

"ICEHOUSE" (COLD) CLIMATES

Earth's climate has changed, within life-sustaining bounds, from warm to cool intervals, on scales from thousands to hundreds of millions of years. In the Phanerozoic Eon there have been three intervals of glaciation (Ordovician, Carboniferous and Cenozoic) lasting tens of millions of years, with ice down to sea level at mid-latitudes (Frakes et al., 1992; Crowell, 1999). These cool "icehouse" intervals were generally times of lower sea level, lower CO₂ percentage in the atmosphere, less net photosynthesis and carbon burial, and less oceanic volcanism than during alternating "greenhouse" intervals (Fischer, 1986). The transitions from Phanerozoic icehouse to greenhouse intervals were synchronous with some biotic crises or mass extinction events, reflecting complex feedbacks between the biosphere and the hydrosphere.

Figure 18 summarizes Earth's entire paleoclimate history, and Figure 19 shows the better-known Phanerozoic Eon, with carbon, strontium and sulfur isotopic ratios that are linked to major climate changes. Figure 110 shows an anti-correlation between atmospheric CO₂ levels and $\delta^{18}\text{O}$ values (proxy for oceanic temperature), which tracks the latitude of ice-rafted glacial debris.

The Cryogenian Period of Neoproterozoic time (about 750–580 Ma) contains rocks deposited in two or more severe Icehouse intervals (Harland, 1964; Knoll, 2000). Laminated cap carbonates with depleted $\delta^{13}\text{C}$ ratios are found on top of glacial marine diamictites in many successions (Kauffman et al., 1997). The sharp juxtaposition of icehouse versus greenhouse deposits has led some to suggest that rapid and extreme climate changes took place in Neoproterozoic time. The Snowball Earth hypothesis proposes that during these Neoproterozoic glaciations, the world ocean froze over. The cap carbonates are thought to have been deposited during a subsequent alkalinity event, caused by rapid warming and supersaturation of sea water on shallow continental shelves (Hoffman et al., 1998; Kennedy et al., 2001; Hoffman and Schrag, 2002).

The Earth's temperature has remained relatively constant for 3.8 by, within a range where life could exist (Figure 111), even though solar luminosity has increased and atmospheric

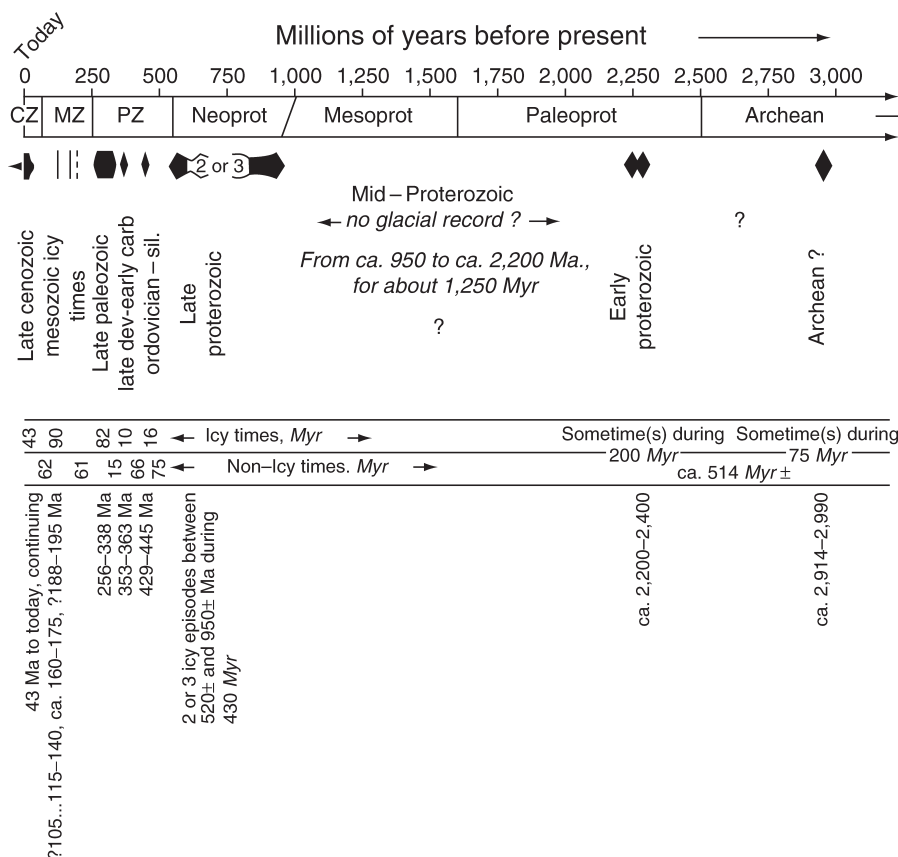


Figure 18 Major ice ages on Earth when the extent of continental glaciers was so great that tongues of ice reached the sea. Duration of icy and non-icy times is shown in the middle of the figure. CZ = Cenozoic; MZ = Mesozoic; PZ = Paleozoic; NEOPROT = Neoproterozoic; MESOPROT = Mesoproterozoic; PALEOPROT = Paleoproterozoic (from Crowell, 1999, Figure 1, used by permission of Geological Society of America).

CO₂ levels have steadily decreased, since Archean time. In the Phanerozoic, the atmospheric CO₂ concentration has varied drastically between icehouse and greenhouse times (Figure I10) (Berner, 1990, 1991; Veizer, 2000; Crowley and Berner, 2001). These concentrations are buffered by feedback loops involving water vapor within the hydrosphere and complex relations in the biosphere (Kump, 2002).

Icehouse characteristics

In general, icehouse conditions occur at times of lower sea level, less cumulative volcanic activity, more vigorous oceanic circulation and bottom water oxidation, less diversity in marine organisms, and deeper levels of the carbonate compensation depth. Relatively warmer, greenhouse intervals are characterized by marine transgression, extensive carbonate bank systems, organic-rich shale basins, marine chert and phosphorite deposition, and nutrient upwelling (Fischer and Arthur, 1977). During icehouse intervals, several or all of the tectonic, geochemical, and astronomical events discussed below coincide. However, determination of the relative cause and effect of each factor, each with its own feedback system, is complex. Both terrestrial and celestial components are involved.

