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Past Climate Variability and Change in the Arctic and at High Latitudes

Chapter 6 Past Rates of Climate Change in the Arctic

Chapter Lead Authors: James White, University of Colorado, Institute of Arctic and Alpine Research
Richard Alley, The Pennsylvania State University

Contributing Authors:

Anne Jennings, University of Colorado
Sigfus Johnsen, University of Copenhagen
Gifford Miller, University of Colorado
Steve Nerem, University of Colorado

ABSTRACT

Climate changes on numerous time scales for varying reasons, and has always done so. In general, longer-lived changes are somewhat larger but much slower than shorter-lived changes. Processes linked to continental drift have affected atmospheric and oceanic currents and the composition of the atmosphere over tens of millions of years; in the Arctic, a global cooling trend has switched conditions from being ice-free year-round near sea level to icy conditions more recently. Within the icy times, variations in Arctic sunshine in response to features of Earth's orbit have caused regular cycles of warming and cooling over tens of thousands of years that were roughly half the size of the continental-drift-linked changes; this "glacial-interglacial" cycling has been amplified by colder times bringing reduced greenhouse gases and greater reflection of sunlight especially from more-extended ice. This glacial-interglacial cycling has been punctuated by sharp-onset, sharp-end (in as little as 1-10 years) millennial oscillations, which near the north Atlantic were roughly half as large as the glacial-interglacial cycling but which were much smaller Arctic-wide and beyond. The current warm period of the glacial-interglacial cycling has been influenced by cooling events from single volcanic eruptions, slower but longer-lasting changes from random fluctuations in occurrence of volcanic eruptions and from weak solar variability, and perhaps by other classes of events. Very recently, human effects are evident, not yet showing both size and duration that exceed peak values of natural fluctuations further in the past, but with some projections indicating that human influences could become anomalous in size and duration, hence in speed.

6.1. Introduction

Climate change, as opposed to change in the weather (the distinction is defined below), occurs on all time scales, ranging from several years to billions of years. The rate of change, typically measured in degrees Celsius (°C) per unit of time (years, decades, centuries, or millennia, for example, if temperature is being considered) is a key determinant of the effect of the change on living things such as plants and animals; collections and webs of living things, such as ecosystems; and humans and human societies. Consider, for example, a 10˚C change in annual average temperature, roughly the equivalent to going from Birmingham, Alabama, to Bangor, Maine. Such a change occurring over thousands of years, as happens when the earth's orbit varies and portions of the planet receive more or less energy from the sun, will alter ecosystems and change the environment including sea level, but the slow change would allow time for adaptation by human societies. A 10˚C change occurring in 10 years, however, is fundamentally different (National Research Council, 2002). Ecosystem adaptation would be very limited, for example, with trees typically unable to migrate that fast by seed dispersal. Human adaptation would be limited as well, with widespread challenges to agriculture, industry, and public utilities in response to changing patterns of precipitation, severe weather events, and others. Such abrupt climate changes on regional scales are well-documented in the paleoclimate record (Alley et al, 2003; National Research Council, 2002). This rate of change is about 100 times as fast as the warming of the last century.

Not all parts of the climate system can change this rapidly. Global temperature change is slowed by the heat capacity of the oceans, for example (e.g., Hegerl et al., 2007); local changes, particularly in continental interiors or where sea-ice changes affect the interaction between ocean and atmosphere, can be faster and larger. Changes in atmospheric circulation are potentially faster than changes in ocean circulation, simply due to the difference in mass and thus inertia of these two circulating systems. This difference in turn influences important climate properties dependent on oceanic or atmospheric circulation. The level of carbon dioxide in the atmosphere, for example, depends in part on ocean circulation, and thus does not naturally vary exceedingly rapidly (e.g., Monnin et al., 2001). Methane concentration in the atmosphere, on the other hand, has increased by more than 50% in decades (Severinghaus et al., 1998), as this gas is more dependent on the distribution of wetlands, which in turn depend on atmospheric circulation to bring rains.

In the following pages we examine past rates of environmental change observed in Arctic paleoclimatic records. We begin with some basic definitions and clarification of concepts. Climate change can be evaluated absolutely, using numerical values of temperature, rainfall, etc., or can be evaluated relative to the impacts caused (National Research Council, 2002), with different groups often having differing views on what constitutes "important." Hence, we begin with a common vocabulary.

