Late Paleozoic tropical climate response to Gondwanan deglaciation

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ABSTRACT

Coupled climate-biome model simulations of the late Paleozoic were developed to determine the response of Pangean tropical climate to Gondwanan deglaciation. The model simulations predict substantial changes over equatorial Pangea including continental drying, a reversal of equatorial winds, warming, heavier $\delta^{18}O$ values of meteoric precipitation, and the expansion of deserts and the contraction of forests. The magnitude of these tropical responses is sensitive to the extent of Gondwana continental ice and the deglacial rise in atmospheric pCO_2, boundary conditions that are not well known for the late Paleozoic. Nonetheless the model predictions are consistent with climatic and environmental trends determined from terrestrial proxy data, implying that the deglaciation of Gondwana was a transformational climate event in tropical Pangea.

Keywords: paleoclimate, Carboniferous, Permian, ice age, general circulation model.

INTRODUCTION

The Permo-Carboniferous glaciation marked the last major pre-Cenozoic ice age. Permo-Carboniferous glacial deposits have been found throughout southern Gondwana and as far north as 30°S (Parrish et al., 1986). Although the extent of the Permo-Carboniferous Gondwanan glaciation is debated (e.g., Isbell et al., 2003; Jones and Fielding, 2004), geological and oxygen isotopic evidence indicates that the maximum ice volume was as great as or greater than that during the Pleistocene glacial maxima (Crowley et al., 1991; Joachimski et al., 2006). During the glacial episodes, atmospheric CO_2 levels were at the lowest levels in the Phanerozoic (Royer, 2006). Given the profound impact of Pleistocene deglaciation on tropical climate, the Gondwanan deglaciation and the concomitant rise in pCO_2 in the early Permian likely had a marked impact on late Paleozoic tropical climate and environment.

In fact the late Carboniferous to early Permian has been recognized as a transitional period in western tropical Pangea marked by a shift in cross-equatorial winds (Soreghan et al., 2002), continental drying (Kessler et al., 2001; Tabor and Montañez, 2002, 2004), marine and terrestrial warming (Tabor and Montañez, 2005; Joachimski et al., 2006), and major floral dominance-diversity shifts (Gastaldo et al., 1996; DiMichele et al., 2001; Cleal and Thomas, 2005). However, many of these changes have been attributed not to Gondwanan deglaciation, but rather to (1) the development of a rain shadow in response to the uplift of an equatorial mountain chain in the Carboniferous and early Permian, or (2) the initiation of Northern Hemisphere monsoonal circulation due to the erosion of the equatorial mountain chain in the early Permian (Rowley et al., 1985; Tabor and Montañez, 2004). In support of these ideas, climate model simulations of the Carboniferous indicate that the uplift of the Ouachita-Appalachian mountain belt could have been important in focusing precipitation in the Tropics (Otto-Bliesner, 2003).

Climate models have been used extensively to explore late Paleozoic climates. A series of studies have focused on the conditions necessary to simulate continental ice-sheets on Gondwana (Crowley et al., 1991; Crowley and Baum, 1992; Hyde et al., 1999) and the role of tropical mountains in promoting coal formation in the Carboniferous (Otto-Bliesner, 2003). Model-proxy comparison studies show promising agreement between Permian simulations and sedimentary climate indicators (Gibbs et al., 2002; Winguth et al., 2002) and have been used to constrain aspects of Pangean paleogeography and atmospheric pCO_2 (Fluteau et al., 2001; Hyde et al., 2006). In this contribution we use a coupled atmosphere-biome general circulation model to simulate the influence of Gondwanan glaciation and pCO_2 on the tropical climate of Pangea.

