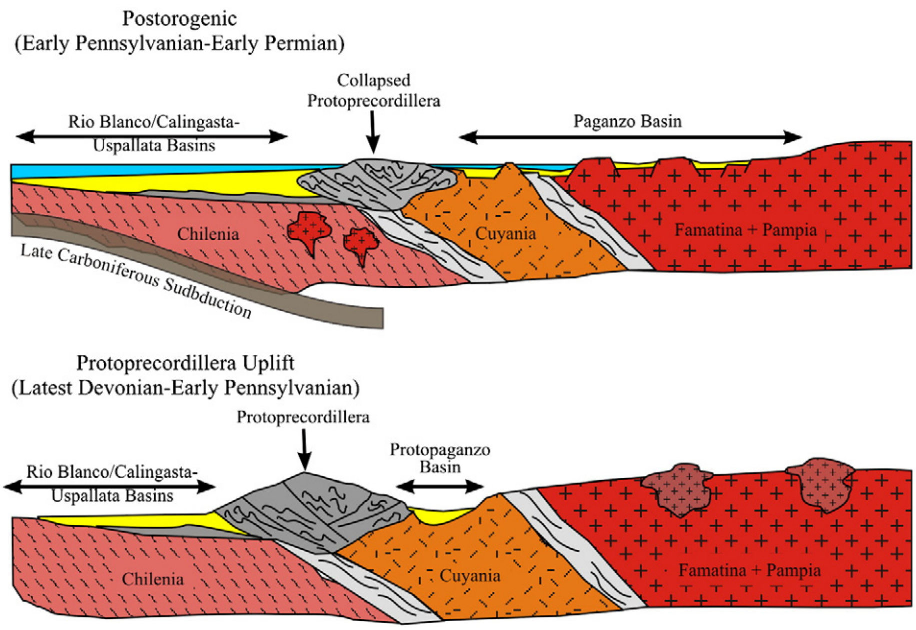


**Fig. 7.** Cross-section showing the basins in western Argentina during formation of the Protoprecordillera and the later collapse of the mountain range in the Late Pennsylvanian–Early Permian. Modified from Limarino et al. (2006), Henry et al. (2010) and Tedesco et al. (2010).

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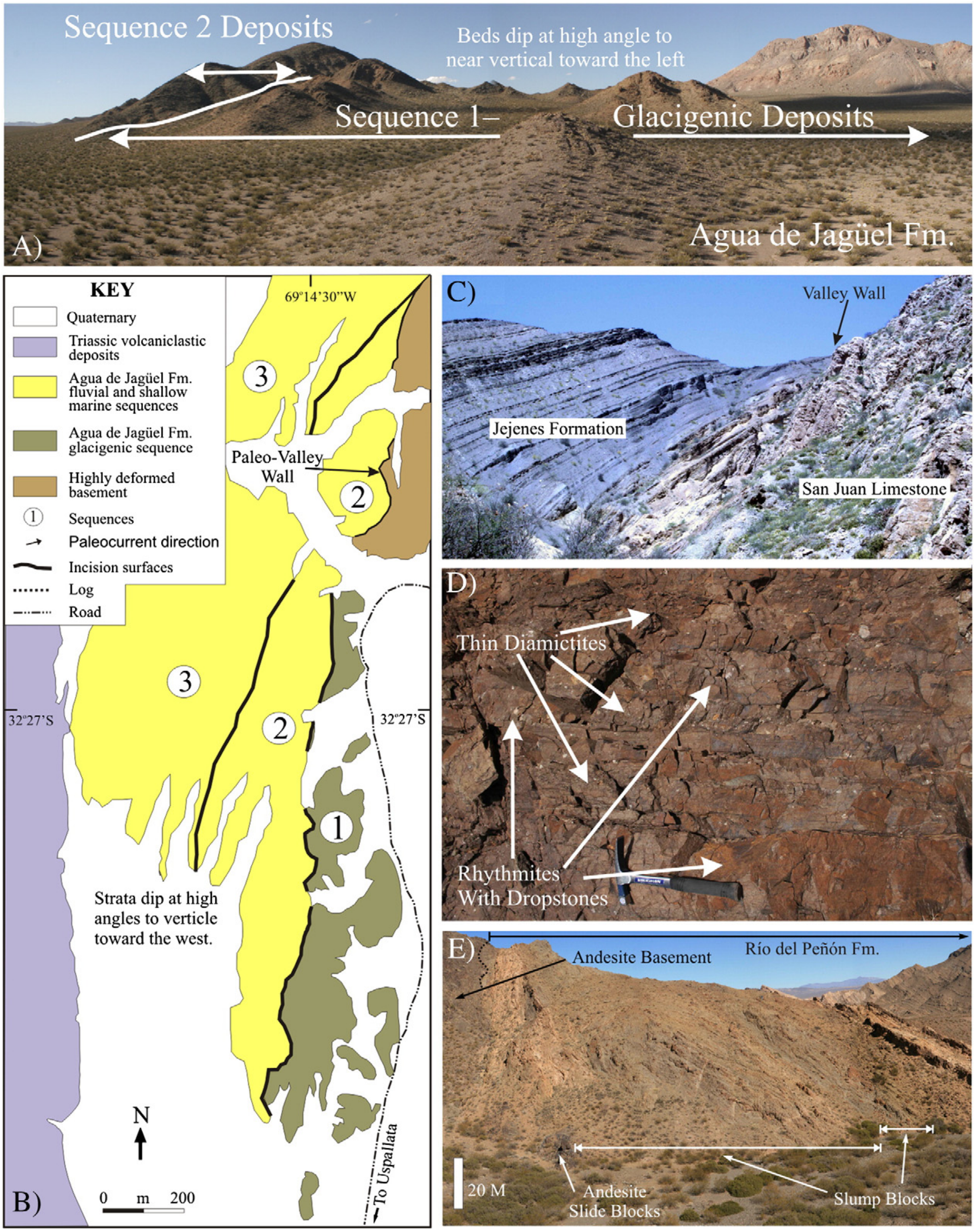


**Fig. 8.** Strata and facies of the deposits in the Protoprecordilleran Basins in western Argentina. A) Glacial and post-glacial deposits of the Agua de Jagüel Formation near Uspallata, Argentina. Outcrop of the glacigenic deposits is ~500 m thick. B) Outcrop belt of the Agua de Jagüel in the Calingasta–Uspallata Basin showing the glacial and post-glacial depositional sequences and their onlap onto basement rock along the margin of the paleo-valley (modified from Henry et al., 2010). C) The Jejenes Formation and the margins of a narrow paleo-valley near San Juan, Argentina. Trees for scale. D) Thin bedded diamictites and rhythmites with dropstones interpreted as debris flows and ice berg rafting deposits in the Agua de Jagüel Formation. Hammer for scale. E) Large-scale slump blocks in the Río del Peñón Formation in the Río Blanco Basin.

