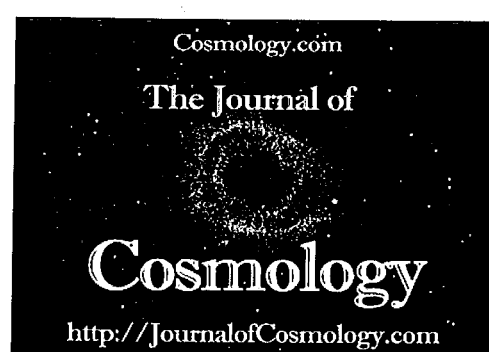
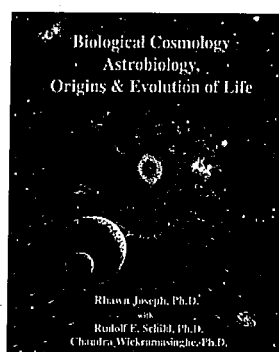


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5. A One Way Mission to Mars
6. Consciousness & the Universe
7. Are Women's Tears a Turn On?

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The World at +4°C Implications of Cainozoic Warm Periods

Andrew Glikson, Ph.D.,

Earth and paleoclimate scientist, School of Archaeology and Anthropology and Research School of Earth Science,
Australian National University, Australia.

Abstract

Climate projections for the 21st century overlap conditions deduced for the mid-Pliocene (3.3-2.8 Ma), mid-Miocene (17-15 Ma) and late Oligocene (25 Ma) warm periods, which paleo-CO₂ and temperature multi-proxy studies estimates suggest were in the range of c.350-410 ppm, 350-520 ppm and 300-750 ppm, respectively. The overlap between current CO₂ level of 389 ppm and upper Cainozoic warm periods suggests current climate change is at a lag stage, tracking toward mean global temperature c.2.4–4.0° C, sea levels 25-40 meters higher than 18th century levels and a near-permanent El-Nino state, as indicated for the mid-Pliocene (Dowsett et al., 1999, 2005; Haywood and Valdes, 2004; Haywood and Williams, 2005; Fedorov et al., 2006; Robinson et al., 2008; Chandler et al., 2008; Pagani et al., 2010). Recent re-calibration of climate sensitivity indices (CS – Mean global temperature rise per doubling of atmospheric CO₂) for the early to mid-Pliocene by Pagani et al. (2010) suggests high CS values in the range of 7.1–9.6° C, similar to the c.8.0° C value of the glacial terminations and to Hansen et al.'s (2008) CS value of c.6° C for slow feedbacks, which include disintegration of ice sheets and change in vegetation cover. These values are a factor of 2 to 3 higher than late Oligocene and mid-Miocene CS values of c.3.0-3.3° C (estimated after Kurschner et al. 2008 and Zachos et al. 2008), within the range of Charney's (1979) 3±1.5° C index. High CS values likely reflect large ice volumes at the outset of deglaciations, which enhance temperature rise through the positive feedback of the ice melt albedo flip process (Hansen et al., 2007, 2008). By distinction from the mid-Pliocene, current climate change is superposed on interglacial conditions with CO₂ level of c.280 ppm, inviting an analogy with mid-Miocene and late Oligocene warm episodes which commenced at CO₂ levels higher than 300 ppm. A role of methane in the late Oligocene and mid-Miocene warm phases is inconsistent with the positive $\delta^{13}C$ anomalies associated with these periods, which possibly reflect the effects of Red Sea rifting volcanism and Columbia Plateau volcanism, respectively. By contrast, the current CO₂ rise rate of c.2.0 ppm/year, more than 4 orders of magnitude faster than mid-Miocene value of 0.00005 (after Kurschner et al., 2008), may result in destabilization of methane from permafrost, arctic sediments and bogs. ty of carbon cycle, methane release and ice melt feedback processes.

1. Climate Sensitivity

"In the Early Pliocene, three to five million years ago, global temperatures were about 3–4° C warmer than today in the low latitudes, and up to 10°C warmer nearer the poles. Climate simulations and reconstructions

of this relatively recent period (geologically speaking) may help constrain realistic magnitudes of future warming." -Schneider and Schneider, 2010.

