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Palaeoclimate across the Late Pennsylvanian–Early Permian tropical palaeolatitudes: A review of climate indicators, their distribution, and relation to palaeophysiographic climate factors

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ABSTRACT

Global-scale compilations of palaeoclimate indicators include records of the temporal and spatial occurrence of coal, laterite, bauxite, Vertisols, calcrete, eolianite, and evaporite at the scale of geological stage. These palaeoclimate indicators provide the primary evidence for palaeoclimate change during the Late Palaeozoic, and have been used to infer a long-term climatic transition from humid to arid conditions on equatorial Pangaea from Late Pennsylvanian through Early Permian time. The cause(s) of Late Pennsylvanian–Early Permian climate trends are unknown but must have resulted from climate factors operating on timescales of tectonic change (10^6–10^7 yr), such as tectonic drift, assembly of Pangaea, orogenesis, and long-term carbon cycling.

Although higher-resolution, local- to regional-scale palaeoclimate reconstructions for the Late Pennsylvanian–Early Permian exist, they generally lack the time control necessary for accurate correlation among sites. Nevertheless, these high-resolution palaeoclimate reconstructions provide details about Permo-Pennsylvanian palaeoclimate that are not perceptible in lower-resolution global datasets. These studies indicate that (1) although the Late Pennsylvanian equatorial latitudes were more humid than the Early Permian tropics, there was also considerable variability in the amount and seasonality of rainfall, and (2) there were several short ($\ll$1–3 Ma) excursions toward relatively more humid climate during the long-term Early Permian transition to aridity in western and central equatorial Pangaea. These higher-resolution climate changes were controlled by climate factors which operated on relatively short timescales (10^4–10^6 yr) such as continental ice-sheet dynamics, sea-level change and associated changes in land–sea distribution, and variations in palaeoatmospheric PCO_2.

Although lithological indicators and geochemical proxies provide the basis for reconstructing past climate, they seldom provide diagnostic evidence to determine which of the possible climate factors were important. To narrow the possible causes of Late Pennsylvanian–Early Permian climate change, we review and evaluate both conceptual and numerical models that have been previously used to explain Late Palaeozoic climate change in light of the detailed spatial and temporal proxy records from across near-equatorial Pangaea. Our ability to test these models is currently limited by our inability to make accurate correlations among proxy sites due to uncertain dating. Nonetheless, we suggest that on tectonic timescales continental drift, increasing atmospheric PCO_2, and deglaciation could explain much of the low-latitude climate record, while changing atmospheric PCO_2 and orbitally-driven glacial–interglacial cycles could account for higher-resolution climate variability on Pangaea.

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1. Introduction

The Late Pennsylvanian–Early Permian was an interval of geological and climatological transition. This interval includes early tectonic post-assembly of the supercontinent Pangaea (Ziegler et al., 1979) which resulted in construction of a wide (~1000 km) and long (5000–7000 km) east–west oriented equatorial Central Pangaean Mountain (CPM) chain of poorly known elevation (Ziegler et al., 1997), both of which (Pangaean assembly and orogenesis) may have contributed toward reorganization of global atmospheric circulation systems into the so-called megamonsoon (Kutzbach and Gallimore, 1989; Dubiel et al., 1991; Parrish, 1993). This interval is also characterized by build-up and ablation of perhaps the most expansive continental ice sheets of

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the Phanerozoic (e.g., Frakes et al., 1992; Isbell et al., 2003). These ice sheets apparently waxed and waned, resulting in sea-level oscillation and large-scale changes in the distribution of land and sea (e.g., Heckel, 1977, 1986, 1990; Ramsbottom, 1979; Busch and Rollins, 1984; Ross and Ross, 1985; Veevers and Powell, 1987; Soreghan, 1994a; Yang, 1996; Rankey, 1997; Olszewski and Patzkowsky, 2003). In addition, geochemical evidence indicates large variations in the concentration of atmospheric CO_2 during Pennsylvanian and Early Permian time (Montañez et al., 2007). All of these climate factors were operative within an ~40–60 million year interval of northward tectonic drift that might also have affected the regional and temporal record of palaeoclimate indicators in tropical Pangaean basins.

The Late Pennsylvanian–Early Permian interval was also a period of tremendous low-latitude environmental change. Palaeoclimate indicators, including records of the temporal and spatial occurrence of coal, laterite, bauxite, Vertisols, calcrete, eolianite, evaporite, and flora, indicate that large regions of near-equatorial Pangaea experienced significant drying over this interval. For example, palaeoclimate indicators of humid tropical climates, such as coal, laterite, and bauxite are common in Pennsylvanian strata, but are virtually non-existent and replaced by indicators of dry climate, such as calcrete and evaporite, in Lower Permian strata of western and central Pangaea (e.g., Mack and James, 1986; Patzkowsy et al., 1991; Kessler et al., 2001; Gibbs et al., 2002; Tabor and Montañez, 2004; Schneider et al., 2006; Tabor et al., in press). The transition from relatively humid to arid climate was rapid in western equatorial Pangaea (Tabor and Montañez, 2004; Montañez et al., 2007; Tabor et al., in press), whereas the transition was protracted over much longer time-scales in central Pangaea (Schneider et al., 2006; Roscher and Schneider, 2006), and on the eastern tropical island blocks palaeoclimate indicators of humidity continued to be deposited through the Late Pennsylvanian and Early Permian (Ikonnikov, 1984; Gibbs et al., 2002; Rees et al., 2002; Yang et al., 2005).

How, and to what extent, did specific tectonic and global climate factors shape palaeoclimate in low-latitude Pangaea? To address this question, we review palaeoclimate indicators from the latest Pennsylvanian (Serpukhovian–Ghzelian) through Earliest Permian (Asselian–Cisuralian) low latitudes of Pangaea. From these palaeoclimate indicators, we develop a regional characterization of climate evolution at stage level, review detailed intrabasinal palaeoclimate reconstructions across low-latitude Pangaea, and evaluate this history in the context of previously proposed explanations for Late Palaeozoic climate change.

2. Background

2.1. Time frame

This contribution focuses upon evidence for atmospheric circulation and climate dynamics preserved in Upper Pennsylvanian (Serpukhovian–Gzhelian) and Lower Permian (Asselian–Cisuralian) terrestrial rocks (Fig. 1). Plate reconstructions (Fig. 2) and palaeoclimate indicators (Fig. 3) used herein are based on a compilation of palaeomagnetic data and palaeogeographic distributions of sedimentological and palaeontological climate indicators (evaporite, coal, eolianite, etc.; Scotese et al., 1999; Blakey, 2007; Boucot and Scotese, in press., see also http://www.scotese.com/climate.htm). These palaeogeographic maps indicate northward drift of Pangaea during Late Pennsylvanian and Early Permian time. It should be noted that Loope et al. (2004) indicate that western equatorial Pangaea remained essentially isolatitudinal

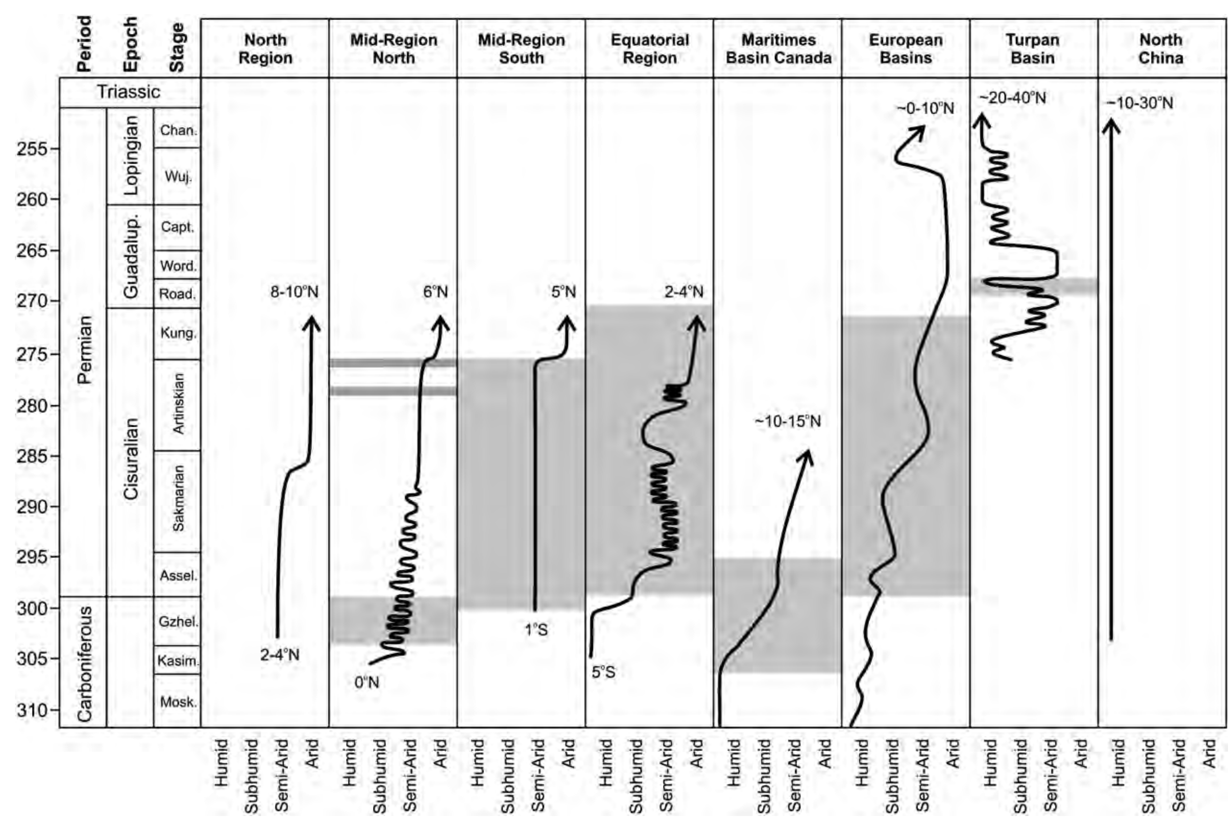


Fig. 1. International geological time-scale according to Gradstein et al. (2004) that is used throughout the text. Also shown are inferred temporal trends in climate based on stratigraphic changes in climate-sensitive lithotypes across tropical Pangaea. "Wet" indicates the presence of climate-sensitive lithotypes corresponding of humid conditions (>1000 mm/yr) and most months (~9) with precipitation in excess of evapotranspiration, including coal, laterite and bauxite. "Dry" indicates the presence of lithotypes corresponding to semi-arid and arid climates with precipitation less than ~760 mm/yr, and less than 5 months/yr with precipitation in excess of evapotranspiration (Cecil et al., 2003). The gray areas indicate stratigraphy with evidence of distinct seasonality, such as vertic palaeosol morphology (e.g., Tabor and Montañez, 2004) and fusain (Falcon-Lang, 2000; Falcon-Lang et al., 2006). The schematic climate curves are constructed from high-resolution, intrabasinal, stratigraphic studies of palaeoclimate indicators cited throughout the text, and references cited therein. See text for discussion.