(Limarino et al., 2006; Net and Limarino, 2006). During glacial deposition, Pampeanas and Famatina basement rocks served as source terrains for much of the Paganzo Basin. However, portions of the Paganzo Basin adjacent to the Protoprecordillera received sediment from the mountain belt located to the west (Net and Limarino, 2006). In the Calingasta–Uspallata and Río Blanco Basins, the metasedimentary clasts and sand-sized rock fragments associated with rocks of the Devonian accretionary prism in the Protoprecordillera served as the primary sediment source. In late-glacial and post-glacial times, volcanic sediment, derived from an arc that had formed to the west, entered the Paganzo

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basin and granitic materials, derived from the Pampeanas and Famatina systems to the east, were deposited in the Calingasta–Uspallata Basin. Desegregation of sediments within the various basins suggests that dispersal systems had breached the Protoprecordillera by late glacial times (cf. Limarino et al., 2006; Net and Limarino et al., 2006).

During the early Bashkirian (early Pennsylvanian), glaciers retreated out of the basins and fjords resulting in glacimarine and glacilacustrine deposits being overlain by organic-rich marine mudrocks (Limarino et al., 2002; Pazos, 2002; López-Gamundí, 2010). The transgression is well documented in the Calingasta–Uspallata and Río Blanco Basins, in the Hoyada Verde, Agua de Jagüel, and Río del Peñón Formations (López-Gamundí et al., 1994; López-Gamundí, 1997; Limarino et al., 2002; Henry et al., 2008, 2010; Gulbranson et al., 2010). Marine incursion into the Paganzo Basin is also indicated by early Pennsylvanian transgressive deposits in the Guandacol, Lagares, Malanzán, and Jejenes Formations (Limarino et al., 2002; Net et al., 2002; Pazos, 2002; Kneller et al., 2004; Dykstra et al., 2006). This mudrock represents the first major transgression into the Paganzo Basin during the Pennsylvanian and indicates that the Protoprecordillera no longer served as a barrier to marine incursions into the basin. Widespread deposition of marine mudrock east of the Protoprecordillera indicates expansion of Paganzo Basin depocenters and suggests that back arc extension was occurring by that time (Limarino et al., 2002, 2006).

Following the early Bashkirian, glaciers disappeared from the Protoprecordilleran region and did not return to west-central Argentina until the Cenozoic. However, Pennsylvanian glaciation either occurred or commenced farther east in the Paraná Basin in Brazil as glaciation in the Protoprecordillera ended (Holz et al., 2008; Rocha-Campos et al., 2008). During the early Pennsylvanian (Bashkirian?), west-central Argentina and the Paraná Basin were located between 40°–60° S latitude (Scotese and Barrett, 1990; Powell and Li, 1994; Torsvik and Cocks, 2004; Blakey, 2008; Lawver et al., 2008). The age of glaciation in the Paraná Basin is controversial (cf. Holz et al., 2008; Rocha-Campos et al., 2008). The conventional age of the glacial strata in the basin based on palynological zonations is thought to range from Moscovian to Early Permian (Asselian–Sakmarian) (Souza and Marques-Toigo, 2005). SHRIMP ages from detrital zircon crystals in the Itararé Group indicate that the deposits are younger than $323.6 \pm 16$ Ma (Mississippian, Serpukhovian), and samples from a tonstein in the coal-bearing Río Bonito Formation, which directly overlies the glacigenic Itararé Group, return an age of $298.5 \pm 2.6$ Ma (earliest Asselian; Rocha-Campos et al., 2008). A major source of ice for the Paraná Basin was the Windhoek highlands in Namibia, Africa (Fig. 1; Rocha-Campos et al., 2008). Paleo-valleys up to 1500 m deep occur along the margins of the highlands. The highland may have originated as a rift shoulder that stood greater than 3000 m above sea level during the Pennsylvanian and early Permian (Visser, 1987; Stollhofen et al., 2008).

### 4.2. Relationship between the ELA and Protoprecordilleran glaciation

At 40° to 60° S latitude, the Carboniferous ELA would have resided well above sea level even in an icehouse world (Figs. 1, 3, and 5; cf. Benn and Evans, 2010). Uplift of the Protoprecordillera during the Viséan would have provided a land surface at high elevation whereby fluctuations in the position of the local ELA at that latitude could have allowed glaciers to nucleate (Fig. 7; Pérez Loinaze et al., 2010). Powell and Veevers (1987) and Eyles (1993) previously reported that orogenesis initiated glaciation in western Argentina. Outcrops of Viséan glacial deposits are relatively small and represent limited glaciation and/or erosional remnants of once more laterally extensive deposits within the mountain belt. Global cooling and uplift likely facilitated glaciation by changing the relationship between the rising land surface and a falling ELA. Fluctuations in the relative position of the ELA during the mid-Carboniferous are suggested by an absence of upper Viséan to lower Serpukhovian glacial deposits followed by later Serpukhovian to early Bashkirian glacial strata. Recently, Balseiro et al. (2009) identified a Serpukhovian fossil flora that contains pteridosperms, sphenophytes, as well as arborescent lycopsids in strata between the two glacial intervals, which suggests that a warm temperate climate existed at that time. High mass balances for the Viséan and Serpukhovian–Bashkirian glaciers allowed them to reach sea level, as indicated by glacimarine deposits contained in paleo-fjords in the Calingasta–Uspallata and Río Blanco Basins (López-Gamundí, 1997; Kneller et al., 2004; Dykstra et al., 2006; Henry et al., 2008, 2010). Evidence supporting the ELA hypothesis includes: 1) glacilacustrine conditions in the Paganzo Basin, which suggest that the mountain belt served as a barrier to marine incursions into the Paganzo Basin; 2) compositional segregation of sandstones in basins located on opposite sides of the Protoprecordillera suggesting that the mountain belt served as a barrier to sediment dispersal; (Net and Limarino, 2006), and paleo-valley relief of greater than 1000 m along the margins of the mountain belt (cf. Kneller et al., 2004; Dykstra et al., 2006; Net and Limarino, 2006; Henry et al., 2008, 2010).

Magmatism began in the Río Blanco Basin and within the Protoprecordillera in the Viséan (Punta del Agua Formation, Mississippian), which was related to compression associated with the Chañic orogeny. Later, postorogenic extensional conditions were established in the Pennsylvanian (Limarino et al., 2006). During the Early Permian (Cisuralian), the eruption of the Choiyoi Group indicate establishment of a volcanic province that extended throughout western Argentina until the Early Triassic (López-Gamundí et al., 1994). This magmatism also signaled a shift to an extensional tectonic regime that caused the collapse of the Protoprecordillera and expansion of the Paganzo Basin to the east (López-Gamundí et al., 1994; Limarino et al., 2006).

