A human-induced hothouse climate?

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ABSTRACT

Hothouse climate has been approached or achieved more than a dozen times in Phanerozoic history. Geologically rapid onset of hothouses in 10^4–10^5 yr occurs as HEATT (haline euxinic acidic thermal transgression) episodes, which generally persist for less than 1 million years. Greenhouse climate preconditions conducive to hothouse development allowed large igneous provinces (LIPs), combined with positive feedback amplifiers, to force the Earth to the hothouse climate state. The two most significant Cenozoic LIPs (Columbia River Basalts and much larger Early Oligocene Ethiopian Highlands) failed to trigger a hothouse climate from icehouse preconditions, suggesting that such preconditions can limit the impact of CO_2 emissions at the levels and rates of those LIPs.

Human burning of fossil fuels can release as much CO_2 in centuries as do LIPs over 10^4–10^5 yr or longer. Although burning fossil fuels to exhaustion over the next several centuries may not suffice to trigger hothouse conditions, such combustion will probably stimulate enough polar ice melting to tip Earth into a greenhouse climate. Long atmospheric CO_2 residence times will maintain that state for tens of thousands of years.

INTRODUCTION

Human fossil-fuel burning injects CO_2 into Earth's atmosphere at geologically unprecedented rates that far outstrip natural rates of change in CO_2 emissions. Evaluating related warming in coming centuries has warranted considerable scientific attention (e.g., IPCC, 2007, and references therein). The geologic record preserves accounts of ancient warmth beyond the range of human experience and allows investigation of humanity's potential to revive such warmth. For instance, Hay (2011) concluded that human activities and related systemic feedbacks could push Earth's climate into a Mesozoic-like greenhouse climate.

Simulations of ancient climate (e.g., Berner, 2004; Park and Royer, 2011) rely on CO_2 as the master climate-controlling greenhouse gas over the long term. On geological time scales, volcanic emissions provide one critical atmospheric input of this gas. Removal of CO_2 by silicate weathering reactions results in cooling only if the carbon is buried as carbonate minerals and/or organic matter. Compensation for monotonically increasing solar luminosity by CO_2-drawdown feedbacks (e.g., Kasting and Ackerman, 1986; Kiehl and Dickinson, 1987) has been important through Earth history, and has probably confined Earth's Phanerozoic temperature range to the icehouse-greenhouse-hothouse climates discussed herein.

Climate simulations using higher temporal resolution that focus on hot climate intervals known from the geological record are increasingly successful as data sets with higher temporal and spatial resolution from ancient climates become available for model validation and calibration. Although such models have tended to underestimate the degree of ancient warmth (Kiehl, 2011), more accurate simulations are emerging. For example, Kiehl and Shields (2005) used an initial condition of 3550 ppmv (~12× the 280 ppm preindustrial CO_2 level) to accurately simulate Late Permian ocean temperatures and to reproduce predicted greenhouse-style thermohaline circulation. Assumptions of ~16× preindustrial CO_2 levels by Winguth et al. (2010) and by Huber and Caballero (2011) approximate warm climates at the Paleocene-Eocene Thermal Maximum (PETM) and in the Early Eocene, respectively.

Kidder and Worsley (2010) proposed more than a dozen geologically brief (<1 Ma) excursions from greenhouse to hothouse climate in the Phanerozoic (Table 1). These include the PETM and some oceanic anoxic event (OAE) pulses (e.g., Leckie et al., 2002), many of which are interpreted as warming intervals (e.g., Jenkyns, 2003) and have also been linked to LIP activity and extinctions (e.g., Keith, 1982; Kerr, 1998; Wignall, 2001; Keller, 2005). These hothouse pulses coincide with peaks in extinction intensity, and all but the oldest pulses are associated with a LIP trigger and related feedbacks (Table 1). Integration of numerous parameters with Earth's biogeochemical record led us (Kidder and Worsley, 2004, 2010) to suggest that a hothouse climate is not just a greenhouse intensification, but that it functionally differs from a greenhouse in ways that leave recognizable geological evidence. Our hothouse model explains the systemic interplay among factors including warmth, rapid sea-level rise, widespread ocean anoxia, ocean euxinia that reaches the photic zone, ocean acidification, nutrient crises, latitudinal expansion of desert belts, intensification and latitudinal expansion of cyclonic storms, and more. Similarly, Emanuel (2002) noted that distinct climate states are governed by critical feedbacks and interplay among factors such as large-scale atmospheric circulation, clouds, water vapor tropical cyclones, oceanic thermohaline circulation, and atmospheric CO_2.

Can rapid human climate-warming activities force the current icehouse climate into a hothouse climate? The intervals characterized by the two best-known Cenozoic LIPs shed light on the potential climate impact of LIPs as compared with human emissions. We assume that warming and cooling feedbacks (Table 2) are built into these examples, and suggest that the warming effects of the Columbia River Basalts and Ethiopian Highlands LIPs (Table 1) were weakened by icehouse preconditions.

Trajectories of human CO_2 atmospheric inputs needed to reach and/or surpass our suggested boundaries for icehouse, greenhouse, and hothouse climates are also explored. If human fossil fuel emissions can substitute for the LIP emissions that appear to have triggered hothouses under suitable ancient preconditions, a hypothetical range of human-induced climate maxima can be

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TABLE 1. HEATT EPISODES (AND TWO NON-HEATTS), LIPs, AND SELECTED HOTHOUSE-RELATED EFFECTS

HEATT and extinction age (Ma)	LIP and age of peak eruption (Ma)	Approximate extinction intensity (%) (1)	Transgression (3rd Order)	Warming	Anoxia	Euxinia (sub-photic)	Euxinia (photic zone)	$\delta^{13}C$	$\delta^{34}S$ (sulfate)	$\delta^{15}N$ <2‰
Mid-Miocene Climate Optimum no HEATT	Columbia River Basalt (15.6–16.0) (2)	5	y(3)	y(4)	y(4)					
No HEATT	Ethiopian Highlands (29–31)	5								
Paleocene-Eocene Thermal Maximum (55)	NAIP (58–62)	5	y	y	lim	y	y	n		y(5)
End-Cretaceous (65)	Deccan Traps (66)	30	y	y	lim	y	y	n		y
Cenomanian-Turonian OAE 2 (93)	Ontong Java II (86–94)	10	y	y	y	y	y	p/n	p	y
Early Aptian OAE 1a (120)	Ontong Java I (119–125)	5	y	y	y	y	y	p/n		y
Toarcian-Pliensbachian (183)	Karoo-Ferrar (179–183)	10	y	y	y	y	y	n	p(6)	y
End-Triassic (200)	CAMP (200)	30	y	y	y	y	y	n	p/n	y
End-Permian (251)	Siberian Traps (249–251)	55	y	y	y	y	y	n	p	y
Hangenberg (359)	East European Platform (365)	20	y(7)		y(7)	y(7)	y(7)	p(8)		
Frasnian-Famennian (374)	Viluy Traps (ca. 373) (9)	25	y	y	y	y	y	p/n	p	y
Late Ordovician (444)		30	y	y	y		y	n		y
SPICE (499)		40	y(10)	y(11)	y(10)	y(10)		p(10)	p(10)	
Botomian (ca. 520)	Antrim (ca. 510)	40	y		y			n	p	
Ediacaran (542)		?	y		y			n		y

Note: HEATT—haline euxinic acidic thermal transgression; LIP—large igneous province; OAE—oceanic anoxic events; NAIP—North Atlantic igneous province; CAMP—Central Atlantic magmatic province. See Kidder and Worsley (2010) or Large Igneous Provinces Commission website (www.largeigneousprovinces.org) for more complete coverage of LIPs. Symbols are defined as follows: y—yes; p—positive; n—negative; lim—limited. Most information listed in the boxes is referenced in Kidder and Worsley (2010) except for some new references which are keyed to numbers in parentheses as follows, including new age dates on the Viluy LIP: (1) Rohde and Muller (2005); (2) Barry et al. (2010); (3) John et al. (2011); (4) Kender et al. (2009); (5) Knies et al. (2008); (6) Owens et al. (2010); (7) Marynowski and Filipiak (2007); (8) Kaiser et al. (2006); (9) Courtillot et al. (2010); (10) Gill et al. (2011); (11) Elrick et al. (2011).