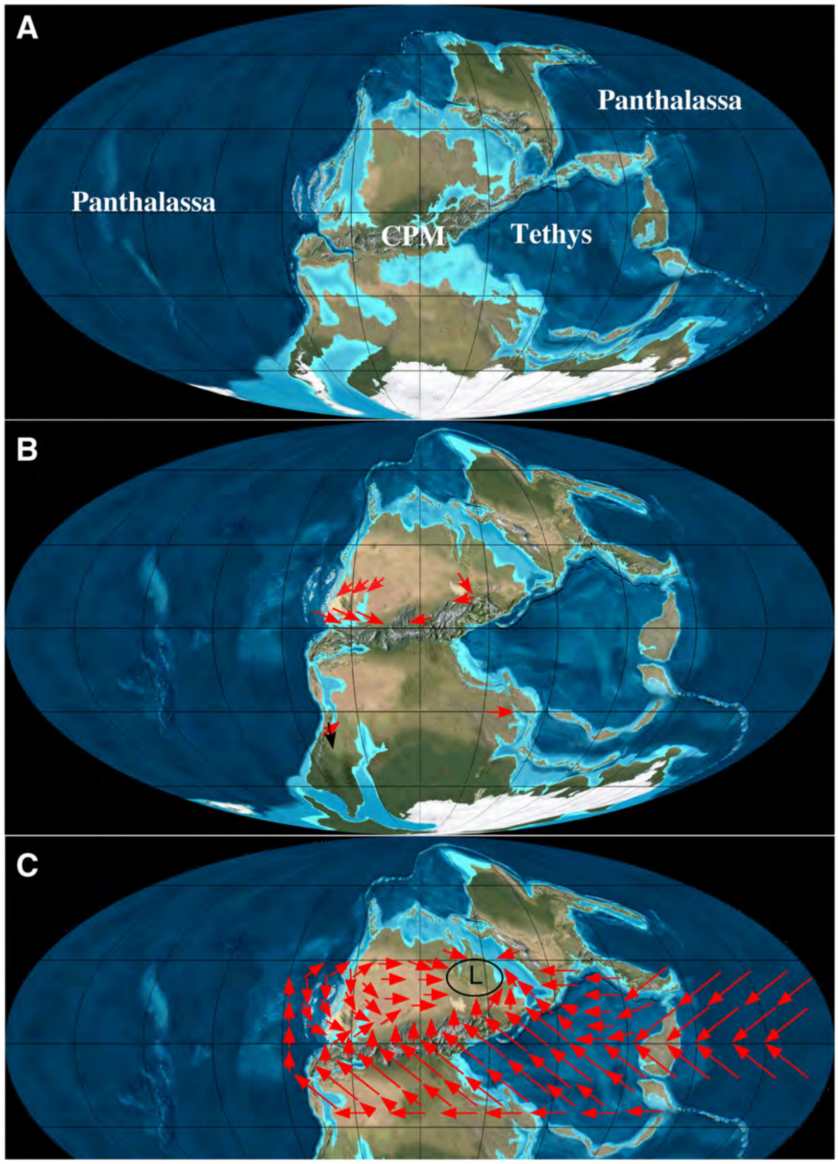


Fig. 2. Mollweide projections of Permo-Carboniferous palaeogeography from Blakey (2007). Dark blue indicates deep water, light blue indicates shallow platform, brown and green indicate exposed continental areas, white indicates continental ice sheet. (A) Projection of Gzhelian time (~300 Ma) showing the location of the Central Pangaean Mountain range as "CPM". (B) Projection of Artinskian time (~280 Ma) with approximate positions of generalized wind directions inferred from Permo-Carboniferous sandstones, wind-wave generated sediments, and chemical indicators of precipitation discussed in the text. See text for discussion. (C) Projection of Artinskian time (~280 Ma) with schematic depiction of hypothesized surface winds under monsoonal atmospheric circulation for boreal summer. Vector length does not correspond to wind speed, and is only meant to indicate generalized atmospheric circulation as determined by conceptual and numerical models discussed throughout the text. Pangaea migrated northward through this interval of time, and the tropical region of Pangaea, between 30°N and S, became progressively more continental from Carboniferous to Permian time as a result of withdrawal of epieric seas. Used here with permission from R. C. Blakey (2007). See text for further discussion. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

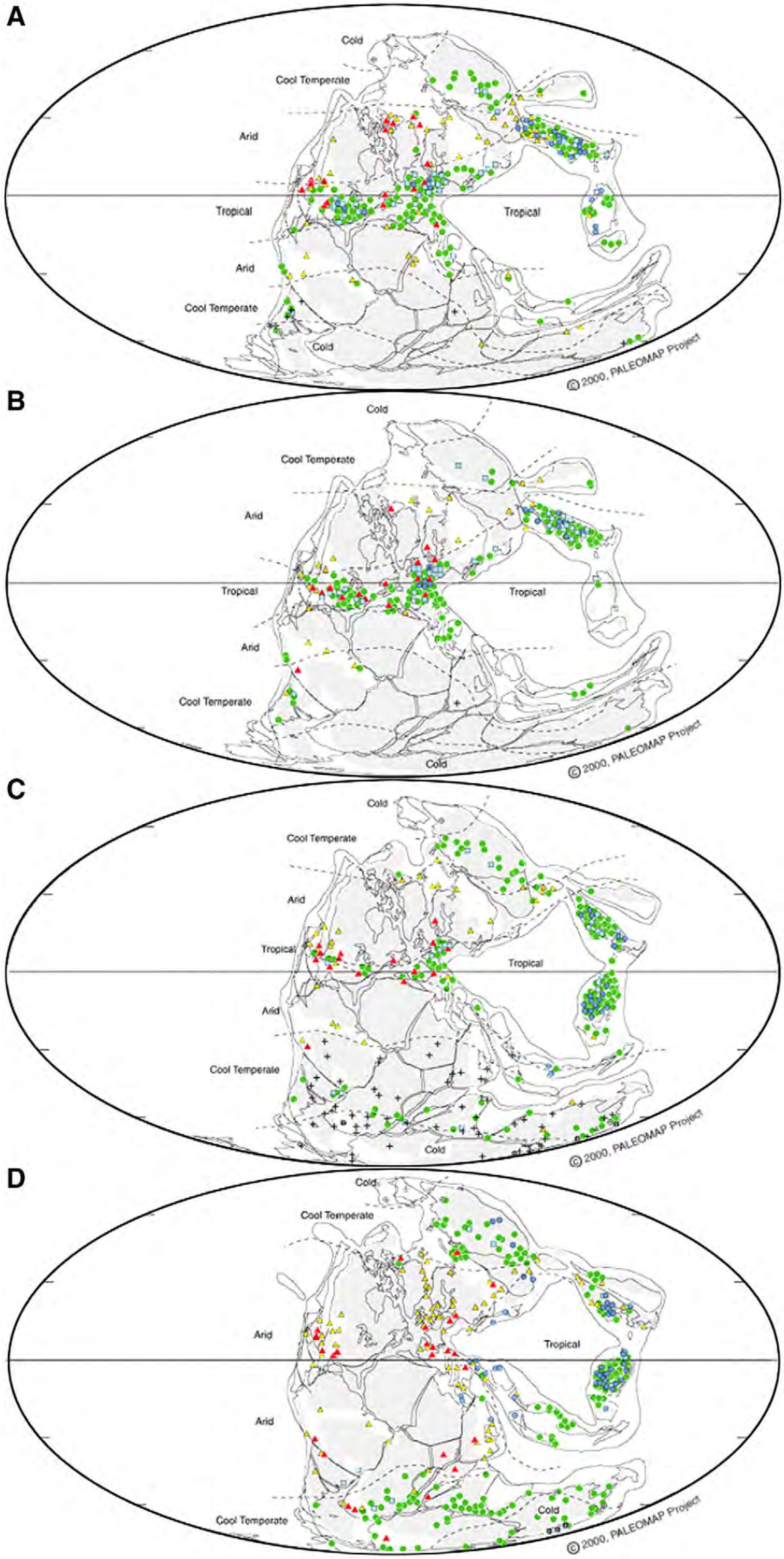
throughout Permo-Carboniferous time, while eastern Pangaea rotated northeastward.

Interbasinal correlation is difficult with terrestrial strata because they typically lack biostratigraphic or chronostratigraphic control needed for high-resolution correlation. However, the (1) Carboniferous Serpukhovian, Kasimovian and Gzhelian (~Virgilian), as well as Early Permian (2) Asselian, Sakmarian, and Artinskian (~Wolfcampian), and (3) Kungurian (~Leonardian) stage boundaries are reasonably well known although finer-scale, interbasinal, intra-stage correlations are not realistic at this time (cf. Lucas, 2006; Schneider et al., 2006; Montañez et al., 2007).

2.2. Late Palaeozoic palaeogeography

Our current understanding of Earth's continental configuration during this time is summarized elsewhere (Scotese and Langford, 1995; Ziegler et al., 1997; Fluteau et al., 2001; Blakey, 2007) and only the most salient features are discussed here. Gondwanaland and Laurasia collided during mid-Carboniferous time (~320 Ma; Scotese and Barret, 1990; Ziegler et al., 1997), resulting in a large horseshoe-shaped landmass, the supercontinent Pangaea, that was roughly divided by the palaeo-equator (Fig. 2). With the exception of a few large islands in the Tethys Ocean, all of Earth's land resided within the Pangaean supercontinent by Pennsylvanian time, and several of the island blocks "docked" against the northeast Tethyan coast of Pangaea during later Permian and Triassic time (e.g., Scotese and Golonka, 1992). Pangaea began with ~68% of its landmass in the Southern Hemisphere during Late Pennsylvanian time, and migrated toward the Northern Hemisphere such that ~62% of its landmass remained in the Southern Hemisphere by Late Permian time (Parrish, 1993).

Earth's oceans were divided into two primary realms during Pennsylvanian and Permian time: (1) the global-scale Panthalassa


A
Cold
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Arid
Tropical
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Ocean and (2) the Tethys Ocean that partially divided northern and southern Pangaea along equatorial- and mid-latitudes (Fig. 2). Numerous epieric seas inundated areas of the Pangaean continent (Fig. 2). Pennsylvanian inland seas occupied large areas of the low-latitude tropics and occupied progressively less surface area of the Pangaean landmass through the Permian (Fig. 2; Table 1).

Accretion of Gondwanaland and Laurasia closed the pre-existing equatorial seaway, and in conjunction with an active subduction zone along the north coast of Tethys, generated a broad (as much as 1000 km wide) and long (5000–7000 km), east–west oriented, equatorial mountain chain and plateau (Central Pangaean Mountains; CPM; Fig. 2A) including the Marathon, Ouachita and Appalachian (U.S.A.); Mauretanide and Atlas (North Africa); and Hercynian (Southern Europe) orogenic belts (Ziegler et al., 1997; Fig. 2). Although the elevations of Permo-Pennsylvanian mountain ranges are largely unknown, the CPM are suspected to have been among the highest ranges of their time, with local palaeo-elevation estimates ranging from <2000 m to ~4500 m (Becq-Giraudon et al., 1996; Ziegler et al., 1997) for the highest portions of the Appalachian and Hercynian orogens. Additional inferred high-altitude (~1000–2000 m; Ziegler et al., 1997) tropical ranges on Late Palaeozoic Pangaea include the proto-Andean mountains of South America (e.g., Sierra Pampeanas and Puna Arch; Lopez-Gamundi, 1997; Coughlin et al., 1998; Haeberlin et al., 2003); the Ancestral Rocky Mountains (ARM; western U.S.A.; Kluth and Coney, 1981; Soreghan et al., 2002a,b, 2007a), Grenville Mountains (east Canada), and Eskimonia Highlands (Greenland; Haller, 1971; Surlyk, 1990) of North America; the Oslo Dome and Scandes Mountains (Norway, Sweden) of northern Europe; the Urals, Kyzylkum uplands and Kaprinsky Swell of eastern Europe (Vinogradov, 1969; Nagornyy and Nagornyy, 1976; Ziegler et al., 1997); and several small ranges in central and North Africa (e.g., Kogbe, 1991; Guiraud et al., 2005).

