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Arctic amplification: can the past constrain the future?

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ABSTRACT

Arctic amplification, the observation that surface air temperature changes in the Arctic exceed those of the Northern Hemisphere as a whole, is a pervasive feature of climate models, and has recently emerged in observational data relative to the warming trend of the past century. The magnitude of Arctic amplification is an important, but poorly constrained variable necessary to estimate global average temperature change over the next century. Here we evaluate the mechanisms responsible for Arctic amplification on Quaternary timescales, and review evidence from four intervals in the past 3 Ma for which sufficient paleoclimate data and model simulations are available to estimate the magnitude of Arctic amplification under climate states both warmer and colder than present. Despite differences in forcings and feedbacks for these reconstructions compared to today, the Arctic temperature change consistently exceeds the Northern Hemisphere average by a factor of 3–4, suggesting that Arctic warming will continue to greatly exceed the global average over the coming century, with concomitant reductions in terrestrial ice masses and, consequently, an increasing rate of sea level rise.

1. Introduction

The Arctic is influenced by a suite of positive feedbacks that amplify the surface air temperature response to climate forcing (Serreze and Francis, 2006; Serreze et al., 2007). The strongest feedbacks are associated with changes in sea ice and snow cover (fast) and terrestrial ice sheets (slow), but changes in sea level, plant ecotonal boundaries and permafrost may also produce positive feedbacks. While changes in atmospheric circulation, cloud cover and other factors may act as negative feedbacks, positive feedbacks appear to dominate. Hence, the concept of Arctic amplification (Manabe and Stouffer, 1980). We define Arctic amplification as the difference between the Arctic-averaged surface air temperature change and the average Northern Hemisphere surface air temperature change when comparing two specific time periods, and we quantify Arctic amplification as the ratio of this temperature difference. In a paleoclimate context, Arctic amplification is typically the ratio of Arctic and Northern Hemisphere average temperature changes under a past climate state different than today's, with a duration of a few hundred to a few thousand years (and in some cases much longer) relative to their 20th century average temperatures.

Arctic amplification is a near-universal feature of climate model simulations forced by increasing concentrations of atmospheric greenhouse gases (e.g., Holland and Bitz, 2003). Available observations indicate that Arctic amplification, tied strongly to reductions in sea ice extent, has already emerged (Serreze et al., 2009). Large uncertainties remain regarding the magnitude of Arctic amplification that can be expected through the 21st century (Holland and Bitz, 2003). To help constrain uncertainties, we utilize the natural experiments of the past to quantify the magnitude of Arctic amplification under a range of forcing scenarios.

2. Arctic feedbacks and Arctic amplification

An amplified temperature response in the Arctic to climate forcing implies the existence of strong positive feedbacks that influence the Arctic to a greater extent than the rest of the planet. The dominant Arctic feedbacks exhibit differences in their seasonal and spatial expressions, and their timescales vary greatly. Seasonal snow cover and sea ice feedback are considered to be fast

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feedbacks, with seasonal response times. Vegetation and permafrost feedbacks operate more slowly, on timescales of decades to centuries. The slowest feedbacks operate on millennial timescales and are related to the growth and decay of continental ice sheets, and the response of the Earth's crust to those changes in mass.

2.1. Ice/snow albedo feedback

The albedo of a surface is defined as the reflectivity of that surface to the wavelengths of solar radiation. Because fresh snow and sea ice have the highest albedos of any widespread surfaces on the planet (Fig. 1), large changes in their seasonal and areal extent will have strong influences on the planetary energy balance (Peixoto and Oort, 1992). Viewed for the planet as a whole, a climate forcing that increases the global mean surface air temperature will lead to a reduction in the areal extent of snow and sea ice, resulting in a reduction in the planetary albedo, stronger absorption of solar radiation, and hence, a further rise in global mean temperature. Conversely, an initial reduction in global mean temperature leads to a greater areal extent of snow and sea ice, a higher albedo, and a further fall in temperature. The albedo feedback is hence positive in that the initial global mean temperature change in response to the forcing is amplified. Given that the Arctic is characterized by its seasonal snow cover and sea ice, it follows that the albedo feedback will be strongly expressed in this region. A contributing factor to the Arctic amplification linked to this feedback is the low-level temperature inversion that characterizes the Arctic region for much of the year. This strong stability limits vertical mixing, which helps to focus the effects of heating near the surface.

There are additional complexities to Arctic amplification (Serreze and Francis, 2006; Serreze et al., 2009). For half of the year the Arctic receives little or no solar radiation, so albedo has little impact on the energy balance during those times. Over land, a more complete view is that initial warming leads to earlier spring melt of the snow cover and later autumn onset of snow. Because the underlying low-albedo surface (earth, tundra, shrubs) is exposed longer, there is a stronger seasonal heating of the atmosphere via longwave radiation and turbulent heat transfers. The temperature change is expected to be especially large during spring when the effects of earlier snowmelt are paired with fairly strong insolation. Adding to the complexity, model-projected Arctic amplification during the 21st century tends to be strongest not over land, but over the Arctic Ocean and during the cold season, when there is little or no solar radiation.

The key to the ocean focus lies in the seasonality of ocean–atmosphere energy exchanges. If there is a climate forcing leading to an increase in temperature, the sea ice melt season becomes longer and stronger. Areas of low-albedo open water will develop earlier in the melt season, and these areas will strongly absorb solar radiation. This will raise the sensible heat content of the ocean mixed layer, and some of this heat will be used to melt more ice, meaning even more dark open water to absorb heat. This is clearly a positive feedback. However, the surface air temperature response over the Arctic Ocean in summer is actually rather small, as most of the solar energy is consumed, melting sea ice and raising the temperature of ocean mixed layer (~50 m thick).

At the close of the melt season and with the setting of the Arctic sun, there is extensive open water and considerable heat in the ocean mixed layer. The development of sea ice in autumn and winter is delayed. Until sea ice forms, creating an insulating barrier between the ocean and the cooling atmosphere, there is a substantial vertical transfer of heat from the ocean mixed layer

