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Past rates of climate change in the Arctic

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ABSTRACT

Climate is continually changing on numerous time scales, driven by a range of factors. In general, longer-lived changes are somewhat larger, but much slower to occur, than shorter-lived changes. Processes linked with continental drift have affected atmospheric circulation, oceanic currents, and the composition of the atmosphere over tens of millions of years. A global cooling trend over the last 60 million years has altered conditions near sea level in the Arctic from ice-free year-round to completely ice covered. Variations in arctic insolation over tens of thousands of years in response to orbital forcing have caused regular cycles of warming and cooling that were roughly half the size of the continental-drift-linked changes. This "glacial-interglacial" cycling was amplified by the reduced greenhouse gases in colder times and by greater surface albedo from more-extensive ice cover. Glacial-interglacial cycling was punctuated by abrupt millennial oscillations, which near the North Atlantic were roughly half as large as the glacial-interglacial cycles, but which were much smaller Arctic-wide and beyond. The current interglaciation, the Holocene, has been influenced by brief cooling events from single volcanic eruptions, slower but longer lasting changes from random fluctuations in the frequency of volcanic eruptions, from weak solar variability, and perhaps by other classes of events. Human-forced climate changes appear similar in size and duration to the fastest natural changes of the past, but future changes may have no natural analog.

1. Introduction

Climate change occurs on all time scales, ranging from years to billions of years. Whether such changes are abrupt depends on both the rate and magnitude of change relative to the variable under consideration. The rate of change is a key determinant of the effect of the change on living things such as plants and animals, on ecosystems, and on humans and human societies. Consider, for example, a 10 °C change in annual average temperature, roughly the equivalent to moving between Birmingham, Alabama and Bangor, Maine. If such a change took place during thousands of years, as happens when variations in the Earth's orbit shift the latitudinal and seasonal distribution solar energy, ecosystems and parts of the physical environment such as sea level would change, but slowly enough to allow human societies to adapt. If a 10 °C change happened in 50 years or less, however, ecosystems would be able to complete only limited adaptation, and human adaptation would be limited as well, with widespread challenges facing agriculture, industry, and public utilities in response to changing patterns of precipitation, temperature, severe weather, and other events. Such abrupt climate changes on regional scales are well documented in the paleoclimate record (National Research Council, 2002; Alley et al., 2003), and these changes were more than ten times faster than the warming of the last century.

Not all parts of the climate system can change at the same rate. Global temperature change is slowed by the heat capacity of the oceans (e.g., Hegerl et al., 2007), for example, and ice sheets grow slowly, regardless of the rate of change. Changes in atmospheric circulation are potentially faster than changes in ocean circulation, owing to the difference in mass and thus inertia of these two

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circulation systems. These differences, in turn, influence important climate properties that depend on the state and movements of oceans and atmosphere. The concentration of carbon dioxide in the atmosphere averaged over a decade depends on ocean exchange, and thus cannot vary rapidly (e.g., Monnin et al., 2001). Methane concentration in the atmosphere, on the other hand, has increased by more than 50% within decades (Severinghaus et al., 1998), as this gas is less dependent on ocean exchange and more dependent on systems that can change rapidly, such as the distribution of wetlands, which in turn depends on rainfall linked to atmospheric circulation.

2. Variability versus change; definitions and clarification of usage

At a time in history when a cold winter, or even a cold month, invokes comments such as "what climate change?", it is important to preface this review of the rates of Arctic change with a few points clarifying the terminology and nature of the changes discussed here.

2.1. Weather versus climate

The globally-averaged temperature difference between an ice age and an interglaciation is 5 to 6 °C (Cuffey and Brook, 2000; Jansen et al., 2007). This is often much smaller than the diurnal temperature change between peak daytime and minimum night-time temperatures. Seasonal temperature changes may also be much larger than that of a glacial-interglacial change (e.g., Trenberth et al., 2007). In assessing the "importance" of a climate change, it is generally accepted that a single change has greater effect on ecosystems and economies, and thus is more "important," if that change is less expected, arrives more rapidly, and stays longer (National Research Council, 2002). In addition, a step change that persists for millennia might become less important than similar-sized changes that occurred repeatedly in opposite directions at random times.

Historically, climate has been taken as a running average of weather conditions at a place or throughout a region. The average is taken for a long enough time interval to largely remove fluctuations caused by "weather." Thirty years is often used for averaging.

Weather, to most observers, implies relatively short term, day-to-day occurrences, which are predictable for only about two weeks. Looking farther ahead than that is limited by the chaotic nature of atmospheric circulation, that is, by the sensitivity of the system to initial conditions (e.g., Lorenz, 1963; Le Treut et al., 2007). To many observers, "weather" refers to those features of Earth's coupled atmosphere-ocean system that are theoretically predictable to two weeks or so, but not beyond.

For many climatologists, however, somewhat longer-term events are often lumped under the general heading of "weather." The year-to-year temperature variability in global average temperature associated with ENSO may be a few tenths of a degree Celsius (e.g., Trenberth et al., 2002), and similar or slightly larger variability can be caused by volcanic eruptions (e.g., Yang and Schlesinger, 2002). The influences of such phenomena are short-lived compared with a 30-year average, but they are long lived compared with the two-week interval described just above. Volcanic eruptions may someday prove to be predictable beyond two weeks. For example, U.S. Geological Survey scientists successfully predicted one of the Mt. St. Helens eruptions more than two weeks in advance (Tilling et al., 1990), and the effects following an eruption certainly are predictable for longer times (Hansen et al., 1992). El Niños are predictable beyond two weeks. However, if one is interested in the climatic conditions at a particular place, a proper estimate would include the average behavior of volcanoes and El Niños, but it would not be influenced by the accident that the starting and ending points of the 30-year averaging period happened to sample a higher or lower number of these events than would be found in an average 30-year period.

The issues of the length of time considered and the starting time chosen are illustrated in Fig. 1. The variability in annual temperatures for the continental United States since 1960 are linked to ENSO, volcanic eruptions, warming related to increased greenhouse gases in the atmosphere, and other factors. Depending on the length of record and portion examined, temperature trends can appear to warm, to cool, or not to change. The warm El Niño years of 1987 and 1988, and the cooling trend in 1992 and 1993 caused by the eruption of Mt. Pinatubo, affect our perception of the time trend, or climate. All possible 30-year (climate) regression lines have a positive slope (warming) with >95% confidence. Thus, all of the short-time-interval lines shown on Fig. 1 are part of a warming climate over a 30-year interval, but clearly reflect weather as well.