Glaciation to sea level at mid latitudes

The empirical definition of an icehouse state as seen in the geologic record is the presence of floating sea-ice at mid-latitudes (30°),

and therefore the presence of ice-rafted debris (dropstones) in marine sediments (Frakes et al., 1992). Glacial ice has likely been present in high mountains throughout the Phanerozoic but the preservation potential of such alpine glacial deposits is low (Crowell, 1999) (see *Ice-rafted debris (IRD)*; *Glacial geomorphology*).

Well-mixed and colder oceans

During an icehouse, the deep ocean floor is oxygenated, and oceans are well-mixed. Icehouse conditions are favored when continental positioning allows north-south ocean circulation to bring warm equatorial water into polar latitudes where it evaporates and generates snowfall. This was the case during the Quaternary. Pleistocene oceanic bottom water temperatures were 15 °C lower than during the Mesozoic greenhouse (Crowley and Berner, 2001).

Carbon dioxide in atmosphere is lower

The percent of CO₂ in the atmosphere is lower in icehouse times, and generally tracks, and partly causes, the net change in temperature (Berner, 1990). Today's CO₂ percentage is in the icehouse range, at 300–370 ppm (Crowley and Berner, 2001). Models show this to be 17% of the Late Cretaceous greenhouse value, and perhaps only 5% of Cambrian Greenhouse CO₂ levels (Figure I10) (Chen and Drake, 1986; Berner, 1990). CO₂ percentage dropped with the Devonian advent of land plants and resulting accelerated silicate weathering.

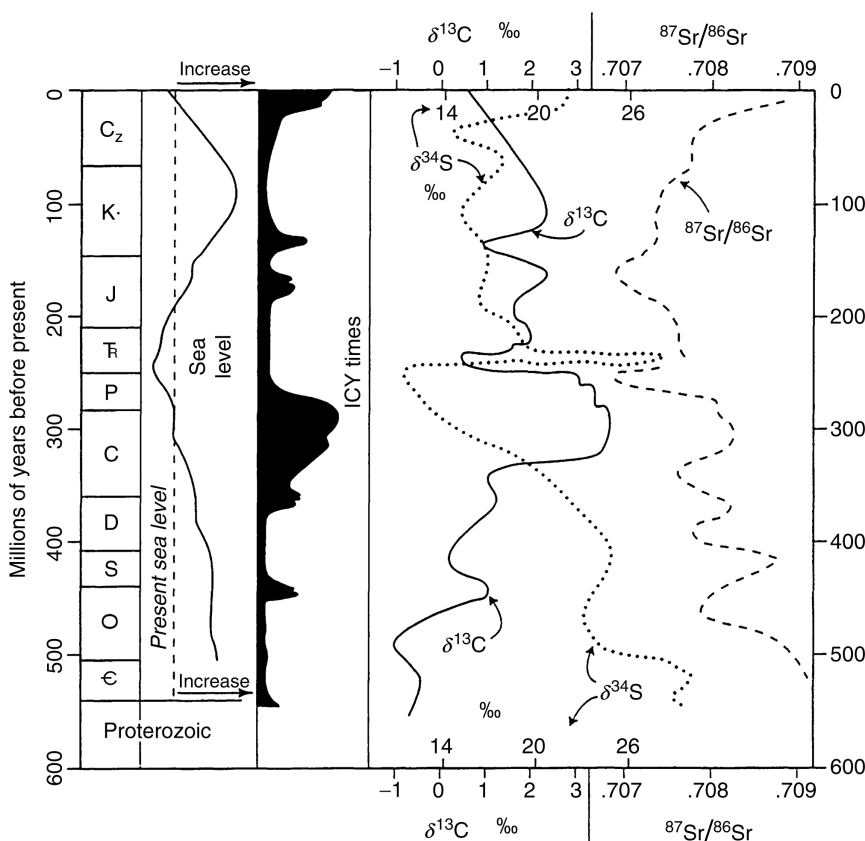


Figure 19 Carbon ($\delta^{13}\text{C}$), strontium ($^{87}\text{Sr}/^{86}\text{Sr}$), and sulfur ($\delta^{34}\text{S}$) isotopic ratios through Phanerozoic time compared with sea level and icy times. Cz = Cenozoic; K = Cretaceous; J = Jurassic; Tr = Triassic; P = Permian; C = Carboniferous; D = Devonian; S = Silurian; O = Ordovician; = Cambrian (from Crowell, 1999, Figure 25; originally modified from Veevers, 1994, Figure 1, used by permission of Geological Society of America).

Carbon isotope values are complex

Carbon is fractionated between organic matter (which is relatively concentrated in ^{12}C) and carbonate (inorganic) carbon (which is heavier in $\delta^{13}\text{C}$). Mantle carbon and methane (organic carbon) are both depleted in ^{13}C . The more sedimentary carbon that is buried as organic matter, the heavier will be the remaining carbonate carbon (Knoll, 1991). Further, increased weathering of sediments that contain isotopically light organic carbon, in orogenic belts, or from exposed continental shelves during times of low sea level, will shift the $\delta^{13}\text{C}$ ratio in carbonate sediments to more negative values.

In the Pleistocene, more negative $\delta^{13}\text{C}$ values correspond with times of greater glaciation. In the Neoproterozoic, major negative $\delta^{13}\text{C}$ spikes are found in carbonate rocks deposited on top of glacial marine sediments.

However, for the Late Paleozoic Ice Age and the Ordovician glaciation, $\delta^{13}\text{C}$ values are heavier than during the intervening greenhouse intervals (Figure 19). This apparent ambiguity has caused much controversy.

Global sea level is lower

Sea level is relatively lower during icehouse modes (Figure 19). At such times, water is stored in glacial ice, causing significant drops in global sea level on timescales of 10^5 yr (100 kyr). The extent of sea ice has an important positive feedback on the

albedo or reflectivity of the Earth, which increases as sea level drops and ice expands. As the albedo increases, the Earth absorbs less solar radiation, and the climate becomes cooler. Conversely, during warmer, more humid times, sea level is high and polar ice is minimal, lowering the Earth's albedo. The ultimate long-term (10^8 years – first order) controls on sea level are tectonic, expressed as the average elevation of water-displacing, thermally inflated ocean floor, which is itself a function of the average rate of sea-floor spreading.

Bioherms and evaporites restricted to low latitude: aragonite seas

During icehouse intervals, bioherms and evaporites are restricted to less than 20° latitude. There is lower invertebrate diversity in high latitudes. Like today, aragonite and high-Mg calcite ooids precipitate in shallow marine environments (as opposed to low-Mg calcite ooids during greenhouse times) (Stanley and Hardie, 1998).