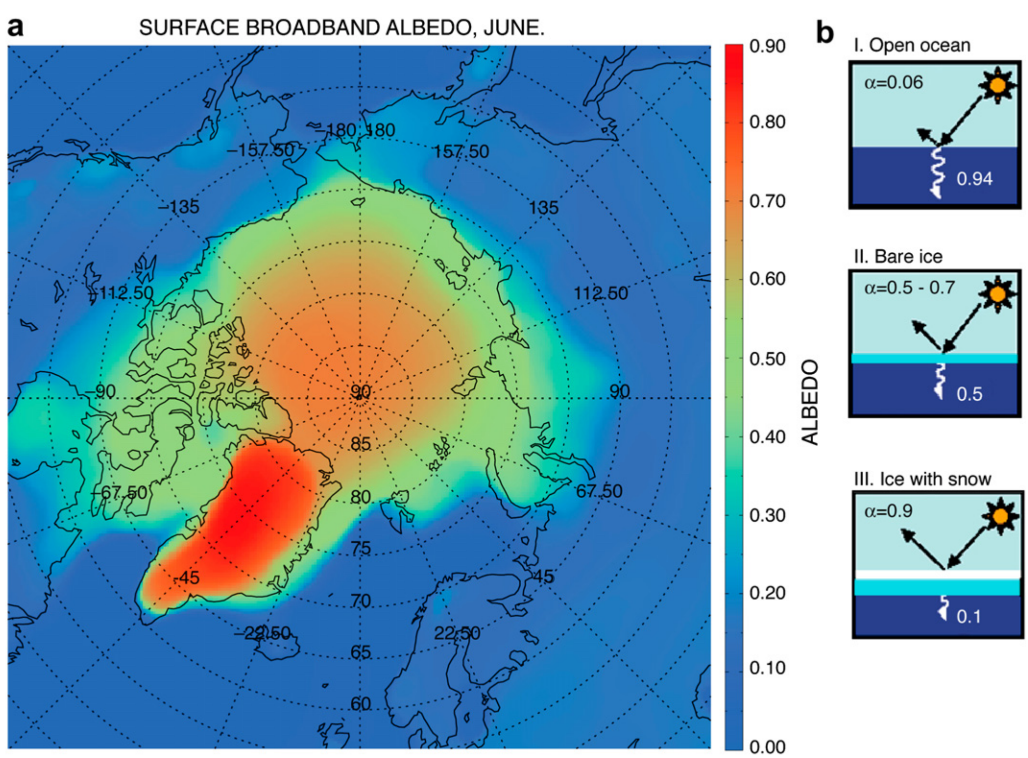

Fig. 1. Albedo values in the Arctic. a. Advanced Very High Resolution Radiometery (AVHRR)-derived Arctic albedo values in June, 1982–2004 multi-year average, showing the strong contrast between snow and ice covered areas (green through red) and open water or land (blue). (Image courtesy of X. Wang, University of Wisconsin–Madison, CIMSS/NOAA). b. Albedo feedbacks. Albedo is the fraction of incident sunlight that is reflected. Snow, ice, and glaciers have high-albedo. Dark objects such as the open ocean, which absorbs some 93% of the Sun's energy, have low-albedo (about 0.07), absorbing some 93% of the Sun's energy. Bare ice has an albedo of 0.5; however, sea ice covered with snow has an albedo of nearly 0.9. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.) Source: http://nsidc.org/seaice/processes/albedo.html.

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back to the atmosphere, acting to warm it. Hence there is a seasonal delay in the temperature expression of the summer albedo feedback. As the climate continues to warm in response to the forcing, the melt season becomes even longer and more intense, meaning a stronger summer albedo feedback, leading to even less ice at the end of summer, and more heat stored in the mixed layer, and a larger heat release back to the atmosphere in winter. The strong increase in autumn and winter air temperatures observed over the Arctic Ocean in recent years is consistent with these processes (Serreze et al., 2009).

2.2. Vegetation feedbacks

Warmer summers and reduced seasonal duration of snow cover lead to a vegetation response. As summers warm, and the growing season lengthens, dark shrub tundra advances poleward, replacing low-growing tundra that, compared to the shrub vegetation, is more easily covered by high-albedo snow. This leads to an albedo feedback that furthers the warming (Fig. 2; Chapin et al., 2005; Sturm et al., 2005; Goetz et al., 2007). The albedo difference between shrub- and low-tundra is most effective in spring, when the solar radiation flux is strong and snow cover is still extensive. The feedback is even more pronounced if warming allows evergreen boreal forest with its dark foliage to advance northward to replace tundra or shrub vegetation. In the case of boreal forest migration, the warming feedback would tend to be partially balanced for a period of time by sequestration of carbon in forest ecosystems, which have more above-ground carbon than shrub ecosystems (Denman et al., 2007). The situation may be different farther south, where the winter solar flux is still substantial. If warming is sufficient to allow deciduous forest replacement of evergreen boreal forest, then there is an increase in the wintertime surface albedo, acting as a negative (cooling) feedback (Bonan et al., 1992; Rivers and Lynch, 2004).

2.3. Permafrost feedbacks

Additional but poorly understood feedbacks in the Arctic involve changes in the extent of permafrost and consequent release of carbon dioxide and methane from the land surface. Feedbacks between permafrost and climate have become widely recognized only in recent decades, building on the works of Kvenvolden (1988, 1993), MacDonald (1990), and Haeberli et al. (1993). As permafrost thaws under a warmer summer climate, it is likely to release carbon dioxide and methane from the decomposition of organic matter previously preserved in a frozen state in permafrost and in widespread Arctic yedoma deposits (Vörösmarty et al., 2001; Thomas et al., 2002; Smith et al., 2004; Archer, 2007; Walter et al., 2007). The increase in atmospheric greenhouse gas concentrations will lead to further warming. Walter et al. (2007) suggested that methane bubbling from the thawing of thermokarst lakes that formed across parts of the Arctic during deglaciation could account for as much as 33–87% of the increase in atmospheric methane measured in ice cores during this period. Such a release would have contributed a strong and rapid positive feedback to warming during the last deglaciation, and it likely continues today (Walter et al., 2006). Although the additional greenhouse gases are well mixed through the global troposphere, consequent warming in the Arctic is amplified through ice-albedo and other high-latitude feedbacks.

2.4. Slow feedbacks during glacial–interglacial cycles

The growth and decay of continental ice sheets and associated isostatic adjustments influence the mean surface temperature,