2.2. Style of change

In rare situations a 30-year climatology appears inappropriate. As recorded in Greenland ice cores, local temperatures fell many degrees Celsius within a few decades about 13 ka ago at the start of the Younger Dryas, a change much larger than the previous inter-annual variability. The temperature remained low for more than a millennium, and then it increased ~10 °C in a few decades, and it has remained substantially elevated since (Cuffey and Clow, 1997;

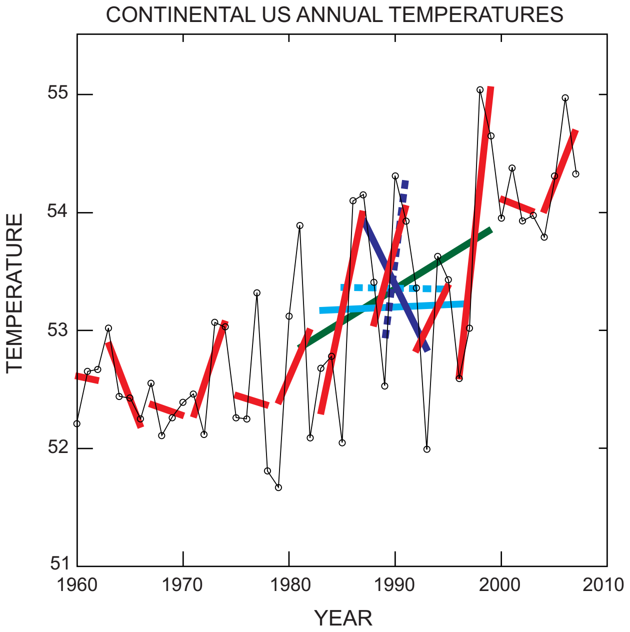


Fig. 1. A "Weather" versus "climate," in annual temperatures for the continental United States, 1960–2007. Red lines are trends for 4-year segments that show how the time period affects whether the trend appears to depict warming, cooling, or no change. Various lines show averages of different number of years, all centered on 1990: Dark blue dash, 3 years; dark blue, 7 years; light blue dash, 11 years; light blue, 15 years; and green, 19 years. The perceived trend can be warming, cooling, or no change depending on the length of time considered. Climate is normally taken as a 30-year average; all 30-year-long intervals (1960–1989 through 1978–2007) warmed significantly (greater than 95% confidence), whereas only 1 of the 45 possible trend-lines (17 are shown) has a slope that is markedly different from zero with more than 95% confidence. Thus, a climate-scale interpretation of these data indicates warming, whereas shorter-term ("weather") interpretations lead to variable but insignificant trends. Data from United States Historical Climatology Network, *http://www.ncdc.noaa.gov/oa/climate/research/cag3/cag3.html* (Easterling et al., 1996).

Severinghaus et al., 1998; Cuffey and Brook, 2000). It is difficult to imagine any observer choosing the temperature average of a 30-year period that included that 10 °C increase and then arguing that this average was a useful representation of the climate. This rapid temperature increase is perhaps the best known and most-representative example of abrupt climate change (National Research Council, 2002; Alley et al., 2003), and the change is ascribed to what is now known colloquially as a "tipping point." Tipping points occur when a slow process reaches a threshold that "tips" the climate system into a new mode of operation (e.g., Alley, 2007). Analogy to a canoe tipping over suddenly in response to the slowly increasing lean of a paddler is appropriate, although the analogy, and the terminology itself, is of limited use. Climate recovers from tipping points; a canoe may sink and thus may not. Indeed, the resilience of climate and its bounded nature that keeps it from tipping to uninhabitable extremes, are hallmarks of the climate often lost in the public discourse on climate change.

Tipping behavior is readily described sufficiently long after the event, but may be less clear while it is happening. Abrupt climate changes are clearly of great societal importance, but their prediction remains elusive. Research on this topic is advancing, and quantitative statements can be made about detection of events, but timely detection may remain difficult (Keller and McInerney, 2007).

2.3. *Spatial characteristics of change*

The Younger Dryas cold event led to prominent cooling around the North Atlantic, weaker cooling around much of the Northern Hemisphere, and weak warming in the far south (reviewed by Alley, 2007). Nonetheless, the most commonly cited records of the Younger Dryas are naturally those that show large signals. Strong local signals can be misinterpreted as global signals. It is essential to recognize the geographic as well as time limitations of climate events and their paleoclimatic records.

Further complicating this discussion is the possibility that an event may start in one region and then require some climatically notable time interval to propagate to other regions. Limited data supported by our basic understanding of how climate processes work suggest that the Younger Dryas cold event began and ended in the north, that much of the response was delayed by decades or longer in the far south, and that transmission of at least some of the signal occurred through the ocean (Steig and Alley, 2003; Stocker and Johnsen, 2003). Cross-dating climate records around the world to the precision and accuracy needed to confirm that relative sequence is a daunting task, especially as both very rapid atmospheric transmission and slower oceanic transmission may have been involved. The mere act of correlating records from different areas then becomes difficult; an understanding of the processes of transmission involved is almost certainly required to support the interpretation. Heinrich (H) events and Dansgaard–Oeschger (D–O) events recorded in ice cores and marine sediment confirm that Younger-Dryas-like abrupt changes occurred frequently during the last glaciation, and these offer additional events that can be targeted to refine our understanding how rapid change initiated in one region of the planet is transmitted to distant regions.

3. Reconstructing rates of change from paleoclimatic indicators

In an ideal world, a discussion of rates of change would not be needed. If climate records were available from all places and all times, with accurate and precise dates, then rate of change would be immediately evident from inspection of those records. However, such a simple interpretation is seldom possible. Uncertainties are always associated with reconstructed climate changes, arising from many sources including measurement uncertainty, relative importance of multiple controls on a given climate proxy, and noise, over time and space. Measurement errors are generally well characterized and usually are relatively small now because of increasingly sophisticated instrumentation and observational protocols. Nonuniqueness of indicators is a more difficult problem. "Noise" may complicate interpretations as well; for example, the beginning and end are difficult to identify exactly for a climate change from one stable but noisy state to another, preventing precise estimation of the rate of change (e.g. Steffensen et al., 2008). In addition, dating uncertainties between records from different sites may preclude highly accurate estimation of the spatial pattern of rate of change.