TABLE 2. EXAMPLES OF WARMING AMPLIFIERS AND COOLING FEEDBACKS TO WARMING

	Estimated time scale (yr)
Warming Amplifiers	
Methane release from tundra, peat, seabed	10^2–10^4
Polar cloud heat retention plus polar precipitation	10^2
Increased mid-latitude insolation as desert belts expand	10^2
Warm-brine sinking	10^2
Polar upwelling of desert-belt generated brine	10^2
High absolute humidity and poleward atmospheric latent heat transport across cloud-free mid-latitudes	10^2
Lower polar albedo with loss of sea ice plus development of polar forests (sea ice was already gone preceding most ancient HEATT episodes)	10^2–10^4
Seafloor geothermal heating-driven polar upwelling via haline mode overturn	10^4
“Tropical” cyclone effects (upwelling, stratospheric water injection, increased poleward heat transport)	10^2–10^4
Cooling Feedbacks to Warming	
Increased silicate weathering and carbonate burial (carbonate burial hampered in acidic oceans)	10^6–10^7
Burial of organic matter in black shales	10^5–10^6

inferred from considering CO_2 emission levels in scenarios such as those of (1) human actions to mitigate climate change, (2) forced mitigation by societal collapse of human economies, and (3) successful rapid exhaustion of fossil-fuel resources.

DEFINING ICEHOUSE, GREENHOUSE, AND HOTHOUSE CLIMATES

Figures 1 and 2 distinguish icehouse, greenhouse, and hothouse climate states. Icehouses have major polar ice caps that calve marine icebergs. A cool greenhouse can have small polar ice caps and Alpine glaciers, but no ice sheets that calve icebergs. Glaciations in the Late Devonian, Late Eocene, and just prior to the Late Ordovician icehouse were probable cool greenhouse climates. A warm greenhouse may have seasonal sea ice as the only polar ice. Thermohaline circulation (Figs. 1 and 2) reflects differing climate states. The “thermal mode” (Zhang et al., 2001) describes the strong pole-driven sinking cold icehouse brines. They acknowledged the weaker polar sinking of brines in the greenhouse climate of the Late Permian as modeled by Hotinski et al. (2001). The “haline mode” describes sinking warm brine driven by evaporation (Zhang et al., 2001). Kidder and Worsley (2004) suggested that such evaporation-driven sinking of brines would be most effective where evaporation in embayments and larger restricted settings (e.g., Mediterranean Sea, Persian Gulf) would feed brines into the deep ocean when pole-driven sinking ceased. Such basins would generate warm brines with increasing potency as transgression expands their surface area (Kidder and Worsley, 2010). Although Zhang et al. (2001) suggested the haline mode was unstable, Kidder and Worsley (2010) proposed that peak LIP forcing can sustain the haline mode.

Other critical changes from icehouse to greenhouse to hothouse (Figs. 1 and 2) include reductions in pole-equator thermal contrast, planetary windbelt velocity, wind shear, and wind erosive power (Kidder and Worsley, 2010). Oceanic anoxia and euxinia expand as climate warms (e.g., Wignall, 2001; Kidder and Worsley, 2004; Wignall et al., 2010), and euxinia moves into the photic zone (e.g., Kump et al., 2005; Kidder and Worsley, 2010) in hothouses. Tropical cyclonic storms strengthen, extend to high

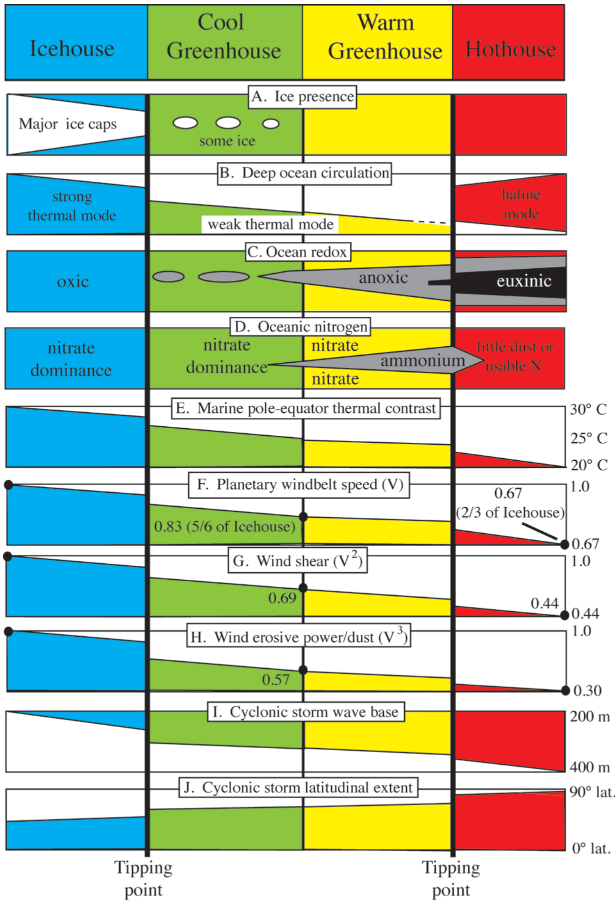

Figure 1. Factors useful in distinguishing icehouse, cool greenhouse, warm greenhouse, and hothouse. Boxes A–D are conceptual. Boxes E–H are based on semiquantitative estimates developed in table 2 of Kidder and Worsley (2010). Approximations in Boxes I and J are supported by modeling of Korty et al. (2008) and by Fedorov et al. (2010), respectively. See text for explanation.

