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The Pennsylvanian–Permian transition in the low-latitude carbonate record and the onset of major Gondwanan glaciation

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Abstract

Recent studies suggest a marked expansion of glacial ice across much of Gondwana beginning in the earliest Permian. Because expansion of glacial ice results in a lowering of sea level, the imprint of ice expansion should be evident worldwide as significant exposure event, hiatuses, or other evidence for sea level drop at or near the Pennsylvanian–Permian boundary. This literature review investigates the signature of an Early Permian expansion of Gondwanan ice through examination of stratigraphic records from eight carbonate-dominated, palaeotropical regions across Pangaea. Tropical carbonate environments are used because most form in tectonically quiescent regions and are sensitive indicators of eustatic change. Correlation between stratigraphic sections is achieved using the most current biostratigraphic and absolute time constraints available. All studied sections show a sequence boundary or basinward shift in facies at or near the Pennsylvanian–Permian boundary, supporting the hypothesis of a significant expansion of glacial ice and global eustatic lowstand beginning in the Early Permian. By contrast, a series of mid-Sakmarian–Kungurian transgression events in the palaeotropics are interpreted to reflect the asynchronous deglaciation of Gondwana. The stratigraphic framework developed herein will allow for better correlations among stratigraphic records from Gondwana and northern Pangaea, which will ultimately improve the understanding of how carbonate systems respond to global icehouse conditions, such as during the late Palaeozoic.

Keywords: Pennsylvanian, Permian, carbonates, eustasy, Late Palaeozoic

1. Introduction

The late Palaeozoic was a time of major climatic and environmental change that saw large-scale fluctuations in the size and distribution of Gondwanan ice sheets. It has been proposed that the Pennsylvanian–Permian transition (hereafter PPT) is a key interval of time during the late Palaeozoic ice age (LPIA) because it represents a period when ice volumes increased dramatically across much of Gondwana (Isbell et al., 2003; Fielding et al., 2008a, 2008b, 2008c). If a significant increase in Gondwanan ice did occur during the earliest Permian, then sedimentary environments should have recorded the major expansion of global ice volume as a series of significant exposure events or hiatuses reflecting eustatic drawdown at or near the Pennsylvanian–Permian boundary. Tropical carbonate platforms including rimmed platforms, epeiric platforms, and ramps are used in this study because they are especially sensitive indicators of eustatic change. Furthermore, these low-latitude environments are not subject to ice-proximal influences such as isostatic loading from glacial ice, and they often occur in tectonically quiescent areas. For these reasons, stacking patterns in tropical carbonates provide ideal proxies for past eustatic events (e.g., transgressions, regressions, and unconformities).

In this paper, the stratigraphic records from eight carbonate-dominated, palaeotropical regions are examined: 1) U.S. Midcontinent; 2) Orogrande Basin, New Mexico; 3) Perm-

ian Basin, Texas; 4) Arctic North America; 5) Bolivia; 6) South China Platform; 7) Russian Platform; and 8) Barents Shelf (Figure 1). These regions are analyzed in terms of lithology/cyclicity changes across the Pennsylvanian–Permian boundary, as well as the stratigraphic position of major surfaces including unconformities (i.e., sequence boundaries). Results support an increase in global ice volume during the earliest Permian, which clarifies how low-latitude carbonate deposition responded to a major shift in global climate conditions. An improved understanding of the palaeotropics during the PPT allows for better global stratigraphic correlations across all of Pangaea during the LPIA.

2. Background

The long-held view of the LPIA was one of a single widespread glaciation event that covered much of Gondwana from the Middle–Late Mississippian through the Early Permian (e.g., Veevers and Powell, 1987). This view, based mainly on information from Euramerican cyclothem deposits, suggested that glacial ice volumes reached a peak during the Late Pennsylvanian (Veevers and Powell, 1987; Heckel, 1994, 2008). However, Isbell et al. (2003) reviewed the stratigraphic data from across Gondwana and demonstrated that the LPIA could be divided into three distinct periods of glaciation (two in the Carboniferous and one spanning the latest Carboniferous–Sakmarian), separated by times of non-glacial conditions. The

Figure 1. Map of locations used in this study (denoted by red circles). Continental configuration based on reconstructions from Golonka and Ford (2000).

claim of discrete glacial intervals separated by warmer time periods during the LPIA was further resolved by work in eastern Australia (Fielding et al., 2008a), where the LPIA can be divided into eight distinct glacial epochs (four in the Carboniferous and four in the Permian), each lasting 1–8 Ma, which are separated by non-glacial intervals of equal duration. Additionally, studies based on compilations of data from other Gondwanan palaeofragments indicate that multiple ice centers expanded during the Early Permian in places like eastern Australia, Antarctica, India, southern Africa, South America, and the Middle East, which suggest an acme in the earliest Permian (Fielding et al., 2008b, 2008c] and references therein).

An Early Permian glacial expansion is also supported by geochemical proxy records of ice volume, $p\text{CO}_2$ (atmospheric carbon dioxide), and temperature. Late Pennsylvanian–Early Permian stable isotopic records from around the world are consistent with a drop in atmospheric $p\text{CO}_2$ in the earliest Permian, as indicated by carbon isotope records from marine carbonates, soil-forming minerals, and sedimentary organic matter (Birgenheier et al., 2010; Ekart et al., 1999; Montañez et al., 2007; see Figure 2). During the PPT, a $\sim 2\text{‰}$ increase in $\delta^{18}\text{O}$ values from low-latitude marine carbonates is consistent with a significant expansion of global ice volume (Frank et al., 2008). As with oxygen isotope compositions, a $\sim 1\text{‰}$ to 2‰ increase in $\delta^{13}\text{C}$ values occurred during the PPT, which is consistent with a drop in $p\text{CO}_2$ and a possible increase in ocean productivity associated with glacial conditions (Frank et al., 2008; Grossman et al., 2008; see Figure 2). Additionally, recent compilations of late Palaeozoic global sea levels (e.g., Rygel et al., 2008) suggest large-scale fluctuations in both magnitude and frequency of eustatic change during the latest Pennsylvanian–earliest Permian, suggesting an increase in global ice volumes.

The notion of a major expansion of Gondwanan glacial ice during the Early Permian is not without contention. The Euramerican cyclothem framework has long been considered an accurate and reliable eustatic record of Gondwanan glacial fluctuations, especially for the Carboniferous (Heckel, 1986, 1994, 2008; Wright and Vanstone, 2001). Disagreement has arisen resulting from the inconsistency among Gondwanan records suggesting a LPIA acme during the Early Permian (e.g., Isbell et al., 2003; Fielding et al., 2008a) and palaeotropical cyclothem deposits, which have been interpreted to suggest a peak in glaciation during the Late Pennsylvanian (e.g., Veevers and Powell, 1987; Heckel, 1994, 2008; Gonzalez-Bonorino and Eyles, 1995; Crow-

ell, 1999). However, the review of Gondwanan stratigraphic data by Isbell et al. (2003) concluded that the two pre-Permian glacial epochs were not extensive enough to have been the primary mechanism that controlled the well documented base level changes in the Euramerican cyclothem record.

3. Methodology and terminology

This study utilizes published stratigraphic data from across tropical and subtropical Pangaea. Since hundreds of papers have been published concerning upper Palaeozoic stratigraphic records, this study focused on sources that describe stratigraphically complete records that possess a high level of chronostratigraphic control. Specific stratigraphic information outlined by individual authors (e.g., palaeogeographic location, lithology/facies, changes in cyclicity, and stratigraphic location of unconformities) was combined with the most robust biostratigraphic and absolute constraints available (e.g., Davydov et al., 1998; Ramezani et al., 2007; Figures 2 & 3). The compiled records were calibrated to the Gradstein et al. (2004) time scale to produce a global stratigraphic framework. Stratigraphic data are plotted against the most recent reconstructions of Gondwanan glacial epochs (Fielding et al., 2008c and references therein), as well as other global climate change proxies such as oxygen and carbon stable isotopic trends (e.g., Frank et al., 2008; Grossman et al., 2008; Birgenheier et al., 2010), $p\text{CO}_2$ trends (Ekart et al., 1999; Montañez et al., 2007), and eustatic records (Rygel et al., 2008).

In the body of this paper, the nature of the Pennsylvanian–Permian boundary for each region is described, along with a brief assessment of the stratigraphic data. Refer to the Appendix for additional details regarding the palaeogeography, systematic stratigraphy, and biostratigraphic correlations for each region.

Although the late Palaeozoic represents a time of extensive global tectonic activity (e.g., Scotese and Langford, 1995; Blakey, 2008), most of the eight regions chosen for this study were tectonically quiescent during the PPT. One exception is the Permian Basin, parts of which underwent uplift and exposure during the PPT, related to the final assembly of Pangaea (Ross, 1986; Ross and Ross, 1986). However, the Permian Basin is still included in this paper because it is one of the best understood in terms of age, depositional environment, and sequence stratigraphy.