3.1. *Regional to pan-Arctic stratigraphic marker horizons*

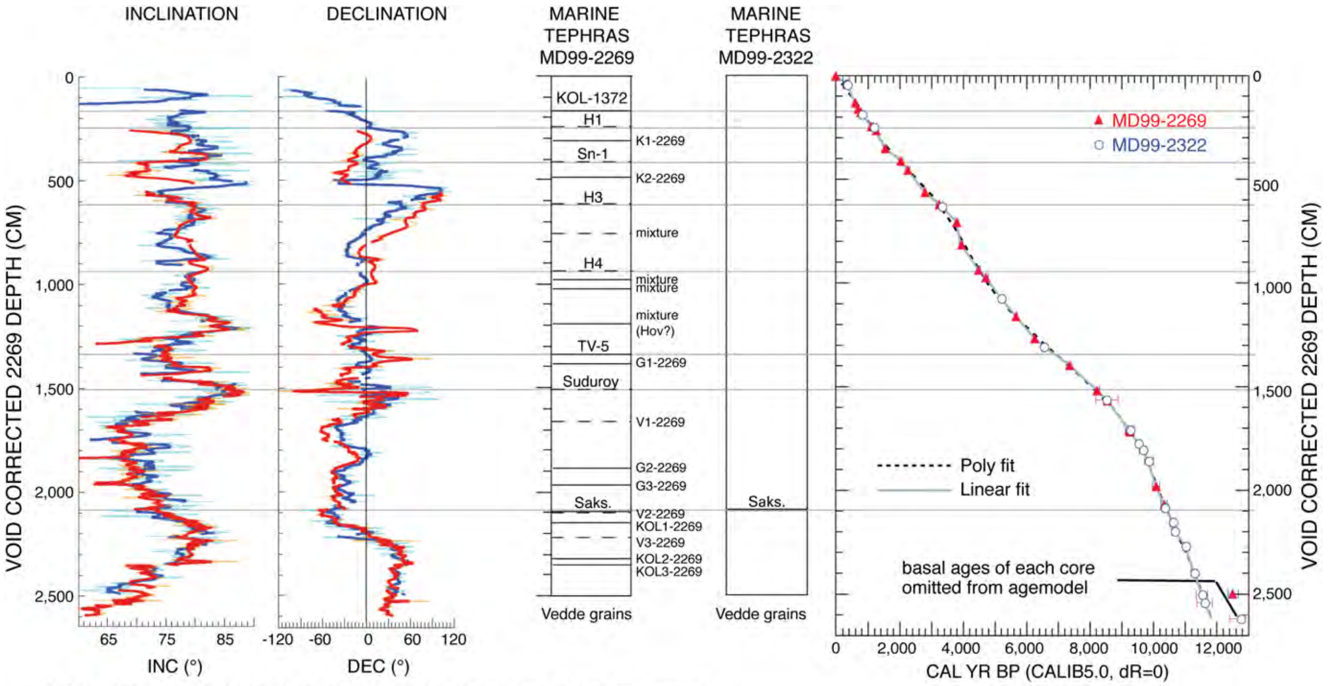
Widely distributed marker horizons can provide a basis to evaluate the spatial pattern of climate change and to derive how the rate of change at one place compares with the rate of change elsewhere.

The Arctic includes major centers of volcanism in the North Atlantic (Iceland) and the North Pacific (Alaska and Kamchatka) sectors. Explosive volcanism from both regions can produce large volumes of source- and time-diagnostic tephra distributed extensively across the Arctic. Volcanic eruptions are discrete events, and major eruptions typically are short lived (hours to days), so that event-specific tephra preserved in lake, bog, and marine sediments and glaciers are almost exactly the same age in all archives. Geochemically distinct tephra layers can be used to synchronize records derived from diverse archives (tephrostratigraphy). Those tephras with diagnostic geochemical signatures that allow them to be securely tied to a specific eruptive event that is independently dated provide chronological constraints for age models (tephrochronology). Synchronizing records with tephrostratigraphy can facilitate the derivation of rates of change. The uncertainties in knowing the time interval between two tephra may be small or large, but whatever the time interval is, it will be the same in all cores containing those two layers.

Paleomagnetic secular variations (PSV) produce systematic sub-centennial to millennial changes in the orientation of magnetic particles in sediment cores (Fig. 2) that may be used to correlate between high-latitude sedimentary archives across regions (Snowball et al., 2007; Stoner et al., 2007). PSV records offer one of the best tools for synchronizing marine and lacustrine sediment cores, avoiding the potential problems of variable marine reservoir ages (Saarinen, 1999; Ojala and Tiljander, 2003; Snowball and Sandgren, 2004).

3.2. *Measurement of rates of change in marine records*

Submillennial climate records in the Arctic are mostly from the continental shelves, where sedimentation rates are sufficiently high. In the central Arctic Ocean perennial sea ice combined with a large distance from the margins results in sedimentation rates less than a few cm ka^{-1} (Polyak et al., 2009, and references therein). In Arctic and Subarctic marine sediment, radiocarbon dating remains the standard technique for obtaining well-dated records during the last 40 to 50 ka. Accelerator mass spectrometry and improved calibration have greatly improved our ability to generate well-constrained age models for high-latitude marine sediment cores, although uncertainties in a variable marine reservoir age remain (e.g., Björck et al., 2003). Arctic cruises mounted in recent years have penetrated deep into the Arctic Ocean with coring systems capable of recovering long (10 to 60 m) sediment cores. Where dates can be obtained from many levels in a core it is



*MD99-2322 adjusted to MD99-2269 depth via synchronization of their PSV records.

Fig. 2. Paleomagnetic secular variations records (left), tephrochronology records (right), and calibrated radiocarbon ages for cores MD99-2269 and -2322 (center) provide a template for Holocene stratigraphy of Denmark Strait (after Stoner et al., 2007, and Kristjansdottir et al., 2007). Solid lines, tephra horizons in core 2269. [Copyright 2007 American Geophysical Union, reproduced by permission American Geophysical Union].

feasible to evaluate centennial and even multidecadal variability (e.g., Ellison et al., 2006; Stoner et al., 2007). However, in the Arctic where cold polar and Arctic water masses dominate the surface water, little carbonate is preserved. These problems hamper development of highly constrained marine chronologies and limit our ability to determine whether high-frequency variability is synchronous or asynchronous between sites.

Where radiocarbon dates can be obtained at the same depth in a core as tephra of known age, deviations of calibrated ages from the tephra age define the marine-reservoir age at that location and time (Eiriksson et al., 2004; Kristjansdottir, 2005; Jennings et al., 2006). For example the Vedde Ash (12.2 ka) is a widely dispersed Icelandic tephra that provides a key Younger Dryas isochron during an interval when marine reservoir ages are poorly constrained and very different from today's. On the North Iceland shelf, changes in the marine reservoir age are associated with shifts in the Arctic and Polar fronts, which have important climatic implications (Eiriksson et al., 2004; Kristjansdottir, 2005).

An approach that uses a combination of PSV and ^{14}C dating allows synchronization of sediment cores and improves the available age control well above the accuracy that each of these methods can achieve on its own. Stoner et al. (2007) applied this technique to two high-accumulation-rate Holocene cores from shelf basins on opposite sides of the Denmark Strait. The large number of tie points between cores provided by the PSV records allowed radiocarbon dates from both cores to be combined to obtain an exceptionally well-dated record of watermass evolution at the gateway to the Arctic (Fig. 2).