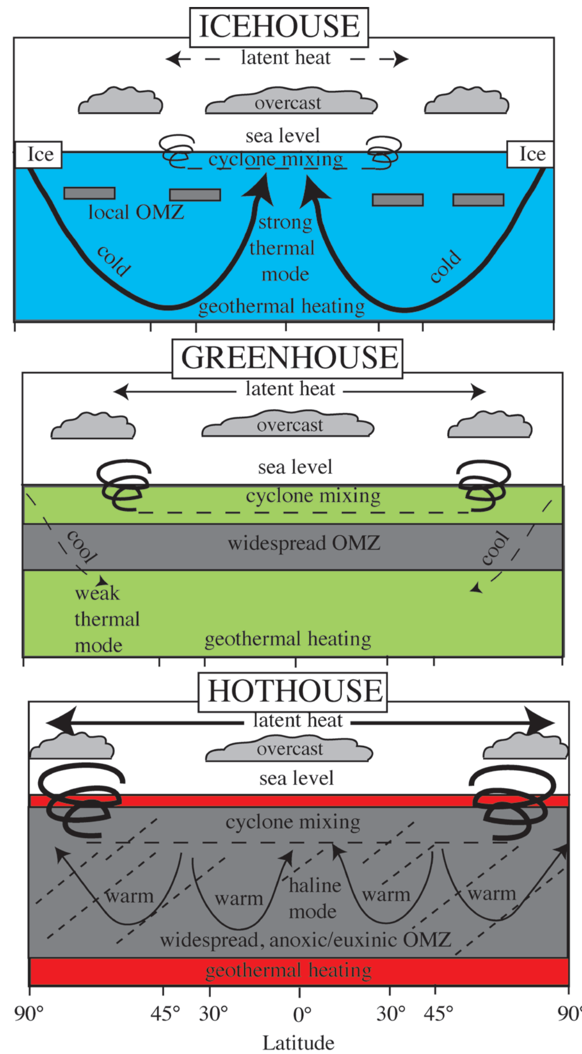

Figure 2. Schematic view of some aspects of icehouse versus greenhouse versus hothouse after Kidder and Worsley (2010). Key factors in the progressive steps from icehouse to hothouse include shifting deep-ocean circulation from thermal to haline mode, expansion of anoxia and euxinia, weakening of planetary windbelts and hence wind-driven upwelling and eolian dust transport to oceans. Also critical is increased cyclonic storm mixing that develops as tropical storms expand their reach to high latitudes and into deeper waters. OMZ—oxygen minimum zone.

latitude, and reach perhaps twice as deeply as those in modern oceans (e.g., Emanuel, 2005; Kidder and Worsley, 2004, 2010; Korty et al., 2008). "Tropical" cyclones that reach polar latitudes help maintain moist and mild climates there by drawing up warm waters via upwelling, thus promoting heat-trapping cloud cover.

Increased polar precipitation generates freshwater runoff (e.g., Kidder and Worsley, 2004, 2010; Sluijs et al., 2011) that can hamper thermal-mode polar deep-water formation. Models for a cap of low-salinity water are consistent with such weakening as polar rainfall and humidity increase in a warming world (e.g., Manabe et al., 1994; Abbot and Tziperman, 2009). Support for such conditions in the geologic record includes a temperate, moist, mid-Pliocene Arctic Ocean (Ballantyne et al., 2010; Fedorov et al., 2010); a warm mid-Miocene Climate Optimum with no coastal ice sheets (e.g., Tripati et al., 2009); a warm Southern Ocean sea surface from mid-Jurassic to Early Cretaceous (Jenkyns et al., 2011); and high-paleolatitude fossil forests at a number of geologic intervals (e.g., Retallack and Alonso-Zarza, 1998; Taylor et al., 2000; Jahren, 2007).

HEATT EPISODES

Kidder and Worsley (2010) proposed that hothouse climates develop via HEATT (haline euxinic acidic thermal transgression) episodes (Fig. 3). The rapid transgression (Table 1; Fig. 3) occurs with deep-ocean warming fed by desert-belt sinking of warm brine, and thermal expansion of ocean water raises relative sea level by up to 20 m. The LIP trigger rapidly emits substantial amounts of carbon dioxide (Fig. 4), but not enough to produce the negative $\delta^{13}C$ excursions (Table 1) that typify most HEATTs (e.g., Erwin, 1993, and references therein). Further warming feedbacks (Table 2) collectively force Earth from HEATT-susceptible warm greenhouse preconditions to a hothouse climate (Kidder and Worsley, 2010).

DID ICEHOUSE PRECONDITIONS WEAKEN THE IMPACT OF TWO CENOZOIC LIPs?

The cooling influences of both collisional orogenesis and the Antarctic circum-polar current and perhaps other icehouse-

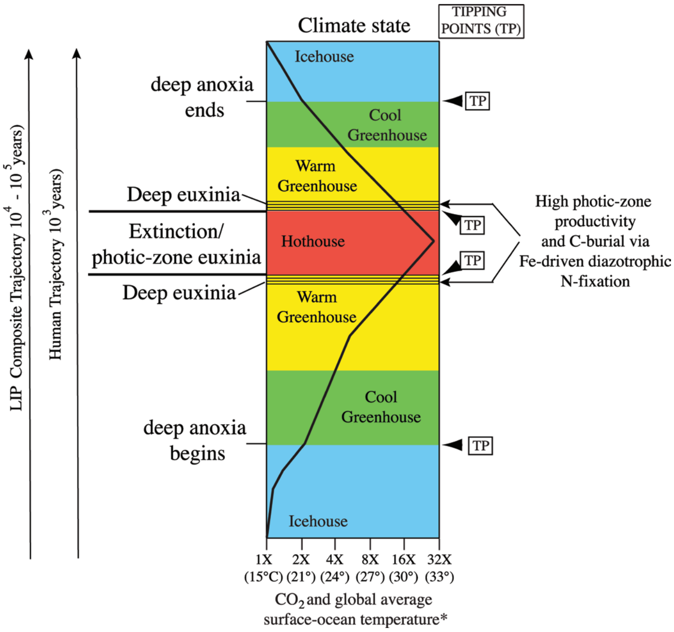

Figure 3. Progression of developments during a HEATT episode after Kidder and Worsley (2010). Icehouse climate sensitivity of Park and Royer (2011) has been adopted. Carbon dioxide thresholds needed for achieving cool greenhouse, warm greenhouse, and hothouse planetary states are suggested using the 280 ppm preindustrial level and today's solar constant. High productivity of diazotrophs and green-algal phytoplankton coupled with increased carbon-burial rate and efficiency as anoxia expands hampers achievement of the HEATT peak unless warming factors and feedbacks can overcome that obstacle. Similar carbon-burial rate after the HEATT peak accelerates cooling from hothouse to warm greenhouse.

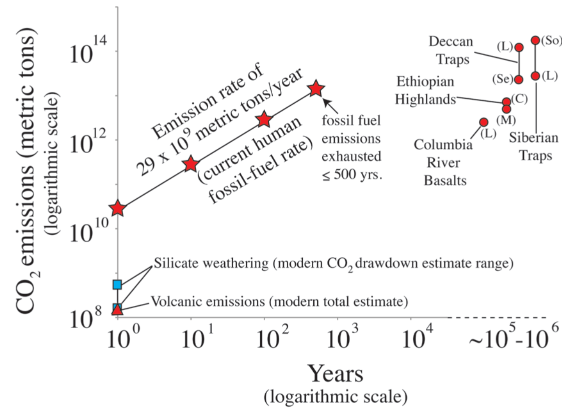

Figure 4. Some CO_2 inputs and outputs to/from Earth's atmosphere. Extrapolated human carbon dioxide emission rates (red stars) (www.eia.gov; U.S. Energy Information Administration, 2010) as compared to selected LIP emission estimates over an assumed duration of 10^5 yr. LIP CO_2 rates were calculated from LIP volume estimates multiplied by 14×10^6 metric tons of CO_2 emitted for each cubic kilometer of basalt (Self et al., 2006). Sources of volume estimates are labeled as follows: L—Large Igneous Province Commission website (www.largeigneousprovinces.org); C—Courtillot et al. (1999); M—Mohr (1983); Se—Self et al. (2006); So—Sobolev et al. (2011). This calculation does not consider amplification by contact metamorphism or clathrate release. Rates of silicate weathering drawdown (blue squares) are from Hilley and Porder (2008). Average annual volcanic emissions (red triangle) are from Williams et al. (1992). Values from Gaillardet and Galy (2008) for silicate weathering drawdown (5.1×10^8 metric tons of CO_2/yr) and volcanic plus metamorphic release (3.0×10^8 metric tons of CO_2/yr) are consistent with values shown in this diagram.

precondition hurdles may have hampered the warming influence of the small Columbia River Basalts (CRB) LIP and the larger Ethiopian Highlands (EH) LIP (Table 1). Larger and older LIPs dwarf these examples in CO_2-emission potential, but both Cenozoic LIPs are closer to potential volumes of human CO_2 emissions (Fig. 4).