METHODS

Late Paleozoic experiments were completed using the GENESIS Earth system model coupled to a terrestrial biosphere model, BIOME4. GENESIS version 2.3 consists of an atmospheric general circulation model (AGCM) coupled to multilayer models of vegetation, soil and land ice, and snow (Thompson and Pollard, 1997). Sea-surface temperatures and sea ice are computed using a 50-m slab oceanic layer with diffusive heat fluxes, and a dynamic sea-ice model. A land-surface transfer model accounts for the physical effects of vegetation, soil, and soil water. The AGCM resolution is spectral T31 (~3.75°) with 18 vertical levels; the surface model grid is 2° × 2°. In our version of GENESIS, water isotopic transport and fractionation processes have been added to the atmospheric physics (Mathieu et al., 2002).

BIOME4 is an equilibrium vegetation model that predicts global vegetation distribution on the basis of physiological considerations (Haxeltine and Prentice, 1996). In our implementation of BIOME, grasses are excluded because they had not evolved by the late Paleozoic. In our coupling of GENESIS and BIOME, GENESIS is run for one year, during which monthly average temperature, precipitation, and insolation are saved. These fields are then used to force BIOME, after which the vegetation is updated in GENESIS. This process is repeated throughout the experiment integration. The implementation of BIOME4 does not substantially influence the large-scale climate. We have performed GENESIS-only experiments with globally uniform biome distributions that essentially show the same results reported here (minus the vegetation predictions).

Five late Paleozoic experiments were completed: two with no ice sheets and CO_2 levels of 355 (LCO_2) and 2800 ppm (HCO_2), and three with small, intermediate, and large ice sheets of 22.8 (ICE-S), 47.2 (ICE-I), and 59.1 (ICE-L) × 10^9 km^2 with mean heights of 1500 m and low (355) pCO_2 (Fig. 1). Atmospheric pCO_2 levels are based on paleosol-carbonate geochemical proxies from Montañez et al. (2007). The Sakmarian paleogeography, topography, and ice extent (in the ICE-S experiment) is based on reconstructions by the Paleogeographic Atlas Project (see http://pgap.uchicago.edu). The land-sea distribution is constant between experiments and does not include any influence of glacioeustasy. All other boundary conditions were identical between experiments and include a reduced solar luminosity (1330.3 Wm^{-2}) based on solar evolution models, 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. The ocean diffusive heat flux was set to a value that provides the best simulation for the modern climate. Though the isotopic concentration of seawater would have changed with ice volume, it was uniformly specified at 0‰ to facilitate comparisons between experiments. The experiments were integrated for 42 model years; results shown here have been averaged over the final ten years.

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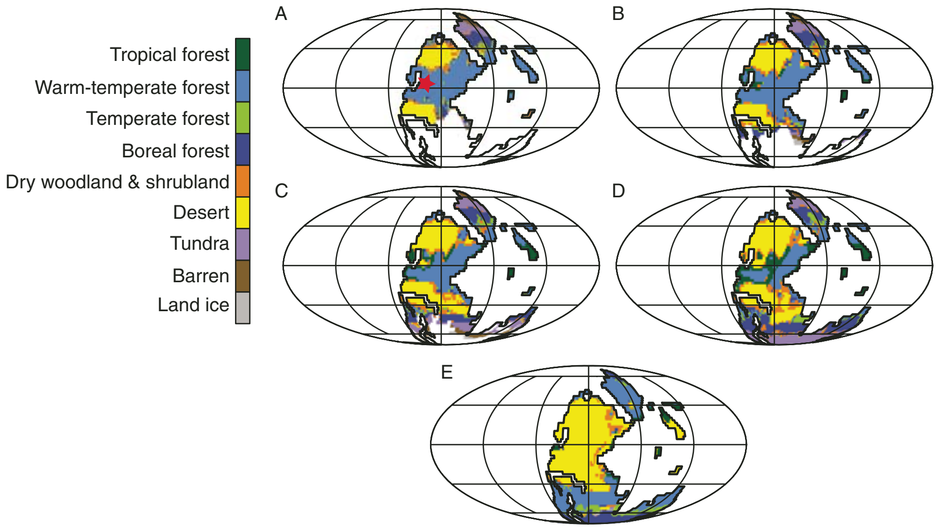

Figure 1. Simulated ecological biomes for ICE-L (A), ICE-I (B), ICE-S (C), LCO_2 (D), and HCO_2 (E) experiments. BIOME4 simulates 27 different ecological biomes; these biomes were condensed into nine mega-biomes according to Harrison and Prentice (2003) to simplify comparisons between experiments. Specified continental ice sheets on Gondwana are shown in white. The intermediate (B) and large (A) ice-sheet reconstructions are idealized, and were created by simply expanding the small ice-sheet margin over land (C). The red star represents the location of western tropical Pangea as discussed in the Discussion and Conclusions.