2.3. Lithological records of Permo-Pennsylvanian surface winds

Although their ages are poorly constrained, the cross-bedding dip-directions of eolian sandstones have been considered the most direct estimates of Permo-Pennsylvanian surface palaeowind directions (See Petersen, 1988; Parrish and Peterson, 1988; Parrish, 1998; Gibbs et al., 2002 for explanations and examples). These deposits are typically limited to regions with actively migrating dunes. On modern Earth such conditions are primarily limited to the dry (<30 mm/yr precipitation), high-pressure subtropical belts near ~25°–35°N and S with sustained wind speeds >6 m/s (Fryberger, 1979; Fryberger and Ahlbrandt, 1979).

Dip vectors for the majority of cross cross-bedded eolian sandstones from Permo-Pennsylvanian strata in the southwestern U.S.A. indicate that surface winds flowed from palaeo-northeast to palaeo-southwest at latitudes ranging from ~8°–15°N, whereas lower-latitude sites from ~5°–10°N indicate palaeo-northwesterly to palaeo-westerly winds over western equatorial Pangaea (Fig. 2B; Petersen, 1988). These results have been corroborated by other lithological and geochemical studies (Kim and Loope, 1995; Tabor and Montañez, 2002; Tabor et al., 2002; Soreghan et al., 2002b, 2007b; Soreghan et al., 2007b; Soreghan and Soreghan, 2007).

Dip vectors of crossbeds from Lower Permian strata in Nova Scotia indicate northeasterly winds in central equatorial Pangaea (~0°; Brisebois, 1981). Middle and Upper Permian eolian strata from NW Europe (Germany, Britain), several hundred kilometers inboard of the Tethyan coastline, indicate northeasterly to southwesterly winds between ~15°–25°N (Glennie, 1983; See also Pochat et al., 2005). Examples of cross-bedded eolian strata in the southern hemisphere (Limarino and Spalletti, 1986; Rees et al., 2002; Ziegler et al., 2003) are extratropical and not discussed further here. Wind-transport directions preserved within the geological record provide information that represents only a small portion of the Permo-Pennsylvanian Earth surface, and significant time averaging must be made among eolian deposits in order to infer regional wind directions (e.g., Parrish and Peterson, 1988). Yet, they constitute an important data set to guide conceptual models and compare with numerical palaeoclimate models.

Table 1
Percentages of total exposed land and sea area between 30°N and S during Permo-Carboniferous time

Time (Ma)	% Land	% Sea
260	22	78
280	22	78
300	16	84

Note. Measurements were taken with a planimeter from Mollweide equal-area projections of palaeogeography by Blakey (2007).

3. Palaeoclimate indicators

3.1. General long term trends

Several different global-scale compilations of climate-sensitive sedimentary lithotypes, mineralogical assemblages, and fossil plant types have been used to reconstruct palaeoenvironmental and palaeoclimatic conditions across Pangaea, and to infer the evolution of palaeoatmospheric circulation, from their spatial and temporal distribution (Nairn and Smithwick, 1976; Parrish et al., 1982; Phillips and Peppers, 1984; Rowley et al., 1985; Ziegler et al., 1987; 2003; Scotese and Barret, 1990; Parrish, 1998; Scotese et al., 1999; Boucot and Scotese, in press). In addition to eolianite (discussed above), these include evaporite, tillite, and climate-sensitive palaeosol morphologies including gypcrete, calcrete, Vertisols, laterite, bauxite, and coal (Table 2). The utility of these lithotypes as palaeoclimate indicators is based upon the observation that such sediments appear to form in a relatively narrow range of climate conditions on modern Earth (Table 2; see also Parrish, 1998; Ziegler et al., 2003). As a result of these climate associations, the distribution of these lithotypes through time and space provides the primary geological evidence for long-term patterns of palaeoclimate and atmospheric circulation. The climate association for each lithology is described, and the Late Pennsylvanian–Early Permian spatial and stratigraphic distribution (Fig. 3) of these lithotypes is reviewed below.

3.1.1. Tillites and other indicators of glaciation

Tillites are polymictites that are deposited directly by glaciers and simply indicate the presence of ice, with no specific information about palaeotemperature (Eyles, 1993). The Pennsylvanian and Permian has have traditionally been associated with continental glaciation (e.g., Frakes et al., 1992). However, Isbell et al. (2003) suggest the distribution of glacial and glacigenic sediments during Late Palaeozoic time may be organized into three discrete and non-overlapping periods of continental ice on southern Gondwanaland. Glacial phases I and II of Isbell et al. (2003) occur in Frasnian–Tournaisian and Bashkirian to lower Moscovian rocks respectively, and are not considered further in this work. Glacial phase III occurs in upper

Fig. 3. Mollweide projections of Permo-Carboniferous palaeogeography with climate-sensitive terrestrial lithotypes projected upon them. Green, blue and "bull's eye" circles indicate coal, bauxite and laterite, respectively, whereas yellow and red triangles indicate evaporites and calcretes, respectively. Blue squares indicate kaolinite deposits, and symbols with crosses indicate glacially derived sediments, neither of which is discussed here. (A) Bashkirian–Moscovian time (318.1–306.5 Ma: Gradstein et al., 2004), (B) Gzhelian (303.9–299.0 Ma; Gradstein et al., 2004), (C) Lower Permian (Cisuralian, 299.0–270.6 Ma; Gradstein et al., 2004), and (D) Middle and Upper Permian (Guadalupian–Lopingian, 270.6–251.0 Ma; Gradstein et al., 2004). See text for discussion. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 2
Palaeoclimate indicators used in this manuscript

Climate indicator	MAT	P>ET	ET>P	MAP (mm)	Reference/comment
Tillite					
	Frigid				Presence of Ice —Parrish (1998), Eyles (1993)
Coal (peat)					
				≥20/ month	Ziegler et al. (1987)
				>1000	Patzkowsy et al. (1991)
		6–12			Cecil et al. (2003)
Laterite					
	>25 °C	11 m/yr	–		Gordon and Tracey (1952), Nicholas and Bildgen (1979)
Bauxite					
	>22 °C	9–11 m/yr	–	>1200	Bárdossy (1993)
	>25 °C	–	–	>1800	Tardy et al. (1990)
Calcrete					
	–		>6 months		Buol et al. (1997)
				<760	Royer (1999) — modern soils
				<1400	Nordt et al. (2006) — modern Vertisols
				>140	Retallack (1994, 2005a,b) — modern soils
				>305	Jenny (1941) — modern soils
Eolianites					
			≥8	<400	Cecil et al. (2003)
Evaporite					
			12	<400	Patzkowsky et al. (1991) — marine evaporites
				<300	Watson (1992) — soil evaporites

Stephanian through Sakmarian strata, and it likely represents the largest and most widespread period of Late Palaeozoic continental glaciation (Isbell et al., 2003; Montañez et al., 2007; Fielding et al., 2008). Although the main phase of Glacial III had ended by Sakmarian–Artinskian time, there is evidence for smaller-scale, discrete periods of continental glaciation and/or climatic cooling in Kungurian through Wordian-age sedimentary strata in eastern Australia (Jones and Fielding, 2004) and Antarctica (Isbell et al., 2003; Montañez et al., 2007).

Regions with evidence of contact with, or near-proximity to, continental ice during glacial phase III include the Paraná basin in Brazil, Paraguay and Uruguay (Eyles et al., 1993; Lopez-Gamundi, 1997) and the adjacent Karoo, Kalahari and Congo basins in South Africa, Namibia, Botswana (Veevers, 1994; Visser, 1997) and the Democratic Republic of Congo (Kar and Bose, 1978) as well as the Transantarctic basin of Antarctica (Miller, 1989; Isbell et al., 1997, 2001, 2003), Officer and Canning basins of western Australia (Veevers, 1984; Eyles and Eyles, 2000), several different basins within Peninsular India (Veevers and Twari, 1995; Wopfner and Casshyap, 1997) and the Arabian plate (Ziegler et al., 1997). Collectively, these data suggest continental ice may have extended from near the southern pole in Antarctica to ~35°S in South America and Africa during Glacial phase III (e.g., Boucot and Scotese, in press; Fig. 3). Estimates of base-level change associated with ice-sheet dynamics range from a few meters to >200 m (e.g., Wilson, 1967; Heckel, 1977; Ross and Ross, 1985, 1987; Goldstein, 1988; Maynard and Leeder, 1992; Soreghan and Giles, 1999; Mack, 2007).

There is no evidence for Late Pennsylvanian–Early Permian continental glaciation on the Pangaean Northern Hemisphere, but lithological evidence for alpine glaciation has been reported from two different sites in the palaeo-tropics. Becq-Giraudon et al. (1996) presented several different types of deposits from the Permian Lodeve basin, France, that were interpreted to have been deposited in an alpine pro-glacial environment very near the palaeo-equator. Assuming a similar climate to modern Earth, Becq-Giraudon et al. (1996) estimated a palaeo-elevation of ~4500 m for the Lodeve basin. Schneider et al. (2006) disputed the glacial interpretation of these deposits and the palaeo-elevation estimates provided by Becq-Giraudon et al. (1996), and argue for a non-glacial origin and low-elevation deposition of these strata.

Strata of the Unaweep Canyon-Paradox basin system, Colorado (U.S.A.) have been interpreted to include both glacial and pro-glacial deposits within and adjacent to an exhumed tropical (~0°N) alpine glacial valley that was cut into the Ancestral Rocky Mountains in the Late Pennsylvanian–Early Permian (Soreghan et al., 2002b, 2007b). However, the palaeoaltitude of these deposits within the ARM during Permo-Pennsylvanian time appears to be rather low; lower than comparable Quaternary examples (Soreghan et al., 2002b, 2007b). If these Pennsylvanian–Permian strata are in fact glacially derived, they suggest a much cooler Late Palaeozoic climate than has previously been considered. It is noteworthy in this regard that stable oxygen and hydrogen isotope data from low-latitude (~5°S), low-elevation Upper Pennsylvanian palaeosol profiles indicate mean annual temperatures of 19±3 °C (Tabor and Yap, 2005; Tabor, 2007), which is ~6 °C cooler than modern low-altitude (<500 m) equatorial sites (e.g., Rozanski et al., 1993).

3.1.2. Coal

Coal is derived from accumulations of peat in environments characterized by high-productivity floras and reduced oxidation of vegetation by aerobic bacteria. Nearly all peat accumulates in humid basinal settings characterized by a high water table, and precipitation is usually >100 cm/yr and exceeds evapotranspiration 10–11 months of the year (Cecil et al., 1985, 2003; Ziegler et al., 1987, 2003; Ziegler, 1990; Patzkowsy et al., 1991; Table 2). Pennsylvanian coals have been reported from most terrestrial equatorial basins between palaeolatitudes from ~12°S and 10°N on the Pangaean continent, as well as many basins along the northern and southern Tethyan coasts and Tethyan island blocks at palaeolatitudes up to ~30–40°N and S (Ronov, 1976; Ziegler et al, 1979; Rowley et al., 1985; Fig. 3). With only a few Early Permian exceptions in the U.S.A. and western Europe (Schowe, 1951; Kehn et al., 1982; Martins, 1998; William DiMichele, pers. comm.), coal disappears from Permian equatorial landscapes in Euramerica (North America, Europe and Central Asia; Fig. 3). However, peat deposition continued along the Tethyan coastlines and on the Tethyan island blocks until end-Permian time (e.g., Sun et al., 2002; Ziegler et al., 2003; Dai, 2006; Yang et al., 2007).