Increased silicate weathering during the Himalayan continental collision has long been considered as a stimulus for the onset and sustenance of the enduring (ca. 35 Ma) Cenozoic icehouse (e.g., Chamberlin, 1899; Raymo, 1991). Likewise, Gondwanaland's collision with Laurasia to form Pangea may have triggered and helped to sustain the even longer-lasting (ca. 70 Ma) late Paleozoic icehouse (Kidder and Worsley, 2010). Temporal correlation of these prolonged orogenies with icehouse climate (Kidder and Worsley, 2010) is prima facie evidence for orogenically driven CO_2 drawdown and carbon burial. Climate cooling via Himalayan silicate weathering has been challenged by reports of high rates of metamorphic degassing of CO_2 in orogenic systems (e.g., Evans et al., 2008; Skelton, 2011). Such arguments against orogenically driven cooling need to offer an alternative mechanism for CO_2 drawdown to explain Cenozoic cooling. That organic carbon burial in the Bengal Fan outstrips estimates of Himalayan silicate weathering (France-Lanord and Derry, 1997) points to carbon burial as the bottom line in cooling. Other aspects of Himalayan orogenesis that favor carbon burial (e.g., nutrient release via silicate weathering, stimulation of iron-dust delivery to oceans, and ocean upwelling by monsoonal winds) need more thorough tracking to better account for the overall impact of the Himalayas on the carbon cycle.

You et al. (2009) noted global average temperature during the Middle Miocene Climate Optimum (MMCO) was ~3 °C warmer than at present, suggesting the MMCO as an analog to predicted warming over the next century. Deep ocean Miocene warming by <2 °C (Zachos et al., 2001) is consistent with a global average model temperature increase of ~1.5 °C (Herold et al., 2012) and a rise in atmospheric CO_2 during the MMCO coincided with the eruption of the CRB LIP (Zachos et al., 2001; Kender et al., 2009; You et al., 2009; Barry et al., 2010). The CO_2 increase may have been only ~50 ppm (Tripati et al., 2009) to 100 ppm (Kürschner et al., 2008) higher than pre-MMCO levels. CO_2 emissions from this LIP were probably insufficient to force a hothouse climate, but the CRB probably emitted as much CO_2 as human fossil-fuel burning will release in the next century (Fig. 4). Miocene Earth-cooling preconditions may have offset the CRB emissions in pulses distributed over >400,000 yr (Self et al., 2006; Barry et al., 2010). The Miocene atmospheric CO_2 gain of 50–100 ppm has already been surpassed by the 110 ppm increase since the nineteenth century.

The larger Ethiopian Highlands (Afar) LIP has an estimated eruptive volume 2×–3× larger than the CRB (Fig. 4). Despite the correspondingly larger volume of calculated CO_2 emissions (Fig. 4), the EH LIP failed to warm climate even at its eruptive peak, which occurred just after the establishment of the Antarctic ice cap. This volcanism began ca. 31 Ma, peaked at ca. 30 Ma, and then declined to lower levels of activity that persist today as part of the Red Sea system (e.g., Courtillot et al., 1999). This LIP volcanism followed sharp cooling at the end of the Eocene that lasted from ca. 34 to 33 Ma (Zachos et al., 2001) as the Antarctic ice cap formed and expanded. That sharp Oi-1 cooling episode

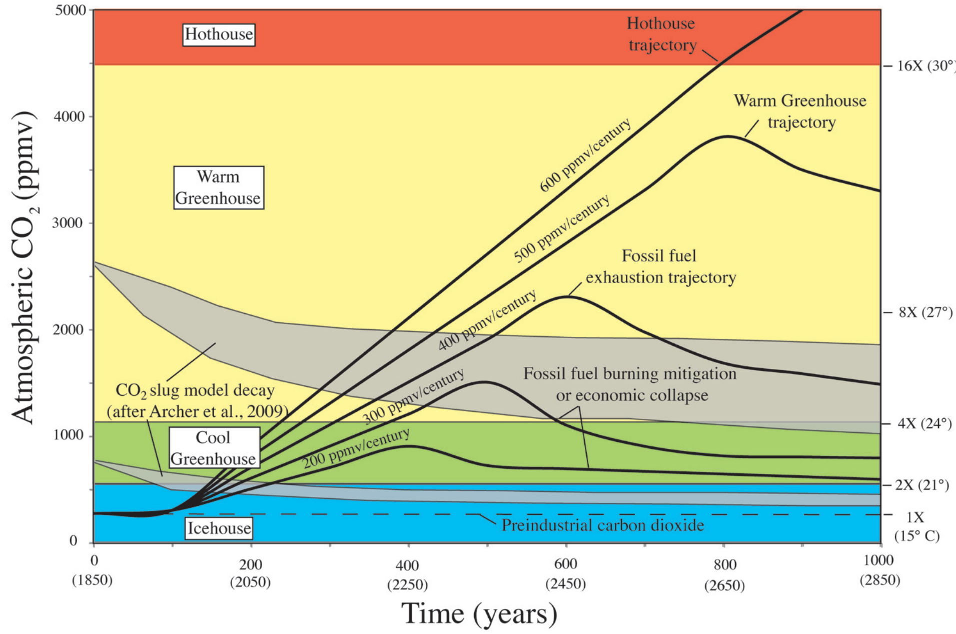

Figure 5. Projected examples of a range of future carbon dioxide emission scenarios plotted against the backdrop of climate state thresholds taken from Figure 3. The carbon dioxide trajectory curves begin with an assumed rate increasing CO_2 by 200 ppmv/century (our modern rate of 2 ppm/yr). The other curves show the hypothetical effect of increasing that rate by increments of 100 ppmv/century when plotted against the icehouse, greenhouse, hothouse thresholds developed herein. We assume various reasons for initiation of curve declines (e.g., human intervention, economic collapse, exhaustion of fossil fuels). Atmospheric declines in CO_2 with time approximate model results of Archer et al. (2009) that show increasing residence time of CO_2 as the size an instantaneous slug (injection) increases. The starting point for the Archer et al. (2009) CO_2 injections is arbitrarily placed at 1850 so as to distinguish those slugs from the slower rates of human injection shown by the trajectory curves. We follow the suggestion of Park and Royer (2011) that temperature sensitivity to CO_2 doublings is more substantial in icehouses (6–8 °C) than in warmer climate states (3–4 °C). We adopt the low end of both sensitivity ranges in this figure.