Figure 2. Chronostratigraphic framework of the eight regions studied for this review (right portion of the diagram). References for individual regions can be found in the text or on Table 1. Dark blue horizontal lines/shading represents lowstands in relative sea level (e.g., sequence boundaries/unconformities); horizontal red lines denote transgression events. Lithology symbols are included on the figure. References for Gondwanan glacial events, stable isotope/atmospheric carbon dioxide trends, and eustatic changes are included on the figure. Key conodont, fusulinid, and absolute ages used for correlation are denoted by green circles, stars, and squares, respectively. See the text and Figure 3 for more information on the biostratigraphy and original references for each section. Time scale based on Gradstein et al. (2004).

Table 1. Summary of stratigraphic data from locations examined in this study.

Region	Formations studied	Palaeolatitude	Nature of Pennsylvanian–Permian boundary	Primary age control	Other stratigraphic information (i.e. cycles, etc.)	Key references
United States Midcontinent	Douglas, Shawnee, Wabbansee, Admire, Council Grove, Chase, and Summer Groups	L. Pennsylvanian: 0–5° N (Golonka and Ford, 2000); E. Permian: 5–10° N (Scotese and Langford, 1995); also Ross and Ross (1990); Witzke (1990)	Conformable boundary between the Glenrock Limestone and the Bennett Shale members in the Red Eagle Limestone; overlying Asselian strata record a distinct basinward shift in facies.	Primarily fusulinids and conodonts	3rd–5th order cycles can be found in most of the strata (i.e. cyclothems). Major flooding events and sequence boundaries throughout the section.	Ritter (1995), West et al. (1997), Mazzullo (1998, 1999), Olszewski and Patzkowsky (2003), Sawin et al. (2006), Mazzullo et al. (2007), and Heckel (2008)
Orogrande Basin	Madera and Hueco Groups, Abo, Yeso, Bursum, and San Andres Formations	L. Pennsylvanian: ± 5° N or S (Golonka and Ford, 2000); E. Permian: ~ 5° N (Scotese and Langford, 1995); also Ross and Ross (1990)	Regional unconformity at the “Bursumian”–Wolfcampian boundary (i.e. Penn–Permian boundary). Hiatus observed across the entire basin as exposure surfaces.	Primarily fusulinids; U–Pb ages	3rd–5th order cycles can be found in most of the strata. Major unconformity at Penn–Permian boundary.	Jordan (1975), Rankey et al. (1999), Candelaria (1988), Raatz (2002), Wahlman and King (2002), and Mack et al. (2003)
Permian Basin (Glass Mountains, Midland Basin, Guadalupe Mountains)	Cisco Group, Gaptank, Neal Ranch, Lenox Hills, Wolfcamp, Hueco, Leonard, Victorio Peak, and Bone-Spring Formations; Clear Fork Group	L. Pennsylvanian: ± 5° N or S (Golonka and Ford, 2000); E. Permian: ~ 5° N (Scotese and Langford, 1995); also Ross and Ross (1990)	Regional unconformity at the Penn–Perm boundary; latest Penn–earliest Permian strata appear to be missing in many places.	Primarily conodonts and fusulinids	3rd–5th order cycles can be found in most of the strata. Major hiatus at Penn–Permian boundary, and in upper Artinskian strata. Sakmarian hiatus in the Glass Mountains portion of the basin.	Ross (1963), Candelaria et al. (1992), Fitch et al. (1995), Mazzullo (1995), Hill (1996), Yang et al. (1998), Saller et al. (1999), Yang and Kominz (1999), Harrell and Lambert (2007), and Tabor et al. (2008)
Arctic North America (i.e. North Alaska Platform)	Sadlerochit and Lisburne Groups; Wahoo and Echooka Formations	L. Pennsylvanian: 30–35° N (Golonka and Ford, 2000); E. Permian: 30–40° N (Scotese and Langford, 1995); also see Scotese and McKerrow (1990)	Major regional unconformity across platforms. Only deep-water/high sediment influx areas do not show an unconformity. Significant emergence event at the Penn–Permian boundary	Fusulinids, forams, brachiopods, corals, ammonoids, conodonts	3rd and 4th order cycles present, mostly in Penn strata. Major unconformities at: Penn–Permian boundary, Sakmarian–Artinskian boundary, Artinskian–Kungurian boundary.	Bamber and Waterhouse (1971), Crowder (1990), Beauchamp (1995), Beauchamp and Baud (2002), and Hanks et al. (2006)
Bolivia	Titicaca Group; Copacabana and Chutani Formations	L. Pennsylvanian: ~ 30° S (Golonka and Ford, 2000); E. Permian: 20–25° S (Isaacson and Diaz-Martinez, 1995)	Major sequence boundary/h hiatus that coincides with the start of the P1 glaciation in eastern Australia (Fielding et al., 2008a).	Fusulinids, conodonts; also, corals, bryozoans	2nd–3rd order cycles throughout L. Penn. and E. Permian. A marked loss of cyclic deposits occurs in the Sakmarian. Major unconformities in Sakmarian, Artinskian strata; also at Penn–Permian boundary, major Sakmarian flooding event.	Dunbar and Newell (1946), Newell et al. (1953), Isaacson and Diaz-Martinez (1995), Sempere (1995), Ottone et al. (1998), and Grader et al. (2008)
South China Platform	Chuanshan and Qixia Formations	L. Pennsylvanian: 0–15° S (Golonka and Ford, 2000); E. Permian: 0–15° S (Scotese and Langford, 1995); also, Nie et al. (1990)	Regional discontinuity or major facies change at Penn–Permian boundary. Evidence suggests platform was mostly exposed during much of Asselian–Sakmarian. Platform flooded in Artinskian.	Fusulinids, conodonts	3rd–5th order cycles can be found in most of the strata. Major sea level lowstand interpreted at Penn–Permian boundary based on facies changes.	Yang et al. (1986), Meyerhoff et al. (1991), Enos (1995), Wang and Jin (2000), and Shi and Chen (2006)
Russian Platform (southern Ural Mountains region)	Russian Platform (near the Moscow Basin)	L. Pennsylvanian: 15–25° N (Golonka and Ford, 2000); E. Permian: 20–30° N (Scotese and Langford, 1995); also Scotese and McKerrow (1990)	Penn–Permian boundary appears to be conformable in many places; significant Gzhelian–Asselian hiatus/exposure events can be traced across the entire Russian Platform.	Conodonts, fusulinids, ammonoids; U–Pb ages	3rd–5th order cycles present throughout the section. Major unconformities in the Gzhelian, Asselian, Sakmarian, and Artinskian strata.	Ross and Ross (1985), Chuvashov (1995), Davydov et al. (1998), Izart et al. (1999), Vennin et al. (2002), Davydov and Leven (2003), Izart et al. (2003), Ramezani et al. (2007), and Vennin (2007)
Barents Shelf (Finnmark Platform; Svalbard)	Gipsdalen, Bjarmeland, and Tempelfjorden Groups	L. Pennsylvanian: 25–30° N (Golonka and Ford, 2000); E. Permian: 30–45° N (Scotese and Langford, 1995); also Scotese and McKerrow (1990)	Major regional unconformity across platforms. Only deep-water/high sediment influx areas do not show an unconformity. Significant emergence event/h hiatus at Penn–Permian boundary. L. Penn–earliest Permian strata missing in many areas.	Fusulinids, conodonts	3rd–5th order cycles from Late Penn. into E. Permian. A marked loss of cyclic deposits (flooding) occurs in late Sakmarian. Major late Artinskian hiatus; transition to cool-water carbonate deposition during the Artinskian.	Heaford (1988), Nakrem (1991), Stemmerik and Worsley (1995), Harland and Geddes (1997), Ehrenberg et al. (1998), Stemmerik (2000), Rafalsen et al. (2008), and Stemmerik (2008)

Many studies included in this paper describe different scales of cyclicity such as 3rd, 4th, and 5th-order cycles. However, a lack of consistency exists in the literature concerning the definitions for different scales of cyclicity. To be consistent, the definitions outlined in Vail et al. (1991), are used herein: 1st-order = 50+ millions of years; 2nd-order = 3–50 Ma; 3rd-order = 0.5–3 Ma; 4th-order = 0.08–0.5 Ma; 5th-order = 0.03–0.08 Ma. Some authors (e.g., Heckel, 2008; Stemmerik, 2008) placed cycle durations within a Milankovitch framework; most commonly a 100,000 or 400,000-year eccentricity cycle; these frameworks were placed within the 4th or 5th-order cycle durations described above.