3.3. *Rates of change in terrestrial records*

Climate proxies in Arctic terrestrial archives are most closely linked to summer temperature, although some proxies also reflect changes in moisture balance. With sufficient age control, environmental proxies extracted from these archives can be used to evaluate rates of change. The most widely used archives are lake sediments and tree rings, with a growing interest in cave deposits (Lauritzen and Lundberg, 2004), all of which add material incrementally over time. Trees extend to relatively high latitudes in Alaska and portions of the Eurasian Arctic, where they contribute high-resolution, usually annually resolved, paleoclimate records of the past several centuries. But they rarely exceed 400 years duration (Overpeck et al., 1997). Most reconstructions rely on climate proxies preserved in lacustrine archives.

Much of the terrestrial Arctic was covered by continental ice sheets during the LGM, and large areas beyond the ice-sheet margins were too cold for lake sediment to accumulate. Consequently, most lake records span the time since deglaciation, typically the past 10 to 15 ka. Continuous lacustrine records >100 ka provide essential information about past environments and about rates of change in the more distant past (e.g., Lozhkin and Anderson, 1995; Brubaker et al., 2005; Hu et al., 2006; Brigham-Grette et al., 2007). In addition to these continuous records, discontinuous lake-sediment archives are found in formerly glaciated regions. These sites provide continuous records spanning several millennia through past warm times. In special settings, usually where the over-riding ice was very cold, slow-moving, and relatively thin, lake basins have preserved past sediment accumulations intact, despite subsequent over-riding by ice sheets during glacial periods (Miller et al., 1999; Briner et al., 2007; Axford et al., 2009). The rarity of continuous terrestrial archives that span the last glaciation hampers our ability to evaluate how rapid, high-magnitude changes seen in ice-core records (D–O events) and marine sediment cores (H events) are manifested in the terrestrial arctic environment.

3.3.1. *Climate proxies and environmental proxies*

Most high-latitude biological proxies record peak or average summer air temperature. Summarizing the timing and magnitude of peak summer warmth during the Holocene across the North American Arctic, Kaufman et al. (2004) noted that most records rely on pollen and plant macrofossils to infer growing-season temperature of terrestrial vegetation. Summer temperature estimates are also derived from larval head capsules of non-biting midges (chironomids) (Walker et al., 1997). Diatom assemblages primarily reflect changes in water chemistry, which also carries a strong environmental signal.

A wide range of physical and geochemical tracers also provide information about past environments. Biogenic silica (mostly produced by diatoms), organic carbon (mostly derived from the decay of aquatic organisms), and the isotopes of carbon and nitrogen in the organic carbon residues allow the generation of closely spaced data – a key requirement for detecting rapid environmental change. Lakes in carbonate terrain may have sufficient authogenic calcite that $\delta^{18}O$ and $\delta^{13}C$ may provide information on past temperatures and precipitation (e.g., Hu et al., 1999b).

Promising new developments in molecular biomarkers (Hu et al., 1999a; Sauer et al., 2001; Huang et al., 2004; D'Andrea and Huang, 2005) offer the potential of a wide suite of new climate proxies that might be measured at relatively high resolution as instrumentation becomes increasingly automated.

3.3.2. *Dating lake sediment*

In addition to the extraction of paleoenvironmental proxies at sufficient resolution to identify rapid environmental changes in the past, a secure geochronology also must be developed for the sedimentary archive. Methods for developing a secure depth-age relationship include: direct dating, identification of key stratigraphic markers dated independently at other sites, and dating by correlation with an established record elsewhere.

The strengths and weaknesses of various dating methods applied to Arctic terrestrial archives have been reviewed recently (Abbott and Stafford, 1996; Oswald et al., 2005; Wolfe et al., 2005). Radiocarbon is the primary dating method for terrestrial archives of LGM and younger age. The primary challenge to the accuracy of radiocarbon dates in Arctic lakes is the low primary productivity of both terrestrial and aquatic vegetation throughout most of the Arctic, coupled with the low rate at which organic matter decomposes on land. In many Arctic lakes the dissolved organic carbon (DOC) incorporated into the sediment fill includes a high proportion of aged terrestrial carbon, resulting in DOC ages that are anomalously old by centuries to millennia, even when the humic acid fraction is isolated from the bulk DOC (Wolfe et al., 2005).

The large and variable reservoir age of DOC has led most researchers to concentrate on identifiable macrofossils, including seeds, shells, and leaves, for which the residence time is relatively short. Aquatic plants are equilibrated with the carbon in the lake water, which for most lakes is equilibrated with the atmosphere, except for hardwater lakes for which limestone dissolution dilutes the ^{14}C activity of lakewater producing a large and variable overestimation of the actual age. Although macrofossil dates have been shown to be more reliable than DOC dates in Arctic lakes, terrestrial macrofossils washed into lake basins are sometimes derived from reservoirs in the landscape and have radiocarbon ages hundreds of years older than the deposition of the enclosing lake sediments.

For young sediment (20th century), ^{210}Pb (age range of 100 to150 years) and the atmospheric nuclear testing spike of the early 1960s (peak abundances of ^{137}Cs, $^{239,240}Pu$ or ^{241}Am) provide a more reliable age control than radiocarbon. Some lakes preserve annual laminations, owing to strong seasonality in either biological or physical parameters. If laminations can be shown to be annual, chronologies can be derived by counting the number of annual laminations, or varves (Hughen et al., 1996; Francus et al., 2002; Snowball et al., 2002).

For late Quaternary sediment beyond the range of radiocarbon, optically stimulated luminescence (OSL) dating, amino acid racemization (AAR) dating, cosmogenic radionuclide (CRN) dating, uranium-series disequilibrium (U-series) dating may be suitable, and, for volcanic sediment, potassium-argon or argon-argon (K-Ar or $^{40/39}Ar$) dating has been applied (e.g., Bradley, 1999; Cronin, 1999). With the exception of U-series dating, none of these methods has the precision to accurately date the timing of rapid changes directly. But these methods are capable of defining the time range of a sediment package and, if reasonable assumptions can be made about sedimentation rates, then the rate at which measured proxies changed can be derived within reasonable uncertainties.

In some instances, very high resolution down-core analytical profiles from sedimentary archives can be correlated with the same proxy in a well-dated record at a distant locality, such as Greenland ice core records, with little uncertainty. Although the best examples of such correlations are not from the Arctic (e.g., Hughen et al., 2004), this method remains a potential tool for providing age control for Arctic lake sediment records.

3.3.3. *Precision vs accuracy*

Arctic lake sediment may be dated with high precision, but with substantial uncertainty in accuracy. For example, systematic errors in the proportion of old carbon incorporated into the humic acid fraction of the DOC used to date the record may allow an abrubt climate change preserved in the sediment to be reliably dated to have occurred within a 100 year interval with little uncertainty, but the absolute age of the start and end of that 100-year interval may have a large uncertainty (Figs. 3 and 4).