Reduced pelagic diversity

Diversity of the pelagic marine realm was reduced (oligotaxy) during icehouse intervals. Such oligotaxic times fostered intense blooms of opportunistic pelagic organisms at times of cumulative lowest diversity (Fischer and Arthur, 1977; Fischer, 1982, 1986).

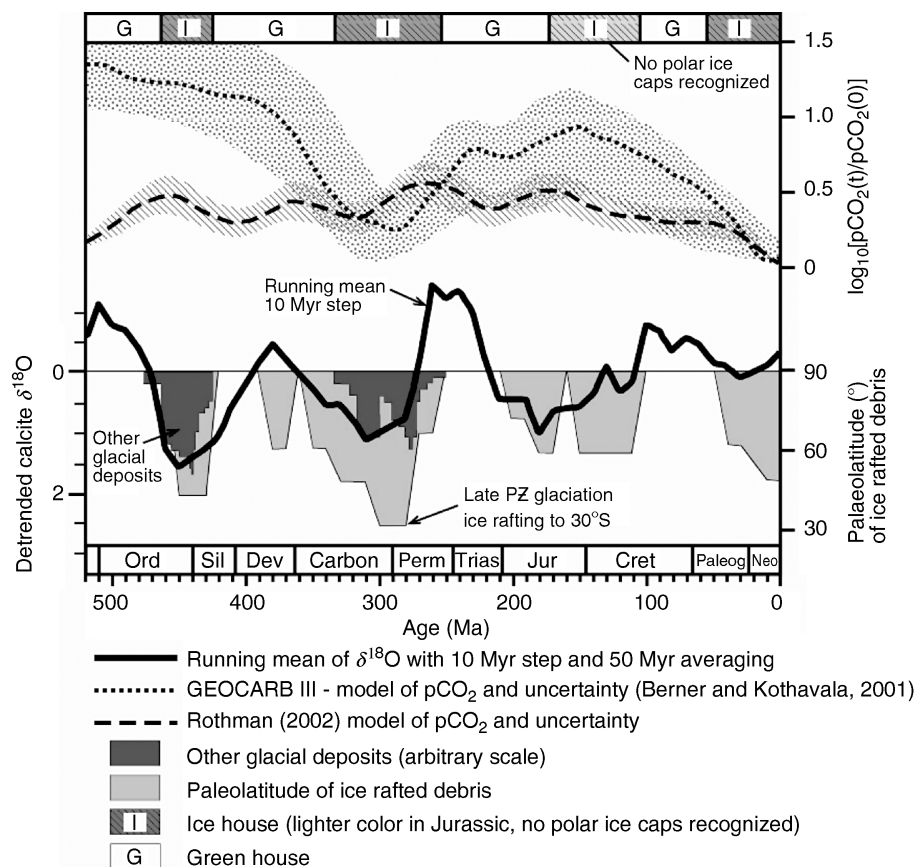


Figure 110 Phanerozoic climatic indicators and reconstructed $p\text{CO}_2$ levels (adapted from Shaviv and Veizer, 2003, Figure 1). Icehouse and greenhouse intervals are shown at the top of the diagram. Upper two curves represent estimated $p\text{CO}_2$ from the GEOCARB III CO_2 model (Berner and Kothavala, 2001) and the model of Rothman (2002). Oxygen isotope values from shells are shown with 10-Myr steps and 50-Myr averages. This smooths the curve and eliminates the highest readings. Paleolatitude of ice rafted debris and qualitative estimate of other glacial deposits are shown in the lower part of the diagram.

Carbonate compensation depth is deeper

During icehouse intervals, the CCD (carbonate compensation depth) is relatively deep within the world ocean, caused by the cooler oceanic temperatures and the resulting greater solubility of carbonate. Such conditions may not, however, apply to the Neoproterozoic, before the advent of pelagic carbonate-secreting organisms (Ridgwell et al., 2003).

Rates of carbon burial are reduced

In general, rates of carbon burial, or sequestration of organic carbon, are reduced in icehouse times, because the oceans are better aerated and the area of dysoxic restricted basins is smaller. The locations of such basins and associated carbon sinks are determined by plate configuration, global and local sea level, and by local organic productivity (controlled by upwelling). Low rates of accumulation of marine organic matter in Late Paleozoic time suggest well-ventilated Icehouse oceans.

Volcanic production of CO_2 is lower

In Phanerozoic cool periods, net production of CO_2 by volcanic activity is lower than during greenhouse events. Plate tectonic and mantle events cause these changes in net volcanic eruption

rates, with decreased rates reflecting periods of supercontinents and slow sea-floor spreading.

Plate tectonic causation

Icehouse climates are favored during times of relatively slow sea-floor spreading, which, in the Phanerozoic, have been times of supercontinent persistence. In general, large continents favor formation of ice sheets (Frakes et al., 1992). For the Late Paleozoic and Late Cenozoic ice ages, a north-south arrangement of continents, fostering longitudinal ocean circulation, was a first-order cause of climate cooling.

A decrease in worldwide volcanic activity, with reduction in CO_2 emission, is controlled by the rate of sea-floor spreading. Such a decrease is favorable for a long-term icehouse event. Slower sea-floor spreading leads to lower global sea level, another icehouse characteristic.

Veevers (1990) suggests that supercontinent cycles are the ultimate cause of icehouse climates, which occur at times of maximum continentality and slower sea-floor spreading. Apparent long-term periodicities in global tectonic phenomena and impact cratering have been linked to motions of the solar system through the galaxy (e.g., Rampino and Stothers, 1986).

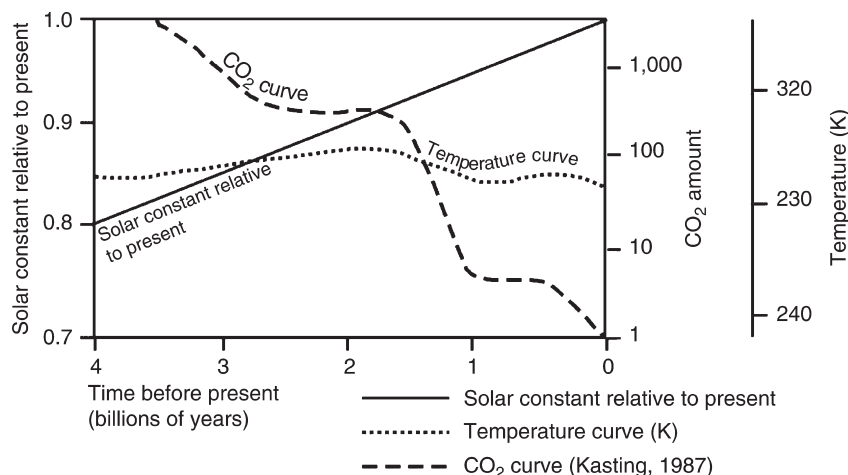


Figure I11 Variation of surface temperature, solar constant and atmospheric CO₂ over geologic time. Temperature curve refers to average surface temperature at 35° latitude. Note that CO₂ percentage has been falling steadily since the Archean, while solar output has been increasing. The net temperature has remained quite constant. Small-scale CO₂ changes, related to icehouse cycles, are superposed on this decreasing secular trend (simplified from Frakes et al., 1992, Figure 1.1).