The issue of biostratigraphic correlation between upper Palaeozoic sections is complicated and resolution varies by region. Conodonts, fusulinids, and ammonoids, the most commonly used taxa for correlation, are used here (Figure 3). To date, Gradstein et al. (2004 and references therein) provides the most robust global biostratigraphic framework for these taxa, with particular regard to their stratigraphic placement relative to official stage boundaries for upper Palaeozoic strata. The Pennsylvanian–Permian boundary discussed herein is based on the Global Stratotype Section and Point (GSSP) for the base of the Permian System, which was placed at a shallow marine shelf section, southeast of the Russian Platform in Kazakhstan (Davydov et al., 1998). This boundary corresponds to the basal occurrence of the conodont *Streptognathodus isolatus*. Refer to the Appendix for more details about the biostratigraphic correlations used in this study.

Although records can be correlated across northern Pangaea with relative confidence due to the presence of cosmopolitan taxa such as conodonts (e.g., Menning et al., 2006), correlation between the northern Pangaea and Gondwana remains difficult. In particular, the absence of conodonts in most Gondwa-

nan sections hinders correlations between hemispheres. In some cases, such as Bolivia (part of Gondwana), fusulinids are present, which represent the primary biostratigraphic tool available to correlate these ice-proximal sections to northern Pangaeian records (Newell et al., 1953; Grader et al., 2008). Absolute age control has solved some of these issues. For example, the Pennsylvanian–Permian boundary in Russia was dated using U–Pb ages from zircons (298.90 ± 0.31 Ma; Ramezani et al., 2007). Additionally, recent isotope ages from Gondwanan sections in eastern Australia (e.g., Fielding et al., 2008a) has been used to constrain glacial epochs, which allows for more robust global correlations between ice-proximal and palaeotropical successions.

3.3. Data limitations

The data sets described in this paper all have potential biases and shortcomings. Age control is the most pertinent issue when performing global stratigraphic correlations. For this study, conodonts and fusulinids are used to determine the stratigraphic timing of individual events (e.g., unconformities and transgressions). However, even with conodont data, there is a certain level of uncertainty in the stratigraphic records; specifically, within individual stages (e.g., Asselian and Sakmarian). This study highlights the continuing need for increased biostratigraphic/radiometric resolution for all upper Palaeozoic stratigraphic sections. Increased age control will allow for the construction of more highly resolved global stratigraphic frameworks.

Additionally, continuing sequence stratigraphic research is needed to determine the global sedimentological impacts resulting from eustatic change. For example, how fast (geologically) can global sequence boundaries form? Why does a eustatic drop create a major sequence boundary in one basin and a basinward shift in facies without a major unconformity in another? Further research could help answer these types of sequence stratigraphic questions, which can be used to help determine how glacioeustatic change is preserved in the stratigraphic record. Nonetheless, given the current state of sequence stratigraphy and biostratigraphic/radiogenic age control, the stratigraphic framework described below represents a robust synthesis of low-latitude stratigraphic data for the PPT.

4. Regional stratigraphic records

4.1. United States Midcontinent

Due to the robust biostratigraphic control (e.g., Thompson, 1954; Ross and Ross, 1988, 1995; Ritter, 1995; Chernykh and Ritter, 1997) and minimal tectonic influence (West et al., 1997; Mazzullo, 1999), the U.S. Midcontinent section is ideal for studying PPT sea level change. In many of the other areas described below (except the Russian Platform), the Pennsylvanian–Permian boundary is an easily identifiable unconformity. However, the Midcontinent section, which is dominated by high-frequency (3rd and 5th-order) cyclic deposits, appears to record a gradual lowering of relative sea level during the Asselian (e.g., West et al., 1997; Olszewski and Patzkowsky, 2003).

The Pennsylvanian–Permian boundary in the U.S. Midcontinent occurs at the base of the Bennett Shale Member of the Red Eagle Limestone in the Red Eagle Sequence, where black shale overlies shallow-water limestone (Ritter, 1995; Olszewski and Patzkowsky, 2003; Figure 2). This abrupt boundary is marked by a maximum transgressive surface within the Red Eagle Sequence (Olszewski and Patzkowsky, 2003; Sawin et al., 2006). The maximum transgression at the base of the Bennett Shale is one of the larger of such events in the Council Grove Group (Ritter, 1995). Stacked paleosols of the Roca Shale overlies the Red Eagle Limestone, which records maximum regression during deposition of the Red Eagle Sequence (Olszewski and Patzkowsky, 2003). As with the Red Eagle, the overlying Grenola Sequence contains deeper water deposits

within the Grenola Limestone, which are overlain by stacked paleosols of the Eskridge Shale. The base of the Eskridge Shale is a sequence boundary that represents the maximum sea-level regression during deposition of the Grenola Sequence (Olszewski and Patzkowsky, 2003; Sawin et al., 2006). Although the Pennsylvanian–Permian boundary in the Midcontinent is associated with a significant transgression event, several studies have documented a prolonged basinward shift in facies from the Red Eagle Limestone to the Beattie Limestone, which overlies the Eskridge Shale (West et al., 1997; Olszewski and Patzkowsky, 2003; Sawin et al., 2006). This claim has also been supported by studies of brachiopod assemblages, which suggest a dramatic loss in taxa due to shallowing of the basin (e.g., Cycle 29 of Olszewski and Patzkowsky, 2001). In general, there is an overall shallowing of facies that occurs throughout the Wolfcampian section of the Midcontinent. For example, typical deep water black shale facies of Pennsylvanian strata are rare to absent in Permian strata, and there is an increase in stacked paleosols throughout the Wolfcampian (West et al., 1997). It has been postulated that this Permian shallowing could have resulted either from increased glacial ice causing a eustatic drop, small-scale tectonically induced thermal uplift associated with the assembly of Pangaea, or some combination of the two (Ross and Ross, 1987; West et al., 1997; Sawin et al., 2006). Gradual sea-level fall is consistent with the slow build-up of glacial ice across Gondwana during the Early Permian.

4.2. Orogrande Basin, New Mexico, USA

Fusulinid biostratigraphy and U–Pb ages provide a fairly robust chronostratigraphic framework for the Orogrande Basin (e.g., Williams, 1966; Steiner and Williams, 1968; Rasbury et al., 1998; Wahlman and King, 2002). An erosional unconformity at the Pennsylvanian–Permian boundary distinctly separates Pennsylvanian (e.g., “Bursumian”) fusulinid faunas from Permian (Nealian) assemblages (Wahlman and King, 2002).

The Pennsylvanian–Permian boundary in south-central New Mexico is expressed as a regional erosional unconformity, suggesting a major withdrawal of the sea (Candelaria, 1988; Raatz, 2002; Wahlman and King, 2002; Figure 2). An erosional unconformity at the Pennsylvanian–Permian boundary is also present across much of the southwestern U.S. (Ross, 1986). It should be noted that Rasbury et al. (1998) reported several U–Pb ages, utilizing paleosol calcite, from the Beeman, Holder, and Laborcita Formations (Sacramento Mountains). Using biostratigraphic positions within cyclothem deposits, they documented the age of the Pennsylvanian–Permian boundary as 302 ± 2.4 Ma, which was within the Laborcita Formation. Reevaluation of the original U–Pb ages, utilizing current GSSP absolute ages and biostratigraphic horizons (e.g., Davydov et al., 1998; Ramezani et al., 2007), would place the System boundary at the top of the Laborcita Formation in the Orogrande Basin. In the Sacramento Mountains an angular unconformity occurs between the Holder and Laborcita Formations (i.e., below the Pennsylvanian–Permian boundary). This angular unconformity is present only in the Sacramento Mountains, which implies a local tectonic influence on sedimentation in the eastern part of the basin (e.g., uplift of the Pedernal Landmass) during the Late Pennsylvanian (Candelaria, 1988; Rankey et al., 1999). Because large-scale tectonic uplift in the region had largely ended in the Late Pennsylvanian (Candelaria, 1988), the formation of an erosional unconformity at the Pennsylvanian–Permian boundary is consistent with a eustatic drop in the earliest Permian.

4.3. Permian Basin, Texas, USA

The stratigraphic record for this region is well-constrained due to conodont and fusulinid biostratigraphic studies (e.g., Ross, 1963; Ross, 1995; Wardlaw and Davydov, 2000), which allows for accurate correlations with the U.S. Midcontinent

(e.g., Wardlaw and Davydov, 2000). An erosional unconformity marks the boundary between Pennsylvanian and Permian strata in the Permian Basin (Candelaria et al., 1992; Fitchen et al., 1995; Mazzullo, 1995; Hill, 1996; Harrell and Lambert, 2007; Figure 2). In places, the boundary is expressed as an angular unconformity, which has been attributed to tectonic activity related to the final phase of thrusting within the Marathon orogenic belt (Ross, 1986; Ross and Ross, 1986; Hill, 1996). A middle Wolfcampian unconformity also occurs in some areas such as the Glass Mountains, which is thought to be related to continuing regional tectonic activity during the Early Permian (Ross, 1986; Fitchen et al., 1995; Hill, 1996).