Evidence supporting collapse of the Protoprecordillera starting in the Bashkirian includes: 1) establishment of the volcanic arc to the west indicating a change in tectonic regime from that present when the Protoprecordillera formed (cf. Net and Limarino, 2006), 2) expansion of the Paganzo Basin to the east due to back-arc extension (cf. Limarino et al., 2006), 3) marine incursion into the Paganzo Basin indicating that the mountain range was breached, allowing marine connections between the eastern basin and basins to the west of the range (cf. Limarino et al., 2002, 2006; Gulbranson et al., 2010; Tedesco et al., 2010), 4) desegregation of the basins with respect to sandstone composition, suggesting that the mountains no longer served as a barrier to sediment dispersal systems (López-Gamundí et al., 1994; Limarino et al., 2002, 2006; Net and Limarino, 2006), 5) an abundance of mass movement deposits in upper glacial and lower post-glacial strata, that may have been initiated by such movements, and 6) onlap of glacial and postglacial strata onto and over the tops of the paleo-valley walls indicating that a change in accommodation pattern from uplift and incision to subsidence and valley-filling occurred (cf. Henry et al., 2008, 2010). Kneller et al. (2004), Dykstra et al. (2006), and Henry et al. (2008, 2010) previously interpreted the mass movement deposits to be due to a marine transgression that destabilized sediment along the walls of the paleo-valleys or to glacial activity.

The isotopic composition of paleo-meteoric water can provide an additional constraint on the existence of orographic moisture barriers, because orographic lift and rainout result in distinct trends in oxygen and hydrogen isotope ratios on the global meteoric water line (Craig, 1961; Rozanski et al., 1993; Jouzel et al., 1997). Here we use paleo-meteoric water estimates available from soil-formed goethite that was collected in the Paganzo Basin and dated to 312.8 Ma (Fig. 9; Gulbranson et al., 2010). $\delta D$ and $\delta^{18}O$ values of a pure goethite end-member from these samples indicate that the goethite formed at a temperature of 15 °C (±3 °C) at a soil depth of at least 10cm (Gulbranson et al., 2011), which would indicate that the mean annual temperature was likely >0 °C (Fig. 9). The oxygen and hydrogen isotopes of paleo-meteoric waters were estimated from the $\delta D$ and $\delta^{18}O$ values corrected for silicates using the equilibrium fractionation factors for the goethite–water system (Yapp, 1993), and suggest a seawater moisture source with little to no isotopic distillation due to rainout or orographic lift (Fig. 9). If the Protoprecordillera was a

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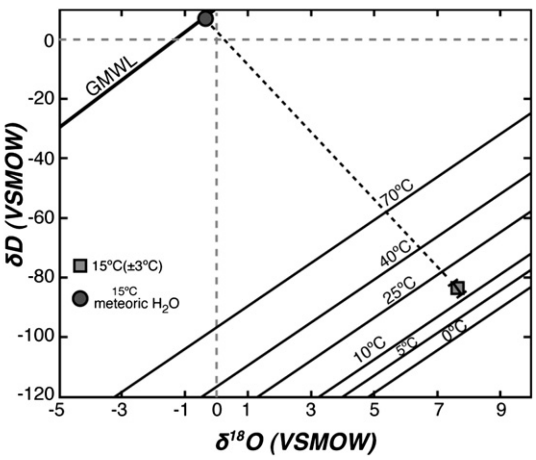


**Fig. 9.** δD versus $\delta^{18}O$ from Moscovian pedogenic goethite. Delta values reflect corrections made from isotope measurements on silicate fraction relative to a mixture of silicate and Fe-(oxy) hydroxide accounting for the mole fraction of each in the bulk sample. Estimated paleotemperature is plotted as a square symbol. The circle symbol represents estimated δD and $\delta^{18}O$ values that were in isotopic equilibrium with the studied goethite at a temperature of 15 °C. The global meteoric water line (GMWL) is plotted from the results of Rozanski et al. (1993). Goethite isotherms (0° to 70 °C) are plotted as a reference to the estimated goethite temperature. It is expected that paleo-meteoric waters would become progressively depleted in D and $^{18}O$ (i.e., more negative) during rainout and orographic lift, the extremely positive delta values estimated here suggest a position very close to the moisture source (i.e., seawater).

prominent topographic feature during the time represented by this soil-formed goethite, then it would be expected that the paleo-meteoric waters would be much more negative in both δD and $\delta^{18}O$ values. The paleo-meteoric water estimates suggest that the Proto-precordillera was not a prominent atmospheric moisture barrier by at least the early Moscovian and that mean annual temperatures were above 0 °C, suggesting that the local ELA during the Moscovian may have risen relative to the Serpukhovian–Bashkirian ELA.

Collapse of the Protoprecordillera mountain belt would have had a profound impact on glaciation (Fig. 7). The loss of altitude would have lowered the land surface below the local ELA, which would have triggered deglaciation in the mountain belt. The rate of collapse of the mountain range may have been of a different magnitude (slower) than the duration of climatic fluctuations that lowered or raised the ELA. However, a net lowering of the Protoprecordillera would have prevented additional glaciations from occurring during later climate cycles, as the land surface would have been lowered below the lowest elevation of the ELA. There is no record of glaciation in the Calingasta–Uspallata, Río Blanco, and Paganzo Basins following the collapse of the Protoprecordillera. In the Early Permian, the climate in western Argentina became increasingly arid as indicated by eolian deposits in the Paganzo Basin (Limarino et al., 2006). Additionally, strata in the Paganzo Basin record a shift from calcic vertisols to calcisols during the middle to late Moscovian, indicating the development of a semi-arid to arid climate (Gulbranson et al., 2010).

The loss of elevation would also explain why LPIA glaciation ended in this region in the Bashkirian while glaciation was beginning or continuing at the same paleolatitude farther east in the Paraná Basin (Figs. 1, 2, and 3; cf. Rocha-Campos et al., 2008). With an elevation of greater than 3000 m, the Windhoek highlands (Namibia) which fed ice into the Paraná Basin, remained at or above the ELA throughout much of Pennsylvanian and Early Permian (Fig. 1).