TROPICAL CLIMATE SENSITIVITY TO DEGLACIATION

The reduction of Gondwana continental ice leads to an increase in global mean annual temperatures (MATs) from 0.4 to 3.8, 8.3, and 11.5 °C in the ICE-S, ICE-I, ICE-L, and LCO_2 experiments, respectively. The primary causes of warming are a reduced global albedo and an increase in greenhouse forcing due to an increase in global specific humidity. The rise in $p\text{CO}_2$ in the HCO_2 experiment raises MAT to 21.9 °C mainly because of an increase in greenhouse forcing. Tropical (30°S to 30°N) continental MATs are also sensitive to the presence and extent of the Gondwana ice sheet, and increase from 11.1 to 15.5, 18.6, and 20.3 °C in the ICE-S, ICE-I, ICE-L, and LCO_2 experiments, respectively. In the HCO_2 experiment, tropical temperatures reach 31.5 °C (Fig. 2A).

Precipitation on Pangea is also sensitive to atmospheric CO_2 and Southern Hemisphere glaciation. In ICE-L and ICE-I cases, tropical precipitation migrates seasonally between hemispheres (Figs. 2B, 2C), leading to a fairly symmetric distribution of annual continental precipitation. The contraction/removal of the Gondwanan ice sheet causes seasonal precipitation to decrease and shift southward into the Southern Hemisphere (Figs. 2B, 2C); as a result, precipitation values decrease by more than 50% over western equatorial Pangea. Tropical precipitation is mainly governed by the position and intensity of the Intertropical Convergence Zone (ITCZ), a region of low pressure characterized by convective updrafts and precipitation. The seasonal migration of the ITCZ follows the annual insolation cycle, but is influenced by interhemispheric temperature contrasts (Broccoli et al., 2006). By reducing cross-equatorial temperature gradients through low-latitude Southern Hemisphere warming, contraction of the Gondwanan ice sheet causes the ITCZ to move southward and weakens the convection that drives tropical precipitation. In the HCO_2 case, the increase in continental temperature and the resulting decrease in soil moisture substantially reduce tropical precipitation across equatorial Pangea (Figs. 2B, 2C).

Deglaciation also influences large-scale tropical circulation patterns. The presence of continental ice and cold temperatures in the ICE-I and ICE-L cases creates a year-round region of relatively high pressure at the surface over southeastern Pangea (~30°S), promoting easterly flow over western equatorial Pangea. In the absence of continental ice, a strong summer monsoon forms over southeastern Pangea, inducing westerlies over western equatorial Pangea (Fig. 2D).

EFFECT OF GLACIATION ON VEGETATION

Although late Paleozoic plant types were distinct from their modern counterparts in similar environments (Gastaldo et al., 1996), the floristic biogeography of the late Paleozoic paralleled that of the present (Ziegler, 1990). Thus we contend that BIOME4 likely provides first-order representation of late Paleozoic biomes.

In the experiments with the most extensive Gondwanan ice sheet, warm-temperate forests dominate the low latitudes, with deserts confined to the subtropical regions (Fig. 1). With the removal of the ice sheet, desert and dry woodland/shrubland biomes invade the lowest latitudes, confining tropical and temperate forests to 0–10°S in western Pangea. The desert expansion reflects the reduced range of and the decreased convective precipitation within the ITCZ. The increase in CO_2 in the HCO_2 case causes further expansion of the desert and dry woodland/shrubland biomes to the near exclusion of forests across equatorial Pangea (Fig. 1).