Robust correlations with the Midcontinent make it possible to infer an overall drop in relative sea level during the Asselian. However, the angular unconformity at the Pennsylvanian–Permian boundary across much of the basin would suggest an overall tectonic cause related to Pangaea assembly (Ross, 1986; Ross and Ross, 1986). Therefore, an earliest Permian drop in relative sea level likely accompanied tectonic influences to create a major regional unconformity across the PPT.

4.4. Arctic North America

Although not as robust as conodonts, fusulinids are the primary biostratigraphic indicator used for correlations in the Arctic region of North America. In this region, they are useful for distinguishing Pennsylvanian and Permian strata (Ross, 1995). The Pennsylvanian–Permian boundary in the North Alaska Platform is an erosional unconformity (i.e., Sub-Sadlerochit unconformity) that separates Wahoo Formation carbonates from Echooka Formation siliciclastics (Bamber and Waterhouse, 1971; Crowder, 1990; Beauchamp, 1995; Hanks et al., 2006; Figure 2). It has been estimated that tens of metres of Wahoo Formation are missing as a result of widespread emergence and erosion of the platform (Crowder, 1990).

The unconformity that separates Pennsylvanian–Permian strata is ubiquitous across the region and does not appear to be angular in nature. Rather, it is erosional, which suggests a relative fall in sea level, either from tectonic emergence or a eustatic drop. It has been suggested that the Early Permian was a time of overall subsidence in the region (e.g., Beauchamp, 1995). Overall subsidence in the basin during the PPT would support a eustatic drop as the causal mechanism for a widespread subaerial/erosional unconformity.

4.5. Bolivia

Both conodont and fusulinid studies have allowed for the construction of a robust biostratigraphic framework (Dunbar and Newell, 1946; Newell et al., 1953; Suárez-Riglos et al., 1987; Ottone et al., 1998). A major sequence boundary occurs at the Pennsylvanian–Permian boundary within the carbonate dominated Copacabana Formation (at 299 Ma; see Figure 2). Grader et al. (2008) correlated this erosional boundary with the beginning of the P1 glacial epoch in eastern Australia, of Fielding et al. (2008a). Higher up in the succession, a regionally extensive latest Sakmarian–early Artinskian flooding surface represents a period of high relative sea level throughout the region (Grader et al., 2008).

The formation of the erosional unconformity across the PPT is consistent with a eustatic drop in the earliest Permian. Additionally, the formation of this sequence boundary does not appear to be related to regional tectonic uplift, which further supports a glacioeustatic control (Grader et al., 2008).

4.6. Southern China

Conodonts and fusulinids are used for regional biostratigraphic correlations across the South China Platform (e.g., Yang et al., 1986; Meyerhoff et al., 1991; Enos, 1995; Wang, 2000; Shi and Chen, 2006). As such, the base of the Permian

system in China can be confidently correlated with the GSSP in Kazakhstan. The boundary between Pennsylvanian and Permian strata in southern China is marked in many areas by a disconformity and a discernable facies shift from open marine bioclastic limestone to a widespread (~500,000 km²) large oncolite-dominated lithofacies up to 12 m thick (Yang et al., 1986; Meyerhoff et al., 1991; Shi and Chen, 2006). This major shift in lithofacies has been attributed to a relative sea level lowstand across the South China Platform during the earliest Permian (Shi and Chen, 2006). Other stratigraphic and sedimentologic data from across southern China also suggest that the stable cratonic South China Platform remained largely emergent during large portions of the Asselian–Sakmarian, and then was regionally flooded in the early Artinskian (Enos, 1995). It should be noted that the PPT exposure event documented in southern China also has been observed in an oceanic atoll succession in Japan (Sano and Kanmera, 1991).

The tectonic stability of the South China Platform during the PPT (Enos, 1995; Shi and Chen, 2006) suggests that base-level change was largely controlled by eustatic mechanisms. The disconformity and unique oncolite facies present in the lowermost Permian strata suggests a significant lowering of relative sea level compared to the uppermost Pennsylvanian open marine bioclastic limestones (Shi and Chen, 2006).

4.7. Russian platform

The excellent chronostratigraphic control make this section the most well constrained record used for this study (Davydov et al., 1998; Izart et al., 1999; Vennin et al., 2002; Davydov and Leven, 2003; Ramezani et al., 2007). This robust age control allows for detailed correlations to be made with other regions, such as the U.S. Midcontinent.

As with the U.S. Midcontinent, the Pennsylvanian–Permian boundary in the Russian section is not associated with a major unconformity, but exhibits a general shallowing of facies through the Asselian Series (Vennin et al., 2002; Vennin, 2007). Lowermost Asselian strata consist of bioclastic wackestones deposited below fair-weather wave base; upper Asselian strata consist of bioclastic packstones deposited in wave agitated conditions (Vennin et al., 2002). Additionally, Asselian strata contain several widespread exposure events that do not appear to be related to a tectonic mechanism (Vennin et al., 2002). As with the U.S. Midcontinent, the Russian Platform succession is herein interpreted as recording a more gradual eustatic drop during the PPT, culminating in a series of Asselian unconformities.

4.8. Barents Shelf (Fimmark Platform); Svalbard

Conodonts and fusulinids are the primary taxa used for biostratigraphic correlations in the Barents Shelf (e.g., Nakrem, 1991; Harland and Geddes, 1997; Nilsson and Davydov, 1997; Anisimov et al., 1998; Stemmerik, 2000). As such, the chronostratigraphic control within this succession is relatively robust.

A major exposure surface occurs at the Pennsylvanian–Permian boundary, which has been well-documented throughout the region (Stemmerik and Worsley, 1995; Ehrenberg et al., 1998; Stemmerik, 2000; Figure 2). The uppermost Pennsylvanian strata are missing in many parts of the Barents Shelf region, which suggests a large-scale drop in relative sea level during the PPT (Stemmerik and Worsley, 1995; Stemmerik, 2000, 2008). Moreover, Gzhelian–Asselian warm-water carbonate units throughout the region show evidence of multiple karsting events, dolomitization, and evaporite deposition, which suggest repeated subaerial exposure during the earliest Permian (Nakrem, 1991; Ehrenberg et al., 1998; Rafaelsen et al., 2008; Stemmerik, 2008). Because this region was tectonically quiescent during the PPT (Stemmerik and Worsley, 1995), repeated karsting during the Asselian was most likely caused by larger amplitudes of relative sea-level change, lower relative sea level compared to the latest Pennsylvanian, or some combination of these factors.

5. Discussion

5.1. Synthesis of global stratigraphic patterns

With the exception of the Russian Platform and the U.S. Midcontinent, the Pennsylvanian–Permian boundary is everywhere represented by a major exposure surface, erosional unconformity, disconformity, or hiatus (Figure 2). The apparent discordance of the U.S. Midcontinent and Russian stratigraphic records requires further evaluation. For the Russian Platform, Vennin et al. (2002) determined that the Asselian–Artinskian interval was dominated by largely regressive deposits associated with a drawdown in sea level. This Asselian–Artinskian “common regression” is evident as widespread exposure surfaces and a general transition from carbonate reef facies to very shallow, wave-influenced palaeoplysinnid boundstones (Vennin et al., 2002). Izart et al. (2003) suggested that high-frequency cyclic sedimentation patterns and subaerial unconformities in these strata most likely resulted from glacioeustatic controls. Although a major unconformity is not present at the Pennsylvanian–Permian boundary in the Russian section, the relative tectonic stability, basinward shift in facies in units overlying the boundary, and karsting events in Asselian strata are interpreted as a protracted drawdown in relative sea level, which is consistent with Gondwanan ice build-up during the Early Permian.

In the U.S. Midcontinent, Ritter (1995) noted that the Pennsylvanian–Permian boundary lies within a major cycle of relative sea-level change, and is associated with a large transgression event. However, multiple studies suggest that facies above the boundary record a significant basinward shift (West et al., 1997; Olszewski and Patzkowsky, 2001, 2003; Sawin et al., 2006). Although there is no subaerial unconformity at the Pennsylvanian–Permian boundary, Asselian strata show an overall shallowing, which records a gradual drawdown in relative sea level, likely from a glacioeustatic mechanism.

For the Russian and U.S. Midcontinent successions, the gradual shallowing of facies during the Early Permian may have been largely influenced by thermal uplift and regional tectonic warping of the cratonic surface associated with the final assembly of Pangaea (West et al., 1997; Watney et al., 2006). As such, high-frequency glacioeustatic cycles (Olszewski and Patzkowsky, 2003) are likely superimposed upon a 2nd or 3rd-order tectonic emergence signal, which may help explain the gradual drawdown of relative sea level during the Asselian.