However, Crowell (1999) does not recognize these tectonic supercycles. The presence or absence of deterministic chrono-tectonic supercontinent cycles is a recurring point of discussion in the study of Earth history.

Records of ancient glaciations

Glaciers are manifested differently at different latitudes, and include ice sheets, valley glacier and piedmont glacier complexes and mountain glaciers. Geological evidence for glaciation and glacial sedimentation are important to paleoclimatologists and sedimentologists, because they can indicate former cold climates (Harland, 1964; Deynoux et al., 1994; Crowell, 1999). Definitive ancient glacial deposits include poorly sorted diamictites, which may contain striated clasts and be associated with dropstone facies in proglacial lakes or seas (Link and Gostin, 1981).

A brief summary of the main icehouse intervals since 1,000 Ma follows. More detail is given in Crowell (1999).

Neoproterozoic (Cryogenian) glacial interval

Late Proterozoic glacial strata are found on all continents. The first evidence for ice sheets at low latitudes was marshaled by Harland (1964) who proposed the Great Infra-Cambrian ice age. Neoproterozoic paleocontinental positions suggest that most landmasses were at low latitudes, in areas of intense silicate weathering (Evans, 2000). This intense weathering traps carbonate and lowers global atmospheric CO₂ percentage (Kirschvink, 1992). Repeated studies have demonstrated low paleolatitudes for Neoproterozoic glacial marine strata (Sohl et al., 1999).

The Cryogenian Period of the Neoproterozoic is named for these icehouse intervals and has been investigated by the International Union of Geological Sciences (Knoll and Walter, 1992; Knoll, 2000; Narbonne, 2003).

Only recently have sufficiently accurate geochronologic data been obtained to place constraints on the Neoproterozoic glacial intervals, so that issues of synchronicity and extent of specific glacial advances remain controversial. Figure I12 shows the time and isotopic constraints, grouped into two broad icehouse

intervals, Sturtian (730–670 Ma) and Marinoan (640–580 Ma), each tens of millions of years long in duration and each with several glaciations. The Marinoan glacial interval is used here in the inclusive sense (Knoll, 2000) to encompass the Varanger, Ghaub, and Gaskiers glaciations and is succeeded by the Ediacaran (Vendian) Period. New radiometric ages from glacial strata reveal multiple ages for ice advance, rather than two synchronous global ice ages (Bowering et al., 2003; Lund et al., 2003; Calver et al., 2004; Fanning and Link, 2004; Hoffmann et al., 2004; Zhou et al., 2004). Purely lithostratigraphic correlations have been a first start, but, to be credible, must be verified by geochronology.

Cap carbonates

Neoproterozoic glacial marine diamictites are often overlain by laminated cap carbonates that have strongly negative $\delta^{13}\text{C}$ values (e.g., Kaufman et al., 1997; Hoffman et al., 1998; Kennedy et al., 2001). Some cap carbonates contain spectacular aragonite crystal fans that grew rapidly on the sea floor. The juxtaposition of glacial (icehouse) deposits with carbonate (greenhouse) deposits is striking. In the Snowball Earth scenario, which proposes that the world oceans froze entirely over (Hoffman et al., 1998; Hoffman and Schrag, 2002), cap carbonates play an integral role and are thought to have been deposited rapidly during a post-glacial marine alkalinity event that was fed by a deeply weathered carbonate-rich regolith formed during the glaciation.

The severity of Cryogenian climate cycles is represented by the wide range of $\delta^{13}\text{C}$ values from Neoproterozoic carbonates (Figure I12) (Kaufman et al., 1997; Lorentz et al., 2004). Documented negative $\delta^{13}\text{C}$ isotopic excursions for carbonates immediately above the glacial rocks are as great as 16‰, going from +10 to –6‰, which exceeds the magnitude of Phanerozoic excursions (Hoffman and Schrag, 2002). Levels of –6‰ suggest near-shutdown of the fractionation of ^{12}C in organic matter, as would happen with rapid deposition of shallow-marine carbonate during post-glacial transgression.

Crowell (1999) points out that these carbonates may not represent immediately warmed ocean water, but rather, slowly