EFFECT OF GLACIATION ON PRECIPITATION $\delta^{18}O$

Paleosol-mineral $\delta^{18}O$ values have proven to be an effective archive for the past $\delta^{18}O$ of meteoric water (e.g., Tabor and Montañez, 2002). The comparison of simulated and paleo-$\delta^{18}O$ is particularly valuable for distinguishing between past boundary conditions that affect atmospheric vapor transport (Poulsen et al., 2007).

In the ICE-L experiment, tropical precipitation $\delta^{18}O$ on land ranges from −17‰ in the Southern Hemisphere to >–3‰ in the Northern Hemisphere (Fig. 3A). The lowest isotopic values coincide with the mean position of the ITCZ and the ice sheet margin, reflecting depletion primarily through increased rainout associated with convection and orographic lifting, respectively. The relatively high values in the Northern Hemisphere represent monsoonal precipitation that has experienced little depletion from its marine source (Fig. 3A).

Deglaciation generally causes tropical precipitation $\delta^{18}O$ to increase. The contraction/removal of the Gondwanan ice sheets causes tropical precipitation $\delta^{18}O$ to increase by several per mil (cf. Figs. 3A, 3C). Because the global average temperature is higher, kinetic effects during evaporation are reduced. Also, in regions with reduced convective precipitation, the amount effect is decreased, resulting in heavier vapor. In comparison, the increase in $p\text{CO}_2$ has a smaller affect on precipitation $\delta^{18}O$. Owing to a reduced kinetic effect, precipitation $\delta^{18}O$ generally increases throughout the Tropics by ~1‰ (cf. Figs. 3C, 3D). The exception is southwestern Pangea (at ~20°S), which experiences precipitation $\delta^{18}O$ increases of up to 4‰ due to a reduction in convective precipitation.

DISCUSSION AND CONCLUSIONS

The late Paleozoic model results predict that the deglaciation of Gondwana would have led to substantial changes in climate over equatorial Pangea including (1) increased aridity throughout the equatorial region, (2) continental warming, (3) displacement of tropical biomes with an expansion of deserts and diminution of forests, (4) development of seasonal cross-equatorial westerlies, and (5) increase in low-latitude precipitation $\delta^{18}O$. The deglacial trends simulated by the GENESIS-BIOME model are consistent