The history of Cainozoic climates (Zachos et al., 2001, 2008) betrays a combination of primary extraterrestrial forcings (Insolation/orbital forcing, asteroid/comet impacts), primary endogenic forcings (plate tectonics, orographic uplift/erosion, volcanism, geothermal rise, magnetic fields), terrestrial and oceanic climate responses (ice sheets formation/disintegration, ocean currents, wind patterns, CO_2 , CH_4 and N_2O cycles, evaporation/precipitation, weathering) and biogenic responses. The respective role of greenhouse gases in the evolution of the atmosphere-ocean-biosphere system, as compared to orbital, geotectonic, orographic and more transient volcanic and extraterrestrial impact factors, remains a subject of debate, with implications for 21st century climate change (Bershanskii, 2010; Harold et al., 2009; Lunt et al., 2008; Miyahara et al., 2010; You et al., 2010; Zachos et al., 2001, 2008). A combination of paleo-climate studies and atmospheric models allows constraints on the effects of CO_2 on temperature and climate under a range of conditions, including glacial and interglacial periods (Charney, 1979; National Academy of Science, 1979). Calibration of natural forcing of temperature is allowed by Greenland and Antarctic ice core studies which suggest temperature rise of $c.5^\circ\text{C}$ for solar forcing of 7.1 W/m^2 , defining climate sensitivity (CS) at $5/7.1 = 0.7^\circ\text{C}$ per 1 W/m^2 or, according to Hansen et al. (2008), $0.75 \pm 0.25^\circ\text{C}$ per 1 W/m^2 . These authors suggest a CS value of $c.6^\circ\text{C}$ for slow feedbacks, which superpose medium-term disintegration of ice sheets and change in vegetation cover on fast feedbacks (sea ice and the hydrological cycle - clouds, evaporation and precipitation). For an increase in radiative forcing due to CO_2 of 4 W/m^2 since the 18th century the predicted temperature rise is 3°C . Recently Pagani et al. (2010) define early and mid-Pliocene climate sensitivities at values in the range of 7.1–9.6, defined by Schneider and Schneider (2010) as "Earth system sensitivity", with implications for 21st century climate projections. The anthropogenic release of >370 billion ton carbon (GtC) to date falls in a unique category. Cainozoic climates The outset of the Cainozoic is marked with recovery from the effects of the 65 Ma-old K-T asteroid bombardment (Chicxulub, $D=170\text{ km}$; Boltysh, $D=24\text{ km}$), whereas the end of the Eocene $c.34\text{ Ma}$ is marked by extraterrestrial bombardment (Popigai, $D=100\text{ km}$, $35.7 \pm 0.2\text{ Ma}$; Chesapeake Bay: $D=85\text{ km}$, $35.3 \pm 0.1\text{ Ma}$; Mount Ashmore, $>50\text{ km}$, end-Eocene). The decline of global temperatures during $c.50\text{--}35\text{ Ma}$ (upper Eocene) has been attributed to weathering-sequestration of CO_2 related to the rise and erosion of the Tibetan Plateau (Ruddiman, 1997, 2003) (Figure. 1).

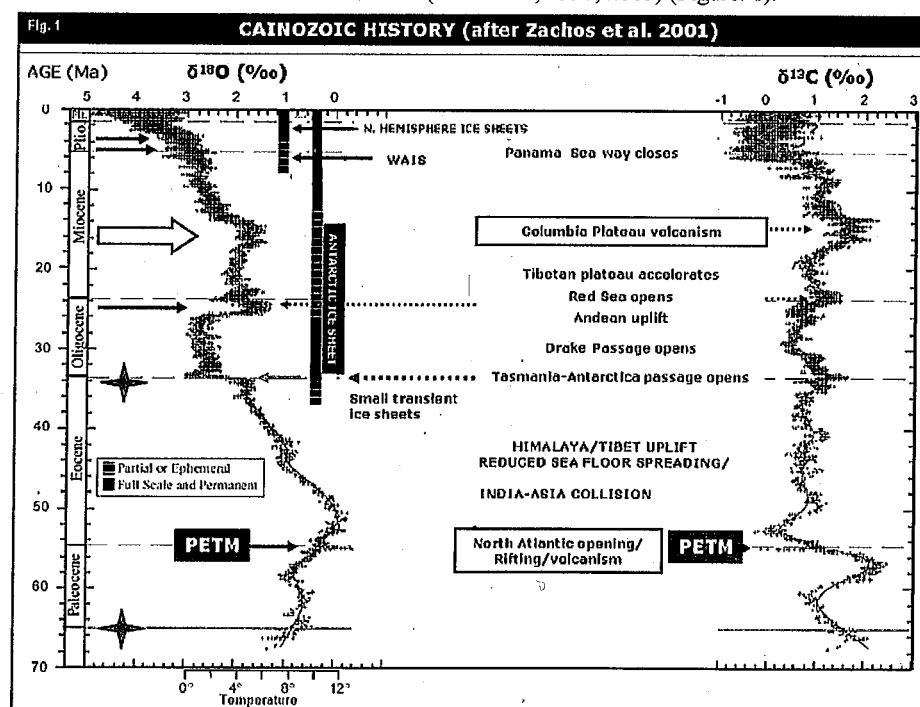


Figure 1. Cainozoic climate history, highlighting principal climate states and transitions based on $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotopic indices and indicating principal tectonic, volcanic and extraterrestrial impact events (K-T boundary, end-Eocene impact cluster). Modified after Zachos et al., 2001. The ocean temperature scale at the bottom relates to ice-free conditions.