3.1.3. Laterite and bauxite

Laterite and bauxite deposits are residual rocks derived from intense chemical weathering at Earth's surface (Bárdossy, 1993). *Laterite* is composed of nearly equal amounts of kaolinite, iron- and aluminum oxyhydroxides, whereas *bauxite* is dominantly composed of aluminum oxyhydroxides (Bárdossy and Aleva, 1990). Modern laterite has been ascribed to climates characterized by precipitation in excess of evapotranspiration ≥11 months/yr and mean annual temperature (MAT; Table 2) >25 °C, whereas bauxite has been ascribed to climates with >1200 mm/yr precipitation, >22 °C MAT, and 9–11 months where precipitation exceeds evapotranspiration (e.g., Bárdossy, 1993).

Neither laterite nor bauxite has been reported from Pennsylvanian rocks in western equatorial Pangaea (Fig. 3A, B), but they do occur in eastern U.S.A. and western Europe, which indicates a narrow belt of intense weathering, and hot and humid climate along palaeo-equatorial latitudes in central Pangaea (e.g., Bárdossy, 1993; Boucot and Scotese, in press). With the possible exception of a Sakmarian laterite in the Sydney basin, Australia (Loughnan, 1975), laterite and bauxite do not occur in Lower Permian rocks from Pangaea (Bárdossy, 1993). However, Tethys Ocean island blocks preserve abundant Pennsylvanian and Permian laterite and bauxite (Ikonnikov, 1984; Pu and Bingwen, 1986; Liacheng and Naixian, 1991; Bárdossy, 1993) that spanned from ~15°S to 30°N palaeolatitude (Fig. 3), indicating continuous humid and hot climate in that region.

3.1.4. Calcrete

Calcrete deposits are continental carbonate-rich layers formed in soils (e.g., Machette, 1985) or from groundwater (Semeniuk, 1981; 1985). Modern calcretes generally form in sub-humid to semi-arid climates where evapotranspiration exceeds precipitation in most months of the year, and monthly precipitation never exceeds evapotranspiration by more than a few tens of mm (Table 2; Buol et al., 1997; Cecil et al., 2003; Alonso-Zarza, 2003). Soil calcretes are particularly common in regions with low mean annual precipitation such as the dry belts of high atmospheric pressure beneath the descending limb of the tropical Hadley cell, near ~20–35°N and S latitude. However, calcretes may occur nearly anywhere in the interiors of large continents and rain shadows (i.e., arid zones) or in sedimentary materials dominated by detrital carbonate such as the Pleistocene-age glacial tills of northern Minnesota (Soil Survey Staff, 1996).

Pennsylvanian calcretes occur in northern areas of western equatorial Pangaea, including the Colorado Plateau region (Goldhammer and Elmore, 1984; Goldstein, 1991; Kenny and Neet, 1993; Ekart et al., 1999; Tabor and Montañez, 2002; Tabor et al., in press) and U.S. Midcontinent (Prather, 1985; Rankey and Farr, 1997; Joeckel, 1999), as well as the Cumberland Basin of Nova Scotia (Tandon and Gibling, 1994; Falcon-Lang, 2004), Oslo trough of Norway (Olaussen et al., 1994), and Sverdrup basin of Ellesmere Island (Theriault and Desrochers, 1993). Calcrete is an important lithology in Lower Permian rocks of the western U.S. (Mack et al., 1991; Kessler et al., 2001; Tabor and Montañez, 2004) western Europe (Steel, 1974; Peryt, 1987, Roscher and Schneider, 2006) and northern Africa (Tabor et al., 2007), suggesting that west and central equatorial Pangaea became drier during the Permian. Calcretes also occur in the upper Lower Permian and Middle Permian rocks of the Junggar basin, Kazahkstan plate, northwest China (Yang et al., 2007), suggesting that drier climates did exist, at least intermittently, in parts of the northwestern coastlines of the Tethys Ocean (Fig. 1).

3.1.5. Evaporite

Evaporite rocks (e.g., gypsum, halite, borates, potash) form in littoral, lacustrine, sabkha, or soil-forming environments (Kendall, 1992). Most modern marine gypsum deposits form in climates limited to regions with precipitation <40 cm/yr, whereas soil-derived evaporite minerals are limited to climates with mean annual precipitation <30 cm/yr (Table 2). The regional climate where conditions are most amenable to evaporite deposition is in the high-pressure subtropical belts associated with the descending limb of the tropical Hadley Cell, between ~20°–35°N and S.

Pennsylvanian marine evaporites occur in western equatorial Pangaea in the Paradox basin (~9°N), northern (~25°–35°N) subtropical latitudes of the Canadian and Greenland archipelago (Stemmerik et al., 1999) and Russian Platform, southern subtropical latitudes of the Amazon basin (Brazil) and peri-Tethyan platforms in Chad and Libya (Fig. 3B). Early Permian evaporite systems were more extensive, extending from ~0° to 20°N in western Pangaea (Texas, New Mexico, Colorado, Wyoming and Idaho), ~25°–35°N along the Uralian seaway, and ~20–30°S in the Amazon basin of South America. Pennsylvanian evaporite-bearing palaeosols have not been reported, although Lower Permian examples from Kansas (McCahon and Miller, 1997), New Mexico (Mack, 2003), Texas and Utah (Tabor et al., in press) indicate development of arid climate in western Pangaea during the Early Permian.

3.1.6. Milankovitch-scale climate oscillations

Meter-scale stacking patterns of Upper Pennsylvanian and Lower Permian marine and mixed marine and terrestrial strata of Euramerica have been attributed (in large part) to periodic glacio-eustatic adjustments of base level (e.g., Wanless and Shepard, 1936; Heckel, 1977, 1986, 1990; Crowell, 1978; Ramsbottom, 1979; Ross and Ross, 1985; Veevers and Powell, 1987; Boardman and Heckel, 1989; West et al., 1993; Soreghan, 1994a; Read, 1995; Yang, 1996; Yang et al., 1998; Soreghan and Giles, 1999, Olszewski and Patzkowsky, 2003, in press). The glacio-eustatic control on magnitude of base-level change, and duration of time represented by each stratigraphic cycle, is interpreted to be Milankovitch-type orbital variations, which periodically changed the distribution of incident solar radiation upon the Late Pennsylvanian and Early Permian Earth, and affected global surface temperatures, continental ice volume, and base level (Perlmutter and Matthews, 1989; Connolly and Stanton, 1992; Read, 1995).

These so-called cyclothems not only furnish a far-field indication of high-latitude climate (i.e., status of continental ice in Gondwanaland) and continental ice-sheet dynamics, but also provide their own suite (s) of regional tropical palaeoclimate indicators that are suggestive of periodic (Milankovitch-scale) climate change. However, the interpretation of regional low-latitude palaeoclimate within the context of high-latitude palaeoclimate (i.e., interglacial = highstand, glacial = lowstand) differs significantly among researchers. For example, some studies interpret strata deposited during base-level lowstands (glacial; cold high-latitudes) to correspond to more arid regional climate (Soreghan, 1994a, Tandon and Gibling, 1994; Yang, 1996; Rankey, 1997), whereas others interpret lowstand strata to correspond to more humid climate (Cecil, 1990; West et al., 1993; Miller et al., 1996, Cecil et al., 2003). Still other studies interpret transgressions as the driest parts of climate cycles (McCahon and Miller, 1997; Olszewski and Patzkowsky, 2003, in press).

3.2. Evidence for seasonal moisture variations

Presence of growth rings in fossil wood (e.g., Chaloner and Creber, 1990) and fusain (fossilized burnt wood; Falcon-Lang, 2000), as well as palaeosol morphologies similar to modern Vertisols (e.g., Birkeland, 1999), provides evidence for seasonal variations in moisture availability and important information about palaeoprecipitation patterns and palaeoatmospheric circulation. Shrink-swell processes dominate in Vertisols as a result of episodic dryness ranging from as short as a few weeks to several years (Buol et al., 1997). Upper Pennsylvanian Vertisols occur in west and central equatorial Pangaea, indicating at least intermittent seasonal moisture deficiency (Goebel et al., 1989; Tandon and Gibling, 1994; Rankey and Farr, 1997; Joeckel, 1999; Tibert and Gibling, 1999; Tabor and Montañez, 2002, 2004; Cecil et al., 2003; Falcon-Lang, 2004). Lower Permian Vertisols are a major stratigraphic component of tropical basins in west and central equatorial Pangaea, as well as the western coasts of Tethys (Fig. 1; Mack et al., 1991; Miller et al., 1996; Kessler et al., 2001; Mack and Dinterman, 2002; Tabor and Montañez, 2002, 2004; Schneider et al., 2006; Yang et al., 2007), suggesting more pronounced seasonal precipitation patterns, over a more vast area, in the Permian tropics than the preceding Pennsylvanian tropics.

The relatively few growth-ring studies of Pennsylvanian and Permian fossil trees from equatorial latitudes (Creber and Chaloner, 1984; Chaloner and Creber, 1990; Tidwell and Munzing, 1995) suggest warm, humid and equable climate with little or no dry season. These biologically-based interpretations of palaeoclimate are generally consistent with lithological indicators of humidity across Late Pennsylvanian lowlands, but *inconsistent* with lithologic indicators of seasonal moisture deficit (Vertisols) in Upper Pennsylvanian and Lower Permian (calcrete, evaporite) rocks. These opposing indications of palaeoclimate might reflect that Upper Palaeozoic floras are biased toward more humid conditions than regional climate (wetland and "wet spot" floras; Mack, 2003; DiMichele et al., 2006; see also DiMichele et al., 2008-this volume). In addition, fossil charcoal (fusain) has been identified from Pennsylvanian-age equatorial

Pangaean upland deposits, suggesting that intermittent dry episodes affected some floras that are otherwise indicative of wet conditions (Falcon-Lang, 2000; Falcon-Lang et al., 2006).