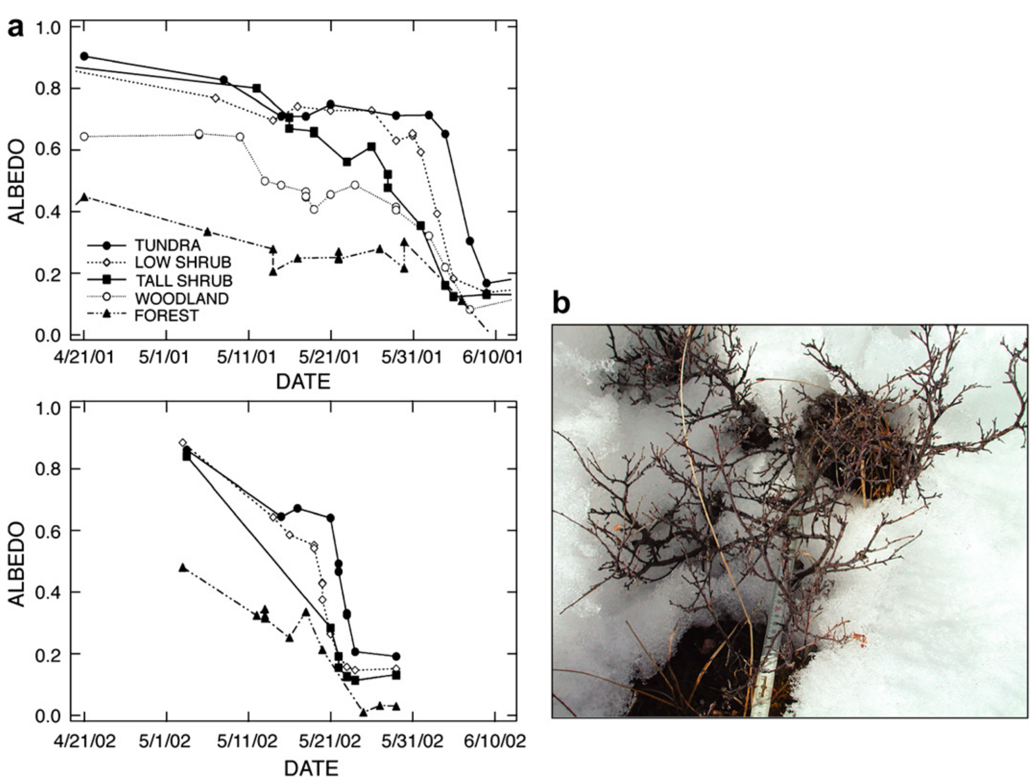

Fig. 2. Changes in vegetation cover throughout the Arctic can influence albedo, as can altering the onset of snowmelt in spring. a) Progression of the melt season in northern Alaska, May 2001 (top) and May 2002 (bottom), demonstrates how areas with exposed shrubs show earlier snowmelt. b) Dark branches against reflective snow alter albedo (Sturm et al., 2005; Photograph courtesy of Matt Sturm). Copyright 2005 American Geophysical Union. Reproduced by permission of American Geophysical Union.

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planetary albedo, freshwater fluxes to the ocean, atmospheric circulation, ocean dynamics, and greenhouse gas storage or release in the ocean (e.g., Rind, 1987). Feedbacks linked to ice sheet growth and decay are considered slow because they operate over millennial timescales (e.g., Edwards et al., 2007).

The growth of ice sheets contributes to Arctic amplification by increasing the reflectivity of the Arctic, amplifying the initial cooling. Furthermore, the great height of these ice sheets across extensive regions produced additional cooling by effectively raising substantial portions of the Arctic surface higher into the troposphere where temperatures are lower. Continental ice sheet growth also contributes positive feedbacks to the global system. Colder ice-age oceans lead to a reduction in atmospheric water vapor, a key greenhouse gas, and drying of the continents. Continental drying, possibly coupled with increased pole-equator pressure gradients, raises atmospheric dust loads, which contribute to additional cooling by blocking sunlight. Complex changes in the ocean–atmosphere gas exchange during glaciations shifted CO_2 from the atmosphere to the ocean and reduced the atmospheric greenhouse effect.

2.5. Freshwater balance feedbacks

The Arctic Ocean is almost completely surrounded by continents (Fig. 3). The largest source of freshwater input to the Arctic Ocean is not from net precipitation over the Arctic Ocean itself, but from river runoff, the majority contributed by the Yenisey, Ob, Lena and Mackenzie rivers (see Vörösmarty et al., 2008). River discharge is key in maintaining low surface salinities along the broad, shallow, and seasonally ice-free continental shelves fringing the Arctic Ocean. The largest of these shelves extends seaward from the Eurasian continent, and it is the dominant region of seasonal sea ice production in the Arctic Ocean (e.g., Barry et al., 1993). Sea ice that forms along the Eurasian shelves drifts toward Fram Strait; its transit time is 2–3 years in the current regime. In the Amerasian sector of the Arctic Ocean, the clockwise-rotating Beaufort Gyre is the dominant ice-drift feature (see Fig. 3).

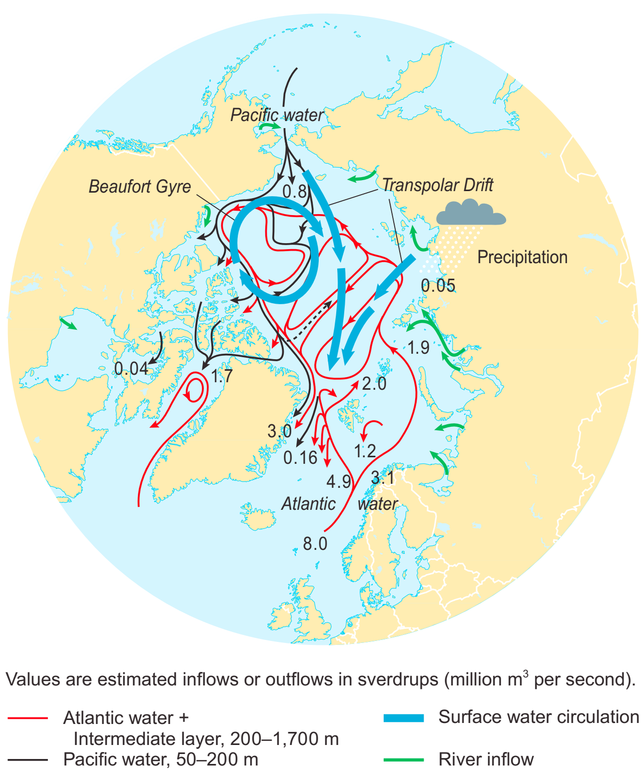

Fig. 3. Inflows and outflows of water in the Arctic Ocean. Red lines, components and paths of the surface and Atlantic Water layer in the Arctic; black arrows, pathways of Pacific water inflow from 50 to 200 m depth; blue arrows, surface water circulation; green, major river inflow; red arrows, movements of density-driven Atlantic water and intermediate water masses into the Arctic (modified from AMAP, 1998, Fig. 3.27). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.) Reproduced by permission Arctic Monitoring and Assessment Program.

Surface currents transport low-salinity surface water (its upper 50 m) and sea ice (essentially freshwater) out of the Arctic Ocean (e.g., Schlosser et al., 2000). The freshwater export is primarily through western Fram Strait, then along the east coast of Greenland into the North Atlantic through Denmark Strait. A smaller volume of surface water and sea ice flows out through the inter-island channels of the Canadian Arctic Archipelago, reaching the North Atlantic through the Labrador Sea. The low-salinity outflow from the Arctic Ocean is compensated by inflow of relatively warm, salty Atlantic water through eastern Fram Strait and the Barents Sea. Despite its warmth, Atlantic water has sufficiently high salt content that its density exceeds that of the cold, low-salinity surface waters. The inflowing relatively dense Atlantic water sinks beneath the colder, fresher surface water upon entering the Arctic Ocean. North of Svalbard, Atlantic water spreads as a boundary current into the Arctic Basin and forms the Atlantic Water Layer (Morison et al., 2000). The strong vertical gradients of salinity and temperature in the Arctic Ocean produce a stable density stratification, which is one of the reasons why sea ice readily forms there.