## 5. Late Pennsylvanian to Early Permian LPIA Maximum

### 5.1. Background

In Gondwana, peak glaciation during the LPIA occurred in the late Gzhelian to early Sakmarian (latest Pennsylvanian–Early Permian) when numerous ice centers occurred across Gondwana (Figs. 1 and 2). The ice sheets were centered over highlands that fed ice into adjacent basins (Visser, 1997a; Isbell et al., 2008b, 2008c; Rocha-Campos et al., 2008; Isbell, 2010). Major basins that received ice included: the Tepuel, Paraná, Chaco–Paraná, and Sauce Grande basins in South America; the Karoo and Kalahari basins in southern Africa; the Transantarctic Basin in Antarctica, the Talchir Basins of greater India; and the Tasmanian, Sydney, Bowen, Gunnedah, Officer, Canning, Perth, and Carnarvon basins of Australia (Fig. 3; Lindsay, 1997; Visser 1997a; Isbell et al., 2003, 2008b, 2008c; Stollhofen et al., 2008; Rocha-Campos et al., 2008; Holz, et al., 2008; Fielding et al., 2008a, 2008b, 2008c; Mory et al., 2008; Isbell, 2010; Koch, 2010; Taboada, 2010; Henry et al., in press). Smaller ice centers or alpine glaciers also occurred on Gondwana and extended into the mid-latitudes. During this interval, $pCO_2$ concentrations were equivalent to modern levels (Fig. 4; ~280 ppmv; Montañez et al., 2007).

During the Gzhelian to the Sakmarian, the South Pole resided in or near the central Transantarctic Mountains, Antarctica (Fig. 3; cf. Domeier et al., 2011). Antarctic strata from this interval contain primarily glacimarine and glacially influenced marine deposits (Fig. 10). However, some successions also contain subglacially deposited units, such as in parts of the Pagoda Formation (Fig. 10). Facies in the glacigenic strata include: 1) sheared diamictites deposited as subglacial tills during grounded ice advance into marine basins (Fig. 10B), 2) wedge-shaped sandstone bodies containing proximal traction (cross-stratification) and distal suspension settling structures deposited as grounding line fans (Fig. 10C), 3) meters to tens of meters thick massive and weakly stratified diamictites deposited from a combination of suspension settling from meltwater plumes and the incorporation of coarse clastic debris rafted by icebergs (Fig. 10A), 4) thin-bedded diamictites deposited from debris flows; and 5) lonestone-bearing mudstones deposited from icebergs and distal settling from suspension (Fig. 10; Matsch and Ojakangas, 1991; Isbell et al., 2008c; Isbell, 2010; Koch, 2010). Glacimarine conditions also occurred throughout the other basins residing within and near the polar circle at that time, including: the Karoo (South Africa), Tepuel (Patagonia, Argentina), and the Tasmanian (Tasmania, Australia) basins (cf. Powell, 1990; Hand, 1993; Visser, 1997a; Isbell et al., 2008b; Fielding et al., 2010; Taboada, 2010; Isbell et al., 2011a; Henry et al., in press).

In eastern Australia, the latest Pennsylvanian to Early Permian glaciation is recorded in glacimarine sediments and proglacial continental deposits in multiple formations in the Bowen, Gunnedah, and Sydney Basins (Figs. 1, 2, and 11; Fielding et al., 2008c, 2008d). Except for the southernmost portion of the Sydney Basin, these basins occurred just outside of the south polar circle, extending from ~52 to 65° S latitude during the LPIA maximum (cf. Domeier et al., 2011). The latest Pennsylvanian–Early Permian glacial interval in eastern Australia has been classified as the P1 glaciation by Fielding et al. (2008c), during which ice sheets, valley glaciers, and ice caps occurred in the region (Jones and Fielding, 2004; Fielding et al., 2008a, 2008c, 2008d). Paleogeographic reconstructions for eastern Australia show ice centered over the Kanimblan or Central Highlands and radiating outward into the adjacent Bowen, Gunnedah, and Sydney Basins (Veevers 2006; Fielding et al., 2008a, 2008b, 2008c, 2008d).

Other large glacigenic basins located outside of the South Polar Circle containing glacimarine sediment include (Fig. 3): the Paraná Basin of Brazil (Holz et al., 2008; Rocha-Campos et al., 2008); the Kalahari Basin of southern Africa (Visser, 1997a); and the Perth, Carnarvon, and Canning Basins of Western Australia (Eyles and Eyles, 2000; Eyles et al., 2001; Mory et al., 2008). Smaller fault-bounded basins containing glacilacustrine and glacial terrestrial deposits also occurred (e.g., Collie, Western Australia; Rafigi, Tanzania; Oman basins).

### 5.2. Relationship between the ELA and maximum LPIA glaciation

During the Late Pennsylvanian (Gzhelian) to Early Permian (Sakmarian), wide-spread glaciation extended from the South Pole

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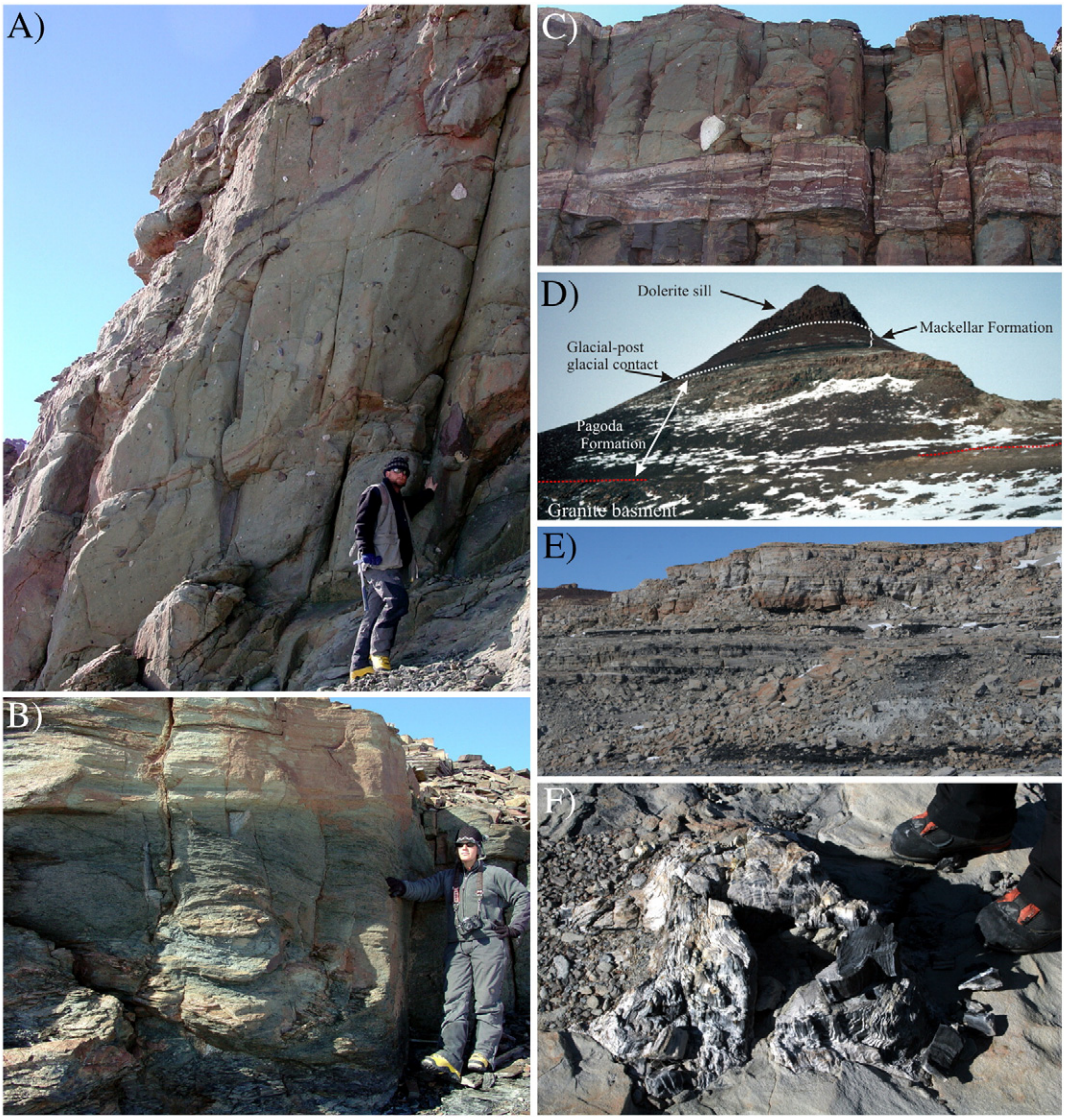