Although not globally synchronous, major transgression events are recorded in many of the Lower Permian stratigraphic records (see Figure 2). Many of these regionally significant transgressions occurred during or after the mid-Sakmarian and are often attributed by authors to eustatic rise following the major deglaciation of Gondwana (Grader et al., 2008; Stemmerik, 2008). In three of the studied successions, the Barents Shelf, North Alaska, and Bolivia, transgressions correspond with a loss in well-defined depositional cyclicity (Beauchamp and Baud, 2002; Grader et al., 2008; Stemmerik, 2008). A lack of global synchronicity implies that these post-Sakmarian transgression events may reflect local/regional changes in sedimentation and tectonics, rather than a single eustatic rise following deglaciation. Additionally, stratigraphic evidence from Gondwana indicates that small-scale glaciation did continue in places like eastern Australia and Africa through the Middle Permian (Isbell et al., 2003; Fielding et al., 2008a), suggesting an asynchronous deglaciation at the end of the LPIA. It is difficult to ascertain why there is not global synchronicity with respect to these post-Sakmarian transgressions (e.g., deglaciations).

5.2. Implications

Despite remaining difficulties in absolute age control, a clear temporal pattern has emerged from this work. Carbonate platforms and ramps across Pangaea record subaerial exposure

and/or a basinward shift in facies during the PPT, which is interpreted as a sequence boundary. Eustatic patterns ultimately reflect a combination of global tectonic influences and fluctuations in continental ice volumes. In the late Palaeozoic, assembly of Pangaea forced changes in mid-ocean ridge volume, subduction zone mechanics, regional subsidence rates, and continental collisions (see additional discussions in Golonka and Ford, 2000; Blakey, 2008). Eustatic changes related to these tectonic mechanisms operate on 1st–3rd-order time scales (e.g., millions of years), whereas glacioeustatic mechanisms work on much shorter time scales (e.g., 4th and 5th-order cycles; see Vail et al., 1991). However, recent work has further complicated this issue by demonstrating the existence of the 3rd-order, long-term eccentricity cycle of ~ 2.4 Ma, which may influence long-term patterns in glacioeustatic cycles (Matthews et al., 1997; Matthews and Frohlich, 2002; Matthews and Al-Husseini, 2010). Both tectonic and glacioeustatic mechanisms can lead to subaerial exposure and the development of unconformities. However, only large-scale increases in continental ice volume work at a temporal scale that could result in the creation of an abrupt and globally synchronous unconformity. That said, the creation of a globally synchronous sequence boundary could be suppressed if regional tectonic events, such as subsidence, acted to diminish the amount of relative sea-level fall associated with a large increase in continental ice volume. If the mechanisms work in concert, rapid eustatic fall (glacioeustasy) combined with slow continental emergence (tectonic) could create a globally synchronous sequence boundary.

The Late Pennsylvanian–Early Permian collision between Gondwana and Euramerica is well documented (e.g., Scotese and Langford, 1995; Golonka and Ford, 2000; Blakey, 2008). As such, it is likely that some of the eustatic drop during the PPT can be explained by tectonic uplift across portions of Pangaea, albeit on much slower time scales than glacioeustasy. This is the case for the Permian Basin where regional uplift within the Marathon orogenic belt created a major hiatus and an angular unconformity in uppermost Pennsylvanian strata (Ross, 1986; Ross and Ross, 1986; Hill, 1996). Additionally, slow continental emergence related to thermal tectonic uplift and cratonic warping may also help to explain the gradual shallowing of facies inferred for the U.S. Midcontinent and Russian sections (West et al., 1997; Watney et al., 2006).

Perhaps the most tectonically stable region studied is the South China Platform. During the PPT, the South China Platform was not attached to Pangaea; it lay on the tectonically passive side of the Yangtze Craton (Enos, 1995; Scotese and Langford, 1995; Golonka and Ford, 2000; Wang and Jin, 2000). Even in this tectonically stable environment, the stratigraphy records a significant drop in relative sea level during the PPT (Yang et al., 1986; Meyerhoff et al., 1991; Shi and Chen, 2006). This evidence implies that the global stratigraphic pattern documented herein cannot be explained by a tectonic mechanism alone.

The global stratigraphic pattern of a eustatic drop documented in this study most likely resulted from a eustatic fall caused by the interplay between higher frequency (3rd–5th-order) glacioeustatic signals superimposed upon long-term global tectonic influences (1st–3rd-order cycles) coinciding with the beginning of increased ice accumulations across much of Gondwana (Isbell et al., 2003; Fielding et al., 2008b). The combination of major tectonic events (e.g., Pangaea assembly) and glacioeustatic fluctuations (e.g., LPIA) helps explain why the lowest sea levels during the Palaeozoic were during the Permian. Since areas of relative tectonic stability (e.g., South China Platform) were also influenced by a major drop in relative sea level during the PPT, it is likely that the ice sheets on Gondwana had a more significant role in controlling eustasy during the PPT.

Given all of the available data, it is hypothesized that variation in Gondwanan ice volume was the dominant control over eustatic change during the latest Pennsylvanian–Sakmarian. Stratigraphic data from carbonate environments across the

palaeotropics and subtropics is consistent with a peak of the LPIA occurring in the very latest Pennsylvanian–Early Permian. Low-latitude carbonate platforms and ramps also record the deglaciation of Gondwana as a series of large-scale asynchronous transgressions beginning in the mid-Sakmarian.

6. Conclusions

Stratigraphic data from latest Pennsylvanian–earliest Permian carbonate platforms and ramps across equatorial Pangaea record the occurrence of a sequence boundary or a distinct basinward shift in facies. This event is present in some form in all eight regions studied for this review, regardless of degree of regional tectonic activity (Stemmerik and Worsley, 1995; Hill, 1996; West et al., 1997; Vennin et al., 2002; Wahlman and King, 2002; Hanks et al., 2006; Shi and Chen, 2006; Grader et al., 2008). Results from this study suggest a common overall cause for the formation of this global stratigraphic pattern, which is consistent with a glacioeustatic drawdown beginning in the earliest Permian.

This study supports emerging evidence that global ice volume reached an acme in the Early Permian (Fielding et al., 2008b, 2008c). A significant increase in ice volume across Gondwana during the PPT resulted in widespread exposure events or a significant basinward shift in facies in low-latitude carbonate environments. The eustatic drop during the PPT was likely enhanced by the tectonically driven eustatic fall associated with the final assembly of Pangaea, which operated on 1st–3rd-order time scales (West et al., 1997; Watney et al., 2006).

Mid-Sakmarian–Kungurian transgression events documented here suggest an end to major Gondwanan glaciation by the mid-Sakmarian. However, the persistence of small ice centers through the Middle Permian suggests an asynchronous deglaciation of Gondwana after the Sakmarian (Isbell et al., 2003; Fielding et al., 2008a, 2008b, 2008c). Further research is needed to document the global nature of these complex post-Sakmarian transgressions and how they relate to the overall demise of the LPIA.

It is intriguing that the inferred eustatic drop during the PPT manifests itself differently in stratigraphic successions across the globe. In many of the records, the eustatic lowstand is represented by a large unconformity. Other records show only a basinward shift in facies during the Asselian, instead of a major exposure surface at the Pennsylvanian–Permian boundary. This study highlights the complicated stratigraphic patterns that can form during periods of increased glaciation, combined with large-scale tectonic processes associated with supercontinent assembly.

The Appendix following the References contains additional stratigraphic information for each region studied (e.g., biostratigraphy, palaeogeography, and systematic stratigraphy).

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Appendix to:

" The Pennsylvanian–Permian transition in the low-latitude carbonate record and the onset of
major Gondwanan glaciation."

Jesse T. Koch and Tracy D. Frank

A1. Biostratigraphic framework

A1.1. Conodonts

Perhaps the best index fossils for correlation of upper Palaeozoic strata are conodonts due to their cosmopolitan nature and rapid evolutionary rates during the late Palaeozoic (Wardlaw, 1995). The Pennsylvanian–Permian boundary discussed herein is based on the Global Stratotype Section and Point (GSSP) for the base of the Permian System, which was placed at a shallow marine shelf section, southeast of the Russian Platform in Kazakhstan (Davydov et al., 1998). This boundary corresponds to the basal occurrence of the conodont *Streptognathodus isolatus*, which is a morphotype of *Streptognathodus* “*wabaunsensis*” (see Fig. 3, Davydov et al., 1998). In the U.S. Midcontinent, this horizon occurs in the basal portion of the Bennett Shale Member of the Red Eagle Limestone (Ritter, 1995). This conodont boundary also has been identified in the Glass Mountains portion of the Permian Basin (Wardlaw and Davydov, 2000), and in southern China (Wang, 2000).

Utilizing conodont zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al., 2004 and references therein; see Fig. 3): the base of the Sakmarian is defined by the basal occurrence of *Sweetognathus merrilli*; the base of the Artinskian is defined by the first appearance of *Sweetognathus whitei*; and the base of the Kungurian is within the *Neostreptognathodus* zone (e.g., *N. pnevi*, *N. exculptus*, *N. pequopenensis*).