The long-term stability of the vertical structure of the Arctic Ocean remains one of the least known, yet most important variables in predicting the future climate, and explaining past changes. Should the stratification weaken so that Atlantic water and Arctic Ocean surface waters mix, sea ice formation would be greatly reduced, leading to a warmer Arctic atmosphere. Observations have shown that the warm Atlantic layer can mix with the colder surface layer in some parts of the Eurasian coastal seas (Rudels et al., 1996; Steele and Boyd, 1998; Schauer et al., 2002), limiting sea ice formation and promoting vertical heat transfer to the Arctic atmosphere in winter.

In recent decades, circum-Arctic glaciers and ice sheets have been losing mass (Dowdeswell et al., 1997; Rignot and Thomas, 2002; Meier et al., 2007), and since the 1930s runoff to the Arctic Ocean from the major Eurasian rivers has broadly increased (Peterson et al., 2002). Changes in river runoff may influence the Arctic Ocean stratification (Steele and Boyd, 1998; Martinson and Steele, 2001; Björk et al., 2002; Boyd et al., 2002; McLaughlin et al., 2002; Schlosser et al., 2002). The net effect of increased river runoff is to promote vertical stratification by enhancing the density contrast between the low-salinity surface layer and denser Atlantic water at depth. A competing effect is an increased flux of Atlantic water to the Arctic Ocean that can more effectively mix with the surface layer, weakening the vertical structure of the surface waters and contributing to sea ice melt and delayed sea ice formation in the autumn (e.g., Polyakov et al., 2005).

2.6. Changes in thermohaline circulation

Wintertime cooling of relatively warm and salty Atlantic surface waters in the Nordic Seas and the Labrador Sea increases their density. The denser waters sink and flow southward at depth to participate in the global thermohaline circulation. Although the term thermohaline circulation and Meridional Overturning Circulation (MOC) are sometimes used interchangeably, the MOC is confined to the Atlantic Ocean and its strength can be quantified using various tracers, whereas thermohaline circulation as typically used refers to a conceptual model of vertical ocean circulation that encompasses the global ocean and is driven by differences in density. Water sinking in the Nordic and Labrador seas (MOC) is replaced by surface inflow from the south, promoting persistent open water in these regions. The open water warms the overlying atmosphere in wintertime over the North Atlantic and extending downwind across Europe and beyond (Seager et al., 2002). Salt rejected from sea ice growing nearby very likely contributes to the density of the adjacent sea water and to its sinking.

Changes in the MOC are among the greatest nonlinearities in the climate system (e.g., Broecker et al., 1985). If the surface waters are made sufficiently less salty by an increase in freshwater from runoff, melting ice, or from direct precipitation (all likely consequences of global warming), then the rate of deep convection may diminish or stop. Results of numerical models indicate that if freshwater runoff into the Arctic Ocean and the North Atlantic increases as surface waters warm in the northern high-latitudes, then the MOC will weaken, cooling the downstream atmosphere in winter (e.g., Rahmstorf, 1996, 2002; Marotzke, 2000; Schmittner, 2005). This provides a negative feedback to Arctic amplification.

Reducing the rate of North Atlantic thermohaline circulation likely has global as well as regional effects (e.g., Obata, 2007). Oceanic overturning is an important mechanism for transferring atmospheric CO_2 to the deep ocean. Reducing the rate of deep convection in the ocean would allow a higher proportion of anthropogenic CO_2 to remain in the atmosphere, a positive feedback for global warming, but not for Arctic amplification.

2.7. How paleoclimate proxies record Arctic amplification

As developed above, Arctic temperature changes due to altered sea ice limits are expected to be most strongly expressed in winter months. However, there are few paleoclimate proxies that reliably record winter air temperatures over the Arctic Ocean. Most paleoclimate proxies in the Arctic are sensitive to summer temperature, and most are recovered from terrestrial archives. Nevertheless, paleoclimate reconstructions document strong Arctic amplification of terrestrial summer air temperatures during times both colder and warmer than present (Section 5, below). There are two compelling reasons why summer temperatures over Arctic lands reflect Arctic amplification despite the expectation that the strongest signal would be in winter over the Arctic Ocean.

2.7.1. Direct effects

The Arctic contains about as much land as it does ocean. When insolation in summer is elevated, there is a strong, direct response over land. This is amplified when snow cover is reduced in the spring season and when vegetation zones migrate northward. Early spring melting of snow cover is enhanced by reduced sea ice because winter air temperatures are warmer. Warmer winters also allow treeline to advance northward, further lowering terrestrial albedo and amplifying warming.

2.7.2. Dynamic effects

Heat is delivered to the Arctic by the convergence of horizontal atmosphere energy and ocean heat transports. Serreze and Barrett (2008) note that the summer circulation over the Arctic is strongly influenced by the differential heating between land and Arctic Ocean, and that this differential heating may, in turn, be influenced by the timing of snowmelt over land or through changes in the atmosphere–ocean heat exchanges. Both factors depend on sea ice distribution and thickness during the preceding winter. Ogi et al. (2003, 2004) demonstrated the dependence of the phase of the summer Northern Annual Mode (NAM) on its corresponding phase during the preceding winter. A negative NAM phase in winter produces more snow on adjacent lands, weakening thermal contrasts the following summer, thereby favoring a negative summer NAM phase and weakened activity along the Arctic frontal zone, linking winter warming with changes in summer. Model simulations also indicate that strong cold season warming over the Arctic Ocean will be spread out across Arctic land areas via the horizontal atmospheric circulation (Lawrence et al., 2008).

Heat is also delivered to the Arctic by ocean currents, principally by the North Atlantic Drift (NAD), which delivers warm, salty water into the Arctic Ocean through Fram Strait. Paleodata document a greater domination of warm Atlantic surface waters around the Arctic during the peak warmth of both the Holocene Thermal Maximum (HTM) and the Last Interglaciation (LIG) than in the recent past (see Polyak et al., this volume). Exactly why the NAD was stronger during past warm times, and its link to the state of Arctic Ocean sea ice, remain uncertain. However, with this additional heat source, sea ice must have been diminished and summer growing seasons on adjacent lands would have been longer.