(Zachos et al., 2001) was followed by warming of deep ocean waters by ~2–3 °C from ca. 33 to 32 Ma. This warming occurred *before* the EH LIP eruptions. Warming did not intensify with the onset of the LIP, suggesting that it could not disrupt the icehouse precondition established with the formation of the Antarctic ice cap. A likely supporting cooling factor was the thermal isolation of Antarctica via development of the circum-polar Antarctic current as the Tasmanian Gateway opened and deepened (Kennett et al., 1974; Katz et al., 2011). An ocean-isolated polar continent is a unique configuration for the Phanerozoic. Its cooling effect plus that of the Himalayan cooling influence discussed earlier may have weakened the climate impact of both the CRB and EH eruptive pulses at their respective rates and magnitudes of CO_2 emission.

TIPPING TOWARD A HOTHOUSE?

Forcing a hothouse requires melting of all polar ice and the breakdown of the thermal mode of oceanic deep-water circulation. Only then can desert-belt evaporation drive the haline mode. Modern polar glaciers are melting unexpectedly rapidly, particularly when water drains beneath them (e.g., Overpeck et al., 2006; Chen et al., 2009). This water accelerates melting and lubricates glacial flow, speeding outlet glaciers toward the sea. The predicted rapid breakup of Antarctic ice shelves (Mercer, 1970) has been under way since the 1990s. Removal of this ice-shelf barrier allows seaward acceleration of glacial flow (e.g., Overpeck et al., 2006). However, as long as sufficient seasonal sea ice forms and evaporative katabatic winds from ice caps are maintained, polar sinking of brines will sustain the thermal circulation mode. Sinking boreal brines will diminish with the loss of the Greenland ice sheet and perennial Arctic sea-ice, leaving the colder and concurrently weakening austral system as the only significant cold-brine generator. The modern circum-polar current that thermally isolates Antarctica from warm surface currents favors a dry polar climate with little cloud cover (Fig. 2), resulting in significant radiative heat loss, which helps keep the polar climate colder than a moist polar atmosphere characterized by higher relative overcast (Fig. 2).

GEOLOGICAL UPTAKE AND EMISSION OF CARBON DIOXIDE

Geological uptake and emission of CO_2 are difficult to measure precisely. Modern volcanic CO_2 is apparently emitted more slowly than silicate weathering draws down CO_2 (Fig. 4), but both sets of estimates are difficult to project because of the very short baseline from which to extrapolate. Nevertheless, geologically rapid injection of CO_2 into the atmosphere by LIPs and associated feedbacks (Table 2) probably overwhelms cooling feedbacks sufficiently to force climate from a greenhouse to hothouse state in some cases.

The CO_2 contribution of LIPs with known volumes can be crudely estimated (Fig. 4) via the Self et al. (2006) suggestion that each cubic kilometer of basalt erupted releases 14 million metric tons (T) of CO_2 to the atmosphere. Self et al. (2006) proposed that much LIP activity may occur as short (10–50 yr) pulses, separated by long intervals. Barry et al. (2010) suggest that eruptive pulses during the 420 ka of the most voluminous phase of CRB outpourings were separated by hiatuses averaging 4 ka.

Atmospheric retention of 15%–35% of a slug (instantaneous model injection) of CO_2 for at least 10,000 yr (Archer et al., 2009) is governed by factors such as seawater uptake, reaction with seawater carbonates, and silicate weathering. Therefore, if successive LIP basalt eruptions average <10^4 yr in recurrence frequency (Barry et al., 2010), total atmospheric CO_2 will build with each successive eruption. The two CO_2 slugs of Archer et al. (2009) are portrayed in Figure 5. The smaller (1000 Pg of carbon) slug represents past human CO_2 emissions plus those expected by the end of the twenty-first century. The larger (5000 Pg C) slug is that expected from burning "the entire reservoir" of fossil fuels. A warming ocean's ability to absorb CO_2 weakens, and its $CaCO_3$ will likely dissolve more slowly than model predictions (e.g., Hay, 2011). Silicate weathering rates can increase in warm, CO_2-rich atmospheres (e.g., Walker and Kasting, 1992; Lenton and Britton, 2006). Still, Lenton and Britton (2006) suggested that >1 million years are needed to return atmospheric CO_2 to the levels present before an emission slug.

Self et al. (2006) proposed that silicate weathering of LIP basalts would minimize warming by quickly drawing down CO_2. However, silicate-weathering rates are probably too slow to draw down atmospheric CO_2 rapidly enough to negate warming effects (e.g., Lenton and Britton, 2006; Archer et al., 2009). Furthermore, much of the LIP basalt will be buried beneath the youngest basalt flows, allowing chemical weathering of only a small fraction of the basalt. Nevertheless, geologically rapid cooling is evident during at least some waning HEATT episodes such as the Cenomanian/Turonian OAE. We suggest that such cooling may be biologically driven as diazotrophic (N-fixing) cyanobacteria capitalize on iron-rich anoxic waters. These and associated green-algal phytoplankton will stimulate a pulse of organic carbon burial and cooling as euxinia reverts to anoxia as sulfide in the ocean's water column is buried (Fig. 3). For example, rapid cooling during the Cenomanian/Turonian OAE (e.g., Jenkyns, 2003) was probably driven by rapid organic carbon burial during waning of a HEATT episode driven by the oceanic Ontong-Java LIP (Table 1) that would weather slowly underwater (e.g., Berner, 2004). However, rapid organic matter burial in the absence of carbonate burial in acidic oceans may only compensate for the temporary loss of carbonate burial and may not greatly increase carbon burial. See Kidder and Worsley (2010) for further discussion of anoxia, euxinia, and N-fixing as applied to onset and decline of hothouse climates.

HOW MUCH CAN HUMANS FORCE CLIMATE?

Human fossil-fuel emissions (even without factors such as methane release, forest destruction, and cement production) can rival, in centuries, the CO_2 that LIPs emit over 10^4–10^5 yr or more (Fig. 4). Continued current rates of CO_2 emission from fossil fuel burning will, in ~100 yr, match the CO_2 release from the entire CRB LIP (Fig. 4). Fossil fuels would be exhausted before their emissions approach the totals of larger LIPs such as the Deccan Traps or the Siberian Traps (Fig. 4). This crude order-of-magnitude discussion shows that human rates of CO_2 emissions outstrip LIP volcanic emission rates by two orders of magnitude and outcompete silicate-weathering rates and organic matter burial feedback, even during the ongoing Himalayan orogeny.