**Fig. 10.** Permian glacial and post-glacial strata in the Transantarctic Basin. A) Massive and weakly stratified diamictite in the Pagoda Formation at Tillite Glacier interpreted as deposited from settling from suspension from meltwater plumes and coarse debris deposited from iceberg rafting. B) Sheared diamictite in the Pagoda Formation at Tillite Glacier interpreted to have been deposited as a subglacial till. C) Wedge-shaped sandstone contained in clast-poor diamictite in the Pagoda Formation at Tillite Glacier interpreted to be the deposit of a grounding line fan. The sandstone body is approximately 2 m thick. D) The glacial/post-glacial contact at Sullivan Nunatak. The succession is approximately 60 m thick. E) The coal-bearing fluvial deposits of the Weller Coal Measures at Allan Hills, Southern Victoria Land. The succession is approximately 45 m thick. F) Permineralized in situ fossil stump in the Weller Coal Measures at Allan Hills.

across Gondwana and into low paleolatitudes (~30° S latitude; Frakes et al., 1992; Wopfner and Casshyap, 1997; Wopfner and Jin, 2009a, 2009b). However, glaciation was characterized by the occurrence of numerous small ice sheets rather than by a single massive glacier (Fig. 1). This interval was also associated with low atmospheric $pCO_2$ concentrations (Montañez et al., 2007). The correlation between widespread glaciation and low atmospheric $pCO_2$ values suggests that this part of the LPIA was initiated by a major global cooling event. The abundance of glacimarine strata in polar Gondwana suggests an ELA at or near sea level, which is similar to that of the lowered ELA in the Polar Regions during the Last Glacial Maximum of the Pleistocene and in present day Antarctica. During the LPIA glacial maximum, the ELA would have been lowered globally, thus allowing glaciers to nucleate on uplands in the mid-latitudes. Such ice centers formed on uplands like the Cargonian and Windhoek Highlands in southern Africa, which fed ice into the Karoo, Kalahari, and Paraná Basins (Brazil). In reference to the 300 Ma pole position of Domeier et al. (2011) (Fig. 3), the Cargonian Highlands were located between 59° and 66° S latitude and the Windhoek Highlands were located at ~53° S latitude. Elevations for the two uplands are estimated to have been >2500 m and >3000 m above sea level respectively (Visser 1987; Stollhofen et al., 2008). Ice sheets were able to form on these uplands located outside the polar circle because they rose above the position of the ELA at their given paleolatitudes.

## 6. Icehouse to Greenhouse transition at the end of the LPIA: Sakmarian to Capitanian/earliest Wuchiapingian glaciations in eastern Australia and ice-free conditions in polar Gondwana

### 6.1. Background

Middle Sakmarian strata in polar Gondwana contain sharp contacts that separate glacigenic deposits below from post-glacial strata above (Fig. 10D). In the central Transantarctic Mountains of Antarctica, which resided at or near the Permian South Pole, diamictites of the Pagoda Formation are sharply overlain by mudrocks of the Sakmarian Mackellar Formation. Strata in the Mackellar Formation only contain extremely rare lonestones in mudrocks just above the glacial/post-glacial contact (cf. Collinson et al., 1994, Seegers-Szablewski and

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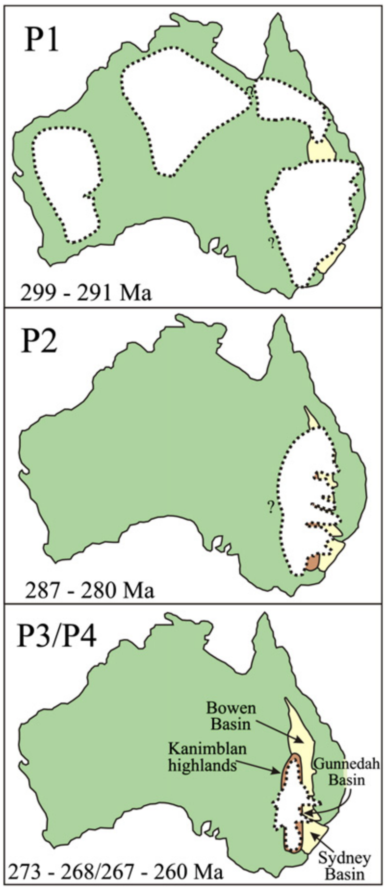


**Fig. 11.** Diagrams showing the extent of ice over Australia during the P1, P2, and P3/P4 glaciations. The Kanimblan highlands provided an uplifted region where glaciers could nucleate through the Permian. Debris from glaciers housed in the Kanimblan highlands were shed into the Bowen, Gunnedah, and Sydney Basins throughout the Carboniferous and Permian (after Veevers, 2006, Fielding et al. 2008a, 2008c, 2008d).