A1.2. Fusulinids

Fusulinids represent a widely used biostratigraphic tool for upper Palaeozoic sections across the palaeotropical and mid-latitude localities (e.g., Ross, 1995; Ross and Ross, 1995), but

their biostratigraphic usefulness is limited by provincialism and homeomorphy with respect to evolution (Davydov, 1996). Despite these issues, fusulinids continue to be utilized as biostratigraphic indicators for upper Palaeozoic studies around the world. At the GSSP, the *Sphaeroschwagerina vulgaris*-*Sphaeroschwagerina fusiformis* zone occurs 6 m above the conodont marker, *Streptognathodus isolatus*. In many regions such as the Orogrande Basin, Barents Shelf, Arctic North America, and Bolivia, conodont data are unavailable and fusulinids are relied upon for biostratigraphic correlations (Dunbar and Newell, 1946; Newell et al., 1953; Ross and Ross, 1987; Ross, 1995; Stemmerik, 2000; Wahlman and King, 2002). In the Boreal and Tethyan regions, the *Sphaeroschwagerina vulgaris*-*Sphaeroschwagerina fusiformis* zones provides the closest approximation for the base of the Permian (Fig. 3; Davydov, 1996). In North America, the base of the Permian is defined by first occurrence of the Nealian fusulinid fauna, which is represented by: *Pseudoschwagerina uddeni*, *P. texana*, *P. beedei*, *Paraschwagerina gigantea*, larger *Leptotriticites* and advanced *Schwagerina* (Fig. 3; Ross, 1963; Wahlman and King, 2002).

Utilizing fusulinid zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al., 2004 and references therein; see Fig. 3): the base of the Sakmarian is defined by the basal occurrence of *Schwagerina moelleri*; the base of the Artinskian is defined by the first appearance of the *Pseudofusulina* genus; and the base of the Kungurian is at the base of the *Brevaxina* zone.

A1.3. Ammonoids

Ammonoids have also been utilized as biostratigraphic indicators for upper Palaeozoic strata, but as with fusulinids, they show a significant amount of provincialism, thus limiting their use for high-resolution biostratigraphy (Gradstein et al., 2004 and references therein). At the GSSP, the *Svetlanoceras* ammonoid zone appears ~27 m above the first occurrence of the conodont marker, *Streptognathodus isolatus* (Davydov et al., 1998). Although less reliable, the boundary between the *Svetlanoceras* and *Shumardites* ammonoid zones marks the Carboniferous–Permian boundary (Fig. 3; Gradstein et al., 2004 and references therein). Utilizing ammonoid zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al., 2004 and references therein; see Fig. 3): the base of the Sakmarian is within the upper part of the *Svetlanoceras* genus; the base of the Artinskian is defined by the first appearance of the *Neopronorites-Metaperrinites* zone; and the base of the Kungurian is within the *Aktubinskia-Artinskia-Neopronorites* zone.

A2. Additional stratigraphic information

A2.1. United States Midcontinent

The Permian System of the Midcontinent extends from Oklahoma to Iowa (including Kansas, Missouri, and Nebraska; it encompasses several basins including the Salina, Forest City, and Sedgwick (Heckel, 1986; Mazzullo, 1998; 1999; Fig. 1; Table 1). Palaeomagnetic data suggest these basins were located between 0–10° N during the PPT (Ross and Ross, 1990; Witzke, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000). The tectonic stability during the late Palaeozoic and lateral continuity of strata across this broad epeiric platform have

allowed for the construction of a robust regional stratigraphic framework (Mazzullo, 1999; Olszewski and Patzkowsky, 2003; Heckel, 2008). Age control for the upper Palaeozoic strata of the Midcontinent is achieved through the use of conodont and fusulinid biostratigraphy (Thompson, 1954; Ross and Ross, 1988; 1995; Ritter, 1995; Fig. 3). The GSSP conodont marker, *Streptognathodus isolatus*, has been observed in the base of the Bennett Shale (Ritter, 1995; Chernykh and Ritter, 1997) and the Nealian fusulinid fauna (described above) occurs in the Neva Limestone (Thompson, 1954).

Upper Palaeozoic strata of the Midcontinent are known for containing cycles of mixed siliciclastic/carbonate deposits. In general, sea level highstands are represented by offshore dark gray/black shales, and sea level lowstands are recorded by shallow marine carbonates, delta deposits, and/or palaeosols (Heckel, 1986; 2008). Individual cycles are traceable across broad regions with relative ease due to a robust biostratigraphic framework and limited tectonic modification.

Several studies have suggested that 3rd and 4th-order cycles dominate in Upper Pennsylvanian/Lower Permian strata (Mazzullo, 1998; Mazzullo, 1999; Olszewski and Patzkowsky, 2003; Heckel, 2008). For PPT strata, high resolution sequence stratigraphic analysis of outcrop data has suggested 5th-order cycles that are superimposed on lower order cycles (Olszewski and Patzkowsky, 2003). However, the style of cyclicity across the Pennsylvanian–Permian boundary remains constant. Although cyclicity appears to have not changed significantly during the PPT, a shift from humid conditions in the Late Pennsylvanian to drier conditions in the Permian has been inferred based on stable isotope compositions of marine carbonates and a shift to more evaporite rich lithofacies (e.g., West et al., 1997; Mazzullo et al.,

2007). This shift to drier conditions during the Permian is consistent with other reports demonstrating a shift to more arid conditions across western Pangea in the Permian (Tabor and Montañez, 2002; Tabor et al., 2008).

A2.2. Orogrande Basin, New Mexico, USA

The Orogrande Basin of south-central New Mexico contains some of the most continuous Pennsylvanian–Permian sections of marine carbonates in western Euramerica (e.g., Robledo Mountains; Wahlman and King, 2002). The basin was located within 5° of the equator during the PPT (Ross and Ross, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000; Fig. 1; Table 1). During Pennsylvanian–Permian time, the Orogrande Basin was a westward extension of the Permian Basin in Texas; it was relatively shallow and elongate, trending generally north to south (Jordan, 1975; Candelaria, 1988). Deposits range from neritic marine carbonate strata in the south to mixed siliciclastic–carbonate strata in the northern portion of the basin (Jordan, 1975; Candelaria, 1988; Rankey et al., 1999). The nomenclature of Upper Pennsylvanian mixed siliciclastic–carbonate strata in the basin is complex and varies with geographic location. In the northern San Andres and Franklin Mountains (northern and southern part of the basin, respectively), Upper Pennsylvanian strata are the Panther Seep and Bursum Formations; coeval strata in the Sacramento Mountains (eastern part of the basin) are the Beeman, Holder, and Laborcita Formations (Kues, 2001; Raatz, 2002). The term Madera Group has been proposed for Upper Pennsylvanian strata in the western part of the basin, such as the Robledo Mountains (Kues, 2001).

Lower Permian strata consist of open marine carbonate-dominated Hueco Group in the southern part of the basin and the terrestrial to marginal marine siliciclastic-dominated Abo Formation in northern areas (Jordan, 1975; Candelaria, 1988; Raatz, 2002; Wahlman and King, 2002; Mack et al., 2003). A transitional zone occurs in the central portion of the basin where Hueco and Abo strata interfinger. In this part of the basin, the marine carbonates of the Hueco Group show characteristics consistent with a more restricted environment (impoverished faunas, etc.) and the siliciclastic strata of the Abo Formation show a more marine influence (e.g., wave ripples, tidal flat facies, etc.) than do lithofacies further to the north (Jordan, 1975; Wahlman and King, 2002; Mack et al., 2003). Since conodont data are unavailable for the Orogrande Basin, PPT deposits have been dated through the use of fusulinid biostratigraphy; the Nealian fusulinid fauna, which represents lowermost Permian, has been found throughout the basin (Williams, 1966; Steiner and Williams, 1968; Wahlman and King, 2002).

Depositional style changed in the Orogrande Basin across the PPT, partially due to a decrease in basin subsidence related to regional tectonic activity (Candelaria, 1988). The Pennsylvanian strata tend to be dominated by high-frequency (3rd and 4th-order) cyclic deposits, especially in the mixed carbonate–siliciclastic successions in the Sacramento Mountains (e.g., Holder and Laborcita Formations). Individual cycles (1–3 metres thick) consist of open marine limestone overlain by terrestrial to marginal marine siliciclastic deposits (Rankey et al., 1999; Raatz, 2002). While 3rd–5th-order cyclic deposits occur in Permian strata, they tend to be less uniform in thickness and do not form classic cyclothems as observed in the underlying Pennsylvanian units (Jordan, 1975; Rankey et al., 1999; Mack et al., 2003). In the southern part of the basin they consist of shallowing upward sequences in open marine carbonate strata. In the

central and northern part of the basin these cycles consist of open to restricted marine carbonates overlain by terrestrial to marginal marine siliciclastic deposits (Jordan, 1975; Mack et al., 2003).