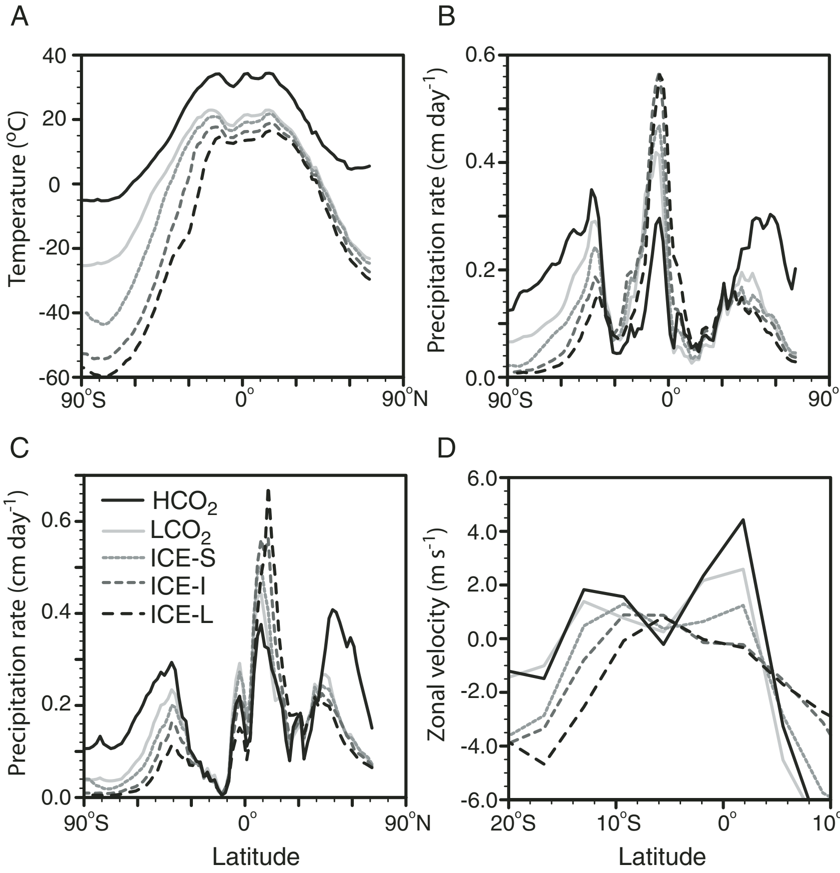

Figure 2. Zonal continental climate averages of annual surface temperature (°C) (A), December-January-February precipitation (cm day^{-1}) (B), June-July-August precipitation (cm day^{-1}) (C) , and December-January-February zonal velocity (m s^{-1}) (D). The legend for all panels is shown in C. In D, the zonal average velocity was estimated over western tropical Pangea. All zonal averages were calculated over land.

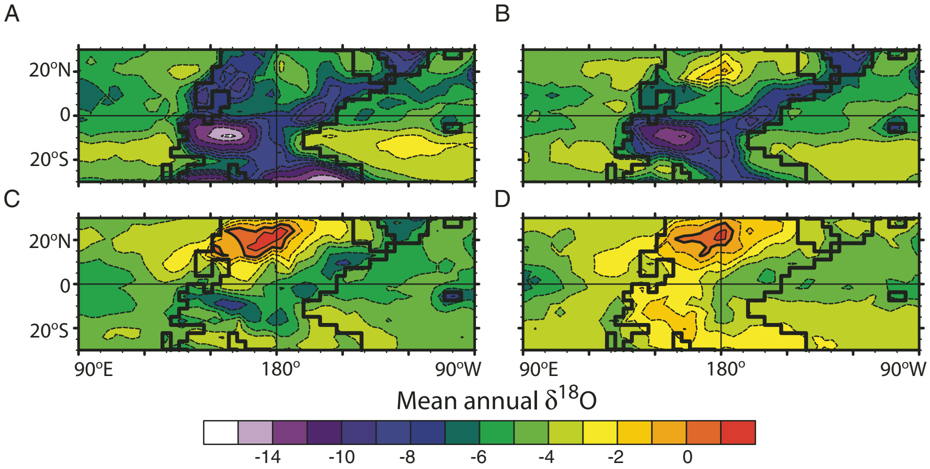

Figure 3. Mean-annual simulated precipitation δ^{18}O (‰) relative to standard mean ocean water. Results are for the ICE-L (A), ICE-I (B), LCO$_2$ (C), and HCO$_2$ (D) experiments. In all experiments, the seawater δ^{18}O is 0‰. In general, deglaciation increases precipitation δ^{18}O. The large δ^{18}O differences between experiments at 10°S are linked to a decrease in convective precipitation, which has low δ^{18}O. The thick line represents the continental outline of Pangea.

with tropical climate changes inferred from the Permo-Carboniferous sedimentary record.

The magnitude of these trends is sensitive to the extent of Gondwanan continental ice prior to deglaciation and the CO$_2$ rise during deglaciation, boundary conditions that are not known with certainty. In this study, we used CO$_2$ levels and ice-sheet extents that bracket a large range of these boundary conditions. Here, we evaluate whether the boundary conditions used in GENESIS are consistent with the late Paleozoic tropical proxy data. Geological evidence from western Pangea (see Fig. 1A) records a large Permo-Carboniferous warming from 22 to 35 °C, a shift in the dominant tropical flora to assemblages with a dry affinity, and variations in meteoric water δ^{18}O from −5.5 to −3.5‰ (Tabor and Montañez, 2005; Montañez et al., 2007). For comparison, in the high CO$_2$ experiment, MATs are 40–42 °C, mean annual precipitation δ^{18}O ranges from −2.5 to −1.5‰, and tropical flora are replaced by desert. In addition, precipitation δ^{18}O in western tropical Pangea increases by less than 1‰ between the high and low CO$_2$ experiments. On the basis of the model-data mismatches, we infer that paleo-CO$_2$ levels were likely lower than 2800 ppm and that an increase in CO$_2$ alone could not account for the observed changes in meteoric δ^{18}O or the shift in low-latitude wind direction.