Figure 5 projects CO_2 trajectories against the backdrop of icehouse-greenhouse-hothouse boundaries shown in Figure 3. Direct human input of CO_2 to the atmosphere will diminish sharply with mitigation, societal collapse, or fossil fuel exhaustion. Feedbacks such as methane emissions will likely amplify warming if they are fast enough, but the hothouse trajectory would probably require more than methane (e.g., Cui et al., 2011; Kump, 2011). Although the maximum potential human emissions of CO_2 will surpass those of the CRB, the duration will be so short in comparison (Fig. 4) that some positive feedbacks in the Earth system (Table 2) may not have time to establish a hothouse. For example, the rapidly initiated PETM did not develop as fully as other HEATTs (Kidder and Worsley, 2010), probably because its trigger was not sustained even amid HEATT-favoring preconditions. Even the 20,000-yr warm-up modeled by Cui et al. (2011) is probably short compared to older HEATTs. So, even if a human-induced hothouse is unlikely, a warm greenhouse may develop as high CO_2 emission rates overwhelm the "protection" exercised by the present icehouse precondition. Pushing the planet from a cool greenhouse to a warm greenhouse will require melting of all Antarctic ice. We speculate that the circum-polar current may hamper this melting, given the failure of the CRB and EH LIPs to melt the smaller-than-modern Antarctic ice cap. Long residence times modeled for atmospheric CO_2 (Archer et al., 2009) would sustain warmth, allowing slow-acting factors such as deep-ocean circulation to adjust. However, such long residence times were in force during and after the CRB and EH LIP eruptions. Furthermore, warming feedbacks (Table 2) would have been active during the CRB and EH eruptions. The higher-than-modern rates of atmospheric CO_2 increase needed to reach a warm greenhouse in centuries (Fig. 5) would require those feedbacks. Figure 5 suggests that, as Earth warms, it becomes increasingly insensitive to CO_2 forcing as atmospheric CO_2 levels rise. For example, doubling CO_2 from 280 to 560 ppm yields an approximate global average temperature increase of 6 °C (Fig. 5). Note that sensitivity to CO_2 doublings is higher in icehouses than in warmer climates (Park and Royer, 2011). Doubling CO_2 at higher values (e.g., 2200–4400 ppm) raises global average temperature by ~3 °C. Even though the rate of warming slows, the higher CO_2 levels ensure that warmth will probably persist for millennia (Archer et al., 2009; Fig. 5).

Finally, humans may not burn all fossil fuels. A hopeful reason is that energy and carbon strategies will reduce atmospheric CO_2 emissions. A pessimistic view is that calamities such as floods, droughts, crop losses, cyclones, and sea level that rises tens of meters will displace populations. Human migrations, conflicts, and economic crises will sharply curtail fossil fuel emissions.

CONCLUSIONS

Humans can raise global atmospheric CO_2 to levels known from much warmer ancient climates (e.g., Hay, 2011; Kiehl, 2011). Conditions in some of those warm climates will probably be achieved if current levels of carbon emissions continue, although precise prediction of the degree and rate of warming is difficult. A cool greenhouse similar to the MMCO in which the tropics and deep sea warm, most northern ice melts, and perhaps half of the Antarctic ice disappears appears possible within centuries. A warm greenhouse is also possible, although reaching it faces steeper precondition hurdles.

We suspect it will be difficult for humans to force Earth from the current icehouse to a hothouse. The likely cool greenhouse in which about half of Antarctica is still ice-covered means devastation from the tens of meters sea level is likely to rise (e.g., Ward, 2010), and poleward shifting of warm climate belts. Although a hothouse may not occur because economic crises or intentional climate-mitigating efforts by humans or fossil-fuel exhaustion limit greenhouse gas emissions, even a cool greenhouse climate will severely disrupt many societies and economies.

Feedbacks (Table 2) and still-unknown amplifiers will ultimately control just how far humans can force climate toward a hothouse. Uncertainties over these feedbacks should not distract us from the likelihood that a cool greenhouse seems imminent within perhaps a century or two. Long atmospheric CO_2 residence times will probably keep Earth from returning to an icehouse for centuries to millennia unless active removal of CO_2 from the atmosphere is undertaken.

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REFERENCES CITED

Abbot, D.S., and Tziperman, E., 2009, Controls on the activation and strength of a high-latitude convective cloud feedback: Journal of the Atmospheric Sciences, v. 66, p. 519–529, doi:10.1175/2008JAS2840.1.

Archer, D., Eby, M., Brovkin, V., Ridgewell, A., Cao, L., Mikolajewicz, U., Caldeira, K., Matsumoto, K., Munhoven, G., Montenegro, A., and Tokos, K., 2009, Atmospheric lifetime of fossil fuel carbon dioxide: Annual Review of Earth and Planetary Sciences, v. 37, p. 117–134, doi:10.1146/annurev.earth.031208.100206.

Ballantyne, A.P., Greenwood, D.R., Sinninghe Damsté, J.S., Csank, A.Z., Eberle, J.J., and Rybczynski, N., 2010, Significantly warmer Arctic surface temperatures during the Pliocene indicated by multiple proxies: Geology, v. 38, p. 603–606, doi:10.1130/G30815.1.

Barry, T.L., Self, S., Kelley, S.P., Reidel, S., Hooper, P., and Widdowson, M., 2010, New $^{40}Ar/^{39}Ar$ dating of the Grande Ronde lavas, Columbia River Basalts, USA: Implications for duration of flood basalt episodes: Lithos, v. 118, p. 213–222, doi:10.1016/j.lithos.2010.03.014.

Berner, R.A., 2004, The Phanerozoic Carbon Cycle: Oxford, Oxford University Press, 150 p.

Chamberlin, T.C., 1899, An attempt to frame a working hypothesis on the cause of glacial periods on an atmospheric basis: The Journal of Geology, v. 7, p. 545–584, 667–685, 751–787, doi:10.1086/608449.

Chen, J.L., Wilson, C.R., Blankenship, D., and Tapley, B.D., 2009, Accelerated Antarctic ice loss from satellite gravity measurements: Nature Geoscience, v. 2, p. 859–862, doi:10.1038/ngeo694.

Courtillot, V., Jaupart, C., Manighetti, I., Tapponnier, P., and Besse, J., 1999, On causal links between flood basalts and continental breakup: Earth and Planetary Science Letters, v. 166, p. 177–195, doi:10.1016/S0012-821X(98)00282-9.

Courtillot, V., Kravchinsky, V.A., Quidelleur, X., Renne, P., and Gladkochub, D.P., 2010, Preliminary dating of the Viluy traps (Eastern Siberia): Eruption at the time of Late Devonian extinction events?: Earth and Planetary Science Letters, v. 300, p. 239–245, doi:10.1016/j.epsl.2010.09.045.

Cui, Y., Kump, L.R., Ridgewell, A.J., Charles, A.J., Junium, C.K., Diefendorf, A.F., Freeman, K.H., Urban, N.M., and Harding, I.C., 2011, Slow release of fossil carbon during the Palaeocene-Eocene thermal maximum: Nature Geoscience, v. 4, p. 481–485, doi:10.1038/ngeo1179.

Elrick, M., Rieboldt, S., Saltzman, M., and McKay, R.M., 2011, Oxygen-isotope trends and seawater temperature changes across the Late Cambrian Steptoean positive carbon-isotope excursion (SPICE event): Geology, v. 39, p. 987–990, doi:10.1130/G32109.1.

Emanuel, K., 2002, A simple model of multiple climate regimes: Journal of Geophysical Research, v. 107, p. 4–1 to 4–10.

Emanuel, K., 2005, Increasing destructiveness of tropical cyclones over the past 30 years: Nature, v. 436, p. 686–688, doi:10.1038/nature03906.

Erwin, D.H., 1993, The Great Paleozoic Crisis: Life and Death in the Permian: New York, Columbia University Press, 327 p.

Evans, M.J., Derry, L.A., and France-Lanord, C., 2008, Degassing of the metamorphic carbon dioxide from the Nepal Himalaya: Geochemistry Geophysics Geosystems, v. 9, Q04021, 18 p., doi:10.1029/2007GC001796.