3.4. *Measurement of rates of change in ice-core records*

Ice-core records have figured prominently in the discussion of rates of change during the time interval for which such records are available. One special advantage of ice cores is that they collect

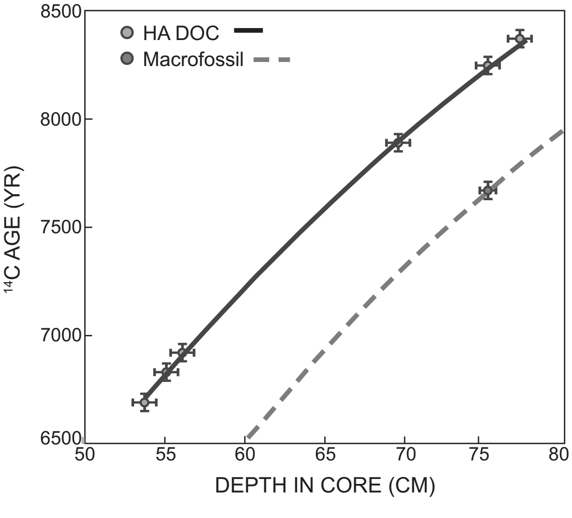


Fig. 3. Precision versus accuracy in radiocarbon dates. Circles are AMS 14C dates on the humic acid (HA) fraction of the total dissolved organic carbon (DOC) extracted from a sediment core from the eastern Canadian Arctic. Square is an AMS 14C date on macrofossil of aquatic moss from 75.6 cm, the same stratigraphic depth as a HA-DOC date. Dashed line is the best estimate of the age-depth model for the core. Samples taken 1to 2 cm apart for HA-DOC dates show a systematic down-core trend suggesting that the precision is within the uncertainty of the measurements (±40 to ±80 years), whereas the discrepancy between macrofossil and HA-DOC dates from the same stratigraphic depth demonstrates an uncertainty in the accuracy of the HA-DOC ages of nearly 600 years. Data from Miller et al. (1999).

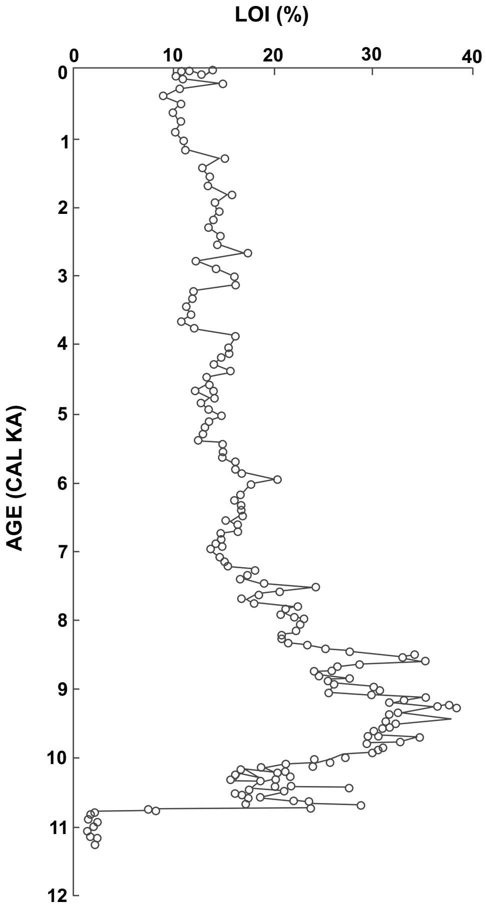


Fig. 4. Down-core changes in organic carbon (measured as loss-on-ignition (LOI)) in a lake sediment core from the eastern Canadian Arctic. At the base of the record, organic carbon increased sharply from about 2% to greater than 20% in less than 100 years, but the age of the rapid change has an uncertainty of 500 years. Data are from Briner et al. (2006). [Copyright 2006, reproduced with permission from Elsevier].

climate indicators from many different regions. In central Greenland, for example, the dust trapped in ice cores has been isotopically and chemically tied to sources in central Asia (Biscaye et al., 1997), the methane has widespread sources in the Arctic and in low latitudes (e.g., Harder et al., 2007), and the snowfall rate and temperature are primarily local indicators (see review by Alley, 2000). This aspect of ice-core records allows one to learn whether climate in widespread regions changed at the same time or different times and to obtain much better time resolution than is available by comparing individual records and accounting for the associated uncertainties in their dating.

Ice cores also exhibit very high time resolution. In many Greenland cores, individual years are recognized so that sub-annual dating is possible. Some care is needed in the interpretation. For example, the template for the history of temperature change in an ice core is typically the stable-isotope composition of the ice. The calibration of this template to actual temperature is achieved in various ways (as reviewed by, e.g., Jouzel et al., 1997; Alley, 2000), but the major changes in the isotopic ratios correlate with major changes in temperature with very high confidence, as discussed there. However, owing to post-depositional processes such as diffusion in firn and ice (Johnsen, 1977; Whillans and Grootes, 1985; Cuffey and Steig, 1998; Johnsen et al., 2000), the resolution of the isotope records does decrease with increasing age and depth. Initially the decrease is due to processes in the porous firn, and later it is due to more rapid diffusion in the warmer ice close to the bottom of the ice sheet. The isotopic resolution may reveal individual storms shortly after deposition but be smeared into several years in ice tens of thousands of years old. Normally in Greenland, accumulation rates of less than about 0.2 m/yr of ice are insufficient to preserve annual cycles for more than a few decades; higher accumulation rates allow the annual layers to survive the transformation of low-density snow to high-density ice, and the cycles then survive for millennia or more before being gradually smoothed.

Records of dust concentration appear to be almost unaffected by smoothing processes, but some chemical constituents seem to be somewhat mobile and thus to have their records smoothed over a few years in older samples (Steffensen et al., 1997; Steffensen and Dahl-Jensen, 1997). Unfortunately, despite important recent progress (Rempel and Wettlaufer, 2003), the processes of chemical diffusion are not as well understood as are diffusive processes of isotopic ratios, so confident modeling of the chemical diffusion is not possible and the degree of smoothing is not well quantified. Persistence of relatively sharp steps in chemical records in old ice that is still in normal stratigraphic order demonstrates that the diffusion is not extensive. The high-resolution features of the dust and chemistry records have been used to date the glacial part of the GISP2 core by using mainly annual cycles of dust (Meese et al., 1997) and the NGRIP core by using annual layers in different ionic constituents together with the visible dust layers (cloudy bands; Fig. 5) back to 42 ka (Andersen et al., 2006; Svensson et al., 2006). Fig. 5 shows the visible cloudy bands in a 72 ka section of the NGRIP core. The cloudy bands are generally assumed to be due to tiny gas bubbles that form on dust particles as the core is brought to surface. During storage of core in the laboratory, these bands fade somewhat. However, the very sharp nature of the bands when the core is recovered suggests that diffusive smoothing has not been important, and that high-time-resolution data are preserved.