3. Climate forcings

Arctic amplification is a response to a particular forcing or set of forcings. Four intervals in the past exhibit very different climates than present but with continental configurations not substantially different from those of the present. For the three younger time intervals, the Holocene Thermal Maximum (HTM, about 9–6 ka ago), the Last Glacial Maximum (LGM, ~21 ka ago), and marine isotope stage 5e, also known as the Last Interglaciation (LIG, about 130–120 ka ago), the climate changes were primarily forced by regular variations in Earth's orbital parameters with some input from changed greenhouse gas concentrations during the LGM (decrease) and LIG (small increase).

During these three intervals, the incoming solar radiation (insolation) anomalies, when averaged across the planet over a full annual cycle, were less than 0.4%. The orbital changes serve primarily to shift insolation around the planet seasonally and/or geographically. Summer insolation anomalies were relatively uniform across the Northern Hemisphere. During the warm times (HTM, LIG) summer insolation anomalies north of 60°N typically were only 10–20% greater than the anomalies for corresponding times averaged across the Northern Hemisphere as a whole (Table 1). For example, at the peak of the LIG (130–125 ka ago), the Arctic (60–90°N) summer

Table 1
Climate forcings for present and past climate states discussed in the text.

Time	Δ Precession	Δ Obliquity	Δ Insolation	Δ GHG (ppmv)
Today	30%	50%	0	+110
PI	30%	50%	0	0
HTM	60%	80%	+32	0
LGM	30%	60%	−6	−100
LIG	85%	90%	+57	+30
MP	No	No	No	+110 ± 30

Δ Precession reflects how close Earth is to the Sun in Northern Hemisphere summer; closest to the Sun (100%) and farthest from Sun (0%).
Δ Obliquity reflects the fractional difference between Earth's actual axial tilt and its shallowest (22.1°; polar insolation least in summer, 0%) and greatest axial tilt (24.5°; polar insolation greatest in summer, 100%).
Δ Insolation is the difference in W m^{-2} for June 60°N relative to present June at 60°N.
Δ GHG is the difference in atmospheric CO_2 concentration relative to the pre-industrial atmosphere concentration of CO_2 in ppmv.
Today = 2009 AD.
Pre-industrial (PI) = 1890 AD.
Holocene Thermal Maximum (HTM) = 8 ka ago.
Last Glacial Maximum (LGM) = 21 ka ago.
Last Interglaciation (LIG) = 129 ka ago.
Middle Pliocene (MP) = 3.5 Ma ago.

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(May–June–July) insolation anomaly was 12.7% above present, while the Northern Hemisphere anomaly was 11.4% above present (Berger and Loutre, 1991). At the same time, the Southern Hemisphere summer (November, December, January) insolation anomaly at 60°S was 6% less than present. Greenhouse gas forcings are globally uniform because of the relatively short mixing time for the troposphere ($\sim$2 years) compared to the long residence times for most greenhouse gases (10–100 years or more).

The well-documented warmth of the middle Pliocene (MP, about 3.5 Ma ago) is not fully explained (e.g., Raymo et al., 1996). The duration of warmth, several hundred thousand years, is much longer than the cycle time of insolation changes resulting from Earth's orbital features ($\sim$20 ka and $\sim$40 ka). Consequently, the warmth of the middle Pliocene was not driven by Earth orbital changes. Solar irradiance has increased steadily since Earth formed, but the rate of increase is slow, and 3.5 Ma ago insolation at the Earth's surface would have been less than 0.1 W m^{-2} lower than at present, notably smaller than the change during a typical 11-year solar cycle, and certainly in the wrong direction. The middle Pliocene is recent enough that continental positions were substantially the same as today. The Antarctic Ice Sheet was smaller, and the Greenland Ice Sheet was smaller or absent, but there are few secure constraints on the actual size of either ice sheet, other than the general guideline provided by sea level estimates. Although uncertainties remain in middle Pliocene sea level reconstructions, the available data suggest substantial fluctuations (ice sheet growth and decay), but with peak sea levels of between 20 and 30 m above present levels (Dwyer and Chandler, 2009).

The most plausible explanation for high-latitude middle Pliocene warmth is an elevated level of atmospheric CO_2, as reconstructed from the stomatal density of fossil leaves (Van der Burgh et al., 1993; Kürschner et al., 1996) and $\delta^{13}C$ of marine organic carbon (Raymo and Rau, 1992; Raymo et al., 1996). Estimates of the atmospheric CO_2 levels during the warmest intervals are $\sim 400 \pm 25$ ppmv (e.g., Kürschner et al., 1996; Raymo et al., 1996; Royer, 2006; Jansen et al., 2007; Tripati et al., 2009; Pagani et al., 2010).

Although the forcing from an atmospheric CO_2 concentration of even 425 ppmv ($\sim$2.5 Wm^{-2}) is sufficient to explain the average planetary temperature, it fails to match the strong regional gradients in the paleoclimate proxies (Crowley, 1996). The most likely additional change is in the poleward ocean heat transport term. Temperature changes were muted in the tropics and large in the Arctic (Haywood et al., 2005; Dowsett and Robinson, 2009), inconsistent with a purely greenhouse gas forced warm interval (Haywood et al., 2009). Thus, it is likely that middle Pliocene warmth originated primarily from changes in greenhouse gas concentrations in the atmosphere, modified by an invigorated Meridional Overturning Circulation, and subsequently amplified in the Arctic by strong positive feedbacks (of which changes in Arctic Ocean sea ice are probably the most important and least constrained), with a slight possibility that other processes also contributed.

4. Relations between forcings and feedbacks

The strength of some feedbacks depends on the duration of the forcing. For example, the positive summer insolation anomaly across the Arctic during the HTM and LIG produced immediate responses from sea ice and snow cover feedbacks, even though winter insolation anomalies were negative. On the other hand, water vapor feedback, one of the strongest positive feedbacks with hemispheric impacts at mid- and low-latitudes, requires longer than seasonal forcing. The transfer of water vapor to the atmosphere from the ocean depends largely on sea-surface temperatures, as well as wind stress. The e-folding time required for the oceanic mixed layer (50–100 m thick) temperatures to equilibrate to a specific forcing is 2–3 years (Lahiff, 1975), much longer than the seasonal change in summer insolation due to orbital terms, which are offset by comparable negative winter anomalies at low- and mid-latitudes. Thus, the water vapor feedback is small in response to purely orbital forcing (Rind, 2006; see also Masson-Delmotte et al., 2006).