Fedorov, A.V., Brierley, C.M., and Emanuel, K., 2010, Tropical cyclones and permanent El Niño in the early Pliocene epoch: Nature, v. 463, p. 1066–1070, doi:10.1038/nature08831.

France-Lanord, C., and Derry, L.A., 1997, Organic carbon burial forcing of the carbon cycle from Himalayan erosion: Nature, v. 390, p. 65–67, doi:10.1038/36324.

Gaillardet, J., and Galy, A., 2008, Himalaya—Carbon sink or source?: Science, v. 320, p. 1727–1728, doi:10.1126/science.1159279.

Gill, B.C., Lyons, T.W., Young, S.A., Kump, L.R., Knoll, A.H., and Saltzman, M.R., 2011, Geochemical evidence for widespread euxinia in the Later Cambrian ocean: Nature, v. 469, p. 80–83, doi:10.1038/nature09700.

Hay, W.W., 2011, Can humans force a return to a "Cretaceous" climate?: Sedimentary Geology, v. 235, p. 5–26, doi:10.1016/j.sedgeo.2010.04.015.

Herold, N., Huber, M., and Müller, R.D., 2012, Modeling the Miocene climate optimum, Part 1: Land and atmosphere: Journal of Climate, doi:10.1175/2011JCLI4035.1 (in press).

Hilley, G.E., and Porder, S., 2008, A framework for predicting global silicate weathering and drawdown rates over geologic time scales: Proceedings of the National Academy of Sciences of the United States of America, v. 105, p. 16,855–16,859, doi:10.1073/pnas.0801462105.

Hotinski, R.M., Bice, K.L., Kump, L.R., Najjar, R.G., and Arthur, M.A., 2001, Ocean stagnation and end-Permian anoxia: Geology, v. 29, p. 7–10, doi:10.1130/0091-7613(2001)029<0007:OSAEPA>2.0.CO;2.

Huber, M., and Caballero, R., 2011, The early Eocene equable climate problem revisited: Climate of the Past Discussions, v. 7, p. 241–304, doi:10.5194/cpd-7-241-2011.

IPCC, 2007, Climate Change 2007: The Physical Science Basis, *in* Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., and Miller, H.L., eds., Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change: Cambridge, Cambridge University Press, 996 p.

Jahren, A.H., 2007, The Arctic forest of the Middle Eocene: Annual Review of Earth and Planetary Sciences, v. 35, p. 509–540, doi:10.1146/annurev.earth.35.031306.140125.

Jenkyns, H.C., 2003, Evidence for rapid climate change in the Mesozoic-Palaeogene greenhouse world: Philosophical Transactions of the Royal Society, v. 361, p. 1885–1916, doi:10.1098/rsta.2003.1240.

Jenkyns, H.C., Schouten-Huibers, L., Schouten, S., and Sinninghe Damsté, J.S., 2011, Middle Jurassic–Early Cretaceous high-latitude sea-surface temperatures from the Southern Ocean: Climate of the Past Discussions, v. 7, p. 1339–1361, doi:10.5194/cpd-7-1339-2011.

John, C.M., Karner, G.D., Browning, E., Leckie, R.M., Mateo, Z., Carson, B., and Lowery, C., 2011, Timing and magnitude of Miocene eustasy derived from the mixed siliciclastic-carbonate stratigraphic record of the northeastern Australian margin: Earth and Planetary Science Letters, v. 304, p. 455–467, doi:10.1016/j.epsl.2011.02.013.

Kaiser, S.I., Steuber, T., Becker, R.T., and Joachimski, M.M., 2006, Geochemical evidence for major environmental change at the Devonian-Carboniferous boundary in the Carnic Alps and Rhenish Massif: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 240, p. 146–160, doi:10.1016/j.palaeo.2006.03.048.

Kasting, J.F., and Ackerman, T.P., 1986, Climatic consequences of very high carbon dioxide levels in Earth's early atmosphere: Science, v. 234, p. 1383–1385, doi:10.1126/science.11539665.

Katz, M.E., Cramer, B.S., Toggweiler, J.R., Esmay, G., Liu, C., Miller, K.G., Rosenthal, Y., Wade, B.S., and Wright, J.D., 2011, Impact of Antarctic circumpolar current development on late Paleogene ocean structure: Science, v. 332, p. 1076–1079, doi:10.1126/science.1202122.

Keith, M.L., 1982, Violent volcanism, stagnant oceans and some inferences regarding petroleum, strata-bound ores and mass extinctions: Geochimica et Cosmochimica Acta, v. 46, p. 2621–2637, doi:10.1016/0016-7037(82)90382-9.

Keller, G., 2005, Impacts, volcanism and mass extinction: Random coincidence or cause and effect: Australian Journal of Earth Sciences, v. 52, p. 725–757, doi:10.1080/08120090500170393.

Kender, S., Peck, V.L., Jones, R.W., and Kaminski, M.A., 2009, Middle Miocene oxygen minimum zone expansion offshore West Africa: Evidence for global