Harold et al. (2009) discuss the mid-Miocene Climate Optimum (MMCO) $c.15\text{ Ma}$ in terms of decrease in the mean level of the Tibetan Plateau from 4700 to 2600 meters, resulting in local temperature increase of up to 9°C , global temperature increase of 0.58°C and weakening of summer monsoon. Marked rises in CO_2 and $\delta^{13}\text{C}$ values during the late Oligocene and MMCO warming episodes may be potentially attributed to large-scale volcanism associated with the opening of the Red Sea rift and Columbia Plateau volcanism, respectively. Changes in sea gates, including the opening of the Tasman-Antarctic and Drake passages, isolated the Antarctic continent from warm currents emanating at low latitudes, resulting in formation of its ice sheets, with fundamental effect on global climate. The

rise of the Panama Cordillera resulted in enhanced cross-latitude circulation in the Pacific Ocean (Humboldt Current) and the Atlantic Ocean (Gulf Current), with effects on the global thermohaline circulation.

Superposed on Cainozoic climate trends are orbital forced oscillations defined by eccentricity (400.10^3 years, 100.10^3 years), obliquity (41.10^3 years) and precession ($19 - 23.10^3$ years). Berger and Loutre (1991) calculate astronomical tuning of ~400 kyr and ~41-kyr Milankovic cycles, closely correlated with high-resolution ^{18}O time series. Effects of c.100 kyr cycles are progressively manifest during the coldest low- CO_2 periods which follow thermal maxima, including 23- 22.3 Ma (post-late Oligocene), 3.3 Ma (post-mid Pliocene) and from c.1.3 Ma (mid to upper Pleistocene). It may follow minimum buffering by greenhouse gases during these periods resulted in enhancement of the global temperature effects of orbital cycles.

Table 1. Principal proxies applied for reconstruction of Cainozoic climate conditions (not including the Holocene). Principal reference: Royer, 2001.

Temperature proxies
Benthic and plankton $\delta^{18}\text{O}$; ^{13}C - ^{18}O bonds in carbonate
Ice cores air bubbles: $\delta^{18}\text{O}$ / deuterium
Mg/Ca ratios in carbonate; pollen
TEX86 paleothermometer based on the relation between number of rings in the membrane lipids of the marine picoplankton Crenarchaeota and temperature.
ALKENONE (KETONE)– Paleo-T and CO_2 proxy.
CO_2 proxies
Stomata pores in fossil plants
Carbon $\delta^{13}\text{C}$ proxy - paleo-soil carbonate
Carbon $\delta^{13}\text{C}$ proxy – phytoplankton
Alkenone paleo- CO_2
Boron/Calcium
Foraminifera $\delta^{11}\text{B}$
$\delta^{44}\text{C}$ and $\delta^{11}\text{B}$ as pH and CO_2 proxies
Organic component of sediments (Sapropel) / N-alkane plant leaf wax / tropical vegetation.
Detrital component of sediments / dust / indicator of mechanical glacial erosion / wind
Boron / salinity / alkalinity
Carbon-sulphur-oxygen mass balance calculations

Progressive refinement of paleo-climate proxies, including temperature proxies ($\delta^{18}\text{O}$, δD , Mg/Ca, alkenone [$\text{C}_{37}-\text{C}_{39}$]) and CO_2 proxies (stomata fossil leaf pores, marine $\delta^{11}\text{B}$, marine phytoplankton $\delta^{13}\text{C}$, soil $\delta^{13}\text{C}$) (Table 1), allows improved constraints on early ocean/atmosphere temperature conditions and climate sensitivity indices (CS: Atmospheric temperature change for doubling of CO_2), with implications for current climate change projections. Periods of interest include the Paleocene-Eocene thermal maximum (PETM), late Oligocene Optimum (25 Ma), mid-Miocene Climate Optimum (MMCO) (c. 17-15 Ma), early Pliocene (c.5.2 Ma), mid-Pliocene (c. 3.3–3.0 Ma) and Pleistocene glacial terminations (Table 2), identified by CO_2 proxy studies (Zachos et al., 2008 [boron proxy]; Kurschner et al., 2008 [stomata proxy]; Royer, 2008 [stomata proxy]; Pagani et al., 2008 [alkenones proxy]; Tripatti et al., 2009 [B/Ca proxy]).

Paleocene-Eocene thermal maximum (PETM: c.55 Ma): The Paleocene– Eocene Thermal Maximum (PETM; c. 55 Ma) constitutes one of several thermal episodes during the early Tertiary, affecting a deep ocean temperature rise of $>5^\circ\text{C}$ over a short geological period of $<10,000$ years (Table 2). Zachos et al. (2001, 2008) attributes this event to the release of some c.2000 billion ton carbon (GtC) as methane, represented by a sharp negative $\delta^{13}\text{C}$ excursion, decrease pH and carbonate sedimentation. Effects of the PETM, lasting less than 170,000 years, include migration of fauna and flora, mass extinction of benthic foraminifera and appearance of new mammal species. Based on the PETM model, Zachos et al. 2008 estimate the effects of anthropogenic injection of 5000 GtC carbon into the atmosphere would last about 10,000 years in terms of CO_2 rise (up to 1800 ppm), ocean surface pH (down to 7.5) and calcite saturation ($\Omega < 2$). The PETM CO_2 growth rate of 0.13 ppm/year exceeds post-Eocene rates by orders of magnitude but is an order of magnitude less than the 21st century rate of c.2.0 ppm/year (Table 2). The rise of CO_2 by c.1300 ppm and temperature by c. 5°C suggest a low climate sensitivity value of about 1.1.