5. Comparing past Arctic temperature anomalies to hemispheric anomalies

During the past 65 Ma, the Arctic has experienced a greater change in temperature, vegetation, and ocean surface characteristics than has any other Northern Hemisphere latitudinal band (e.g., Sewall and Sloan, 2001; Bice et al., 2006; Zachos et al., 2008). Those times when the Arctic was unusually warm or cold offer insights into the feedbacks within the Arctic system that can amplify changes imposed from outside the Arctic. To assess the geographic differences in the climate response to relatively uniform hemispheric changes in seasonal insolation forcing, Arctic summer temperature anomalies can be compared to the Northern Hemisphere average summer temperature anomalies for the HTM, LGM, and LIG because of the similar forcing in the Arctic and Northern Hemisphere.

A difficulty in developing Northern Hemisphere and Arctic temperature anomaly comparisons is that for most intervals only a limited number of sites are available with quantitative estimates of past temperatures, and the vast majority of these sites reflect only summer temperature estimates. Several recent syntheses have attempted to derive estimates of summer temperature anomalies during warm times for large portions of the Arctic. However, fewer hemispheric syntheses are available. To obtain hemispheric estimates of past temperature anomalies we include climate model simulations driven by the known forcings. These simulations show considerable fidelity in reproducing the global anomalies indicated by a wide range of site-specific summer temperature proxies for the relevant times, and hemispheric anomalies can be assessed within these models. The hemispheric anomalies so produced are consistent with the available paleoclimate data, and so they are used here.

The Paleoclimate Modeling Intercomparison Project (PMIP2; Harrison et al., 2002, and see http://pmip2.lsce.ipsl.fr/) coordinates an international effort to compare paleoclimate simulations produced by a range of climate models with data-based paleoclimate reconstructions for the HTM (PMIP-defined as 6 ka ago) and for the Last Glacial Maximum (LGM; $\sim$21 ka ago). As part of PMIP, Harrison et al. (1998) compared global (mostly Northern Hemisphere) vegetation patterns reconstructed from the output of 10 different climate model simulations for 6 ka. The 6 ka vegetation maps closely agreed with the vegetation reconstructed from paleoclimate records. Similar comparisons on a regional basis for the Northern Hemisphere north of 55°N (Kaplan et al., 2003), the Arctic (CAPE Project Members, 2001), Europe (Brewer et al., 2007), and North America (Bartlein et al., 1998) also showed close matches between paleoclimate data and models for the HTM. Comparison of models and data for the Last Glacial Maximum (Bartlein et al., 1998; Kaplan et al., 2003), and Last Interglaciation (CAPE–Last Interglacial Project Members, 2006; Otto-Bliesner et al., 2006) reached similar conclusions (also see Pollard and Thompson, 1997; Farrera et al., 1999; Pinot et al., 1999; Kageyama et al., 2001). Paleoclimate data correspond reasonably well with model simulations of the HTM and LIG warmth, and LGM cold, although climate model simulations generally underestimate the magnitude of change, and there are significant variations between different models. However, the

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PMIP2 experiments, which include an interactive ocean, more closely match the paleodata for both 6 and 21 ka ago than did the original PMIP experiments (Masson-Delmotte et al., 2006; Braconnot et al., 2007). The general agreement between data and models provides confidence that climate model simulations of past times may be compared with paleoclimate-based reconstructions of summer temperatures for the Arctic. Clearly, however, additional data and additional analyses of existing data sets as well as new data would improve confidence in the results and reduce uncertainties. Notably, current understanding of sea ice conditions in the Arctic Ocean is clearly insufficient and requires both new data generation and a development of better sea ice proxies for paleo records.

Intervals when the Arctic was warmer than present in the recent past as reconstructed from proxy data and independently supported by climate model experiments remain imperfect analogues for future greenhouse gas warming because the forcings are different. This point was stressed in Chapter 9 Section 9.6.2 on p. 724 of the 2007 IPCC Fourth Assessment WG1 Report (IPCC, 2007): "*As with analyses of the instrumental record discussed in Section 9.6.2, some studies using palaeoclimatic data have also estimated PDFs [climate sensitivity probability density function] for ECS [equilibrium climate sensitivity] by varying model parameters. Inferences about ECS made through direct comparisons between radiative forcing and climate response, without using climate models, show large uncertainties since climate feedbacks, and thus sensitivity, may be different for different climatic background states and for different seasonal characteristics of forcing (e.g.,* Montoya et al., 2000*). Thus, sensitivity to forcing during these periods cannot be directly compared to that for atmospheric CO_2 doubling.*" Nevertheless, paleoclimate reconstructions provide essential examples of how the planetary climate system responds to a range of forcings and constrain the relative importance of a range of strong climate feedback mechanisms. In an Arctic context, paleoclimatic reconstructions allow a measure of the effectiveness of Arctic amplification across a range of different climate forcings and boundary conditions, and serve as targets for climate model experiments.

6. Arctic amplification in four case studies

We evaluate paleoclimate estimates of Arctic summer temperature anomalies relative to recent, and the appropriate Northern Hemisphere or global summer temperature anomalies, together with their uncertainties, for the following time periods: the Holocene Thermal Maximum, the Last Glacial Maximum, Last Interglaciation, and the middle Pliocene.

6.1. Holocene Thermal Maximum (HTM)

Arctic $\Delta T = +1.7 \pm 0.8$ °C; Northern Hemisphere $\Delta T = +0.5 \pm 0.3$ °C; global $\Delta T = 0° \pm 0.5$ °C.

A recent summary of summer temperature anomalies in the western Arctic (Kaufman et al., 2004) built on earlier summaries (Kerwin et al., 1999; CAPE Project Members, 2001) and is consistent with more-recent reconstructions (Kaplan and Wolfe, 2006; Flowers et al., 2008). Although the Kaufman et al. (2004) summary considered only the western half of the Arctic, the earlier summaries by Kerwin et al. (1999) and CAPE Project Members (2001) indicated that similar anomalies characterized the eastern Arctic; all syntheses report the largest anomalies in the North Atlantic sector. Although few data are available for the central Arctic Ocean, the circumpolar data set provides an adequate reflection of air temperatures over the Arctic landmasses and adjacent shallow seas.

Climate models suggest that the average planetary summer temperature anomaly was concentrated over the Northern Hemisphere. Braconnot et al. (2007) summarized the results from 10 different climate model contributions to the PMIP2 project that compared simulated summer temperatures 6 ka ago with recent temperatures. The global average summer temperature anomaly 6 ka ago was 0 ± 0.5 °C, whereas the Northern Hemisphere anomaly was $+0.5 \pm 0.3$ °C. These patterns are similar to those in model results described by Hewitt and Mitchell (1998) and by Kitoh and Murakami (2002) for 6 ka ago, and a global simulation for 9 ka ago (Renssen et al., 2006). All simulate little difference in summer temperature outside the Arctic when those temperatures are compared with pre-industrial temperatures.

Because the positive summer insolation forcing is restricted to the Northern Hemisphere and there is no greenhouse gas forcing, Arctic amplification cannot be compared to the global summer temperature anomaly because the global insolation forcing is close to zero and there is no simulated global temperature anomaly (Masson-Delmotte et al., 2006).