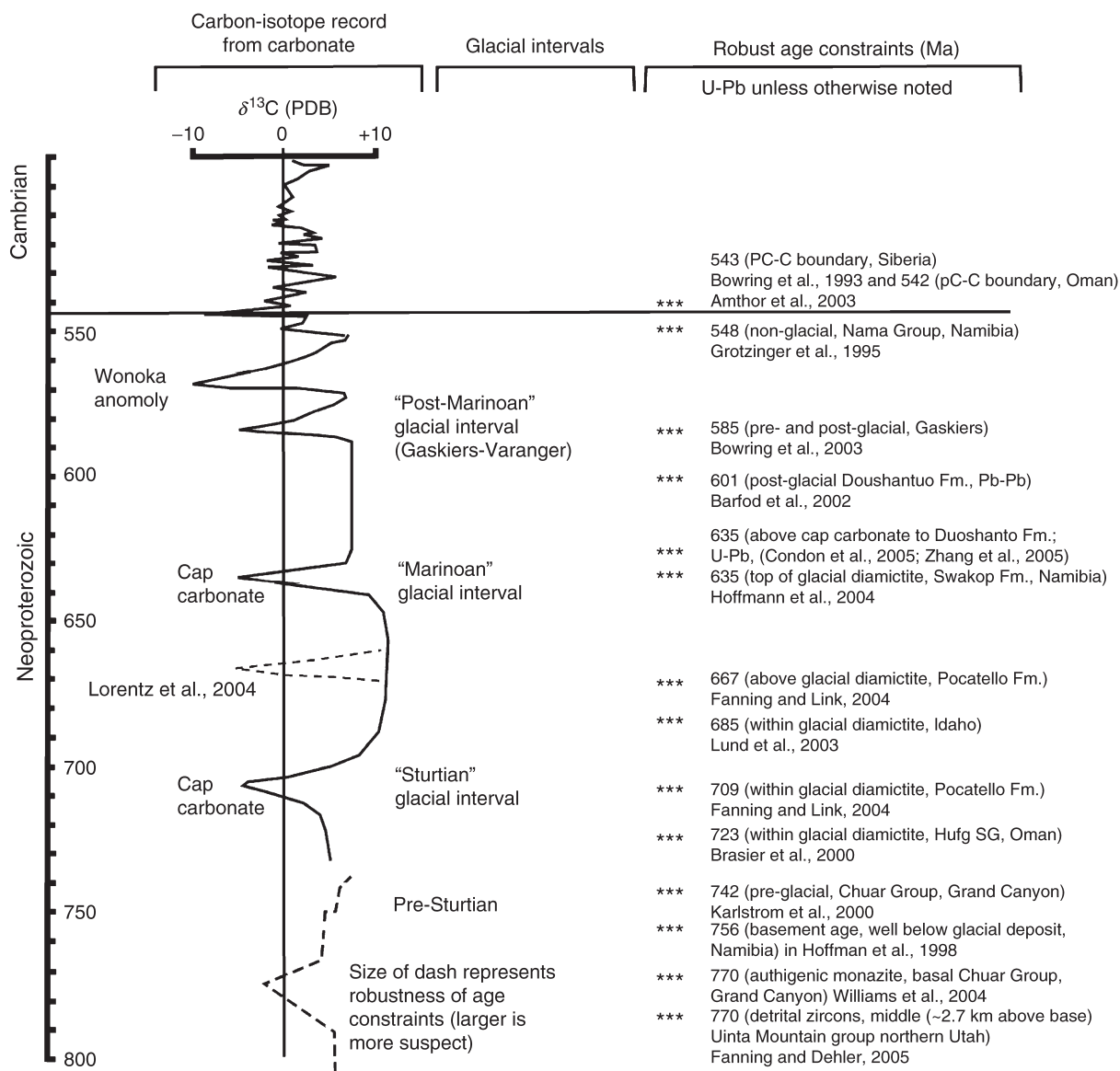


Figure 112 Late Neoproterozoic glacial intervals, carbon isotope values (Kaufman et al., 1997; Kennedy et al., 2001), and geochronologic constraints. Figure prepared by F. A. Corsetti, University of Southern California.

warming transgressive shelves above marine tillites. At least one post-glacial cap carbonate in south China has extremely low $\delta^{13}\text{C}$ values of -45‰ , suggesting local methane release caused by post-glacial sea level rise and flooding of continental shelves (Jiang et al., 2003).

Proposed explanations of low-latitude Cryogenian glaciation

Low latitude glaciation has been explained by several hypotheses:

1. A climate system similar to that of today with unusual paleogeographic and tectonic factors combining to allow low-latitude glaciation;
2. A Snowball Earth with an ice shelf covering the ocean. (Note that a modification into a "Slushball Earth" may be

necessary to account for evidence of meltwater during Neoproterozoic glaciations).

3. A greater inclination of Earth's spin axis such that polar regions have more equable climates than low-latitude positions (Williams, 1975);
4. True polar wander, where the lithosphere decouples from the asthenosphere and moves with respect to the magnetic field. This would cause inconsistent paleomagnetic pole positions on different continents;
5. Extraterrestrial influences, which may trigger climate cooling by the presence of abundant air-borne ejecta.

Late Ordovician and Early Silurian glaciations

During the Late Ordovician to early Silurian (~ 458 – 421 Ma), short but severe glaciations affected areas of South Africa,

the Saharan region of North Africa, and South America as the continents drifted over the South Pole. These glaciations were primarily due to continental positioning, and did not represent a global icehouse event. Atmospheric CO₂ percentage was high (Figure I10), but its effects were overridden by favorable continental positioning of Gondwanaland near the South Pole (Crowley and Berner, 2001).

Late Devonian and Early Carboniferous glaciations

Late Devonian and Early Carboniferous ice ages are recognized in isolated places in Brazil, in what were then high southern polar latitudes. Crowell (1999) includes these in his major episodes of glaciation, though Frakes et al. (1992) do not do so.

Late Paleozoic ice age

The Late Paleozoic ice age, from early Carboniferous to late Permian time (Figure I13) (~350–253 Ma) is the period of the

well-known Gondwanan glaciations. It produced striated surfaces and tillites on all Gondwanan continents (Australia, India, Asia, Antarctica, South America and Africa). Glaciers extended away from the South Pole as far as 40° S latitude. In most places, several advance and retreat cycles can be recognized, and they are diachronous on different continents, tracking movement through high southern latitudes. In the Northern Hemisphere, glacio-eustatic cyclothem are recognized in the North American mid-continent. Gondwanan glacial deposits were part of the evidence used by Wegener for continental drift. Stones derived from southern and western Africa are found in glacial deposits in Brazil and Argentina.

The Late Paleozoic ice ages ended at least 5 Myr before the mass extinction at the end of the Permian. The warming process may have been facilitated by the outpouring of the Siberian trap lavas and associated CO₂ emission (see *Glaciations, pre-Quaternary; Late Paleozoic climates*).