4. Classes of changes and their rates

The day-to-night and summer-to-winter temperature changes are typically larger – but have less climate-persistence – than long-lived features such as ice ages. This observation suggests that it is wise to separate rates of change on the basis of persistence. Similarly the slow increase in solar irradiance can be discounted on time scales <100 Ma (e.g., Kasting, 2005). For climate change on sub-decadal to millennial timescales special reliance is placed on Greenland ice-core records in the Arctic because of their high time resolution and confident paleothermometery, although an increasing number of marine and lacustrine records at high resolution are expected to expand the spatial reconstructions of abrupt climate change.

4.1. Tectonic time scales

Plate tectonics and related slow shifts in global biogeochemical cycling, together with evolving life forms, can have profound local

Fig. 5. A linescan image of NGRIP ice core interval 2528.35–2530.0 m depth. Gray layers, annual cloudy bands; annual layers are about 1.5 cm thick. Age of this interval is about 72 ka, which corresponds with Greenland Interstadial 19. (Svensson et al., 2005) [Copyright 2005 American Geophysical Union, reproduced by permission American Geophysical Union].

and global effects on climate on $>10^7$ years (e.g., Donnadieu et al., 2006). The primary direct effects of continental drift on the Arctic during the Cenozoic have been to increase the land mass at high latitudes and to modify the degree to which the Arctic Ocean exchanges water with the global ocean by altering the oceanic gateways between landmasses (Moran et al., 2006).

The Arctic 50 Ma ago appears to have been ice free, at least near sea level, with a mean annual temperature of about 15 °C (Miller et al., this volume), compared with recent values closer to −15 °C. This suggests a cooling of roughly 30 °C over the past 50 Ma, or 0.7 °C Ma^{-1}. Although some time intervals may have experienced little or no temperature change, and during other intervals the rate of change may have been larger, an average temperature change of about 1 °C Ma^{-1} is a realistic “tectonic” rate of change.

4.2. *Orbital time scales*

Features of Earth’s orbit cause small changes (<1%) in globally averaged incoming solar radiation but large changes (>10%) in local insolation. Orbital changes serve primarily to redistribute solar energy seasonally from north to south and back (precession) or geographically from poles to equator and back (tilt). The leading interpretation (e.g., Imbrie et al., 1993) is that ice sheets grew when summer insolation at high northern latitudes was at a minimum, allowing survival of snow through the summer melt season. Ice sheets melted when summer insolation at high northern latitudes was at a maximum. Because the globally averaged forcing is nearly zero but the globally averaged response is large (e.g., Jansen et al., 2007), the Earth system must have strong amplifying processes, or feedbacks. Changes in greenhouse-gas concentrations (especially water vapor and carbon dioxide), changes in albedo (via changes in ice and snow), plus changes in vegetation and blocking of the Sun by dust, all provide strong positive feedbacks (Jansen et al., 2007).

The globally averaged glacial/interglacial change is estimated to be 5 to 6 °C (Jansen et al., 2007). Changes in the Arctic were larger. In central Greenland, typical glacial and interglacial temperatures differed by about 15 °C, and the maximum warming from the most-recent ice age was about 23 °C (Cuffey et al., 1995). Very large changes occurred where ice sheets grew during the ice age and melted during the subsequent warming, related to the cooling effect of the higher elevation of the ice sheets, but the elevation change is not the same as a climatic effect.

An order-of-magnitude estimate for the temperature change across the Arctic associated with the end of an ice age is ∼15 °C in ∼15 ka, or about 1 °C ka^{-1}, and peak rates were perhaps twice that. The ice-age cycle of the last few hundred thousand years is often described as consisting of about 90 ka of cooling followed by a rapid transition to peak warmth that persisted for about 10 ka. This implies faster warming than cooling (e.g., Imbrie et al., 1993; Jansen et al., 2007). Thus, cooling rates were probably notably slower on average than 1 to 2 °C ka^{-1} warming rates.

Kaufman et al. (2004) analyzed the timing and magnitude of the peak Holocene warmth throughout broad regions of the Arctic; near and downwind from the melting Laurentide Ice Sheet, peak warmth was delayed until most of the ice was gone (after 8 ka), whereas far from the ice sheet peak warmth was ∼12 ka, closer to the timing of peak summer insolation.

Throughout the Holocene, Arctic summer temperatures have broadly decreased, consistent with the reduction in summer insolation (Kaufman et al., 2004). The magnitude of this cooling was generally less in the Pacific sector (<1 °C) and greater in the Atlantic sector (up to 5 °C); thus, the orbital signal during the Holocene has been 0.1 to 0.5 °C ka^{-1}.

4.3. *Millennial or abrupt climate changes*

Exceptional attention has been focused on the abrupt climate changes recorded in Greenland ice cores and in many other records spanning the same intervals (e.g. Dansgaard et al., 1989; National Research Council, 2002; Alley et al., 2003; Alley, 2007). The most recent of these abrupt changes, the Younger Dryas, has been well known for decades from lacustrine, riverine and bog sediments, and the moraines left by retreating ice sheets and mountain glaciers, primarily in Europe.

The first deep ice core through the Greenland Ice Sheet, at Camp Century (1966), produced a δ^{18}O isotope profile that showed unexpectedly rapid and strong climatic shifts through the entire last glaciation (Dansgaard et al., 1969, 1971; Johnsen et al., 1972). The fastest transitions occurred within centuries, much faster than Milankovitch timescales. Other archives pointed to the same possibility of large and rapid climate changes during the last ice age. For example, the Grand Pile pollen profile (Woillard, 1978, 1979) showed that the last interglacial (MIS 5) ended rapidly, within only 150 ± 75 years, and pointed to many sharp warming events during the subsequent glacial cycle.

The next deep core in Greenland (Dye-3) also produced rapid climatic oscillations that matched the Camp Century results. The cause for these oscillations had already been hinted at by Ruddiman and Glover (1975) and Ruddiman and McIntyre (1981), who assigned the cause for strong climatic anomalies to thermohaline circulation changes combined with strong zonal winds partly driving the surface currents in the North Atlantic; these forces drove sharp north-south shifts of the polar front. In light of the new ice-core data, the oscillations around the Younger Dryas became part of a much longer series of similar events, which Dansgaard et al. (1984) and Oeschger et al. (1984) assigned to circulation changes in the North Atlantic. Broecker et al. (1985) argued for bi-stable North Atlantic circulation as the cause for the Greenland climatic jumps.

Year-by-year sampling of the warming after the Younger Dryas in Dye-3 and subsequent ice cores, redefined the rate of climate change, and pushed the definition of “abrupt” from century time scales to decadal and nearly annual scales (Dansgaard et al., 1989). Alley et al. (1993) suggested the possibility that much of an abrupt climate change was completed in a single year for at least one climatic variable (snow accumulation at the GISP2 site). These studies were notable in that they moved the subject of large climate changes from the century and millennial timescales that are too slow for most policy and societal interests, to timescales of years to decades that are of direct societal importance.