6.2. Last Glacial Maximum (LGM)

Arctic $\Delta T = -18 \pm 7$ °C; global and Northern Hemisphere $\Delta T = -5 \pm 2$ °C.

Quantitative estimates of annual or seasonal temperature reductions during the peak of the Last Glacial Maximum are rare for the Arctic. Ice core borehole temperatures, which offer the most compelling evidence (Cuffey et al., 1995; Dahl-Jensen et al., 1998), reflect mean annual temperatures. Greenland paleotemperature estimates cannot be partitioned by season, and it is not yet possible to estimate how much lower summer temperatures were during the LGM from the Greenland proxies, but to partition most of the cooling into the winter months results in unrealistically low winter temperatures, hence we assume equal reductions in all seasons.

There are few terrestrial sites in the Arctic that were ice-free during the LGM and that contain secure paleotemperature proxies for summer (or winter) temperatures. One such site in Beringia suggests summer temperatures were ca 20 °C lower during the LGM (Elias et al., 1996). The LGM limits of the Eurasian Ice Sheet (Svendsen et al., 1999) were used by Siegert et al. (1999) to constrain an ice sheet model that could produce an ice margin compatible with these limits only if the region around the Kara Sea was characterized by extreme polar desert conditions. LGM climate model simulations suggest that the glacial anticyclone over the Laurentide Ice Sheet delivered relatively warm air to much of Beringia, but delivered cold, polar air to much of the northern North Atlantic and downstream across NW Europe, Scandinavia and into the Eurasian Arctic. Because of the limited data sets for temperature reduction in the Arctic during the LGM, a large uncertainty is specified.

The average global temperature decrease during the LGM, based on paleoclimate proxy data, was 5–6 °C, with little difference between the Northern and Southern Hemispheres (Farrera et al., 1999; Braconnot et al., 2007). A similar temperature anomaly is derived from climate model simulations (Otto-Bliesner et al., 2007). Masson-Delmotte et al. (2006) calculated LGM Arctic amplification by comparing mean annual temperatures derived from Greenland ice core data with model-simulated mean annual temperatures. They derived estimates of Arctic amplification of about 2, after correcting for changes in ice sheet height.

The forcings responsible for LGM cold are primarily minimum summer insolation at high northern latitudes and reduction in greenhouse gases (both CO_2 and CH_4). These reductions began much earlier, at the end of the Last Interglaciation (ca 120 ka ago), and contributed to ice sheet growth across much of northern North

America and Eurasia. Slow feedbacks associated with ice sheet growth (ice-elevation and glacial-isostasy; Rind, 1987), coupled with fast feedbacks related to increased albedo, expanded sea ice coverage and reduced planetary water vapor (colder ocean surface waters), amplified the cooling, which reached a maximum during the LGM.

6.3. Last Interglaciation (LIG)

Arctic $\Delta T = +5 \pm 1$ °C; global and Northern Hemisphere $\Delta T = +1 \pm 1$ °C.

A recent summary of all available quantitative reconstructions of summer temperature anomalies for the Arctic during peak Last Interglaciation warmth shows a spatial pattern similar to that shown by HTM reconstructions. The largest anomalies are in the North Atlantic sector and the smallest anomalies are in the North Pacific sector, but the anomalies are substantially larger (5 ± 1 °C) than they were during the HTM (CAPE–Last Interglacial Project Members, 2006). A similar pattern of LIG summer temperature anomalies is apparent in climate model simulations (Otto-Bliesner et al., 2006). Global and Northern Hemisphere summer temperature anomalies are derived from summaries in CLIMAP Project Members (1984), Crowley (1990), Montoya et al. (2000), and Bauch and Erlenkeuser (2003).

The primary forcings responsible for LIG Northern Hemisphere summer warmth include the unusual alignment of precession and obliquity terms in Earth's orbit such that Earth was closest to the Sun in Northern Hemisphere summer when tilt was at a maximum. This produced strong positive summer insolation anomalies across the Northern Hemisphere. These anomalies were amplified by a modest increase in greenhouse gases. Unlike the early Holocene, deglaciation from the preceding glacial maximum was rapid, and sea level reached modern levels 129 ± 1 ka ago, at the same time that the summer insolation maximum was attained. During the Holocene, deglaciation was slower, with sea level reaching modern only 5 ka ago, 6 ka after the insolation maximum. The combination of early deglaciation and greater axial tilt during the LIG relative to the Holocene produced stronger LIG summer temperature anomalies across the Northern Hemisphere.

Increased greenhouse gases and a net positive Northern Hemisphere annual insolation anomaly at the peak of the LIG may have produced modest additional positive feedbacks from increased atmospheric water vapor as the ocean surface temperatures warmed slightly, and from vegetation as boreal forests expanded to the Arctic Ocean coast across large areas.

6.4. Middle Pliocene

Arctic annual $\Delta T = +12° \pm 3$ °C; hemispheric and global annual $\Delta T = +4° \pm 2$ °C.

Widespread forests throughout the Arctic in the middle Pliocene offer a glimpse of a notably warm time in the Arctic, which had essentially modern continental configurations and connections between the Arctic Ocean and the global ocean. The most recent reviews of the Pliocene Arctic Ocean are provided by Matthiessen et al. (2009) and Polyak et al. (this volume). Reconstructed terrestrial and shallow coastal Arctic temperature anomalies are available from several sites that show much warmth and probably no summer sea ice in the Arctic Ocean. These sites include the Canadian Arctic Archipelago (Dowsett et al., 1994; Elias and Matthews, 2002; Ballantyne et al., 2006), Iceland (Buchardt and Símonarson, 2003), and the North Pacific (Heusser and Morley, 1996). A global summary of middle Pliocene biomes by Salzmann et al. (2008) concluded that Arctic mean annual temperature anomalies were in excess of 10 °C; some sites indicate temperature anomalies of as much as 15 °C. Estimates of global sea-surface temperature anomalies are from Dowsett (2007).

Global reconstructions of mid-Pliocene temperature anomalies from proxy data and general circulation models show an average warming of 4 ± 2 °C, with greater warming over land than oceans, but with substantial warming of sea-surface temperatures even at low- to mid-latitudes (Budyko et al., 1985; Chandler et al., 1994a; Raymo et al., 1996; Sloan et al., 1996; Dowsett et al., 1999; Haywood and Valdes, 2004, 2006; Jiang et al., 2005; Salzmann et al., 2008).