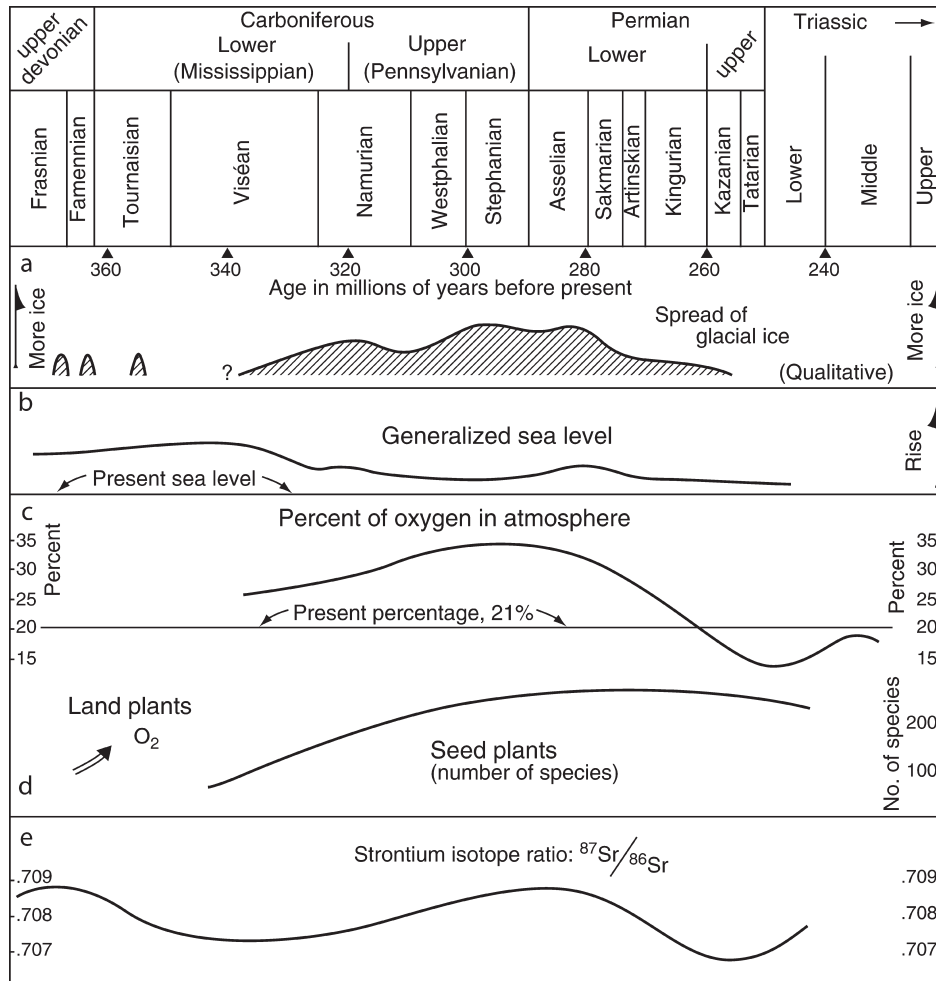


Figure I13 Time plot of glacial expansions during the Late Paleozoic ice age from Devonian through Triassic time. Figure shows several possible cause and effect factors related to glacial advances. Geologic time scale calibrations from Harland et al. (1990), generalized sea-level curve from Vail et al. (1977), percent of oxygen in atmosphere from Berner (1990), number of species of land plants from Niklas et al. (1985), strontium isotopic ratios simplified from Burke et al. (1982) and Holser et al. (1988) (from Crowell, 1999, Figure 39; used with permission of the Geological Society of America).

Late Cenozoic ice age

The Late Cenozoic, Eocene to Recent glaciation (~55–0 Ma) (Frakes et al., 1992) began slowly and locally in the Antarctic. Ice-rafted sediment is present in 35 Ma Antarctic sediments. Worldwide cooling ensued by early Miocene time (22 Ma). Northern Hemisphere climates got progressively colder, as reflected by paleobotanical records of the southward migration of the extent of broad-leaved evergreen vegetation. The Cenozoic rise of the Himalayas intensified the Indian monsoon, and affected climate in numerous ways, including favoring Arctic cooling. The collision of India and Asia also created a CO₂ sink, due to enhanced chemical weathering of the uplifted Himalayan orogenic belt, which removed CO₂ from the atmosphere (see *Mountain uplift and climate change*). The greatest extent of glaciation was reached in the Pleistocene. Pleistocene climate cyclicity is recorded in $\delta^{18}\text{O}$ curves as measured from CaCO₃ shells of foraminifera, with the $\delta^{18}\text{O}$ values inversely related to ocean-water temperature (i.e., positive values indicate cold ocean temperatures and vice-versa; see *Oxygen isotopes*). Pleistocene glacial cyclicity generally follows Milankovich astronomical controls (see *Astronomical theory of climate change*; *Pleistocene climates*; *Glaciations, Quaternary*).

Summary

Earth has undergone several Icehouse periods, with the Pleistocene being the most recent. The controlling factors are both cyclic (external or astronomical) and secular (internal to the Earth), largely the product of plate tectonics (see *Climate change, causes*). In his synthesis of pre-Mesozoic ice ages, Crowell (1999) sees an irregular pattern of major glaciations, caused by changes in paleogeography and driven by plate tectonics, with secondary atmospheric and astronomical influence.

Earth's orbital variations are surely cyclic: "Allocyclic sedimentary rhythms – driven by orbital reaction of the Earth with this moon, its sibling planets, and the Sun, and transmitted through climate – are real" (Fischer, 1986). Luckily for us, Earth's history of extraterrestrial interactions has not fundamentally disrupted the life-sustaining Earth-Ocean-Lithosphere system. As shown in Figure I11, Earth's surface temperature has remained relatively constant within limits since the Archean, even though the luminosity of the Sun has increased and the amount of CO₂ in the atmosphere has decreased. Water vapor has acted as a buffer and climate moderator (Kump, 2002; Shaviv and Veizer, 2003).

CO₂ cycling through the biosphere is internal to the climate system, and thus cannot drive a secular change through time, but does affect feedbacks (see *Carbon cycle*). Net volcanism, related to sea-floor spreading rate, is an external forcing factor to the climate system, and had a major effect on the levels of CO₂ during the Mesozoic Greenhouse period (Kump, 2002). Even though the burning of fossil fuels has caused atmospheric CO₂ concentration to rise steadily since observations on Mauna Kea began in 1958 (Chen and Drake, 1986), changes in climate owing to human activity are geologically ephemeral. In taking a long-time view of the Earth's history, they represent but a quick excursion.

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Bibliography

Amthor, J.E., Grotzinger, J.P., Schroeder, S., Bowring, S.A., Ramezani, J., Martin, M.W., and Matter, A., 2003. Extinction of *Cloudina* and *Namacalathus* at the Precambrian-Cambrian boundary in Oman. *Geology*, **31**, 431–434.

Barfod, G.H., Albarede, F., Knoll, A.H., Xiao, S., Telouk, P., Frei, R., and Baker, J., 2002. New Lu-Hf and Pb-Pb age constraints on the earliest animal fossils. *Earth Planet. Sci. Lett.*, **201**, 203–212.

Berger, W.H., 1982. Climate steps in ocean history—Lessons from the Pleistocene. In Berger, W.H., Crowell, J.C., et al. (eds), *Climate in Earth History, Studies in Geophysics*. Washington, DC: National Academy Press, pp. 43–54.