Isbell, 1998; Isbell et al., 2008c; Isbell, 2010; Koch, 2010; Miller and Isbell, 2010). The 150-m-thick Mackellar Formation consists of multiple coarsening-upward successions of mudrock and thin-bedded sandstones deposited as turbidites. Thick deltaic and fluvial sandstone of the 220-m-thick Lower to Middle Permian Fairchild Formation and interbedded fluvial sandstone, floodplain mudrock, coal, and lacustrine siltstone of the 750(+)-m-thick Middle to Upper Permian Buckley Formation (and correlative units throughout the Transantarctic Mountains; Fig. 10E) successively overlie the Mackellar Formation (Barrett, et al., 1986; Collinson et al., 1994; Isbell et al., 1997). During deposition of strata in the Fairchild and Buckley Formations, the central Transantarctic Mountains were located within a few degrees of the Permian South Pole (Fig. 3; cf. Domeier et al., 2011). The Buckley Formation and equivalent units, including the Weller Coal Measures, contain a high-abundant low-diversity fossil flora including numerous permineralized fossil forest horizons (Fig. 10F). Tree rings in these fossilized stumps show no evidence of frost damage (Taylor et al., 1992). However, rare pebble-sized lonestones, which suggest transport by river/lake or anchor ice, occur within floodplain and lacustrine deposits near the base of the Buckley Formation and in equivalent units elsewhere in the Transantarctic Mountains.

Glacial/post-glacial signatures similar to that of the Antarctic succession occur throughout Gondwana and in strata deposited within or near the South Polar circle of that time (Figs. 1, 2, and 3). In the Karoo Basin of South Africa, diamictites of the Dwyka Group are abruptly overlain by siltstone and mudstone of the mid-Sakmarian to Artinskian Prince Albert Formation at the base of Ecca Group (cf. Visser, 1997a; Catuneanu et al., 1998, 2005; Branch et al., 2007; Herbert and Compton, 2007; Isbell et al., 2008b; Stollhofen et al., 2008; López-Gamundí, 2010), and marine strata of the Ecca Group are conformably overlain by fluvial sandstones and floodplain mudrocks of the Upper Permian to Middle Triassic Beaufort Group (Catuneanu et al., 1998). However, no other glacial indicators have been found in the Karoo Basin higher in the section than those found in the Dwyka Group strata (cf. López-Gamundí, 2010). In Tasmania, diamictites of the Wynyard Formation are overlain by pebbly siltstones of the Inglis Formation (cf. Clarke and Forsyth, 1989). Although lonestones, tentatively interpreted as iceberg rafted dropstones, occur throughout much of the Permian strata, diamictites in the Wynyard Formation are the last known ice contact deposit within the Tasmanian succession (cf. Hand, 1993; Fielding et al., 2010; Henry et al., in press). In eastern Australia, which was located just outside of the Asselian–Sakmarian Polar Circle (Figs. 2 and 3), strata in the latest Pennsylvanian–Early Permian P1 glacial interval are sharply overlain by mid-Sakmarian mudstone with rare pebbles and sandstones (e.g., Boonderoo Beds) or rhythmically laminated siltstone (e.g., Youlambie Conglomerate; Jones and Fielding, 2004; Fielding et al., 2008a, 2008c, 2008d). Additionally, middle to upper Sakmarian, non-glacial strata in Gondwana are correlated with a dramatic rise in atmospheric $pCO_2$ values (Fig. 4; Montañez et al., 2007)

Above the Sakmarian glacial/post-glacial contact, there are no known ice contact deposits in polar and mid- to high-latitude Gondwana (Figs. 2, 10D, 10E, and 10F; Antarctica, Tasmania, Patagonia, Brazil, and South Africa,), except in eastern Australia (Figs. 1, 2, 3, and 11), which straddled the polar circle. There, Permian strata contain late Sakmarian to Early Artinskian (P2; 287–280 Ma), late Kungurian to Roadian (P3; 273–268 Ma), and latest Wordian to late Capitanian/earliest Wuchiapingian (P4; 267–260 Ma) glacial records (Fielding et al., 2008c, 2008d). At those times, the Bowen, Gunnedah, and Sydney Basins were located between ~52 and 70° S latitude, with the southern Sydney Basin having its closest approach to the South Pole at ~275 Ma (cf. Domeier et al., 2011). The P2 glacial interval is characterized by outsized clasts and diamictites in these basins, and the base of the P2 strata is marked by a flooding surface in the southern Sydney Basin that is attributed to glacial isostatic loading from advance of an adjacent ice sheet. P2 glaciers are interpreted to have been ice sheets, based on interpreted isostatic loading and broad geographic distribution of glacigenic sediments (Fielding et al., 2008c, 2008d) (Figs. 2 and 11). The P3 glacial event is identified by outsized clasts in siltstones in the Sydney and Gunnedah Basins, and also by a flooding surface in the basins attributed to glacial isostatic loading. The glaciers of P3 are hypothesized to have been ice sheets as the geographic distribution of glacigenic sediments is similar to that of P2 (Fielding et al., 2008c, 2008d) (Figs. 2 and 11). The P4 glaciation is identified by outsized clast- and glendonite-bearing mudrock in the Bowen and Sydney Basins, and by a flooding surface in the Bowen Basin. Glaciers are interpreted to have been smaller during the P4 event than those during P3. At that time they may have been represented by small ice sheets or ice caps, because dropstones are the only glacial signature for this interval, suggesting deposition distal to a glacial front (Fig. 11; Fielding et al., 2008c, 2008d). A correlation exists between the P2 to P4 events in eastern Australia with the Middle to Upper Permian low $pCO_2$ concentrations (Fig. 4; Royer, 2006; Montañez et al., 2007).

### *6.2. Relationship between the ELA and mid Sakmarian to Capitanian/earliest Wuchiapingian glaciation*

No ice-contact glacial deposits are known from polar Gondwana during the mid-Sakmarian to the end of the Permian (Figs. 1, 2, and 3) except in the southern Sydney Basin which straddled the South Polar Circle. Instead, the polar record is of climate warming in Antarctica, which consists of lonestone-free marine/basinal strata and fossil plant-bearing fluvial deposits including fluvial coal measures. The absence of frost damage within permineralized wood (Fig. 10F) indicates

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warm conditions during the growing season (Taylor et al., 1992). However, lonestones contained within floodplain and lacustrine deposits at the base of the Buckley Formation and equivalent units throughout the Transantarctic Mountains suggest seasonal lake and river ice for a short interval during the Mid Permian (cf. Kempema et al., 2001, 2002). Although lonestones occur throughout the marine post-Wynyard Permian succession in Tasmania, no ice contact deposits are known. These dropstones have been interpreted as ice rafted debris from either far-traveled icebergs or sedimentation from sea ice (Fielding et al., 2010). These findings are significant as they indicate that the ELA in polar Gondwana resided well above sea level and well above the known paleo-land surface. This suggests that following the Early Permian glaciations, major global warming had occurred and that mean annual temperatures remained above freezing throughout the remainder of the Permian at the South Pole. Following the Early Permian glacial event, Montañez et al. (2007) reported increasing $pCO_2$ levels in the atmosphere (Figs. 2 and 4). However, later decreases in $pCO_2$ levels did not cause glaciation in polar Gondwana.