Comparisons between GENESIS climate predictions for the late Carboniferous glaciation and tropical proxy data indicate that the intermediate ice-sheet reconstruction may provide the best agreement. In the ICE-S experiment, MATs and precipitation δ^{18}O are too high, 26–28 °C and −3.5 to −2.5‰ respectively in western Pangea. (Note that 1.5‰ was added to the model simulated precipitation δ^{18}O to account for an increase in mean seawater δ^{18}O during this time of greater ice volume [Joachimski et al., 2006]). Moreover, GENESIS simulates only minor differences in tropical precipitation and vegetation between the ICE-S and LCO$_2$ experiments, a prediction that is at odds with Carboniferous evidence of glacial-interglacial wet-dry cycles and floral dominance oscillations (DiMichele et al., 2001; Kessler et al., 2001; Cecil et al., 2003). In the ICE-L experiment, MATs in western tropical Pangea are likely too low, 18–20 °C.

We emphasize that these model-data comparisons are based on limited tropical data, and that additional quantitative proxy data from across tropical Pangea are needed. Nonetheless, these comparisons suggest that (1) neither the small ice-sheet reconstruction nor the high CO$_2$ level is supported by paleoclimate and paleofloral evidence, and (2) neither the Gondwanan ice sheet removal nor the CO$_2$ rise could independently account for the Permo-Carboniferous climate proxy record. However, in combination, ice-sheet contraction and pCO$_2$ rise can account for all of the major climate trends on equatorial Pangea.

An alternative interpretation is that the model-data differences arise from deficiencies in the GENESIS-BIOME model. In particular, the specification of present-day ocean heat transport in GENESIS may bias our results. Although we acknowledge the potential for model shortcomings, we suspect that the large uncertainty in boundary conditions is likely the primary cause for the mismatch. GENESIS's climate sensitivity, e.g., annual global warming of 2.5 °C with a doubling of CO_2, is similar to that of other GCMs (Thompson and Pollard, 1997). Late Paleozoic paleo-CO_2 estimates have an uncertainty of approximately ± 1200 ppm (Montañez et al., 2007), and the timing and extent of Gondwanan glaciation is debated.

In conclusion, this study supports the hypothesis that the Gondwanan deglaciation had a major influence on tropical Pangea. Our late Paleozoic simulations provide specific predictions that are consistent with the Permo-Carboniferous sedimentary record of tropical change, and that may be useful in constraining paleo-CO_2 levels and the extent of continental ice on Gondwana.

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REFERENCES CITED

Broccoli, A.J., Dahl, K.A., and Stouffer, R.J., 2006, Response of the ITCZ to Northern Hemisphere cooling: Geophysical Research Letters, v. 33, doi: 10.1029/2005GL024546.

Cecil, B.C., Dulong, F.T., West, R.R., Stamm, R., Wardlaw, B., and Edgar, N.T., 2003, Climate controls on the stratigraphy of a Middle Pennsylvanian cyclothem in North America, *in* Cecil, C.B., and Edgar, N.T., eds., Climate controls on stratigraphy: SEPM Special Publication 77, p. 151–182.

Cleal, C.J., and Thomas, B.A., 2005, Palaeozoic tropical rainforests and their effect on global climates: Is the past the key to the present?: Geobiology, v. 3, p. 13–31, doi: 10.1111/j.1472–4669.2005.00043.x.

Crowley, T.J., and Baum, S.K., 1992, Modeling late Paleozoic glaciation: Geology, v. 20, p. 507–510, doi: 10.1130/0091–7613(1992)020<0507:MLPG>2.3.CO;2.