The forcing of the warmth of the middle Pliocene remains uncertain. Orbital oscillations cannot be the primary cause, because Pliocene warmth persisted through many orbital cycles ($\gg$100 ka). The most likely explanation is an elevated level of CO_2 (400 ± 25 ppmv, but with notable uncertainties as shown by Jansen et al., 2007; Fig. 6.1), coupled with smaller Greenland and Antarctic Ice Sheets (Haywood and Valdes, 2004; Jansen et al., 2007) and reduced Arctic Ocean sea ice. A possible role for altered oceanic circulation is also considered in some studies (Chandler et al., 1994b; also Jansen et al., 2007 and references therein).

6.5. Earlier warm times

The trend of larger Arctic anomalies was already well established during the early Cenozoic peak warming and of the Cretaceous before that. Somewhat greater uncertainty is attached to these more ancient times in which continental configurations differed significantly from present. Barron et al. (1995) estimated global average temperatures about 6 °C warmer in the Cretaceous than recently. Subsequent work suggests upward revision of tropical sea-surface temperatures by as much as a few degrees (Alley, 2003; Bice et al., 2006). The Cretaceous peak warmth seems to have been somewhat higher than early Cenozoic values, or perhaps similar (Zachos et al., 2001). Early Cenozoic summertime temperatures of ~18 °C in the Arctic Ocean and ~17 °C on adjacent Arctic

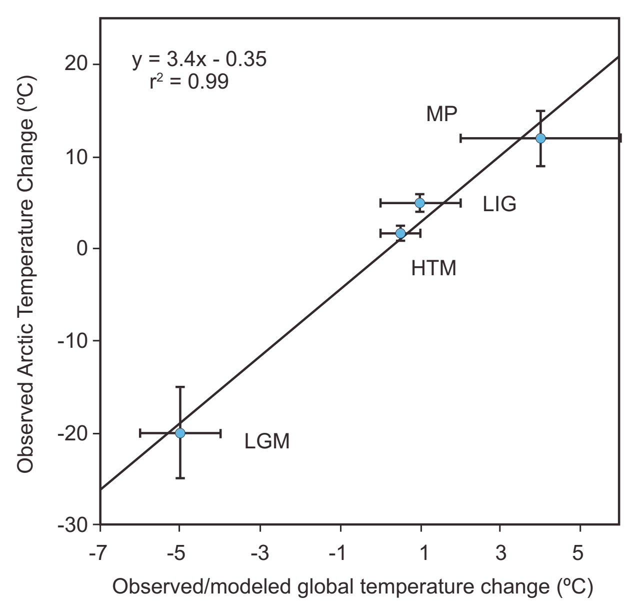

Fig. 4. Paleoclimate data quantify the magnitude of Arctic amplification. Shown are paleoclimate estimates of Arctic summer temperature anomalies relative to recent, and the appropriate Northern Hemisphere or global summer temperature anomalies, together with their uncertainties, for the following: the Last Glacial Maximum (LGM; ~20 ka ago), Holocene Thermal Maximum (HTM; ~8 ka ago), Last Interglaciation (LIG; 130–125 ka ago) and middle Pliocene (~3.5 Ma ago). The trend line suggests that summer temperature changes are amplified 3–4 times in the Arctic.

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lands, were followed during the short-lived Paleocene–Eocene Thermal Maximum by warming to about 23 °C in the summer Arctic Ocean and 25 °C on adjacent lands (Moran et al., 2006; Sluijs et al., 2006, 2008; Weijers et al., 2007). No evidence of wintertime sea ice exists, and temperatures very likely remained higher than during the mid-Pliocene. In both periods, temperature changes in the Arctic were much larger than the globally averaged change. The Cretaceous and early Cenozoic warmth was apparently forced primarily by increased greenhouse gas concentrations (e.g., Donnadieu et al., 2006; Jansen et al., 2007; Royer et al., 2007).

7. Summary and conclusions

Based on four specific case studies, and supported by less well defined estimates from the early Cenozoic and Cretaceous, Arctic amplification appears to have operated during times both colder and warmer than present, and across a wide range of forcing mechanisms. This conclusion is expected; at least some of the strong Arctic feedbacks that serve to amplify temperature change do so without regard to causation – warmer summer temperatures melt reflective snow and ice, regardless of whether the warmth came from changing solar output, orbital configuration, greenhouse gas concentrations, or other causes. Global warmth and an ice-free Arctic during the early Eocene occurred without albedo feedbacks at the same time that the tropics experienced sustained warmth (Pearson et al., 2007). The magnitude of Arctic amplification may depend on the extent to which slow vs fast feedbacks are engaged, and whether hemispherically uniform feedbacks (water vapor, greenhouse gases) are triggered.

The four case studies described above indicate that Arctic temperature anomalies were much larger than global ones in all instances. To compare the magnitude of Arctic amplification across this wide range of forcings and temperature anomalies, the change in Arctic summer temperature anomalies is plotted against the Northern Hemisphere or global temperature anomaly for all four time periods (Fig. 4). A linear regression through these data has a slope of 3.4 ± 0.6, suggesting that the change in Arctic summer temperatures tends to be 3–4 times as large as the global change.

The similarity in the magnitude of Arctic amplification across a range of climate forcing scenarios is surprising and perhaps somewhat fortuitous. For example, HTM warmth was forced almost exclusively by the orbitally driven extra summer insolation across the Northern Hemisphere. This positive summer insolation anomaly was effectively balanced at low- and mid-latitudes by a similar reduction in winter insolation. With the net annual insolation anomaly only slightly greater than zero, there was insufficient time seasonally to warm the ocean's surface water. Consequently, the water vapor feedback was not activated, whereas sea ice and snow cover feedbacks respond efficiently to seasonal insolation anomalies. In contrast, during both the LGM and MP, slow feedbacks were activated that would have strongly altered atmospheric water vapor, yet the magnitude of Arctic amplification is similar in both of those cases to the two interglacial reconstructions.

Circumstances in the past provide imperfect analogues for the near future, when greenhouse gases alone are expected to dominate the forcings, and timescales are too short for slow feedbacks to have a strong contribution. Furthermore, the paleoclimatic data summarized above are from climate states that were relatively stable over millennia or longer, and do not provide detailed information on the path by which Arctic and other regions reached those climate states (Serreze and Francis, 2006). Nevertheless, an emerging secure scenario is that over any forcing, or combination of forcings experienced during the Cenozoic, when the Earth warmed, the Arctic warmed even more. As the Arctic warms, ice (both terrestrial ice and snow, and sea ice) melts, amplifying warmth. And with the loss of terrestrial ice, sea level rises. Because the feedback processes responsible for the observed Arctic amplification in the past remain active today, it is very likely that Arctic amplification will continue for the foreseeable future, if greenhouse gases continue to rise. With this amplification, sea ice will continue to contract, and glaciers and ice sheets will experience accelerated melting, with concomitant increases in the rate of sea level rise.

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