In contrast to polar Gondwana, Fielding et al. (2008c, 2008d) reported that glaciers continued to feed debris into the Bowen, Gunnedah, and Sydney Basins in eastern Australia until the end of the Capitanian/earliest Wuchiapingian which were located between 52 and 70° S latitude (Figs. 2, 4, and 11). The occurrence of dropstones in these strata suggests that the glaciers reached the sea. However, it should be noted that clasts can be transported by other mechanisms (Gilbert, 1990). Ice is hypothesized to have been most extensive during the P2 event with glaciations depicted by Fielding et al. (2008c, 2008d) (Figs. 2 and 11) as an ice sheet that covered most of eastern Australia. However, Fielding et al. (2008c, 2008d) suggested that the ice centers diminished in size during the subsequent P3 and P4 events (Fig. 11). The P3 and P4 events represent a more distal record of glaciation than do strata defining the P2 event. The P2–P4 glaciations correlate with intervals of low paleo-atmospheric $CO_2$ concentration (Fig. 3; Royer, 2006; Montañez et al., 2007), whereas the non-glacial interval separating the P2 and P3 events is characterized by higher $pCO_2$ levels. Although the P2 to P4 glacial events in eastern Australia correlate with low $pCO_2$ values, neither these glacial events or the $pCO_2$ values are associated with Gondwana-wide glaciation or the glaciation of Antarctica. Glacial indicators correlative with P2 have not been identified elsewhere in Gondwana, with the possible exception of outsized clasts in the Liffey Group and Cascades Group in the nearby Tasmania Basin, which are hypothesized to have been derived from far-traveled icebergs possibly from eastern Australia (cf. Fielding et al., 2010). Rather, an absence of glacial indicators from higher latitude strata in Antarctica suggests that the P2 to P4 glacial conditions were unique to eastern Australia, which therefore indicates that the ELA for this latitude was locally lowered relative to the global ELA for that time. Otherwise, a lowering of the ELA due to a major global cooling event would have promoted the outbreak of glaciers at higher latitudes in Gondwana.

Why would glaciation resume in eastern Australia in the late Sakmarian and continue into the Capitanian/earliest Wuchiapingian (earliest Late Permian), but not return to Gondwanan basins located at higher latitudes? The answer to this question is currently unknown and is likely a complicated combination of factors that resulted in changes to the local ELA relative to the eastern Australian land surface. Four factors may have played a role in producing the P2 to P4 glaciations in eastern Australia. These include: 1) anomalously cold conditions for this sector of Gondwana during the Middle to Late Permian due to upwelling of cold bottom waters (Jones et al., 2006), 2) high paleotopography, 3) fluctuations in $pCO_2$ levels large enough to cause minor fluctuations to the local ELA but not major enough to have promoted polar glaciation, and perhaps 4) their location adjacent to the Panthalasan Ocean and in a subpolar low pressure convergent zone located between the Polar Easterlies and the Mid-latitude Westerlies.

Jones et al. (2006) suggested that cold upwelling waters allowed anomalously cold conditions to persist into the Late Permian along eastern Australia, whereas the rest of Gondwana remained relatively warm. This hypothesis of upwelling of cold, nutrient rich, deep ocean water is used to support the presence of glendonites in the Australian strata, which formed originally in the sediments as the mineral ikaite (Domack et al., 1993; Jones et al., 2006). Ikaite formation is prompted by the upwelling of cold ($-1.9$ to $+7$ °C), high alkalinity waters that interact with organic rich sediment (Suess et al., 1982; Bischoff et al., 1993; Jones et al., 2006) like those contained within Permian strata deposited in offshore environments along the paleo-coastline of eastern Australia. Wind and ocean circulation reconstructions for the region suggest offshore directed winds, which support the concept of the upwelling of colder deep oceanic waters (cf. Gibbs et al., 2002; Winguth et al., 2002; Jones et al., 2006). However, cold waters alone cannot explain glaciations, otherwise glaciers would be forming along the entire length of the Arctic and sub-Arctic coastline at the present time, which they are not.

In addition to the possible influence of the upwelling of cold ocean waters creating cool to cold conditions, paleotopography is also required to have fostered the P2–P4 glacial intervals in eastern Australia. Because the ELA is closest to sea level at the poles and rises toward the equator, substantial elevations are required to have raised the land surface above the local ELA at mid-latitudes and to have allowed glaciers to form. Glaciation in eastern Australia during P2, P3, and P4 was likely fostered by the presence and later remnants of the Kanimblan highlands (Lachlan/Thomson Fold Belts) located to the west and southwest of the Bowen, Gunnedah, and Sydney Basins (Fig. 11; cf. Veevers, 2006). Reconstructions of the glacial centers for eastern Australia are located over this upland (Fig. 11; cf. Fielding et al., 2008a, 2008c, 2008d). The Kanimblan Highlands were uplifted by east–west compression in the latest Devonian to early Carboniferous, and subsequent north–south compression during the Serpukhovian–Bashkirian (mid-Carboniferous) related to the collision of Gondwana and Laurasia further uplifted the Kanimblan Highlands (Powell, 1984; Powell and Veevers, 1987; Veevers, 2006). Later in the Pennsylvanian, intrusion of granites and the merger of the margin with the Currububula Volcanic Arc allowed the Kanimblan Highlands to maintain their elevation (Scheibner and Veevers, 2000). Erosion dominated the uplands during the P1 glacial interval (Veevers et al., 1994a). The tectonic regime later shifted to extension, which occurred from the Asselian to Kungurian (Early Permian, including P1 and P2; Veevers et al., 1994b; Fielding et al., 2001). During this time, the eastern parts of the Kanimblan uplands collapsed to form the various basins in eastern Australia (Veevers et al., 1994a). However, eastward directed paleocurrent orientations within coarse-grained alluvial fan and fluvial deposits in the basins indicate upland sources in at least remnants of the Kanimblan uplands (cf. Veevers et al., 1994a, 1994b; Fielding et al., 2001). Then, during the Roadian to early Capitanian (following the onset of P3), passive thermal subsidence and associated marine transgression occurred. During this time, paleocurrent orientations within the strata show transport from west to east, away from the former Kanimblan highlands (cf. Fielding et al., 2001). Foreland loading took place during P4 from the Capitanian to the Wuchiapingian (266–258 Ma; Late Permian), with uplift of the New England Fold Belt to the north and the development of an uplifted volcanic arc to the east. This orogenic activity may have elevated the forebulge on the craton, thereby uplifting the former Kanimblan highlands during P4. Clastics continued to enter the basin from the west from the direction of the former Kanimblan uplands during the P4 glacial interval.