A2.3. Permian Basin, Texas, USA

The Permian Basin in western Texas and southeast New Mexico comprises several subbasins (e.g., Delaware, Midland, etc.) that existed as a segmented foreland basin during the late Palaeozoic. These relatively deep basins contain some of the most complete Permian sections in northern Pangea (Fitchen et al., 1995; Mazzullo, 1995; Hill, 1996). During the PPT, the region was located within 5° of the palaeoequator (Ross and Ross, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000; Fig. 1; Table 1). Strata contain mixed siliciclastic–carbonate facies, which resulted from the collision of the North and South American plates during the late Palaeozoic (Ross, 1986; Fitchen et al., 1995). The best outcrops of PPT strata occur in the Glass Mountains region of the Delaware Basin, where the Upper Pennsylvanian Gaptank Formation consists of open marine carbonates and is unconformably overlain by Permian (Wolfcampian) strata (Hill, 1996). Lower Permian strata in the Glass Mountains consist of open marine carbonate shelf facies that are divided into the Neal Ranch Formation, which is unconformably overlain by the Lenox Hills Formation (Ross, 1963; Hill, 1996). Outside of the Glass Mountains the uppermost Pennsylvanian mixed siliciclastic–carbonate strata are called the Cisco Formation (Group), and the lowermost Permian mixed siliciclastic–carbonate strata are assigned to either the Wolfcamp Formation or Hueco Group (Fitchen et al., 1995; Mazzullo, 1995; Hill, 1996; Yang et al., 1998). In the Glass Mountains, the conodont genus *Streptognathodus* has been found in the Grey Limestone Member, which places the

Pennsylvanian–Permian boundary near the top of the Gaptank Formation (Wardlaw and Davydov, 2000). Other key conodont markers such as, *Sweetognathus merrilli*, *Sweetognathus whitei*, and *Neostreptognathodus pequopensis*, define Sakmarian, Artinskian, and Kungurian deposits, respectively (Fig. 3). The fusulinid biostratigraphy also supports the conodont data, with the Nealian fusulinid fauna present in Lower Permian strata (Ross, 1963; 1995; Fig. 3). However, the biostratigraphic resolution of the fusulinid data in the Glass Mountains is not as highly resolved as the conodont data (Wardlaw and Davydov, 2000).

There is evidence of multiple scales of depositional cyclicity throughout the basin (e.g., 3rd and 4th-order, etc.) in both Pennsylvanian and Permian strata, which are recorded as alternating beds of shallow-water carbonates and terrestrial to basinal siliciclastics (Mazzullo, 1995; Yang et al., 1998; Saller et al., 1999; Yang and Kominz, 1999). Although subtle changes in facies and cycle thickness are evident, significant changes in depositional style do not appear to be present across the PPT.

A2.4. Arctic North America

Northern Alaska moved from 30–35° N during the Pennsylvanian to 30–40° N during the Permian (Scotese and McKerrow, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000; Bensing et al., 2008; Fig. 1; Table 1). Rifting dissected Arctic Alaska and the North Alaska Platform developed on horst-structures in a shallow-water marine environment (Beauchamp, 1995). The North Alaska Platform was chosen over the adjacent Sverdrup Basin because tectonic rifting was less developed in this region (Beauchamp, 1995). Upper Pennsylvanian strata consist of the Wahoo Limestone (upper Lisburne Group), which is overlain by mixed

carbonate–siliciclastic strata of the Permian-age Echooka Formation/lower Sadlerochit Group (Bamber and Waterhouse, 1971; Crowder, 1990; Beauchamp, 1995; Hanks et al., 2006). The uppermost Mississippian–Upper Pennsylvanian Wahoo Limestone was deposited under open marine conditions on a shallow carbonate ramp and contains a diverse fauna of brachiopods, fusulinids, conodonts, foraminifers, and corals (Bamber and Waterhouse, 1971; Crowder, 1990; Hanks et al., 2006). The overlying Echooka Formation (informal basal conglomerate member and the Joe Creek Member) contains a mix of conglomerate, restricted carbonates, and nearshore siliciclastic strata (Bamber and Waterhouse, 1971; Crowder, 1990; Beauchamp, 1995; Hanks et al., 2006). Upper Palaeozoic strata from the North Alaska Platform are dated primarily through the use of fusulinids (Ross, 1995; Fig. 3). Only Carboniferous–Early Permian fusulinids are known from northern Alaska, which is attributed to the tectonic drift into more northerly latitudes during the mid-Permian (Beauchamp, 1995; Ross, 1995). Although Permian strata are predominantly siliciclastics (e.g., Echooka Formation), coeval open marine carbonate strata in the adjacent Sverdrup Basin contain key Permian fusulinid markers such as:

Sphaeroschwagerina vulgaris-fusififormis (e.g., Pennsylvanian–Permian boundary); *Schwagerina moelleri* for the base of the Sakmarian; *Pseudofusulina* genus for the base of the Artinskian; and the *Parafusulina* genus for upper Artinskian–lower Kungurian strata (Ross, 1995; see Fig. 3).

Cyclic deposits (3rd and 4th-order) are known, consisting of typical shoaling upward cycles that dominate the Wahoo Formation (Bamber and Waterhouse, 1971). Fining upward cycles, interpreted as tempestites, are common in the Echooka Formation (Crowder, 1990). Even though cyclic deposits have been documented, there has been only a broad attempt to relate these changes to global palaeoclimatic and eustatic variations. Beauchamp and Baud (2002)

determined that high-frequency depositional cyclicity ended in the latest Sakmarian, which coincides with a regional maximum flooding event.

A2.5. Bolivia

The Bolivia/Peru region of South America was situated at $\sim 30^\circ$ S during the Mississippian and Early Pennsylvanian (Golonka and Ford, 2000; Fig. 1; Table 1). Northward movement into more tropical latitudes ($20\text{--}25^\circ$ S) during the Late Pennsylvanian–Early Permian allowed for the development of extensive warm-water carbonate platforms (Isaacson and Díaz-Martínez, 1995; Sempere, 1995). This region is different from the other localities discussed in this paper because Bolivia was part of Gondwana, and more proximal to the extensive glacial activity that occurred further to the south. Upper Pennsylvanian–Lower Permian deposits in Bolivia are represented by the Copacabana Formation, which consists of mostly shallow-water carbonates interbedded with terrigenous siliciclastic units (Grader et al., 2008). The Copacabana Formation occurs in a series of basins and subbasins throughout Bolivia and parts of central and southern Peru (Newell et al., 1953; Grader et al., 2008). The mixed siliciclastic–carbonate–volcanic Chutani Formation unconformably overlies the Copacabana Formation throughout most of the region.

Although conodont studies exist (e.g., Suárez-Riglos et al., 1987), fusulinid biostratigraphy remains the most widespread taxa used for age determinations of Bolivian PPT strata (Dunbar and Newell, 1946; Newell et al., 1953; Ottone et al., 1998; Fig. 3). The fusulinid faunas described from Bolivia (e.g., Nealian fusulinid fauna) can be correlated with those from the southwestern United States (e.g., Ross, 1963; Wahlman and King, 2002) because they

contain similar taxa (Dunbar and Newell, 1946; Newell et al., 1953). Corals and bryozoans also have been used for age control in the region (Wilson, 1990; Sakagami, 1995).

A major sequence boundary occurs at the Pennsylvanian–Permian boundary within the Copacabana Formation (299 Ma; see Fig. 2). Grader et al. (2008) correlated this boundary with the beginning of the P1 glacial epoch in eastern Australia, of Fielding et al. (2008). Higher up in the succession, a regionally extensive latest Sakmarian–early Artinskian flooding surface represents a period of high relative sea level (Grader et al., 2008).

The Copacabana Formation contains high-frequency cycles throughout, which have been interpreted as 2nd, 3rd, and 4th-order sequences (Grader et al., 2008). Individual cycles consist of lowstand and transgressive siliciclastic deposits that are capped by highstand open marine carbonate strata. The tops of cycles often show signs of exposure (Isaacson and Díaz-Martínez, 1995). As in the Barents Shelf and North Alaska regions, the higher-frequency cycles disappear from the record after the late Sakmarian (Beauchamp and Baud, 2002; Grader et al., 2008; Stemmerik, 2008).

The formation of the erosional unconformity across the PPT is consistent with a eustatic drop in the earliest Permian. Additionally, the formation of this sequence boundary does not appear to be related to regional tectonic uplift, which further supports a eustatic control (Grader et al., 2008).