precursor events: Geology, v. 37, p. 699–702, doi:10.1130/G30070A.1.
Kennett, J.P., Houtz, R.E., Andrews, P.B., Edwards, A.R., Gostin, V.A., Hajos, M., Hampton, M.A., Jenkins, D.G., Margolis, S.V., Ovenshine, A.T., and Perch-Nielsen, K., 1974, Development of the Circum-Antarctic current: Science, v. 186, p. 144–147, doi:10.1126/science.186.4159.144.
Kerr, A.C., 1998, Oceanic plateau formation: A cause of mass extinction and black shale deposition around the Cenomanian-Turonian boundary: Journal of the Geological Society of London, v. 155, p. 619–626, doi:10.1144/gsjgs.155.4.0619.
Kidder, D.L., and Worsley, T.R., 2004, Causes and consequences of extreme Permo-Triassic warming to globally equable climate and relation to the Permo-Triassic extinction and recovery: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 203, p. 207–237, doi:10.1016/S0031-0182(03)00667-9.
Kidder, D.L., and Worsley, T.R., 2010, Phanerozoic Large Igneous Provinces (LIPs), HEATT (Haline Euxinic Acidic Thermal Transgression) episodes, and mass extinctions: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 295, p. 162–191, doi:10.1016/j.palaeo.2010.05.036.
Kiehl, J.T., 2011, Lessons from Earth's Past: Science, v. 331, p. 158–159, doi:10.1126/science.1199380.
Kiehl, J.T., and Dickinson, R.E., 1987, A study of the radiative effects of enhanced atmospheric CO_2 and CH_4 on early Earth surface temperatures: Journal of Geophysical Research, v. 92, p. 2991–2998, doi:10.1029/JD092iD03p02991.
Kiehl, J.T., and Shields, C.A., 2005, Climate simulation of the latest Permian: Implications for mass extinction: Geology, v. 33, p. 757–760, doi:10.1130/G21654.1.
Knies, J., Mann, U., Popp, B.N., Stein, R. and Brumsack, H.-J., 2008, Surface water productivity and paleoceanographic implications in the Cenozoic Arctic Ocean: Paleoceanography, v. 23, PA1S16, 12 p., doi:10.1029/2007PA001455.
Korty, R.L., Emanuel, K.A., and Scott, J.R., 2008, Tropical cyclone-induced upper-ocean mixing and climate: Application to equable climates: Journal of Climate, v. 21, p. 638–654, doi:10.1175/2007JCLI1659.1.
Kump, L.R., 2011, The last great global warming: Scientific American, v. 305, p. 56–61, doi:10.1038/scientificamerican0711-56.
Kump, L.R., Pavlov, A., and Arthur, M.A., 2005, Massive release of hydrogen sulfide to the surface ocean and atmosphere during intervals of oceanic anoxia: Geology, v. 33, p. 397–400, doi:10.1130/G21295.1.
Kürschner, W.M., Kvacek, Z., and Dilcher, D.L., 2008, The impact of Miocene atmospheric carbon dioxide fluctuations on climate and the evolution of terrestrial ecosystems: Proceedings of the National Academy of Sciences of the United States of America, v. 105, p. 449–453, doi:10.1073/pnas.0708588105.
Leckie, R.M., Bralower, T.J., and Cashman, R., 2002, Oceanic anoxic events and plankton evolution: Biotic response to tectonic forcing during the mid-Cretaceous: Paleoceanography, v. 17, p. 13-1–13-29.
Lenton, T.M., and Britton, C., 2006, Enhanced carbonate and silicate weathering accelerates recover from fossil fuel CO_2 perturbations: Global Biogeochemical Cycles, v. 20, GB3009, doi:10.1029/2005GB002678.
Manabe, S., Stouffer, R.J., and Spelman, M.J., 1994, Response of a coupled ocean-atmosphere model to increasing atmospheric carbon dioxide: Ambio, v. 23, p. 44–49.
Marynowski, L., and Filipiak, P., 2007, Water column euxinia and wildfire evidence during deposition of the Upper Famennian Hangenberg event horizon from the Holy Cross Mountains (central Poland): Geological Magazine, v. 144, p. 569–595, doi:10.1017/S0016756807003317.
Mercer, J.H., 1970, Antarctic ice and interglacial high sea levels: Science, v. 168, p. 1605–1606, doi:10.1126/science.168.3939.1605-a.
Mohr, P., 1983, Ethiopian flood basalt province: Nature, v. 303, p. 577–584, doi:10.1038/303577a0.
Overpeck, J.T., Otto-Bliesner, B.L., Miller, G.H., Muhs, D.R., Alley, R.B., and Kiehl, J.T., 2006, Paleoclimatic evidence for future ice-sheet instability and rapid sea-level rise: Science, v. 311, p. 1747–1750, doi:10.1126/science.1115159.
Owens, J., Gill, B.C., Lyons, T.W., and Jenkyns, H., 2010, Perturbations to the global sulfur cycle during ocean anoxic events: Geological Society of America Abstracts with Programs, v. 42, no. 5, p. 298.
Park, J., and Royer, D.L., 2011, Geologic constraints on the glacial amplification of Phanerozoic climate sensitivity: American Journal of Science, v. 311, p. 1–26, doi:10.2475/01.2011.01.
Raymo, M.E., 1991, Geochemical evidence supporting T.C. Chamberlin's theory of glaciation: Geology, v. 19, p. 344–347, doi:10.1130/0091-7613(1991)019<0344:GESTCC>2.3.CO;2.
Retallack, G.J., and Alonso-Zarza, A.M., 1998, Middle Triassic paleosols and paleoclimate of Antarctica: Journal of Sedimentary Research, v. 68, p. 169–184.
Rohde, R.A., and Muller, R.A., 2005, Cycles in fossil diversity: Nature, v. 434, p. 208–210, doi:10.1038/nature03339.
Self, S., Widdowson, M., Thordarson, T., and Jay, A.E., 2006, Volatile fluxes during flood basalt eruptions and potential effects on the global environment: A Deccan perspective: Earth and Planetary Science Letters, v. 248, p. 518–532, doi:10.1016/j.epsl.2006.05.041.
Skelton, A., 2011, Flux rates for water and carbon during greenschist facies metamorphism: Geology, v. 39, p. 43–46, doi:10.1130/G31328.1.
Sluijs, A., Bijl, P.K., Schouten, S., Röhl, U., Reichart, G.-J., and Brinkhuis, H., 2011, Southern ocean warming, sea level and hydrological change during the Paleocene-Eocene thermal maximum: Climate of the Past Discussions, v. 7, p. 47–61.
Sobolev, S.V., Sobolev, A.V., Kuzmin, D.V., Krivolutskaya, N.A., Petrunin, A.G., Arndt, N.T., Radko, V.A., and Vasiliev, Y.R., 2011, Linking mantle plumes, large igneous provinces and environmental catastrophes: Nature, v. 477, p. 312–316, doi:10.1038/nature10385.
Taylor, E.L., Taylor, T.N., and Cúneo, N.R., 2000, Permian and Triassic high latitude paleoclimates: Evidence from fossil biotas, *in* Huber, B.T., MacLeod, K.G., and Wing, S.L., eds., Warm Climates in Earth History: Cambridge, Cambridge University Press, p. 321–350.
Tripati, A.K., Roberts, C.D., and Eagle, R.A., 2009, Coupling of CO_2 and ice sheet stability over major climate transitions of the last 20 million years: Science, v. 326, p. 1394–1397, doi:10.1126/science.1178296.
Walker, J.C.G., and Kasting, J.F., 1992, Effects of fuel and forest conservation on future levels of atmospheric carbon dioxide: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 97, p. 151–189, doi:10.1016/0031-0182(92)90207-L.
Ward, P.D., 2010, The Flooded Earth: Our Future in a World without Ice Caps: New York, Basic Books, 281 p.
Wignall, P.B., 2001, Large igneous provinces and mass extinctions: Earth-Science Reviews, v. 53, p. 1–33, doi:10.1016/S0012-8252(00)00037-4.
Wignall, P.B., Bond, D.P.G., Kuwahara, K., Kakuwa, Y., Newton, R.J., and Poulton, S.W., 2010, An 80 million year oceanic redox history from Permian to Jurassic pelagic sediments of the Mino-Tamba terrane, SW Japan, and the origin of four mass extinctions: Global and Planetary Change, v. 71, p. 109–123, doi:10.1016/j.gloplacha.2010.01.022.
Williams, S.N., Schaefer, S.J., Calvache, M.L., and López, D., 1992, Global carbon dioxide emission to the atmosphere by volcanoes: Geochimica et Cosmochimica Acta, v. 56, p. 1765–1770, doi:10.1016/0016-7037(92)90243-C.
Winguth, A., Shellito, C., Shields, C., and Winguth, C., 2010, Climate response at the Paleocene-Eocene Thermal Maximum to greenhouse gas forcing—A model study with CCSM3: Journal of Climate, v. 23, p. 2562–2584, doi:10.1175/2009JCLI3113.1.
You, Y., Huber, M., Müller, R.D., Poulsen, C.J., and Ribbe, J., 2009, Simulation of the Middle Miocene Climate Optimum: Geophysical Research Letters, v. 36, L04702, doi:10.1029/2008GL036571.
Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001, Trends, rhythms, and aberrations in global climate 65 Ma to present: Science, v. 292, p. 686–693, doi:10.1126/science.1059412.
Zhang, R., Follows, M.J., Grotzinger, J.P., and Marshall, J., 2001, Could the Late Permian deep ocean have been anoxic?: Paleoceanography, v. 16, p. 317–329, doi:10.1029/2000PA000522.

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