The Younger Dryas cold reversal is well represented in European/North Atlantic records, and continental records down wind. Although a Younger Dryas signal has been detected in the Pacific sector (see Peteet, 1995a,b; Hajdas et al., 1998), the magnitude is substantially reduced. Surprisingly, the Younger Dryas is poorly represented in glacial records from Arctic sites (Miller et al., 2005; Mangerud and Landvik, 2007), possibly because the cold excursion was accompanied by a reduction in precipitation so that Arctic glaciers failed to respond by significant re-advances. And despite the strong signal indicative of rapid, dramatic Younger Dryas cooling in Greenland ice cores, no definitive records document or refute accompanying glacier expansion or cold around the edge of the Greenland Ice Sheet (Funder and Hansen, 1996; Björck et al., 2002; Kelly et al., 2008; Alley et al., this volume). These observations are consistent with the cold excursions being primarily a wintertime phenomenon, whereas paleoclimatic proxies are most sensitive to summertime conditions. Cold excursions are related to changes in the North Atlantic, and their amplitude is reduced away from the core region (Denton et al., 2005; Alley, 2007; also see

Björck et al., 2002). The dependency of magnitude on geographic separation from the North Atlantic is poorly constrained, but ranges from perhaps only one-tenth as large in many parts of the Arctic, to zero temperature change in places, to regions of weak warming in the Southern Hemisphere. The globally averaged signal in precipitation change was weak, although in some regions rainfall seems to have changed markedly (e.g., Cai et al., 2008).

Subsequent ice cores from Greenland (GISP2, GRIP, NGRIP) have provided subannual resolution for several geochemical parameters, and these data have been used for the North-GRIP core to provide absolute dating back to 60 ka (Svensson et al., 2005; Rasmussen et al., 2006; Vinther et al., 2006). The GISP2 and GRIP ice cores have also been synchronized with the NGRIP core through MIS 2 (Rasmussen et al., 2006) and during the Holocene (Vinther et al., 2009), refining our ability to quantify rates of abrupt change. The temperature shifts into warm intervals on millennial timescales (D–O events; Johnsen et al., 1992; Dansgaard et al., 1993) are large, ranging from 10 to 16 °C based on borehole thermometry (Cuffey et al., 1995; Johnsen et al., 1995; Jouzel et al., 1997), and on studies of the isotopic effect of thermal firn diffusion on gas isotopes (Severinghaus et al., 1998; Lang et al., 1999; Leuenberger et al., 1999; Landais et al., 2004; Huber et al., 2006). NGRIP, the most recent of the Greenland deep cores, shows that the rapid warmings into interstadials occur in only 20 years during MIS 3 and 2; this information indicates temperature changes of 0.5 °C a^{-1} or faster.

The continuous annual resolution provided by Greenland ice cores demonstrates that even within the Holocene abrupt climate change occurred. The 160-year cold excursion centered on 8.2 ka, produced 4 to 5 °C cooling in Greenland (Leuenberger et al., 1999). The initial cooling, which occurred over <20 years, and perhaps much less, is believed to have been caused by the emptying of Lake Agassiz in North America (Alley and Ágústsdóttir, 2005). The 8.2 ka cold event is well represented in many North Atlantic Arctic records (e.g., Alley and Ágústsdóttir, 2005; Alley, 2007), where it is registered by glacier readvances due to colder summers, without significant reductions in winter snowfall (Miller et al., 2005), unlike the aridity of the Younger Dryas in the High Arctic.

4.4. Higher-frequency events in the Holocene

The final substantial climate influence of the waning continental ice sheets occurred early in the Holocene, at 8.2 ka. Subsequently, Holocene climate change across the Arctic, although muted compared to glacial times, is not entirely without variability. Slow cooling during the Holocene is linked to orbital forcing modulated by the decay of continental ice sheets. Superimposed on this cooling trend are oscillations of roughly 1 °C or less, at various temporal spacings. Great effort has been expended in determining what is signal versus noise in these records, because the signals are small, and issues of whether events are broadly synchronous or not become important.