3.3. Synopsis of Late Pennsylvanian–Early Permian climate patterns

Several important Permo-Pennsylvanian tropical climate trends may be inferred from high-resolution intrabasinal palaeoclimate reconstructions (Fig. 1) and long-term regional stratigraphic patterns of climate-sensitive lithotypes across equatorial Pangaea (Fig. 3): (1) Northwestern equatorial Pangaea was persistently dry through Late Pennsylvanian–Early Permian time (Mack, 2003; Tabor and Montañez, 2002; Tabor et al., in press; however, see point (3) below). (2) Western and central equatorial Pangaea experienced a long-term (~10–20 Ma) drying trend (Robinson, 1973; Mack and James, 1986; Parrish, 1993; West et al., 1993; Kessler et al., 2001; Mack, 2003; Tabor and Montañez, 2004); aridification was rapid in the west, occurring in <2 Ma near the Permo-Pennsylvanian boundary (Montañez et al., 2007; Tabor et al., in press), whereas drying was more gradual (~20 Ma) in central Pangaea (Ziegler et al., 2003; Roscher and Schneider, 2006; Fig. 1). (3) Shorter-duration (<1 to 5 Ma) stratigraphic trends in North American and European basins suggest ~4 to 6 climate reversals toward more humid conditions during the long-term drying trend. These so-called "pluvial" events, or relatively humid intervals, may also be present in other portions of Pangaea, such as Permian fluvial–eolian deposits of the Colorado Plateau (Blakey and Middleton, 1983; Loope, 1985; Blakey, 1990; 1996; Soreghan et al., 1997; Retallack, 2005a; Mountney, 2006), Permian cyclothems of the U.S. midcontinent (West et al., 1993; Calder, 1994; Tandon and Gibling, 1994; Miller et al., 1996) and Permian eolian strata of the Pennines basin (Clemmensen, 1991; Frederiksen et al., 1998). It is also possible that perceived pluvial events in fluvio-eolian strata might correspond to changes in the position of the groundwater table, and concomitant landscape stability, in the absence of any significant regional palaeoclimatic or palaeoatmospheric change (Besly and Fielding, 1989; Parrish, 1993, 1998; Tabor et al., in press). If these pluvials represent widespread and contemporaneous regional climate change, they may be ultimately useful for interbasinal correlation of terrestrial strata. At this time, however, the chronology of the strata is too poorly known, and the correlation too coarse across tropical Pangaean basins to test such a hypothesis. (4) Distinct seasonality, either in precipitation or moisture availability (or both), began in western and central equatorial Pangaea during Late Pennsylvanian time (Cecil, 1990; Mack et al., 1991; West et al., 1993; Soreghan, 1994a; Yang, 1996; Rankey and Farr, 1997; Kessler et al., 2001; Tabor and Montañez, 2004; Tabor et al., in press), and continued into Early Permian time (Fig. 1). (5) Eastern Pangaea (Kazahstan Plate, Turpan basin) experienced climatic oscillations between "semi-arid" and "humid" episodes, nonetheless this region remained more humid than any contemporaneous sites in western or central Pangaea, and pronounced seasonal climate is evident for only a brief period in Middle Permian (Roadian) time (Yang et al., 2007). (6) The Tethyan island blocks remained humid, with little or no evidence of seasonality, through the Late Pennsylvanian and Early Permian (Ikonnikov, 1984; Liacheng and Naixian, 1991; Rees et al., 2002; Ziegler et al., 2003).

4. Factors of tropical climate variation

What factors controlled tropical climate in equatorial Pangaea? To answer this, we can draw upon nearly thirty years of scientific investigation into the climate of Pangaea (e.g., Parrish, 1982, 1993, 1998; Parrish and Peterson, 1988; Kutzbach and Gallimore, 1989; Witzke, 1990; Ziegler, 1990; Dubiel et al., 1991; Crowley and Baum, 1992; Kutzbach et al., 1993; Crowley et al., 1993; Otto-Bliesner, 1993, 1996, 1998, 2003; Kutzbach, 1994; Crowley, 1994; Hyde et al., 1999; 2006; Kent and Olsen, 2000; Olsen and Kent, 2000; Fluteau et al., 2001; Gibbs et al., 2002; Rees et al., 2002; Ziegler et al., 2003; Kiehl and Shields, 2005; Horton et al., 2007). Although only a few palaeoclimate models have focused specifically on the Late Pennsylvanian–Early Permian transition (Poulsen et al., 2007; Peyser and Poulsen, 2008-this volume), this large body of Pangaean palaeoclimate studies provides general insights into the major controls on Pangaean climate. We present below seven climate factors (summarized in Table 4) that may have had a substantial impact on Late Pennsylvanian–Early Permian tropical climate, summarize the likely influence of each climate factor on low-latitude Pangaea, and evaluate whether these changes are consistent with climate trends inferred from palaeoclimate indicators.

4.1. Climate Factor 1: Tectonic drift

Pennsylvanian through Early Permian plate tectonic reconstructions used here indicate that most of Pangaea was moving northward between 320 and 280 Ma (e.g., Table 3, Figs. 2 and 3). At least some portions of eastern Pangaea had a very complex history of northward and southward movement (see for example Urumiqi, China in Table 3). However, western and central Pangaea moved progressively northward through ~10°–14° of latitude (Table 3). Based on this, several studies attribute long-term tropical aridification to migration of tropical basins from humid climate zones associated with the Intertropical Convergence Zone into an adjacent northern subtropical zone (Witzke, 1990; Ziegler et al., 1977; Gibbs et al., 2002; Rees et al., 2002; Tabor et al., in press). Considering modern climate patterns, an ~14° northward movement from the latitude circle for peak equatorial rainfall (~6°N) results in precipitation change from ~2000 mm/yr to ~840 mm/yr (e.g., Barron and Moore, 1994). Thus, it is possible that much of the Pennsylvanian–Permian record of palaeoclimatic change is related to tectonic drift through climate zones. However, Loope et al. (2004) suggested, based on previous palaeomagnetic studies of *Mesozoic* strata from the Colorado Plateau, U.S.A. (Steiner, 1983, 2003; May and Butler; 1986; Bazard and Butler, 1991; Steiner and Lucas, 2000), that western equatorial Pangaea remained stable through Permian time, and did not move northward until Late Triassic time. If west and central Pangaea remained stable through the Late Palaeozoic, then continental drift is not a meaningful climate factor.

In addition, continental drift does not explain several aspects of the palaeoclimate record. Northward continental drift does not account for: (1) regional differences in the onset of aridification or the rapid change from humid to arid conditions in western equatorial Pangaea (Tabor and Montañez, 2002, 2004, 2005; Mack, 2003; Tabor 2007; Tabor et al., in press); (2) significant, relatively short-term (10^4 to

Table 3
Estimated Pennsylvanian and Permian palaeolatitudes of major cities that are upon or near Pennsylvanian and Permian basins, as well as resulting distance of drift, in degrees of latitude, between 320 and 280 Ma

Place	Flagstaff, AZ	Denver, CO	Wichita Falls, TX	Washington, D.C.	Halifax, N.S.	London, England	Montpellier, FR	Prague, Czech	Agadez, Niger	Urumiqi, China
280 Ma	8.8°	9.5°	1.7°	−0.5°	5.1°	9.8°	2.6°	10.4°	−22.4°	37.6°
300 Ma	2.9°	3.1°	−4.7°	−7.7°	−2.2°	2.7°	−4.3°	3.8°	−30.1°	40.6°
320 Ma	−1.1°	−1.4°	−9.2°	−14.2°	−7.6°	−3.0°	−9.8°	−1.7°	−35.7°	34.3°
° of drift	9.9 N	10.9 N	10.9 N	13.7 N	12.7 N	12.8 N	12.4 N	12.1 N	13.3 N	3.3°

Palaeolatitude estimates are taken from "Point Tracker (v. 4) for Windows" software program developed by C.R. Scotese

10^6 yr) tropical climate changes inferred from Upper Pennsylvanian and Lower Permian indicators (Loope, 1980, 1984, 1985; Blakey and Middleton, 1983; Langford and Chan, 1988; West et al., 1993; Calder, 1994; Tandon and Gibling, 1994; Miller et al., 1996; Rankey, 1997; Soreghan et al., 1997, 2007b; Soreghan et al., 2002a,b, 2007; Olszewski and Patzkowsky, 2003, in press; Falcon-Lang, 2004; Tramp et al., 2004; Mountney, 2006; Soreghan et al., 2007a,b); (3) persistent humid climate in eastern tropical Pangaea (Rees et al., 2002; Ziegler et al., 2003); and (4) reduction and expansion of the physical space of different climate zones, such as the expansion of the arid climate belt at expense of the humid tropical climate belt from Late Pennsylvanian to Early Permian time (as depicted in Fig. 3).

4.2. Climate Factor 2: Land–sea distribution

Palaeogeographic reconstructions in Blakey (2007) indicate that the total area of Earth between 30°N and S was occupied by ~16% land and ~84% water (Table 1, Fig. 2) during the Late Pennsylvanian (~300 Ma), whereas Artinskian (~280 Ma) Earth was ~22% land and 78% water. The increase of land area is due to withdrawal of epeiric seas. Although Late Pennsylvanian tropical land–sea distribution varied significantly and frequently in response to glacio-eustatic cycles (e.g., Heckel, 1977, 1980, 1984, 1986, 1990, 1991, 2002; Ramsbottom, 1979; Veevers and Powell, 1987; Klein, 1993; West et al., 1993; Cecil et al., 2003; Olszewski and Patzkowsky, 2003), sedimentary infill and regression of those seaways during Early Permian time resulted in a contiguous tropical landmass, no less than 5000 km wide. Therefore, the long-term pattern of terrestrial tropical physiography would have been characterized by increasing distance from moisture sources through Permo-Pennsylvanian time.

Thus, long-term tropical aridification might reflect withdrawal of tropical inland seas and increasing continentality (Ziegler et al., 2002; 2003; Rees et al., 2002; Tabor and Montañez, 2002). Palaeoclimate models demonstrate that changes in sea-level and land–sea distribution can have significant influence on climate and ocean circulation (e.g. Barron and Washington, 1984; Poulsen et al., 1998, 2003; Herrmann et al., 2004), including the distribution and magnitude of precipitation (Poulsen et al., 1999; Gibbs et al., 2002). Yet, the role of sea-level fall during the Late Pennsylvanian–Early Permian has not been specifically investigated.

To evaluate whether sea-level fall in the Early Permian led to drying of equatorial Pangaea, we have developed two atmosphere-biome model experiments with Late Pennsylvanian (from R. Blakey, http://jan.ucc.nau.edu/~rcb7/globehighres.html) and Early Permian (from the Palaeogeographic Atlas Project, http://pgap.uchicago.edu) palaeogeographic scenarios using GENESIS 3.0, an earth system model composed of an atmospheric general circulation model coupled to models of vegetation, soil or land ice, snow, and a 50-m slab ocean layer (Thompson and Pollard, 1997). The only difference between these experiments is palaeogeography; primarily sea sea-level and land–sea distribution.

In GENESIS, Late Palaeozoic palaeogeography does influence simulated continental precipitation over tropical Pangaea. In the Late Pennsylvanian experiment, the low-latitude band of high precipitation is slightly more extensive. This difference is most noticeable between 0° to 10°N and 10°S to 20°S where precipitation rates are greater by up to ~1 mm day^{-1} (Fig. 4). Higher precipitation rates in the Pennsylvanian experiment are most likely due to the presence of proximal moisture sources. Support for this conclusion comes from seasonal precipitation distribution (Fig. 5), where maximum precipitation rates are greatest in the Pennsylvanian simulation in low-latitude regions adjacent to epeiric seas. Moreover, seasonal distribution of low-latitude continental precipitation is similar between experiments, indicating that the position of the ITCZ has not changed substantially.

The withdrawal of epeiric seas from Pangaea does not cause a large enough change to account for the long-term aridification of near-equatorial Pangaea based on precipitation predictions from GENESIS. Although precipitation rates decrease near the equator, they remain high (>3.5 mm day^{-1}) in both experiments. Moreover, while average low-latitude continental precipitation rates decrease, GENESIS predicts that some near-equatorial regions, including western equatorial Pangaea, would become wetter in the Permian (Fig. 5). This result is at odds with palaeoclimate indicators.

4.3. Climate Factor 3: Supercontinentality

Aridification of low-latitude Pangaea has been linked to assembly of Pangaea. In this view, west and central Pangaea became progressively drier as the supercontinent coalesced because the interior and lee side of the equatorial supercontinent became distant from the low-latitude moisture source, namely western Tethys (e.g., Hay et al., 1982; Parrish, 1993; Perlmutter and Matthews, 1989). Neither modern climatology nor Palaeozoic climate model simulations support the supercontinentality hypothesis.