Table 2. Ages (black), mean temperatures (red), CO_2 levels (green), CO_2 growth rates, climate sensitivities and sea levels (blue) during principal Cainozoic warm periods.

Temperature proxies
Benthic and plankton $\delta^{18}\text{O}$; 13C - 18O bonds in carbonate
Ice cores air bubbles: $\delta^{18}\text{O}$ / deuterium
Mg/Ca ratios in carbonate; pollen
TEX86 paleothermometer based on the relation between number of rings in the membrane lipids of the marine picoplankton Crenarchaeota and temperature.
ALKENONE (KETONE)– Paleo-T and CO_2 proxy.
CO_2 proxies
Stomata pores in fossil plants
Carbon $\delta^{13}\text{C}$ proxy - paleo-soil carbonate
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Foraminifera $\delta^{11}\text{B}$
$\delta^{44}\text{C}$ and $\delta^{11}\text{B}$ as pH and CO_2 proxies
Organic component of sediments (Sapropel) / N-alkane plant leaf wax / tropical vegetation.
Detrital component of sediments / dust / indicator of mechanical glacial erosion / wind
Boron / salinity / alkalinity
Carbon-sulphur-oxygen mass balance calculations

Oligocene Optimum (c.25 Ma): Following Zachos et al. (2008) alkenone proxy evidence, a late-Oligocene CO_2 rise of c.700 ppm was associated with temperature rise of c.3° C under ice free conditions, yielding a low climate sensitivity of c.1.5° C (Table 2). Stomata evidence (Royer, 2006) yields a CS value of 2.0° C. Kurschner et al. (2008) stomata data suggests an optimum Oligocene CO_2 levels of 550-650 ppm, indicating a sharp decline of CO_2 to c.340 ppm associated with the Oligocene to Miocene transition. Kurschner et al. 2008 describe late Oligocene to early Miocene climates in terms of:

"a closed forest with palm and bamboo understory, whereas early Miocene plant communities were characterized by a mix of C3 grasses and herbs, forming savannas or open woodland habitats. Recently, phytolith studies from the eastern Mediterranean region reveal that relatively open, grass-dominated habitats were established by at least the early Miocene".

Mid-Miocene: Stomata proxy studies by Kurschner et al. (2008) indicate CO_2 rise from early Miocene values of c. 340 irregularly to 400-500 ppm in the mid- Miocene c.15.5 Ma, followed by a sharp decrease to 280 ppm toward 14 Ma and an increase to 340 ppm in the late Miocene. Correlation with atmospheric temperature rise by c.1.2° C yields a low CS value of c.3.3° C. Model studies by Harold et al. (2009) consider factors underlying the mid-Miocene climate optimum (MMCO), comparing atmospheric paleo- CO_2 levels (Royer et al. 2001 [18-14 Ma: ~300-370 ppm CO_2]; Pagani et al., 1999 [pre-10 Ma: <300 ppm CO_2]) and paleo-temperatures (Flower & Kennett, 1994: +6° C) with 20-21st century climate parameters. On this basis the authors suggest a "deficiency in our understanding of the global climate system", proceeding to test the effects of ocean-atmosphere heat exchange, plate tectonics, subduction, orographic elevation, uplift and erosion on the MMCO, using general circulation models. Revised estimates by You et al. (2009) refer to MMCO CO_2 level of 460-580 ppm, narrowed from earlier estimates of 300-600 ppm. These values are consistent with foraminiferal B/Ca proxies, indicate MMCO CO_2 levels of 430 ppm and temperatures ~3-6° C warmer than the present (Tripathi et al., 2009).

Pliocene: Detailed multi-proxy studies of the early to mid-Pliocene (c.5.2–3.0 Ma), when continent-ocean patterns were similar to the present, allow correlations of atmospheric CO_2 levels, mean global temperatures, sea levels and related parameters, with implications for current climate change trajectories. Based on paleoclimate studies (Chandler et al., 1994, 2008; Chandler, 1997; Dowsett et al., 1999, 2005; deMenocal, 2004; Cronin et al., 2005; Fedorov et al., 2006; Robinson et al., 2008; Tripathi et al., 2009; Pagani et al., 2010), Pliocene temperatures about 2.4-4.0 degrees C higher than pre-industrial levels ensued in sea levels about 25+/-12 meters, a major shift in latitudinal climate zones toward the poles (Figures 2 and 3), melting of at least one third of the polar ice sheets, total melt of mountain glaciers and increased thermal energy of the atmosphere expressed by extreme weather events.