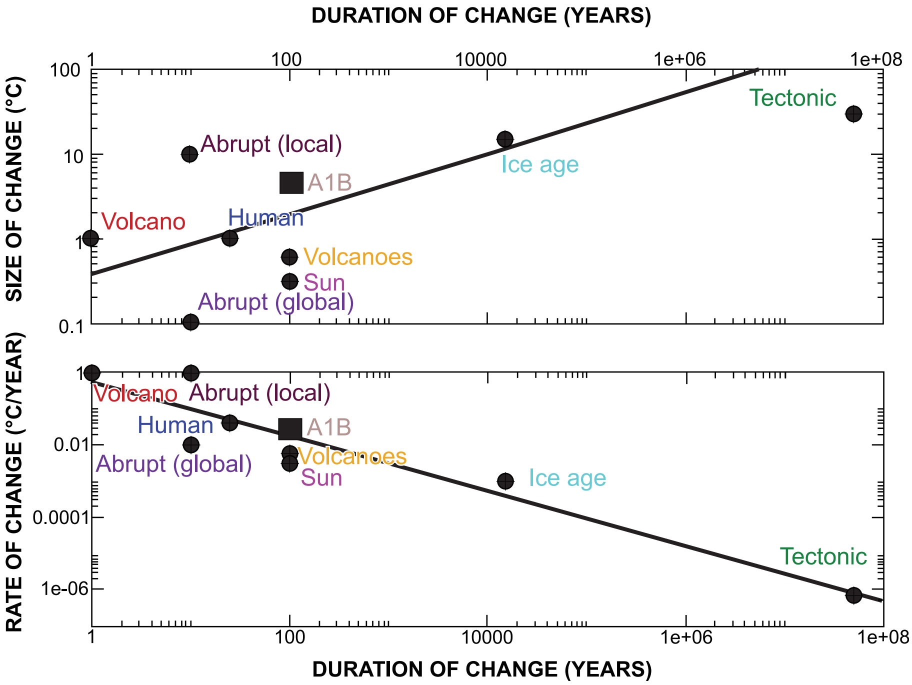


Fig. 6. Summary of estimated peak rates of change and sizes of changes associated with various classes of cause. Error bars are not provided because of difficulty of quantifying them, but high precision is not implied. Logarithmic axes allow the wide range of behaviors to be shown in a single figure. The natural changes during the Little Ice Age–Medieval Climate Anomaly have been somewhat arbitrarily partitioned as 0.6 °C for changes in volcanic-eruption frequency (labeled "volcanoes" to differentiate from the effects of a single eruption, labeled "volcano"), and 0.3 °C for solar forcing to provide an upper limit on solar causes; a larger volcanic role and smaller solar role would be easy to defend (Hegerl et al., 2007), but a larger solar role is precluded by available data and interpretations. The abrupt climate changes are shown for local Greenland values and for a poorly constrained global estimate of 0.1 °C. These numbers are intended to represent the Arctic as a whole, but much Greenland ice-core data have been used in determinations. The instrumental record has been used to assess human effects (e.g. Delworth and Knutson, 2000; Hegerl et al., 2007). The "human" contribution may have been overestimated and natural fluctuations may have contributed to late-20th-century change, but the possibility that the "human" contribution was larger than shown here and that natural variability offset some of the change cannot be excluded. The ability of climate models to explain widespread changes in climate primarily on the basis of human forcing, and the evidence that there is little natural warming forcing during the latter 20th century (Hegerl et al., 2007), motivate the plot as shown. Also included for scaling is the projection for the next century (from 1980–1999 to 2080–2099 means) for the IPCC SRES A1B emissions scenario (one often termed "middle of the road") scaled from Fig. 10.7 of Meehl et al. (2007); see also Chapman and Walsh (2007). This scenario is shown as the black square labeled A1B; a different symbol was chosen to show the fundamental difference of this scenario-based projection from data-based interpretations for the other results on the figure. Human changes could be smaller or larger than shown as A1B, and they may continue to possibly much larger values farther into the future. There is no guarantee that human disturbance will end before the end of the 21st Century, as plotted here. The regression lines pass through tectonic, ice-age, solar, volcano, and volcanoes; they are included solely to guide the eye and not to imply mechanisms.

Ice-core records from Greenland show the forcing and response of individual volcanic eruptions. A large explosive eruption caused a cooling of roughly 1 °C in Greenland, and the cooling and then warming each lasted roughly 1 year (Grootes and Stuiver, 1997; Stuiver et al., 1997), although a cool "tail" lasted longer. In a study of 34 volcanic eruptions over the last 720 years recorded in the GISP2 ice core, cooling lasted 1 to 2 years, and the average cooling was 0.7 °C, with a somewhat stronger, 1.8 °C, cooling seen for the largest events (White et al., 1997). Thus, the temperature changes associated with volcanic eruptions are strong, on the order of 1 °C a^{-1}, but not sustained. Because volcanic eruptions are essentially random in time, accidental clustering in time can influence longer term trends stochastically.

The possible role of solar variability in Holocene changes (and in older changes; e.g., Braun et al., 2005) is of considerable interest. Ice-core records are prominent in reconstruction of solar forcing (e.g., Bard et al., 2007; Muscheler et al., 2007). Identification of climate variability correlated with solar variability then allows assessment of the solar influence and the rates of change caused by the solar variability.

Much study has focused on the role of the Sun in the oscillations from Medieval times, through the Little Ice Age and into the 20th Century (Hegerl et al., 2007). In Greenland, the temperatures during this interval varied by ~1 °C. Attribution exercises show that much of this amplitude can be explained by volcanic forcing in response to the changing frequency of large eruptions (Hegerl et al., 2007). In addition, some of this temperature change might reflect oceanic changes (Broecker, 2000; Renssen et al., 2006), and some fraction is probably attributable to solar forcing (Hegerl et al., 2007). Human influences on the environment were measurable, and thus land cover changes and small changes to greenhouse gases such as methane may have also played a role. Although the total range of temperature variations over the past millennium is about 1 °C, rapid warmings and coolings with that interval suggest that the changes closer to 1 °C/century occurred; some fraction of that change is attributable to solar forcing and some to volcanic and perhaps to oceanic processes. Because recent studies tend to indicate greater importance for volcanic forcing than for solar forcing (Hegerl et al., 2007), changes of 0.3 °C/century may be a reasonable estimate of an upper limit for the solar forcing observed (but with notable uncertainty). Weak variations of the ice-core isotopic ratios that correlate with the sunspot cycles and other inferred solar periodicities similarly indicate a weak solar influence (Stuiver et al., 1997; Grootes and Stuiver, 1997). Whether a weak solar influence acting on millennial time scales is evident in poorly quantified paleoclimatic indicators (Bond et al., 2001) remains a hotly debated topic. The ability to explain the climate of the past millennium without appeal to such a periodicity and the evidently very small role of any solar forcing in those events largely excludes a major role for such millennial oscillations in the Holocene.

The warming from the Little Ice Age extends into the instrumental record. In the instrumental data (Parker et al., 1994; also see Delworth and Knutson, 2000), the Arctic sections, particularly the North Atlantic sector, show warming of roughly 1 °C in the first half of the 20th century (and with peak warming rates of twice that average). The warming likely arose from some combination of volcanic, solar, and human (McConnell et al., 2007) forcing, and perhaps some oceanic forcing. The warming was followed by weak cooling and then a similar warming in the latter 20th century (about 1 °C per 30 years) primarily attributable to human forcing with little, and perhaps opposing, natural forcing (Hegerl et al., 2007).

The lack of correlation between indicators of climate and indicators of past magnetic-field strength, or between indicators of climate and indicators of in-fall rate of extraterrestrial materials, (see above; also see Bard and Delaygue, 2008) means that any role of these possible forcings must be minor and perhaps truly zero.

5. Summary

The peak rates of climate change associated with different causes are plotted in a summary fashion in Fig. 6. On longer times, the magnitude of change increases, but the rate of change decreases. This observation suggests that climate does not, on a regular basis, change rapidly by the largest amounts observed on long time scales. The largest forcings, tectonic and large ice sheets, for example, are inherently slow. To illustrate this concept, regression lines were added through the tectonic, ice-age, volcano, volcanoes, and solar points; abrupt climate changes and human-caused changes appear on Fig. 6, but were omitted from this regression because of difficulty in estimating an Arctic-wide value.

The local to regional effects of the abrupt climate changes in the North Atlantic are clearly anomalous compared with the general trend of the regression lines, because these abrupt changes regionally were both large and rapid. These events have commanded much scientific attention for precisely this reason. However, globally averaged, these events are unimpressive: they fall well below the regression lines, thus demonstrating clearly the difference between global and regional behavior. An Arctic-wide assessment of abrupt climate changes would yield rates of change that would plot closer to the regression lines than do either the local Greenland or global values.

Thus far, human influence does not stand out relative to other, natural causes of climate change. However, the projected changes can easily rise above those trends, especially if human influence continues for more than a hundred years and rises above the IPCC "mid-range" A1B scenario. No generally accepted way exists to formally assess the effects or importance of size versus rate of climate change, so no strong conclusions should be drawn from the observations here.

The data clearly show that strong natural variability has been characteristic of the Arctic at all time scales considered. The data suggest the twin hypotheses that the human influence on rate and size of climate change thus far does not stand out strongly from other causes of climate change, but that projected human changes in the future may do so.

The discussion here relied much more heavily on ice-core data from Greenland than is ideal in assessing Arctic-wide changes. Great opportunities exist for generation and synthesis of other data sets to improve and extend the results here, using the techniques described in this paper. If widely applied, such research could remove the over-reliance on Greenland data.

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