Crowley, T.J., Baum, S.K., and Hyde, W.T., 1991, Climate model comparison of Gondwanan and Laurentide glaciations: Journal of Geophysical Research, v. 96, p. 9217–9226.

DiMichele, W.A., Pfefferkorn, H., and Gastaldo, R.A., 2001, Response of the Late Carboniferous and Early Permian plant communities to climate change: Annual Review of Earth and Planetary Sciences, v. 29, p. 461–487, doi: 10.1146/annurev.earth.29.1.461.

Fluteau, F., Besse, J., Broutin, J., and Ramstein, G., 2001, The Late Permian climate: What can be inferred from climate modeling concerning Pangea scenarios and Hercynian range altitude?: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 167, p. 39–71, doi: 10.1016/S0031–0182(00)00230–3.

Gastaldo, R.A., DiMichele, W.A., and Pfefferkorn, H.W., 1996, Out of the icehouse and into the greenhouse: A Late Paleozoic analog for modern global vegetational change: GSA Today, v. 6, p. 1–7.

Gibbs, M.T., Rees, P.M., Kutzbach, J.E., Ziegler, A.M., Behling, P.J., and Rowley, D.B., 2002, Simulations of Permian climate and comparisons with climate-sensitive sediments: Journal of Geology, v. 110, p. 33–55, doi: 10.1086/324204.

Harrison, S.P., and Prentice, C.I., 2003, Climate and CO_2 controls on global vegetation distribution at the last glacial maximum: Analysis based on paleovegetation data, biome modeling and paleoclimate simulations: Global Change Biology, v. 9, p. 983–1004, doi: 10.1046/j.1365–2486.2003.00640.x.

Haxeltine, A., and Prentice, I.C., 1996, BIOME3: An equilibrium terrestrial biosphere model based on ecophysiological constraints, resource availability and competition among plant functional types: Global Biogeochemical Cycles, v. 10, p. 693–709, doi: 10.1029/96GB02344.

Hyde, W.T., Crowley, T.J., Tarasov, L., and Peltier, W.R., 1999, The Pangean ice age: Studies with coupled climate-ice sheet model: Climate Dynamics, v. 15, p. 619–629, doi: 10.1007/s003820050305.

Hyde, W.T., Grossman, E.L., Crowley, T.J., Pollard, D., and Scotese, C.R., 2006, Siberian glaciation as a constraint on Permian-Carboniferous CO_2 levels: Geology, v. 34, p. 421–424, doi: 10.1130/G22108.1.

Isbell, J.L., Lenaker, P.A., Askin, R.A., Miller, M.F., and Babcock, L.E., 2003, Reevaluation of the timing and extent of late Paleozoic glaciation in Gondwana: Role of the Transantarctic Mountains: Geology, v. 31, p. 977–980, doi: 10.1130/G19810.1.

Joachimski, M.M., von Bitter, P.H., and Buggisch, W., 2006, Constraints on Pennsylvanian glacioeustatic sea-level changes using oxygen isotopes on conodont apatite: Geology, v. 34, p. 277–280, doi: 10.1130/G22198.1.

Jones, A.T., and Fielding, C.R., 2004, Sedimentological record of the late Paleozoic glaciation in Queensland, Australia: Geology, v. 32, p. 153–156, doi: 10.1130/G20112.1.

Kessler, J.L.P., Soreghan, G.S., and Wacker, H.J., 2001, Equatorial aridity in western Pangea: Lower Permian loessite and dolomitic paleosols in northeastern New Mexico, U.S.A: Journal of Sedimentary Research, v. 71, p. 817–832.

Mathieu, R.D., Pollard, D., Cole, J.E., White, J.W.C, Webb, R.S., and Thompson, S.L., 2002, Simulation of stable water isotope variations by the GENESIS GCM for modern conditions: Journal of Geophysical Research, v. 107 (D4), p. 2-1–2-18.

Montañez, I.P., Tabor, N.J., Niemeier, D., DiMichele, W.A., Frank, T.D., Fielding, C.R., Isbell, J.L., Birgenheier, L.P., and Rygel, M.C., 2007, CO_2-forced climate and vegetation instability during Late Paleozoic deglaciation: Science, v. 315, p. 87–91, doi: 10.1126/science.1134207.