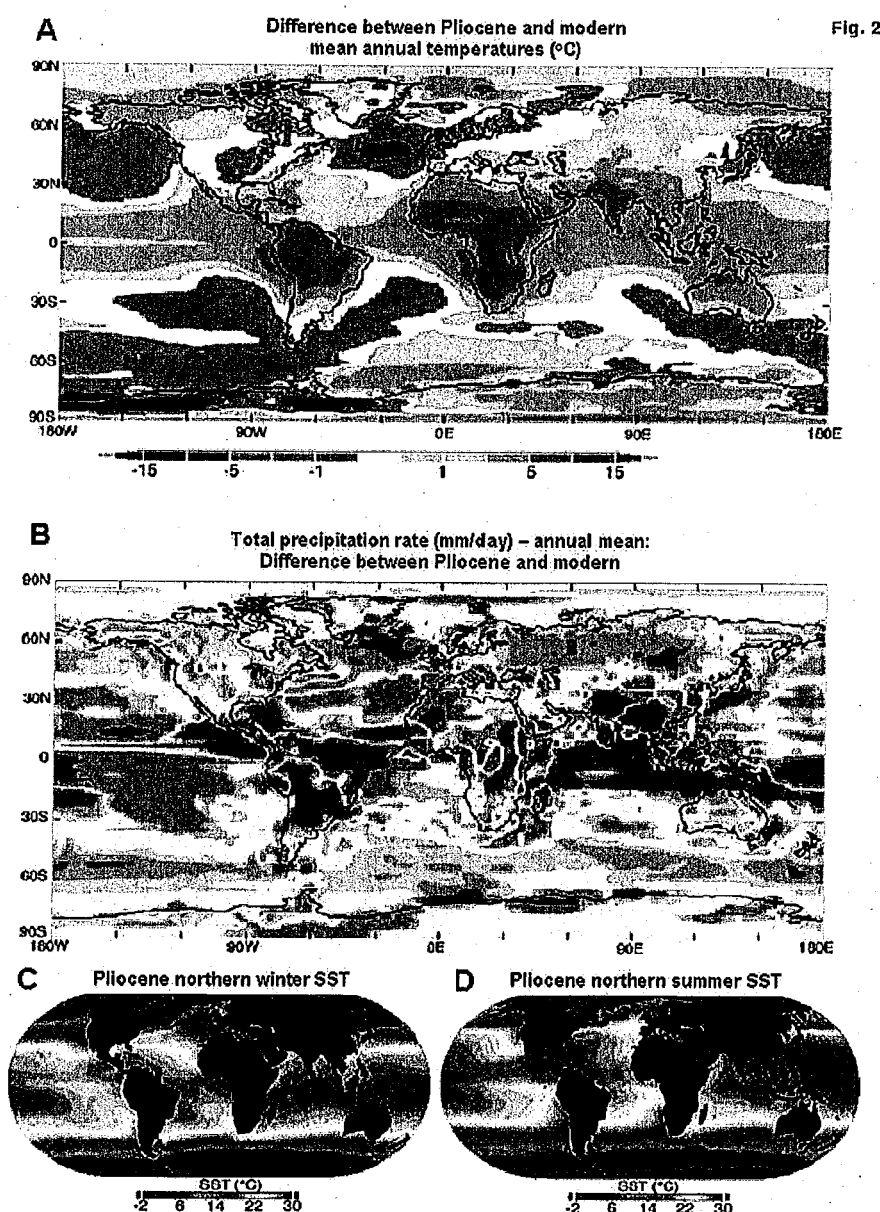


Figure 2. Differences between mid-Pliocene and modern annual mean surface temperature, precipitation rates and ocean temperatures (a) Difference between Pliocene and modern mean annual temperatures based on a fully coupled ocean atmosphere General Circulation Simulation using HadCM3 and an atmospheric general circulation model (GCM) simulation using fixed PRISM2 SSTs and the HadAM3 GCM. The coupled ocean atmosphere model HadCM3 predicts higher surface temperature in the tropics and relatively cool conditions in parts of the mid-latitudes, particularly over the North Atlantic and Pacific Oceans (modified from Haywood and Valdes, 2004); (b) Difference in annual mean total precipitation rate (mm/day) between the GISS and the Hadley Centre GCMs for the mid-Pliocene. Note the reduction in precipitation in storm track regions in the GISS GCM and the shift in location of the Indian Monsoon. (c) Pliocene February Sea Surface Temperatures (SST); (d) Pliocene August SST.