Powell and Veevers (1987), Eyles (1993), and Veevers (2006, 2009) argued that, from the Serpukhovian to the Asselian, the highlands provided an uplifted land surface where glaciers nucleated. However, attention was not directed towards glacial intervals P2–P4 by these authors. Due to the fact that these ice centers continued to form in this region, and because coarse clastic from the west entered the eastern

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Australian basins throughout the Permian (Veevers et al., 1994a; Veevers, 2000; Fielding et al., 2001, 2008c, 2008d), it is reasonable to hypothesize that the former Kanimblan highlands and emplaced volcanic formations within the highlands still provided adequately elevated land surfaces from the Sakmarian into the Capitanian/earliest Wuchiapingian. Because glaciers appear to have nucleated there, the land surface must have resided above the local ELA in that region. Although the elevation of the Kanimblan Highlands in eastern Australia has not been constrained for the Pennsylvanian and Permian, Stephenson and Lambeck (1985) suggested that at the end of the Permian, this upland had elevations of between 2000 and 3000 m.

The APWP of Domeier et al. (2011) place the Kanimblan Highlands and the adjacent Bowen, Gunnedah, and Sydney Basins in a north south belt that extended from 52 to 70° S latitude. This position would have placed the uplands and basins in an area of convergent wind patterns between the Polar Easterlies and the Mid-latitude Westerlies. At the present time, the region at ~60° S is characterized by subpolar low-pressure cells, which tend to result in cool to cold wet weather (Rasmussen and Turner, 2003). Such a zone, in combination with relatively high elevations, may have produced depression of the ELA on to the Kanimblan Highlands due to increased precipitation, which may have produced conditions suitable for the nucleation of glaciers.

The combination of the highlands, ocean circulation patterns that promoted upwelling, and atmospheric circulation patterns may have resulted in an anomalously cold, wet climate in eastern Australia compared to other regions in Gondwana. Orographic lift of air masses over the highlands likely contributed to increased precipitation, thus influencing glacial mass balance. Additionally, $p\mathrm{CO_2}$ levels dropped during the P2 and again during the P3 glacial events and remained low (~300 ppmv) during the initiation of P4 (Figs. 2 and 4; cf. Royer, 2006; Montañez et al., 2007). These combined conditions may have allowed for depression of the ELA in eastern Australia and the formation of glaciers during the P2, P3, and P4 glacial events.

$p\mathrm{CO_2}$ levels were low during P2 and P3, but glaciation was not occurring elsewhere in Gondwana, even in regions closer to the South Pole, so $p\mathrm{CO_2}$ levels were not the chief influence driving nucleation of glaciers during these intervals (Figs. 2, 3, and 4). During P4, $p\mathrm{CO_2}$ levels rose and fluctuated (cf. Royer, 2006; Figs. 2 and 4) and perhaps induced warmer temperatures that resulted in smaller glaciers and a waning glacial signature (cf. Fielding et al., 2008c, 2008d). Foreland loading during this time may have uplifted the Kanimblan highlands as part of the forebulge, and this uplift may have been sufficient to trigger glaciation due to a relative drop in the ELA. Ultimately, glaciation across eastern Australia was driven by a complex interplay of drivers that shifted the ELA up and down within the former Kanimblan highlands, but the glacial signatures of P2 to P4 indicate that the ELA fell to altitudes below the upper elevations of the highlands during those intervals, allowing glaciers to form there.

## 7. Discussion and conclusions

In addition to the classically cited drivers of LPIA glaciation such as paleo-latitude and $p\mathrm{CO_2}$, the position of the ELA relative to the land surface should be taken into account when considering the initiation and demise of glacial intervals of the LPIA. The ELA vs. paleotopography is a built-in 'driver' for glaciation: elevations must be above the ELA for glaciers to form, and this precondition should be remembered when drivers of glaciation and deglaciation are discussed.

The presence of available land surfaces above the ELA where glaciers could nucleate was an important controlling factor for glaciation during the LPIA. Fluctuations in the position of the ELA then resulted in either deglaciation during a rise in the ELA relative to the land surface, or glaciation due to a fall in the ELA. Changes in the relative position of the ELA were facilitated by tectonism, global temperature changes associated with fluctuations in greenhouse gasses, or factors that promoted local fluctuations in temperature or precipitation as snow. In western Argentina, glaciation occurred over the Protoprecordillera, a fold–thrust belt, in the Viséan and in the Serpukhovian–Bashkirian due to uplift of the mountain range above the local ELA. It is reasonable to conclude that the Protoprecordillera provided the necessary elevation for glaciers to maintain positive mass balance ratios during that glacial interval. However, the Protoprecordillera collapsed below the ELA in the early Pennsylvanian, and glaciation did not resume in that region following deglaciation in the Bashkirian. On the contrary, glaciation continued in eastern and southern South America during the rest of the Pennsylvanian, and was likely controlled by a combination of altitude and paleolatitude over the uplands that fed ice into the various basins.

Lowering of the global latitudinal distribution of the ELA during the Gzhelian to Sakmarian allowed for widespread glaciation across Gondwana and for glaciers to extend into the mid-latitudes. Glacimarine deposits in polar Gondwana indicate that the ELA resided at or near sea level at that time, while in the mid-latitudes, glaciers nucleated on uplands areas. Abrupt mid-Sakmarian transitions from glacial to post-glacial deposition including extensive coal deposits and an absence of glacial deposits in Antarctica, indicate a rise in the ELA in the Polar Regions and a rise in the ELA globally. The ELA remained high and well above sea level in the Polar Regions throughout the rest of the Paleozoic.

In eastern Australia, glacial intervals continued into the Middle to earliest Late Permian, even though glaciation had ceased in other higher latitude regions in Gondwana by the mid-Sakmarian. The continued glaciation in eastern Australia was likely fostered by elevation provided by the Kanimblan highlands. However, due to its mid- to high-latitude location, eastern Australia likely required additional factors to allow lowering of the local ELA. These factors may have included anomalously cool conditions promoted by upwelling of cold waters along the adjacent coastline, occurrence of the area within the belt of subpolar low-pressure cells, and minor fluctuations in $p\mathrm{CO_2}$, which may have been enough to trigger eastern Australian glaciation, but not enough to have trigger glaciation at the Permian South Pole. The middle Sakmarian to Capitanian/earliest Wuchiapingian is a complicated interval during the LPIA with ice free poles and mid- to high-latitude eastern Australian glaciers. Causes for this paradox are problematic. However factors specific to eastern Australia as outlined above may have lowered the ELA in that region.

Continued investigations of Carboniferous and Permian strata promise to provide a better understanding of how Earth transitioned out of the LPIA and into the Late Permian–Mesozoic greenhouse state. Such deep-time studies are valuable as they provide insight into current climate change on Earth.

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