In the modern climate, for example, humid conditions with precipitation rates in excess of 1500 mm yr^{-1} persist across western equatorial Africa due to the seasonal migration of the ITCZ, which is strongly influenced by the African monsoon. Similarly, high precipitation rates are predicted over western equatorial Pangaea (Gibbs et al., 2002; Peyser and Poulsen, 2008-this volume; Fig. 5). As in west Africa, rainfall in western equatorial Pangaea is delivered from the Panthalassan Ocean by westerly winds (See evidence for this in Section 2.3) that result from summer low pressure on the subtropical continent.

Moreover, the supercontinentality hypothesis hinges upon a tectonic mechanism; that accretion of the supercontinent Pangaea and orogenic construction of the CPM affected climate. Such a tectonic mechanism is expected to be operative over very long timescales (10^7 yr; e.g., Parrish, 1993). Pangaea had assembled through collision of Laurasia and Gondwanaland in Late Mississippian to Early Pennsylvanian time (Scotese et al., 1979; Scotese and Golonka, 1992). If the supercontinentality hypothesis is valid, aridity should be evident throughout Upper Pennsylvanian deposits in western and central equatorial Pangaea. Upper Pennsylvanian rocks from western and central Pangaea preserve some evidence of moisture deficiency and semi-arid to arid palaeoclimate, such as calcretes, Vertisols and calcic Vertisols (e.g., Cecil et al., 1985; Cecil, 1990; Kessler et al., 2001; Tabor et al., in press). However, these Pennsylvanian arid climate indicators are intercalated stratigraphically and juxtaposed spatially with abundant indicators of humid climate, such as coals, deeply weathered palaeosols (Ultisols), laterites and bauxites in western and central tropical Pangaea (Cecil, 1990; Tabor and Montañez, 2002; 2004; Ziegler et al., 2003; Tabor et al., in press; Fig. 3). Intercalation of humid and arid indicators does not support supercontinentality as an important palaeoclimate factor.

4.4. Climate Factor 4: Monsoon

The climate of Pangaea has long been characterized as monsoonal (Robinson, 1973; Parrish et al., 1982; Rowley et al., 1985; Parrish and Peterson, 1988; Dubiel et al., 1991; Parrish, 1993; Kessler et al., 2001; Tabor and Montañez, 2002) based on sedimentologic and palaeontologic evidence of extreme seasonality and palaeowind directions. Pangean palaeoclimate model simulations predict intense monsoonal circulation, and nearly complete breakdown of zonal atmospheric circulation over Pangaea, due to extreme continentality (Kutzbach and Gallimore, 1989). Lithological and palaeontological evidence suggests that monsoonal circulation over western equatorial Pangaea intensified through the Late Palaeozoic and early Mesozoic as the supercontinent migrated northward, reaching its maximum intensity in the Late Triassic (Dubiel et al., 1991; Parrish, 1993). However, Olsen and Kent (2000; Kent and Olsen, 2000) dispute the presence of monsoonal atmospheric circulation based on evidence for zonal atmospheric circulation from climate-sensitive lithologies distributed over ~20° of

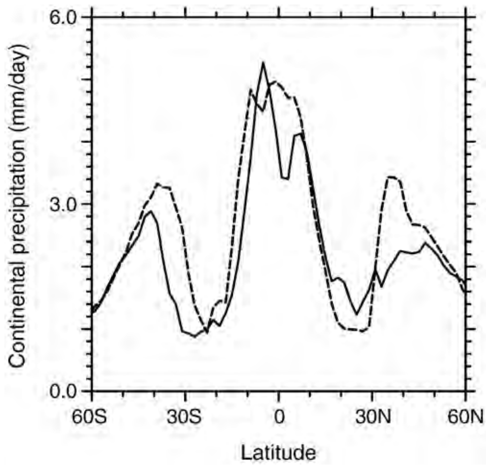

Fig. 4. Zonal-average continental precipitation rate (mm day^{-1}) simulated using the GENESIS 3.0 Earth system model. Annual precipitation rates for Late Pennsylvanian (dashed) and Early Permian (solid) simulations are shown. Note that precipitation rates are almost uniformly higher in the Late Pennsylvanian simulation than the Early Permian simulation. The increase in low-latitude precipitation is due to palaeogeographic differences as described in Section 4.2. In mid-latitudes, continental precipitation rates are also higher in the Pennsylvanian simulation than the Permian simulation. This is likely the consequence of both the closer proximity of the moisture source and an increase in global hydrologic cycling resulting to an increase in global temperature due to an enhanced greenhouse effect (by water vapor) and a reduction in surface albedo (with an increase in marine area). Besides palaeogeography, all other boundary conditions remain constant between experiments and include a reduced solar luminosity (1330.3 W m^{-2}); 2× pre-industrial CO_2 levels (560 ppm); pre-industrial concentrations of CH_4 (0.650 ppm) and N_2O (0.285 ppm); and a circular orbit with an average (23.5°) obliquity similar to modern.

palaeolatitude in Upper Triassic and Lower Jurassic fluvial-lacustrine rocks in eastern North America.

Monsoonal circulation in northern Pangaea was likely established by Late Pennsylvanian time, as evidenced by geological and geochemical indicators of a reversal of wind directions over western equatorial Pangaea (~5°S–8°N) from predominantly northeasterly to northwesterly (see summary in Section 2.3). Onset and intensification of the northern hemisphere monsoon has been suggested as a possible cause of Late Pennsylvanian– early Early Permian climate trends, particularly long-term aridification, across western equatorial Pangaea (Parrish, 1993; Kessler et al., 2001; Soreghan et al., 2002a,b; Tabor and Montañez, 2002; Tabor et al., 2002; Tabor and Montañez, 2004). Due to diversion of the ITCZ away from the equator and high summer temperatures, a strong monsoon (i.e. megamonsoon) led to arid conditions across equatorial Pangaea (Fig. 2C) (Kutzbach and Gallimore, 1989). Although a Permo-Pennsylvanian northern hemisphere monsoon was apparently operative, it likely was much weaker than its Triassic descendent due to the southward position of Pangaea (Parrish, 1993). Early Permian climate models predict a strong southern monsoon, but weaker northern monsoon (Patzkowsy et al., 1991; Gibbs et al., 2002, see Fig. 6; Peyser and Poulsen, 2008-this volume). In this context, intensification of the northern monsoon probably had a substantial influence on western equatorial climate, but at a time later than the Early Permian.

Pangaean monsoon intensity may have varied on timescales shorter than continental drift. Perlmutter and Plotnick (2003) suggest that Pangaean monsoon intensity increased during sea-level lowstand (glacials) when continentality was amplified, resulting in higher precipitation in northeastern tropical Pangaea, and decreased during sea-level highstand (interglacials). GCM experiments also support the possibility of glacial–interglacial variations in monsoon intensity due to the extent of continental ice in Gondwanaland. A strong southern hemisphere monsoon is predicted over an ice-sheet-free Gondwanaland resulting in seasonal reversal of wind patterns (i.e., westerlies) over western equatorial Pangaea. However, with the addition of continental ice on southern Gondwanaland, the summer monsoon weakens and winds over western equatorial Pangaea become predominantly easterly during December, January, and February (Peyser and Poulsen, 2008-this volume; see their Fig. 8).

4.5. Climate Factor 5: Uplift/collapse of the Central Pangaean Mountains

Mid-Carboniferous collision of Gondwanaland and Laurasia resulted in construction of the CPM slightly south of the palaeo-

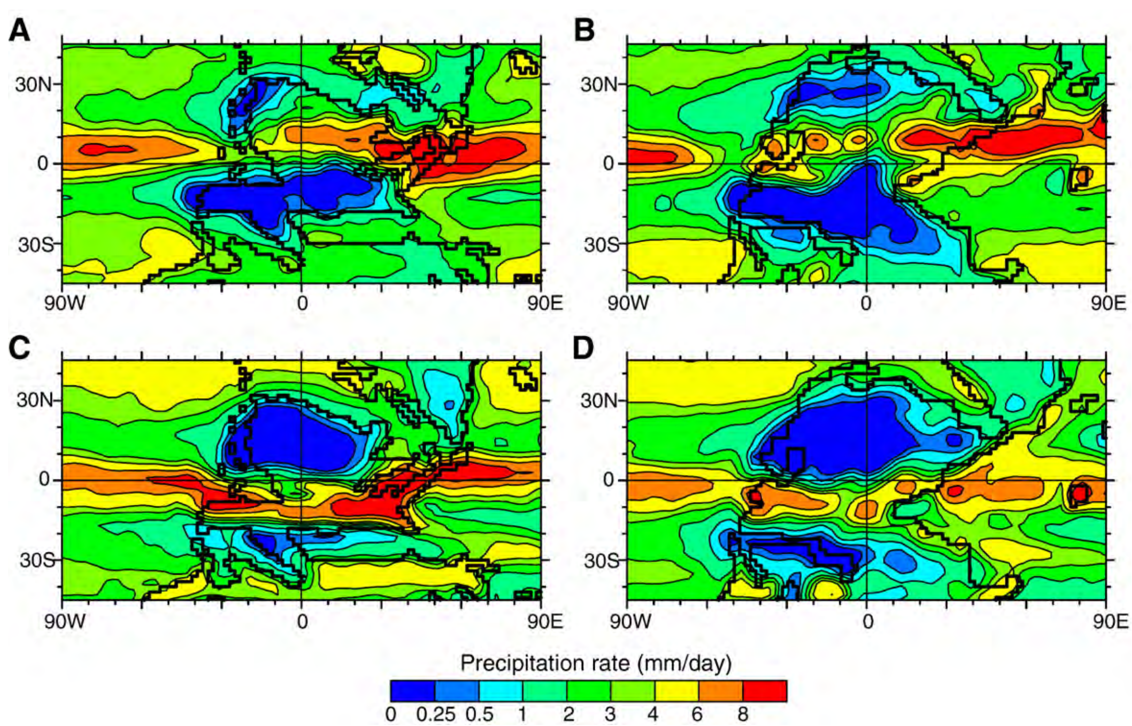

Fig. 5. Late Pennsylvanian and Early Permian seasonal precipitation rates (mm day^{-1}) simulated using the GENESIS 3.0 Earth system model. Average June–July–August and December–January–February precipitation rates are plotted for the Late Pennsylvanian (A, C) and Early Permian (B, D). The thick black line represents the continental outline.