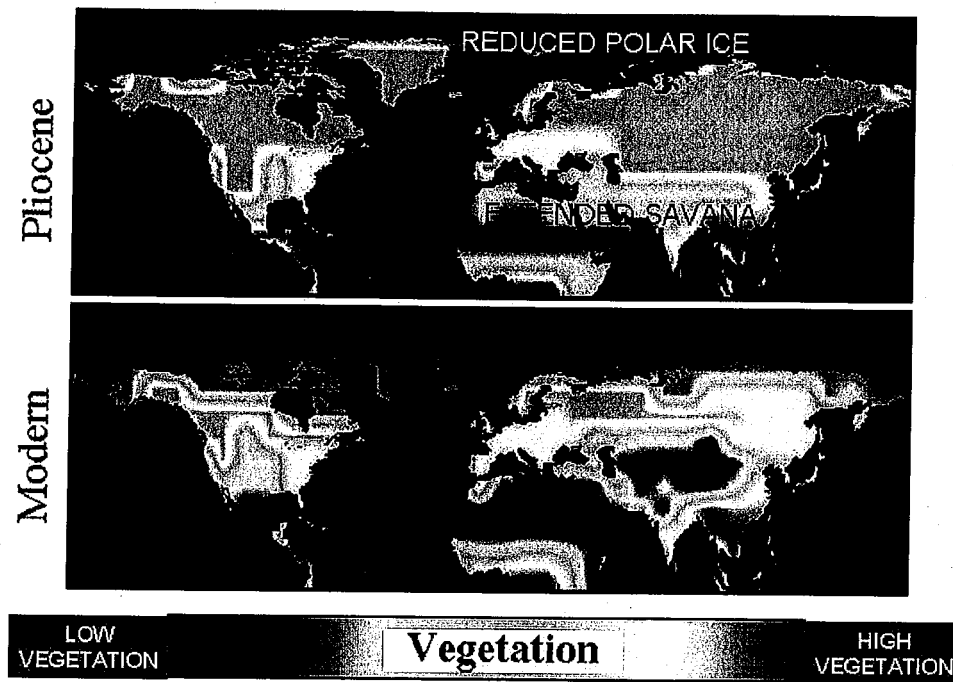


Figure 3. Albedo changes model for the mid-Pliocene (Chandler, 1997) http://www.giss.nasa.gov/research/features/199704_pliocene/fig2.gif Note the larger extent of the deserts (Sahara, Gobbi and Mexico) in the Holocene and of vegetated savanna and of boreal forest in the sub-Arctic and Greenland in the Pliocene.

Prior to c.4.2 Ma open passage between the Pacific and Atlantic oceans and warm climates resulted in a near-permanent El-Nino state (Figure 4), lower crosslatitudinal temperature gradients and thereby weaker current and wind activity. The closure of the Pacific and Atlantic Ocean basins due to the rise of the Panama Cordillera isolated the Pacific and Atlantic Ocean basins from each other and enhanced cross-latitude currents, including the Humboldt and California cold currents and the Gulf Stream. This accentuated the ENSO cycle, enhancing the La Nina phases.

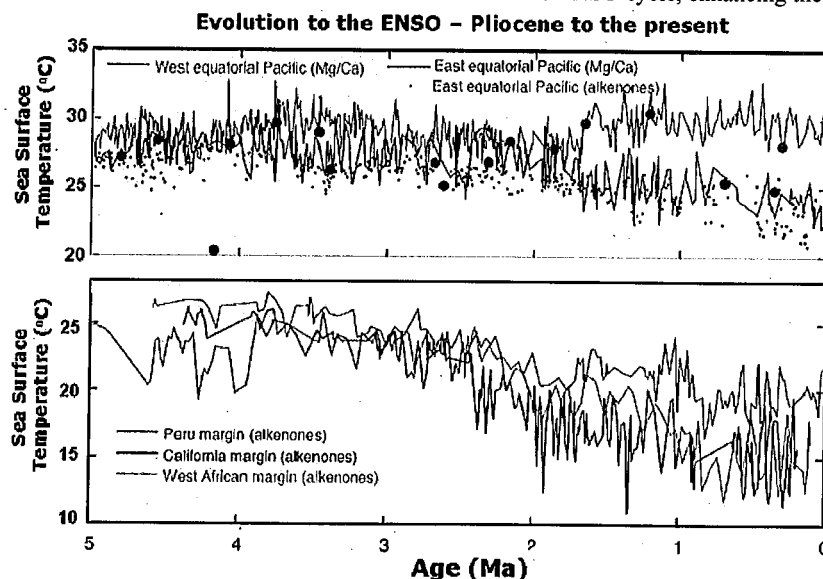


Figure 4. Pliocene to present sea surface temperature (°C) records in the western equatorial Pacific (red line, ODP site 806) and in the eastern equatorial Pacific (blue line, site 847), both based on Mg/Ca, and for the eastern Pacific based on alkenones (green dots). Pink shading denotes the early Pliocene. Fedorov et al. 2006. Note the temporal divergence of west and east Pacific temperatures, indicating increased role of the La Nina ENSO phase.

Elements of early to middle Pliocene climate include:

1. Mean global temperatures ranging between 2.4-4.0 degrees C and CO₂ levels between 350 – 412 ppm

- (Pagani et al., 2010). Continental and oceanic regions were 1 to 5 degrees C warmer than at present (Figure 2).
2. An estimate based on a rise of CO₂ level to about 400 ppm and temperatures by 3° C higher than at present (Lunt et al. 2009), assuming background CO₂ levels of about 250-300 ppm, would yield a high CS value of about 10° C, consistent with Pagani et al.'s (2010) climate sensitivity estimate of 9.6±1.4° C for the early Pliocene (4.2 Ma) and in the range of 7.1±1.0° C – 8.7±1.3° C for the mid Pliocene (~3.3 Ma).
 3. The extent of the Greenland and Antarctic ice sheets was about 50% for the mid-Pliocene and 33% for the early Pliocene, relative to the present.
 4. The sub-Arctic experienced greater precipitation and the tundra was occupied by forests (Figure 2).
 5. Climate zones were located closer to the poles relative to the present (Figure 1A). This involved higher precipitation over large parts of presentday deserts, including the Sahara and Gobi deserts, but lower precipitation over most regions affected by the El-Nino as well as mid-high latitudes zones (Figure 2B).
 6. A permanent El-Nino state dominated tropical and subtropical climates (Figure 4) associated with torrential rains along the west coast of the Americas and in equatorial Africa, and droughts in the southwestern Pacific, with effects on the Indian monsoon.
 7. Enhancement of the thermohaline circulation in the North Atlantic resulted in (A) elevated temperature and (B) enhanced evaporation and thereby snow fall, glacier movement and re-melting. Low temperature anomalies in the North Atlantic (Figure 2) and other high-latitude ocean regions may in part reflect Greenland ice melt cooling of the ocean water.
 8. Elevated precipitation in the Sahara enhanced spread of savannah conditions (Figure 2), allowing further migration of pre-historic humans.
 9. Ocean warming effects occurred to depth of about 2000 meters.
 10. Sea levels were 25±12 meters higher than in the late Holocene. The extent to which these conditions may apply to 21st century climates and beyond is discussed later.

Pleistocene intra-glacial variations and glacial terminations: Pleistocene intra-glacial climate cycles and glacial termination records militate for an extreme sensitivity of the atmosphere-ocean-cryosphere system. c. 1470 years-long cycles during the last 110-20 kyr glacial period (Dansgaard-Oeschger [D-O] cycles) involved rapid temperature changes in Greenland, with marked effects in the tropics (Ganopolski and Rahmstorf, 2002; Broecker, 2000). 21 D-O cycles are recorded in Greenland ice cores, some showing sharp global temperature rises of 3-4° C in 100 years, a rate commensurate with climate change projections for the 21st century. Sharp climate changes of several degrees C in a few years associated with the Younger Dryas (12.9 – 11.7 kyr), represented by δD (Deuterium) and oxygen isotope anomalies in Greenland ice cores (Steffensen et al., 2008; Kobashi et al., 2008), signify extreme instability of North Hemisphere ice sheets.

Estimates based on the last glacial termination (180–280 ppm CO₂; c.4.5° C: Hansen et al. 2007) yield a CS value of c.8° C, likely reflecting the amplifying effect of melting of the extensive Laurentian and Fennoscandian ice sheets through the ice melt feedback loop, i.e. the so-called albedo-flip effect—loss of ice reflection and gain of infrared absorption by opened water, which precedes CO₂ rise (Hansen et al. 2007). The more than doubling of the CS index during Pleistocene glacial terminations, relative to the Charney (1979) index, reflects the greater extent of ice sheets during the glacial periods upon the terminations and thus stronger role of the albedo-flip effect.

2. The Anthropocene and 21st Century Climate Projections

The effects of *H. sapiens* on the terrestrial climate, defined as the Anthropocene from the beginning of the industrial age about 1750 (Steffen et al., 2007), involves emission of over 370 billion ton carbon and the rise of CO₂ from 280 to 389 ppm, leading to a rise of mean global temperature by c.0.8° C since 1750AD (IPCC 2007-AR4), plus about 0.5° C currently masked by sulfur aerosols, yielding a climate sensitivity value of >3.4° C, not accounting for the full effects of current albedo loss due to melting cryosphere. Incipient effects of the Anthropocene may have commenced with the rise of early civilizations, extensive agricultural cultivation and livestock husbandry, indicated by the rise of methane by about 100 ppb from c.7000 years-ago and CO₂ by 20 ppm from c.5000 years-ago (IPCC-2007-AR4 Report, Figure SPM.1; Ruddiman, 2003, 2007; Kutzbach et al., 2010).

Holocene climates display an irregular decline in CO₂ and temperatures from the Holocene Optimum about 8200 years-ago, through a warm period at 950-1150AD (Medieval Warm Period), to a low at 1600-1700AD (Little Ice Age, related to minimum sunspot activity) to an irregular rise by 1.3° C (0.8° C + 0.5° C albedo masking effect) from the early 20th century, followed by a sharp rise from about 1975, currently at a rate of c.2 ppm CO₂/year (Table 2) (IPCC-2007-AR4). As distinct from Pliocene warm periods, the current rise rate of CO₂ may exceed the capacity of many species to adapt to environmental changes. The sensitivity of the atmosphere to this level of forcing is underscored by the pace of CO₂ rise, which exceeds the last glacial termination referred to above by near

two orders of magnitude. An association of further global warming with transient cool phases, consequent on the cooling effects of melting ice from the Greenland and Antarctica ice sheets on adjacent oceans, as during the Younger Dryas and the Holocene Optimum (8200 years-ago), is possible and may in part account for current climate projections for the North Atlantic (Keenlyside et al., 2008).