Otto-Bliesner, B.L., 2003, The role of mountains, polar ice, and vegetation in determining the tropical climate during the Middle Pennsylvanian: Climate model simulations, *in* Cecil, C.B., and Edgar, N.T., eds., Climate controls on stratigraphy: SEPM Special Publication 77, p. 227–237.

Parrish, J.M., Parrish, J.T., and Ziegler, A.M., 1986, Permian-Triassic paleogeography and paleoclimatology and implications for Therapsid distribution, *in* Hotton, N., II, McLean, P.D., Roth, J.J., and Roth, E.C., eds., The ecology and biology of mammal-like reptiles: Smithsonian Institution Press, p. 109–131.

Poulsen, C.J., Pollard, D., and White, T.S., 2007, GCM simulation of the isotopic concentration of precipitation in the middle Cretaceous: A model-proxy comparison: Geology, v. 35, p. 199–202, doi: 10.1130/G23343A.1.

Rowley, D.B., Raymond, A., Parrish, J.T., Lottes, A.L., Scotese, C.R., and Ziegler, A.M., 1985, Carboniferous paleogeographic, phytogeographic, and paleoclimatic reconstructions: International Journal of Coal Geology, v. 5, p. 7–42, doi: 10.1016/0166–5162(85)90009–6.

Royer, D.L., 2006, CO_2-forced climate thresholds during the Phanerozoic: Geochimica et Cosmochimica Acta, v. 70, p. 5665–5675, doi: 10.1016/j.gca.2005.11.031.

Soreghan, M.J., Soreghan, G.S., and Hamilton, M.A., 2002, Paleowinds inferred from detrital-zircon geochronology of upper Paleozoic loessite, western equatorial Pacific: Geology, v. 30, p. 695–698, doi: 10.1130/0091–7613(2002)030<0695:PIFDZG>2.0.CO;2.

Tabor, N.J., and Montañez, I.P., 2002, Shifts in late Paleozoic atmospheric circulation over western equatorial Pangea: Insights from pedogenic mineral $\delta^{18}O$ compositions: Geology, v. 30, p. 1127–1130, doi: 10.1130/0091–7613(2002)030<1127:SILPAC>2.0.CO;2.

Tabor, N.J., and Montañez, I.P., 2004, Morphology and distribution of fossil soils in the Permo-Pennsylvanian Wichita and Bowie Groups, north-central Texas, USA: Implications for western equatorial Pangean palaeoclimate during icehouse-greenhouse transition: Sedimentology, v. 51, p. 851–884, doi: 10.1111/j.1365–3091.2004.00655.x.

Tabor, N.J., and Montañez, I.P., 2005, Oxygen and hydrogen isotope compositions of Permian pedogenic phyllosilicates: Development of modern surface domain arrays and implications for paleotemperature reconstructions: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 223, p. 127–146, doi: 10.1016/j.palaeo.2005.04.009.

Thompson, S.L., and Pollard, D., 1997, Greenland and Antarctic mass balances for present and doubled CO_2 from GENESIS version 2 global climate model: Journal of Climate, v. 10, p. 871–900, doi: 10.1175/1520–0442(1997)010<0871:GAAMBF>2.0.CO;2.

Winguth, A.M.E., Heinze, C., Kutzbach, J.E., Maier-Reimer, E., Mikolajewicz, U., Rowley, D., Rees, A., and Ziegler, A.M., 2002, Simulated warm polar currents during the middle Permian: Paleoceanography, v. 17, p. 1057, doi: 10.1029/2001PA000646, doi: 10.1029/2001PA000646.

Ziegler, A.M., 1990, Phytogeographic patterns and continental configurations during the Permian Period, *in* McKerrow, W.S., and Scotese, C.R., eds., Palaeozoic palaeogeography and biogeography: Geological Society [London] Memoir 12, p. 363–379.

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