Berner, R.A., 1990. Atmospheric carbon dioxide levels over Phanerozoic time. *Science*, **249**, 1382–1386.

Berner, R.A., 1991. A model of atmospheric CO₂ over Phanerozoic time. *Am. J. Sci.*, **291**, 339–376.

Berner, R.A., and Kothavala, Z., 2001. GEOCARB III: A revised model of atmospheric CO₂ over Phanerozoic time. *Am. J. Sci.*, **301**, 182–204.

Bowring, S., Myrow, P., Landing, E., Ramezani, J., and Grotzinger, J., 2003. Geochronological constraints on terminal Neoproterozoic events and the rise of Metazoans. *Geophys. Res. Abstr.*, **5**, 13,219.

Brasier, M., McCarron, G., Tucker, R., Leather, J., Allen, P., and Shields, G., 2000. New U-Pb zircon dates for the Neoproterozoic Ghubrah glaciation and for the top of the Huqf Supergroup, Oman. *Geology*, **28**, 175–178.

Burke, W.H., Denison, R.E., Hetherington, E.A., Koepnick, R.B., Nelson, H.F., and Otto, J.B., 1982. Variations of seawater ⁸⁷Sr/⁸⁶Sr throughout Phanerozoic time. *Geology*, **10**, 516–519.

Calver, C.R., Black, L.P., Everard, J.L., and Seymour, D.B., 2004. U-Pb zircon age constraints on late Neoproterozoic glaciation in Tasmania. *Geology*, **32**, 893–896.

Chen, C.-T.A., and Drake, E.T., 1986. Carbon dioxide increase in the atmosphere and oceans and possible effects on climate. *Annu. Rev. Earth Planet. Sci.*, **14**, 201–236.

Condon, D., Zhu, M., Bowring, S., Wang, W., Yang, A., and Jin, Y., 2005. U-Pb ages from the Neoproterozoic Doushantuo Formation, China. *Science*, **308**, 95–98.

Crowell, J.C., 1999. *Pre-Mesozoic Ice Ages: Their Bearing on Understanding the Climate System*. 192, Boulder, CO: Geological Society of America Memoir 106pp.

Crowley, T.J., and Berner, R.A., 2001. CO₂ and climate change. *Science*, **292**, 870–872.

Deynoux, M., Miller, J.M.G., Domack, E.W., Eyles, N., Fairchild, I.J., and Young, G.M. (eds.), 1994. *Earth's Glacial Record*. Cambridge, UK: Cambridge University Press, 266pp.

Evans, D.A.D., 2000. Stratigraphic, geochronological, and paleomagnetic constraints upon the Neoproterozoic climatic paradox. *Am. J. Sci.*, **300**, 347–433.

Fanning, C.M., and Dehler, C.M., 2005. Constraining depositional ages for Neoproterozoic siliciclastic sequences through detrital zircon ages: A ca. 770 maximum age for the lower Uinta Mountain Group. *Geological Society of America Abstracts with Programs*, **37**.

Fanning, C.M., and Link, P.K., 2004. 700 Ma U-Pb SHRIMP ages for Neoproterozoic (Sturtian) glaciogenic Pocatello Formation, southeastern Idaho. *Geology*, **32**, 881–884.

Fischer, A.G., 1982. Long-term climatic oscillations recorded in stratigraphy. In Berger, W. (ed.), *Climate in Earth History*. National Research Council, Studies in Geophysics, Washington, DC: National Academy Press, pp. 97–104.

Fischer, A.G., 1986. Climatic rhythms recorded in strata. *Annu. Rev. Earth Planet. Sci.*, **14**, 351–376.

Fischer, A.G., and Arthur, M.A., 1977. Secular variations in the pelagic realm. In Cook, H.C., and Enos, P. (eds.), *Deep Water Carbonate Environments*. SEPM Special Publication 25, pp. 18–50.

Frakes, L.A., Francis, J.E., and Syktus, J.I., 1992. *Climate modes of the Phanerozoic*. New York: Cambridge University Press, 274pp.

Grotzinger, J.P., Bowring, S.A., Saylor, B.Z., and Kaufman, A.J., 1995. Biostratigraphic and geochronologic constraints on early animal evolution. *Science*, **270**, 598–604.

Harland, W.B., 1964. Critical evidence for a great infra-Cambrian glaciation. *Geologische Rundschau*, **54**, 45–91.

Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., and Smith, D.G. (eds.), 1990. *A Geologic Time Scale 1989*. Cambridge, UK: Cambridge University Press, 263pp.

Hoffman, P.F., and Schrag, D.P., 2002. The snowball Earth hypothesis: testing the limits of global change. *Terra Nova*, **14**, 129–155.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P., and Schrag, D.P., 1998. A Neoproterozoic snowball Earth. *Science*, **281**, 1342–1346.