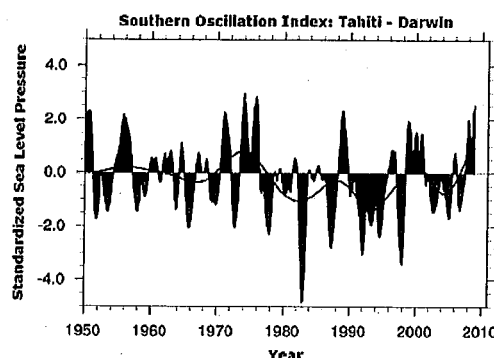


Figure 5. Monthly Southern Oscillation Index (SOI) based on standardized sea level pressure (SLP) difference data measured between Tahiti and Darwin, Australia from 1950 to the present. Note the increase in intensity and frequency El-Nino states from about 1983.

http://gcmd.nasa.gov/records/GCMD_NOAA_NWS_CPC_SOI.html.

An indication current climate trends are approaching Pliocene-like conditions is provided by the extensive melting of the large ice sheets (Climate Progress, 2009; msnbc, 2009; Overpeck et al., 2006) and increased frequency of El Nino events relative to La Nina events, in particular since about 1982 (Figure 5). The accentuation in the amplitude of mean global temperature fluctuations associated with the ENSO cycle, displayed by peak El Nino (1983, 1987, 1992, 1998, 2005) and La Nina (1974-76, 1989, 1999-2000, 2009) events indicates increasing atmosphere energy levels and related climate variability, reflected by extreme weather events. Climate tipping points may be preceded by quiet lulls, as indicated by intervals ahead of glacial terminations (Dakos et al., 2008). Whereas the timing of a range of potential tipping points (Lenton et al., 2008) can not be accurately predicted, as CO₂ is added to the atmosphere and associated feedbacks carbon cycle and ice melt/albedo flip, the likelihood of extreme weather events increases.

The current CO₂ concentration of 389 ppm is within the range indicated for the early to mid-Pliocene (4.2–3.0 Ma) (365–415 ppm CO₂; 2.4 – 4° C and sea level 25±12 meters above pre-industrial (Pagani et al., 2010), and also the range of the mid-Miocene climate optimum (16 Ma; 350–430 ppm CO₂; 3–6° C; sea level 25–40 meter) (Tripathi et al. 2009; Kurschner et al. 2008; Royer, 2008). The superposition of these warm periods above periods characterized with CO₂ levels of about 300 ppm invites comparison with current rise above Holocene conditions.

These observations imply current climate change is at a lag stage tracking toward conditions induced by mean global temperature of c.3.0–4.0° C, which in the Pliocene resulted in melting of large parts of the Greenland and west Antarctic ice sheets and sea level rise of 25+/-12 meters (Haywood and Williams, 2005). The trend results in migration of climate zones toward the poles, weakening of the polar vortex, weakening of cross-latitude cold ocean currents and weakening of the La Nina phases of the ENSO cycle, by analogy to mid-Pliocene (>2.8 Ma) conditions. Depending on the rate of future carbon emission, the possibility that conditions similar to the Paleocene-Eocene Thermal Maximum (PETM) 55 million years-ago may be reached can not be discounted. At that stage the release of c.2000 GtC as methane (Zachos et al., 2008), less than the current fossil fuel reserves (c.6000 billion ton carbon), resulted in >5° C temperature rise and in extinction of species. Factors relevant to this scenario include feedbacks from the carbon cycle and ice/warm water interactions, reduction of the ocean's CO₂ absorption capacity, the scale of deforestation and release of methane from permafrost, bogs and shallow sediments.

Civilization developed along river valleys where agriculture was made possible by the regulation of river flow by the seasonal fluctuations of mountain ice and glaciers. Cold periods resulted in reduction of river flow, with consequent famines, while warm periods resulted in floods and erosion of fertile river terraces, as recorded along the Nile (Krom et al., 2002), Euphrates, Indus and the Yellow River (Lamberg-Karlovsky, 1974). Whereas river valley agriculture benefits from seasonal re-fertilization of soil, extensive cultivation of grain in hill and open plain terrains leads to depletion of soils, rendering human habitats vulnerable to climate and environmental changes (Diamond, 2005). Feeble attempts by civilization to mitigate the climate are drowning in a tide of misinformation and opposition by vested interests and those with their own peculiar agendas (Hogan, 2009; Hamilton, 2010). There is nowhere the 6.5 billion of contemporary humans of this planet can go. Even escape to other planets is not feasible. By the outset of the 2nd decade of the new millennium, urgent global mitigation and adaptation measures are needed if *H. sapiens* chooses to restore the conditions which allowed the rise of civilization beginning some 10,000 years ago. Depending on future levels of carbon emissions and feedbacks from the oceans, drying biosphere and methane deposits, future scale of mass extinction of species analogous to those recorded in the geological past (Elewa and Joseph, 2009; Glikson, 2005, 2008; Keller, 2005; Stanley, 1987; Sepkoski, 1996; Veron, 2008; Ward,

1994, 2007) defies contemplation.

"We're simply talking about the very life support system of this planet" -Joachim Schellnhuber, Director, Potsdam Climate Impacts Institute, advisor to the German government (Beyond 4 Degrees, 2009)

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