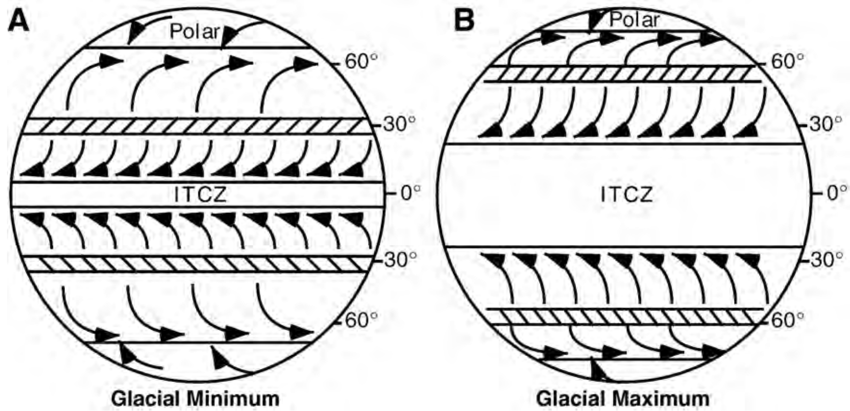

Fig. 6. Hadley circulation patterns proposed by Perlmutter and Matthews (1989) for glacial maximum (A) and glacial minimum (B). See text for discussion.

equator, which subsequently drifted into the Northern Hemisphere during Permian time (Ziegler, 1990; Ziegler et al., 1997). Stage-level compilations of coal, laterite and bauxite from Middle and Upper Pennsylvanian strata define a broad (~12°S to ~10°N) equatorial band of humid climate on Pangaea (Ronov, 1976; Parrish 1998; Fig. 3) that essentially flanked the northern and southeastern sides of the Central Pangaean Mountains (Fig. 3). This lead Rowley et al. (1985) to suggest the CPM might have been the location of a large, high-altitude, low pressure zone which drew air and moisture from both the easterly ITCZ over Tethys and westerly winds over Panthalassa. Subsequent studies of Permo-Pennsylvanian surface wind transport (Petersen, 1988; Parrish and Peterson, 1988; Kim and Loope, 1995; Kessler et al., 2001; Soreghan et al., 2002a,b; Tramp et al., 2004; Soreghan and Soreghan, 2007; Soreghan, 2007) and precipitation patterns (Tabor and Montañez, 2002; Tabor et al., 2002) provide evidence for a band (from ~5°S–8°N) of reverse (northwesterly) equatorial airflow over western Pangaea, suggesting that orography of the CPM may have exerted an important control upon atmospheric circulation and precipitation patterns over equatorial Pangaea.

Climate model simulations of the Carboniferous support the idea that the CPM focused precipitation over equatorial Pangaea (Otto-Bliesner, 1998, 2003; Peyser and Poulsen, 2008-this volume). The presence of a high (1–3 km) CPM: (1) contributes toward production of a large, low atmospheric pressure zone over equatorial Pangaea, (2) results in a strong component of reverse equatorial flow over western Pangaea, (3) acts to impede the boreal summer excursion of the ITCZ toward the Northern Hemisphere and in so doing, (4) results in everwet conditions from ~15°S to 10°N. Note, however, that monsoonal climate, as described in Climate Factor 4 (above), might also result in reverse equatorial flow over western equatorial Pangaea *in the absence of any CPMs*. Therefore palaeoclimate indicators of reverse equatorial flow (point (2) above) in western equatorial Pangaea do not affirm the presence of a high-altitude CPM.

The change from humid palaeoclimate indicators in Pennsylvanian strata to arid and seasonal indicators in Permian strata may also be related to the evolution of the CPM. Ziegler (1990; Ziegler et al., 1997) suggested that the CPM weathered to lower elevations and migrated into the Northern Hemisphere in Permian time. As a result, the easterly ITCZ became free to migrate northward and southward into the boreal and austral summer hemispheres, respectively. This scenario results in generally drier climate, greater seasonal precipitation, and possible onset of monsoonal circulation over equatorial Pangaea (Rowley et al., 1985; Parrish, 1993; however, see Section 4.4). Nevertheless, GCMs predict low pressure cells continued to develop near a low-lying Permian CPM, and result in ≤25% greater summer precipitation over central Pangaea than palaeogeography with no CPM (Kutzbach and Ziegler, 1993; Kutzbach, 1994).

There are three major shortcomings of Hypothesis 4. (1) The elevations of the CPM ranges are poorly known, such that it is difficult to assess how large of an impact the CPM exerted as a high-altitude heat source, and orographic barrier, at any time or place across the Late Pennsylvanian–Early Permian tropics. (2) Stage-level compilations of coal distribution across tropical Pangaea suggest humid conditions were present across the tropics, but it is not clear that a contiguous humid belt existed at any given interval of time because of relatively coarse time control and correlation of palaeoclimate indicators. In fact, study of a single *Middle* Pennsylvanian cyclothem across the U.S.A. (Cecil et al., 2003) indicates a climate gradient from humid in the east, to dry/arid in the west, and thus suggests that a contiguous *Late* Pennsylvanian humid belt across western equatorial Pangaea may be unrealistic. And, (3) stage level compilations of palaeoclimate indicators for the Permo-Carboniferous tropics depict relatively rapid aridification (<2–3 million years; Fig. 3) across western (and parts of central) equatorial Pangaea, and there is persuasive evidence for short short-term (~10^4–10^5 yr) oscillations between relatively humid and semi-arid to arid tropical climates during both Pennsylvanian and Early Permian time (Loope, 1980, 1984, 1985; Blakey and Middleton, 1983; Cecil et al., 1985, 2003; Langford and Chan, 1988; Cecil, 1990; West et al., 1993; Calder, 1994; Tandon and Gibling, 1994; Miller et al., 1996; Rankey, 1997; Soreghan et al., 1997, 2007a,b; Olszewski and Patzkowsky, 2003; Falcon-Lang, 2004; Tabor and Montañez, 2004; Mountney, 2006). Therefore, elements of Permo-Pennsylvanian tropical climate change, and its possible relationship(s) to atmospheric circulation systems, cannot be reconciled with rates of tectonic processes (10^6–10^7 yr) associated with the evolution of the CPM.

4.6. Climate Factor 6: Waxing and waning of ice sheets in Gondwanaland

Waxing and waning of ice sheets on Gondwanaland is recognized as a possible cause of climate variability over low-latitude Pangaea through their influence on large-scale atmospheric circulation (Ziegler et al., 1987; Perlmutter and Matthews, 1989; Cecil, 1990; Miller and West, 1993; Miller et al., 1996; Soreghan et al., 1997; Cecil et al., 2003; Perlmutter and Plotnick, 2003). Numerical climate model results also support this possibility (Poulsen et al., 2007; Peyser and Poulsen, 2008-this volume). However, there is no consensus concerning how the large-scale circulation and tropical climate responded to ice sheets in Gondwanaland; several very different ideas have been proposed which reside within three major categories:

(1) According to Perlmutter and Matthews (1989), ice sheets affected tropical climate through their influence on the extent of large-scale circulation patterns. During inter- and non-glacial episodes (base-level highstand), polar high-pressure cells shrank, causing the ITCZ to expand poleward, perhaps as far as ~10°N and S, to produce a broad, humid tropical region (Fig. 6). During episodes of glaciation (base-level lowstand),

polar high-pressure cells expanded, causing the ITCZ to shrink to a thin band near the equator. Collectively in this model, glacial episodes are suspected to have resulted in generally drier and cooler tropical environments, and inter- and non-glacial episodes resulted in generally more warm and humid environments. This style of orbital-climate forcing is consistent with interpretations of palaeoclimate indicators from (A) Earth's tropics over the past ~21,000 yr (Perlmutter and Matthews, 1989; Anhuf et al., 2006), and (B) conceptual models developed to interpret Pennsylvanian and Permian cyclothems in many areas of North America (Soreghan, 1994a, 1997; Tandon and Gibling, 1994; Yang, 1996; Rankey, 1997; Olszewski and Patzkowsky, 2003, in press) and Permo-Pennsylvanian eolianites in southwestern U.S.A. (Loope, 1980, 1984, 1985; Blakey and Middleton, 1983; Langford and Chan, 1988) and U.K. (Clemmensen, 1991; Frederiksen et al., 1998).

(2) Cecil (1990; Cecil et al., 2003) and West et al. (1993; Miller and West, 1993; Miller et al., 1996) developed various conceptual cyclostratigraphic models which implicate glacio-eustasy, Milankovitch-scale orbital frequencies, variations in global atmospheric circulation systems and resulting climate change across equatorial Pangaea in order to explain stacking patterns of palaeoclimate indicators preserved within Permo-Pennsylvanian "cyclothem" deposits from western and central equatorial Pangaea. These models posit atmospheric high-pressure zones existed over the Permo-Pennsylvanian ice

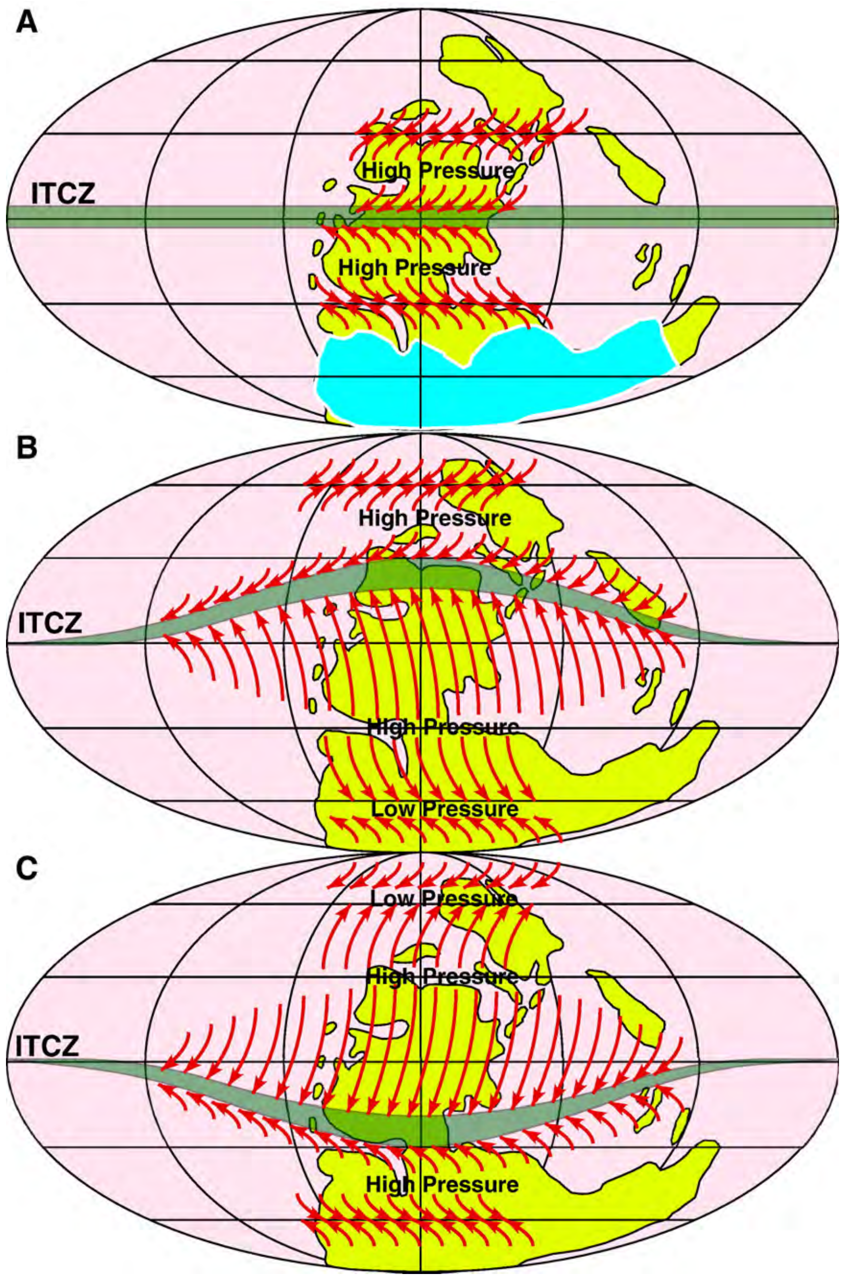

Fig. 7. Westphalian (305 Ma–or, Kasimovian; Gradstein et al., 2004) palaeogeographic reconstruction redrawn from Otto-Bliesner (2003) displaying area of continents (yellow) and modern continental names, the approximate position of the CPM, and proposed of continental ice (blue). Red arrows indicate surface wind directions and green bands indicate position of ITCZ according to the conceptual model of Cecil et al. (2003; Miller and West, 1993; Miller et al., 1996). (A) Depicts intervals of Gondwanan glaciation, which is supposed to have minimized latitudinal excursions toward the summer hemisphere, creating a narrow ITCZ over the palaeo-equator. During interglacials or non-glacials the ITCZ is supposed to have migrated toward to the north during boreal summer (B) and south during austral summer (C). The excursions of the ITCZ resulted in increased seasonality and generally drier climate across the equatorial tropics. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

sheets in Gondwanaland that suppressed seasonal excursions of the ITCZ and produced a stable, humid and warm equatorial climate belt capable of sustaining peat deposition and coal formation during glacial episodes of sea-level lowstand and exposure of the Pangaean tropics (Fig. 7). Polar high-pressure cells during inter- and non-glacials were minimized and permitted seasonal diversion of the ITCZ toward the summer-hemisphere mid-latitudes, similar to the monsoonal atmospheric circulation described by Parrish (1993; see Section 4.4). Therefore, this set of conceptual models predicts that episodes of low ice volume, and sea-level highstand, produced generally drier tropics, with more seasonal precipitation, and only a narrow equatorial band of very humid conditions.