- Hoffmann, K.-H., Condon, D.J., Bowring, S.A., and Crowley, J.L., 2004. U-Pb zircon date from the Neoproterozoic Ghaub Formation, Namibia: Constraints on Marinoan glaciation. *Geology*, **32**, 817–820.
- Holser, W.T., Schidlowski, M., Mackenzie, F.T., and Maynard, J.B., 1988. Biogeochemical cycles of carbon and sulfur. In Gregor, C.B., Garrels, R.M., Mackenzie, F.T., and Maynard, J.B. (eds.), *Chemical Cycles in the Evolution of the Earth*. New York: Wiley, pp. 105–173.
- Jiang, G., Kennedy, M.J., and Christie-Blick, N., 2003. Stable isotopic evidence for methane seeps in Neoproterozoic postglacial cap carbonates. *Nature*, **426**, 822–826.
- Karlstrom, K.E., Bowring, S.A., Dehler, C.M., Knoll, A.H., Porter, S.M., DesMarais, D.J., Weil, A.B., Sharp, Z.D., Geissman, J.W., Elrick, M.B., Timmons, J.M., Crossey, L.J., and Davidek, K.L., 2000. Chuar Group of the Grand Canyon: Record of breakup of Rodinia, associated change in the global carbon cycle, and ecosystem expansion by 740 Ma. *Geology*, **28**, 619–622.
- Kasting, J.F., 1987. Theoretical constraints on oxygen and carbon dioxide concentrations in the Precambrian atmosphere. *Precambrian Res.*, **34**, 205–229.
- Kaufman, A.J., Knoll, A.H., and Narbonne, G.M., 1997. Isotopes, ice ages, and terminal Proterozoic Earth history. *Proc. Natl. Acad. Sci.*, **94**, 6600–6605.
- Kennedy, M.J., Christie-Blick, N., and Sohl, L.E., 2001. Are Proterozoic cap carbonates and isotopic excursions a record of gas hydrate destabilization following Earth's coldest intervals? *Geology*, **29**, 443–446.
- Kirschvink, J.L., 1992. Late Proterozoic low-latitude global glaciation: The snowball Earth. In Schopf, J.W., and Klein, C. (eds.), *The Proterozoic biosphere*. New York: Cambridge University Press, pp. 51–52.
- Knoll, A.K., 1991. End of the Proterozoic Eon. *Sci. Am.*, **265**, 64–73.
- Knoll, A.K., 2000. Learning to tell Neoproterozoic time. *Precambrian Res.*, **100**, 3–20.
- Knoll, A.H., and Walter, M.R., 1992. Latest Proterozoic stratigraphy and Earth history. *Nature*, **356**, 673–677.
- Kump, L.R., 2002. Reducing uncertainty about carbon dioxide as a climate driver. *Nature*, **419**, 188–190.
- Link, P.K., and Gostin, V.A., 1981. Facies and paleogeography of Sturtian glacial strata (Late Precambrian), South Australia. *Am. J. Sci.*, **281**, 353–374.
- Lorentz, N.J., Corsetti, F.A., and Link, P.K., 2004. Seafloor precipitates and C-isotope stratigraphy from the Neoproterozoic Scout Mountain Member of the Pocatello Formation, southeast Idaho: implications for Earth System behavior. *Precambrian Res.*, **130**, 57–70.
- Lund, K., Aleinikoff, J.N., Evans, K.V., and Fanning, C.M., 2003. SHRIMP U-Pb geochronology of Neoproterozoic Windermere Supergroup, central Idaho: Implications for rifting of western Laurentia and synchronicity of Sturtian glacial deposits. *Geol. Soc. Am. Bull.*, **115**, 349–372.
- Narbonne, G.M., 2003. I.U.G.S. Subcommittee on the Terminal Proterozoic System, 18th circular (September 2003). <http://geol.queensu.ca/people/narbonne/trm-prot/> (June 2004).
- Niklas, K.J., Tiffney, B.H., and Knoll, A.H., 1985. Patterns in vascular land plant diversification: an analysis at the species level. In Valentine, J.W. (ed.), *Phanerozoic Diversity Patterns: Profiles in macroevolution*. Princeton, NJ: Princeton University Press, pp. 97–128.
- Parrish, J.T., 1982. Upwelling and petroleum source beds, with reference to the Paleozoic. *AAPG Bull.*, **66**, 750–774.
- Rampino, R., and Stothers, R.B., 1986. Geological periodicities and the galaxy. In Smoluchowski, R., Bahcall, J.N., and Matthews, M.S. (eds.), *The Galaxy and the Solar System*. Tucson, AZ: University of Arizona Press, pp. 241–259.
- Ridgwell, A.J., Kennedy, M.J., and Calderia, K., 2003. Carbonate deposition, climate stability, and Neoproterozoic ice ages. *Science*, **302**, 859–862.
- Rothman, D.H., 2002. Atmospheric carbon dioxide levels for the last 500 million years. *Proc. Natl. Acad. Sci.*, **99**, 4167–4171.
- Shaviv, N.R., and Veizer, J., 2003. Celestial driver of Phanerozoic climate? *GSA Today*, **13**, 4–10.
- Sohl, L.E., Christie-Blick, N.M., and Kent, D.V., 1999. Paleomagnetic polarity reversals in Marinoan (ca. 600 Ma) glacial deposits of Australia: implications for the duration of low-latitude glaciations in Neoproterozoic time. *Geol. Soc. Am. Bull.*, **111**, 1120–1139.
- Stanley, S.M., and Hardie, L.A., 1998. Secular oscillations in the carbonate mineralogy of reef-building and sediment-producing organisms driven by tectonically forced shifts in seawater chemistry. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **144**, 3–19.
- Vail, P.R., Mitchum, R.M. Jr., Todd, R.G., Widmier, J.M., Thompson, S., III, Sangree, J.B., Bubb, J.N., and Hatlelid, W.G., 1977. Seismic stratigraphy and global changes of sea level. In Payton, C.E. (ed.), *26, Seismic stratigraphy—Application to Hydrocarbon Exploration*. Tulsa, OK: American Association of Petroleum Geologists Memoir pp. 29–212.
- Veivers, J.J., 1990. Tectonic-climatic supercycle in the billion-year plate-tectonic eon: Permian Pangean icehouse alternates with Cretaceous dispersed continent Greenhouse. *Sediment. Geol.*, **68**, 1–16.
- Veivers, J.J., 1994. Pangea: Evolution of a supercontinent and its consequences for Earth's paleoclimate and sedimentary environments. In Klein, G.D. (ed.), *Geological Society of America Special Paper 288, Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent*. Boulder, CO: Geological Society of America, pp. 13–23.
- Veizer, J., Godderis, Y., and Francois, L.M., 2000. Evidence for decoupling of atmospheric CO₂ and global climate during the Phanerozoic eon. *Nature*, **408**, 698–701.
- Williams, G.E., 1975. Late Precambrian glacial climate and the Earth's obliquity. *Geol. Mag.*, **112**, 441–444.
- Williams, M. and 2004. Dating sedimentary sequences: in situ U/Th-Pb microprobe dating of early diagenetic monazite and Ar–Ar dating of marcasite nodules: Case studies from Neoproterozoic black shales in the southwestern U.S. *Geological Society of America Abstracts with Programs*, **35**, 595.
- Worsley, T.R., Nance, R.D., and Moody, J.B., 1986. Tectonic cycles and the history of the Earth's biogeochemical and paleoceanic record. *Paleoceanography*, **1**, 233–263.
- Zhang, S., Jiang, G., Zhang, J., Song, B., Kennedy, M.J., and Christie-Blick, N., 2005. U-Pb sensitive high-resolution ion microprobe ages from the Doushantuo Formation in south China: Constraints on late Neoproterozoic glaciations. *Geology*, **33**, 473–476.
- Zhou, C., Tucker, R.D., Xiao, S., Peng, Z., Yuan, X., and Chen, Z., 2004. New constraints on the ages of Neoproterozoic glaciations in south China. *Geology*, **32**, 437–440.

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