A2.6. Southern China

During the late Palaeozoic, the Yangtze Craton (e.g., South China block) was situated between 0–15° S in the Palaeo-Tethys Ocean (Nie et al., 1990; Scotese and Langford, 1995;

Golonka and Ford, 2000; Fig. 1; Table 1). Southern China, a broad and relatively flat terrain during the late Palaeozoic, did not collide with northern China (Sino-Korean Craton) until the Mesozoic (Enos, 1995; Wang and Jin, 2000). This palaeogeographic configuration allowed epicontinental seas to periodically cover southern China, leading to the development of epeiric platforms in the Dian-Qian-Gui Basin during portions of the Pennsylvanian–Permian (Wang and Jin, 2000). PPT strata in the region consist of the Chuanshan Formation, an oncoid-bearing carbonate unit which is disconformably overlain by the late Cisuralian carbonate-dominated Qixia Formation (Wang and Jin, 2000; Shi and Chen, 2006).

Upper Palaeozoic strata from the Palaeo-Tethys region are dated using conodont and fusulinid biostratigraphy (Yang, 1986; Meyerhoff et al., 1991; Enos, 1995; Wang, 2000; Shi and Chen, 2006; Fig. 3). The key conodont species, *Streptognathodus isolatus*, which defines the base of the Permian System, is widely distributed throughout southern China (Wang, 2000). This conodont marker coincides with a key fusulinid zone (*Sphaeroschwagerina vulgaris*–*Sphaeroschwagerina fusiformis*), which also aids in Pennsylvanian–Permian boundary correlations (Ross, 1995). Other faunas such as, brachiopods, ammonoids, and corals provide useful age constraints in many places (Meyerhoff et al., 1991).

Whereas depositional cyclicity has been observed in many sections across the South China Platform, there have been few attempts to correlate individual cycles and sequences to other global stratigraphic frameworks (e.g., Midcontinent, Russian Platform, etc.). Shi and Chen (2006) recognized high-frequency (3rd and 4th-order?) cycles in sedimentary stacking patterns of the Chuanshan Formation and related these cycles to glacioeustatic fluctuations.

A2.7. Russian Platform

During the Late Pennsylvanian, most of eastern Russia was situated between 15–25° N in a tropical to subtropical environment (Golonka and Ford, 2000; Fig. 1; Table 1). The region moved north during the Early Permian to 20–30° N (Scotese and McKerrow, 1990; Scotese and Langford, 1995). The Late Pennsylvanian–Early Permian Russian Platform was dominated by warm-water carbonate deposition; phylloid algal mound development reached a peak in the Asselian–Sakmarian (Vennin et al., 2002; Vennin, 2007). The Russian nomenclature system utilizes a set of horizons instead of formations to differentiate lithostratigraphic units. Horizon determinations are based on lithostratigraphic changes as well as fusulinid/conodont assemblages. Gzhelian deposits across the Russian Platform are generally separated into the Rusavkian, Pavlovoposadian, Noginian, and Melekhovian Horizons (Ross and Ross, 1985; Davydov and Leven, 2003). Asselian strata are separated into the Sjuranian, Uskalykian, and Shikhanian Horizons, and Sakmarian deposits are divided into the Tastubian and Sterlitamakian Horizons (Chuvashov, 1995; Vennin et al., 2002). Artinskian strata are divided into the Burtsevian, Irginian, and Sarginian Horizons, and Kungurian strata defined by the Saranian, Fillipovian, and Irenian Horizons (Chuvashov, 1995).

The chronostratigraphic framework is extremely well constrained across the Russian Platform. Although conodonts are used to identify the official Pennsylvanian–Permian boundary (e.g., first appearance of *Streptognathodus isolatus*), fusulinid and ammonoid biostratigraphy remain extremely useful (Davydov et al., 1998; Izart et al., 1999; Vennin et al., 2002; Davydov and Leven, 2003; Fig. 3). Additionally, volcanic tuffs interbedded with conodont-rich facies

have allowed for U-Pb dating of GSSP-equivalent strata, which yielded an absolute age of 298.90 ± 0.31 Ma (Ramezani et al., 2007).

Cyclic deposits are very common in carbonate strata of the Russian Platform. Many of these cycles have been interpreted as 3rd and 4th-order (high-frequency) in nature (e.g., Izart et al., 1999). Many are bounded by subaerial unconformities, including a widespread exposure event at the Sakmarian–Artinskian boundary (Ross and Ross, 1985; Vennin et al., 2002; Vennin, 2007; Fig. 2). Additionally, documented deeper water marl deposits overlying shallow water carbonate reef facies in upper Artinskian strata have been interpreted as a regional transgression and platform drowning (e.g., Vennin et al., 2002).

A2.8. Barents Shelf (Finnmark Platform); Svalbard

The present-day Barents Shelf region was situated 25–30° N during most of the Pennsylvanian (Golonka and Ford, 2000; Stemmerik, 2000; Fig. 1; Table 1). The region moved progressively northward to 35–45° N during the Late Pennsylvanian and Early Permian (Scotese and McKerrow, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000; Stemmerik, 2000). This northward shift led to dramatic climate change in the region, from subtropical warm and dry during the Gzhelian–early Sakmarian to cool temperate conditions during the late Sakmarian (Ehrenberg et al., 1998; Stemmerik, 2000; Rafaelsen et al., 2008; Stemmerik, 2008). On the Finnmark Platform, photozoan-dominated warm-water carbonates of the upper Gipsdalen Group (Ørn Formation) were deposited during the Gzhelian through mid-Sakmarian (Ehrenberg et al., 1998; Stemmerik, 2000; Rafaelsen et al., 2008; Stemmerik, 2008). The Gipsdalen Group is overlain by heterozoan-dominated cool-water carbonates of the Bjarmeland Group (e.g., Isbjørn

and Polarrev Formations; see Rafaelsen et al., 2008). Kungurian through Wuchiapingian strata are deeper water carbonates, including the chert-dominated Tempelfjorden Group (e.g., Røye Formation; see Rafaelsen et al., 2008).

The upper Palaeozoic stratigraphy on Svalbard is similar to that of the Finnmark Platform. The Upper Pennsylvanian–Lower Permian Ørn Formation on the Finnmark Platform is coeval with the Wordiekammen Formation and Finnlayfjellet Members of the Gipsdalen Group (Harland and Geddes, 1997; Rafaelsen et al., 2008; Stemmerik, 2008). The Bjarmeland Group from the Finnmark Platform is not recognized, partially due to an erosional unconformity that removed much of the Artinskian strata (Stemmerik and Worsley, 1995; Rafaelsen et al., 2008). Lower Bjarmeland Group equivalent strata on Svalbard are represented by the Gipshuken Formation of the upper Gipsdalen Group (Rafaelsen et al., 2008; Stemmerik, 2008). On Svalbard the Gipsdalen Group is unconformably overlain by the Kapp Starostin Formation of the Tempelfjorden Group (Heafford, 1988; Stemmerik and Worsley, 1995; Rafaelsen et al., 2008).

Age constraints for Pennsylvanian–Artinskian strata of the Barents Shelf region come primarily from fusulinid and conodont biostratigraphy (Nakrem, 1991; Harland and Geddes, 1997; Nilsson and Davydov, 1997; Anisimov et al., 1998; Stemmerik, 2000 and references therein; Fig. 3). Cooler conditions during the late Early Permian were not conducive to fusulinid communities, and the record disappears by the late Artinskian. As such, ages for upper Artinskian and younger strata are based on available conodont biostratigraphy (Nakrem, 1991). Lowermost Permian strata contain the fusulinid, *Sphaeroschwagerina vulgaris*, and the conodont, *Streptognathodus constrictus*, which suggests an early Asselian age (Nakrem, 1991; Nilsson and Davydov, 1998; Stemmerik, 2000 and references therein; Figs. 2 and 3). The

fusulinids *Schwagerina moelleri* and *Eoparafusulina paralinear* have been found in the upper Kapp Duner Formation (e.g., upper Wordiekammen), suggesting a Sakmarian age (Nilsson and Davydov, 1998; Stemmerik, 2000). The fusulinid, *Parafusulina jenkinsi*, and the conodonts *Neostreptognathodus pequopensis* and *Sweetognathus whitei*, are known from the Hambergfjellet Formation (e.g., Gipshuken Formation), which suggests an Artinskian age (Nakrem, 1991; Stemmerik, 2000 and references therein; Figs. 2 and 3). The Tempelfjorden Group (Fig. 2) contains the conodont, *Neostreptognathodus idahoensis*, suggesting a Kungurian age (Nakrem, 1991).

Cyclic deposits are common and 3rd, 4th, and 5th-order cycles have been inferred in both Upper Pennsylvanian and Lower Permian strata (e.g., Stemmerik, 2008). Shoaling upward cycles of warm-water carbonates are often capped by exposure surfaces (Ehrenberg et al., 1998; Rafaelsen et al., 2008; Stemmerik, 2008). There does not appear to be a major change in depositional style across the Pennsylvanian–Permian boundary. However, a major change in sedimentation occurs in the mid-Sakmarian, where a significant flooding event and loss of depositional cyclicity occurs (Stemmerik and Worsley, 1995; Stemmerik, 2008; Fig. 2).

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