(3) Using numerical climate model experiments, Peyser and Poulsen (2008-this volume) suggest that changes in the low-latitude surface temperature gradient, rather than high-latitude sea-level pressure, controlled low-latitude precipitation. During times of extensive glaciation in Gondwanaland, the cross-hemispheric (Northern to Southern Hemisphere) temperature gradient would have enhanced, driving more vigorous Hadley circulation and larger precipitation rates between 5°S and 30°N. In contrast to conceptual models, numerical climate models do not predict significant shifts in the seasonal extent of the ITCZ in response to variations in the size of the ice sheet in Gondwanaland.

4.7. Climate Factor 7: Atmospheric PCO_2

Geochemical models and CO_2 proxies indicate that atmospheric PCO_2 was at or near the Phanerozoic minimum in the Late Palaeozoic (Berner and Kothavala, 2001; Royer et al., 2007). Using palaeosol carbonate and fossil plant organic matter $\delta^{13}C$ values, Montañez et al. (2007) report a long-term rise in Late Pennsylvanian–Early Permian PCO_2 from <1000 to ~3500 ppmV with shorter-term variations of up to ~2000 ppmV. The likely influence of atmospheric PCO_2 variations on Pangaean surface temperature and ice volume is widely recognized (e.g. Crowley and Baum, 1992; Hyde et al., 1999; Huynh and Poulsen, 2005; Montañez et al., 2007; Horton et al., 2007). Yet, atmospheric PCO_2 has largely been overlooked as a control on tropical precipitation over equatorial Pangaea. Poulsen et al. (2007) and Peyser and Poulsen (2008-this volume) report climate model results that suggest a strong link between PCO_2 and tropical precipitation. In their simulations, increasing PCO_2 (from 1 to 8× pre-industrial levels; ~355 to 2800 ppmV) enhances continental evaporation, leading to a decrease in soil moisture over low-latitude Pangaea. On Pangaea, as in the modern Amazon basin, soil moisture is an important source of water for evapotranspiration. In the numeric climate model (Peyser and Poulsen, 2008-this volume), a reduction in soil moisture results in a decrease in precipitation over equatorial Pangaea and an increase in surface temperature.

Assuming this PCO_2–precipitation relationship persisted throughout the Late Pennsylvanian–Early Permian, the long-term increase in atmospheric PCO_2 may have led to long-term aridification in

Table 4
Matrix of climate effects vs. climate factors discussed throughout the text

Climate effect	Climate factor							Provisos/explanations
	Tectonic drift	Land:sea area	Supercontinentality	Monsoon	Orography	Ice sheets	Atmospheric PCO_2	
Long-term aridification	2	2	2	2	3	3	2*	*Tropical aridity arises if atm. PCO_2 increases from IP to P. time.
Short-term climate oscillation	3	2*	3	2	3	1†	1°	*Monsoon slightly strengthened, and seasonal precipitation slightly increased over tropical Pangaea, during sea-level regressions. †More vigorous convective precipitation (and increased rain) in tropical ITCZ during maximum ice. °High CO_2 results in N. Hem. Summer monsoon and aridity; low CO_2 results in no monsoon and relative humidity in the tropics.
Monsoon	3	4*	1	–	3	3†	1°	*Monsoon slightly strengthened during low stands. †Size of ice sheets (1) may correlate w/ behavior of monsoon (2) b/c of variations in atmospheric PCO_2, but (1) and (2) do not affect one another directly. °High CO_2 results in N. Hem. Summer monsoon; low CO_2 results in no monsoon.
Reverse equatorial flow	3	3	1	1	2*	3	1°	*GCM's indicate higher CPM elevation results in stronger reverse equatorial flow, but CPM are not necessary to generate reverse flow. °High CO_2 results in N. Hem Summer monsoon and westerly winds over tropical Pangea; low CO_2 results in no monsoon and no westerly winds.
Latitudinal migration of ITCZ	4	4	4	4	4	4	4	Variation in atmospheric CO_2, and concomitant changes in ice sheet extent, effect power and amount of precipitation generated by the ITCZ, but not its position/concentration.
Eastward aridification of tropics through Permian time	2*	2	2	2	2†	3	1°	*The most southerly basins in central Pangaea (e.g. Appalachian, Lodeve) were situated slightly further south than western Pangean basins during Permo-Pennsylvanian time, and so may have migrated out of more humid regions (ITCZ) to arid regions at a later time than western Pangean basins. †Reverse equatorial flow associated with times of monsoonal circulation provided a source of moisture in western Pangea °Presence of CPM, even at low elevation, resulted in higher rainfall in central Pangaea than without any CPM. However, Permian weathering of CPM resulted in less rainfall. @Tropical aridity arises if atm. PCO_2 increases from IP to P. time.

1 = Climate factor is likely very important, 2 = Climate factor is of moderate/secondary important, 3 = Climate factor is not important, 4 = Climate effect or climate factor is not necessary to explain the distribution of palaeoclimate indicators.
*, †, °, @ = Provisos explaining why a climate factor was, or was not, likely important.
Numerals within the matrix reflect the considered importance of a particular climate factor in driving a corresponding climate effect. Provisional statements and explanations are also given.

equatorial Pangaea. This long-term trend is consistent with lithological indicators of climate. However, on shorter timescales, the stratigraphic record may not wholly support this relationship; for example, continental drying in western equatorial Pangaea predates the rise in atmospheric PCO_2 (Montañez et al., 2007). A rise in atmospheric PCO_2 may also explain the strengthening of equatorial Westerlies; Peyser and Poulsen (2008-this volume) describe an intensification of the Northern Hemisphere summer monsoon and westerly flow over western equatorial Pangaea under high atmospheric PCO_2.

There are two major shortcomings of the CO_2 hypothesis: (1) The PCO_2-record has very large uncertainties and coarse temporal resolution. (2) Except possibly in western equatorial Pangaea, the temporal resolution and dating of lithological indicators of moisture are currently insufficient to adequately test correlations with PCO_2.

5. Summary

Several conclusions may be drawn from our analysis of the temporal and spatial distribution of climate factors, and the various hypotheses that have been proposed to explain Pangaean climate evolution:

(1) Late Pennsylvanian–Early Permian climate evolution, as recorded by lithological indicators, was complex and is unlikely to have a single explanation. Moreover, the stratigraphic record demonstrates climate change at multiple temporal scales that were unlikely to have a single, common cause. In particular, climate factors related to tectonic evolution, including orographic uplift and erosion and continental drift, are unlikely to influence climate effects on timescales of less than 10^6–10^7 yr.

(2) There is now abundant lithologic (Petersen, 1988; Loope et al., 2004; Soreghan et al., 2007b), geochemical (Soreghan et al., 2002b; Tabor and Montañez, 2002; Tabor et al., 2002) and GCM (Gibbs et al., 2002; Peyser and Poulsen, 2008-this volume) evidence that precipitation over western equatorial Pangaea likely had a Panthalassan (rather than Tethyan) origin, as a consequence of seasonal migration of the ITCZ. The large size of the Pangaean supercontinent (i.e., supercontinentality and land–sea distribution, Sections 4.2 and 4.3 respectively; Table 4) is probably the direct cause for this climate effect. Consequently, climate factors such as supercontinentality (i.e., long down-wind transport) and orographic blocking, which presumably reduced the Tethyan-derived water vapor transported from east-to-west across equatorial Pangaea, did not likely influence precipitation over western equatorial Pangaea to this western source of precipitation. Moreover, this tropical monsoon pattern does not require additional factors, such as orography or the development/intensification of a megamonsoon, to explain westerly flow across western equatorial Pangaea.

(3) Several factors could have influenced tropical precipitation over equatorial Pangaea, including sea-level, land–sea distribution, atmospheric PCO_2, orography, and glaciation in Gondwanaland. Of these factors, the influence of the uplift and erosion of the CPMs on tropical precipitation has received particular attention in the literature (e.g. Rowley et al., 1985; Ziegler, 1990; Parrish, 1993; Ziegler et al., 1997; Tabor and Montañez, 2002; 2004; Tabor et al., 2002). This attention is surprising given the large uncertainties in the CPM elevation and uplift history. The rates of CPM uplift and erosion (~10^7 yr) are too slow to explain most of the relatively rapid (10^4–10^6 yr) climate changes that have been interpreted from Late Pennsylvanian–Early Permian strata in tropical environments. Moreover, climate model results suggest that while the CPMs can affect tropical precipitation, their influence is smaller than that of other factors such as atmospheric PCO_2 and continental glaciation.

(4) Numerous conceptual models have been developed which include broad latitudinal migration of the Intertropical Convergence Zone, up to 30°N and S, as a key climate factor for equatorial drying (Ziegler et al., 1987; Parrish, 1993; Miller and West, 1993; Tabor and Montañez, 2002; Cecil et al., 2003). Conversely, the results of GCM experiments indicate that the position of Hadley cells remained within 10° of the palaeo-equator. Furthermore, GCM results indicate that differences in cross-hemispheric temperature gradients control the distribution of rainfall across the tropics. Specifically, Hadley cell convection is more vigorous and rainfall is greater in the tropics during glacial episodes in Gondwanaland. Migration of the ITCZ is unnecessary to explain palaeoclimate changes during the Late Pennsylvanian or Early Permian.

(5) The primary limitation to more accurate determination of which climate factors were important climate drivers during Late Pennsylvanian–Early Permian time is the relatively coarse resolution, and poor correlation, of lithostratigraphic palaeoclimate indicators. As a consequence of these limitations, it is presently difficult to adequately test the influence of climate controls such as atmospheric PCO_2 and continental ice-sheet variability on Pangaean climate. Higher Higher-resolution, chronostratigraphically controlled, palaeoclimate studies have become increasingly abundant in the past decade. As these studies continue to fill the spatial and temporal knowledge gaps in Pangaean tropical climate evolution, and interbasinal correlation becomes better resolved (e.g., Roscher and Schneider, 2006; Montañez et al., 2007), more robust and satisfactory models of Permo-Pennsylvanian atmospheric circulation will be generated.

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