6.2. Variability versus change; definitions and clarification of usage

Climate scientists and weather forecasters are familiar with opposite sides of very common questions. Does this hot day (or month, or year) prove that global warming is occurring, or, does this cold day (or month, or year) prove that global warming is not occurring? Does global warming mean that tomorrow (or next month, or next year) will be hot, or does the latest argument against global warming mean that tomorrow (or next month, or next year) will be cold? Has the climate changed? When will we know that the climate has changed? To people accustomed to seven-day weather forecasts, in which the forecast beyond the first few days is not very accurate, the answers are often not very satisfying. The next sections briefly discuss some of the issues involved.

6.2.1. Weather vs. climate

The globally averaged temperature difference between an ice age and an interglacial is about 5-6^{o}C (Cuffey and Brook, 2000; Jansen et al, 2007). The 12-hour temperature change between peak daytime and minimum nighttime temperatures at a given place, or the 24-hour change, or the seasonal change, may be much larger than that 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 (NRC, 2002). In addition, a step change that then persists for millennia might become less important than similar-sized changes that occurred repeatedly in opposite directions at randomly varying times.

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

Weather, to most observers, implies day-to-day occurrences, which are predictable for only about two weeks. Looking further ahead than that is limited by the chaotic nature of the atmospheric system; that is, by the sensitivity of the system to initial conditions (e.g., Lorenz, 1963; Le Treut et al., 2007), as described next. All thermometers have uncertainties, even if only a fraction of a degree, and all measurements by thermometers are taken at particular places and not in-between. All temperature estimates at and between thermometers are thus subject to some uncertainty. A weather-forecasting model can correctly be started from a range of possible starting conditions that differ by an amount equal to or less than the measurement uncertainties. For short times of hours or even days, the different starting conditions provided by the modern observational system typically have little effect on the weather forecast; vary the starting data within the known uncertainties, and the output of the model will not be affected much. However, if the model is run for times beyond a few days to perhaps a couple of weeks, the different starting conditions produce very different forecasts. The forecasts are "bounded"—they do not produce blizzards in the tropics or tropical temperatures in the Arctic wintertime, for example, instead producing "forecasts" recognizably possible for all regions covered—but the forecasts differ greatly in the details of where and when convective thunderstorms or frontal systems occur, and how much precipitation will be produced over what time period. To many observers, "weather" refers to those features of the Earth's coupled atmosphere-ocean system that are 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 associated with the El Nino/La Nina phenomenon 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 to a thirty-year average, but long-lived compared to the two-week interval described just above. Volcanic eruptions may prove to be predictable beyond two weeks, and the effects following an eruption certainly are predictable over longer times. El Ninos are predictable beyond two weeks. However, if one is interested in the climatic conditions at a place, a proper estimate would include the average behavior of volcanoes and El Ninos, but it would not be influenced by the accident that the starting and ending points of the thirty-year averaging period happened to sample a higher or lower number of these events than would be found in an average thirty-year period.

6.2.2. Style of change

In some situations a thirty-year climatology appears inappropriate. As recorded in Greenland ice cores, the local temperatures fell many degrees Celsius over a few decades about 13 ka ago during the Younger Dryas time, a larger change than the interannual variability. The temperature then remained at low values for more than a millennium, and then jumped upward about 10^{o}C in about a decade or less, and has remained substantially elevated since (Severinghaus et al., 1998; Cuffey and Clow, 1997; Alley, 2000). It is difficult to imagine any observer choosing the temperature average of a thirty-year period that included that 10^{o}C jump, and arguing that this average was a useful representation of the climate. The jump 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 (Alley, 2007). Analogy to a canoe tipping over suddenly in response to the slowly increasing lean of a paddler is appropriate.

Tipping behavior is readily described sufficiently long after the event, although it is much less evident that a climate scientist could have predicted the event just before it occurred, or that a scientist experiencing the event could have stated with confidence that conditions had tipped. 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).

6.2.3. How to talk about rates of change

The term "abrupt climate change" has been defined with some authority in the report of the US National Research Council (2002). However, many additional terms such as "tipping point" remain colloquial, although arguably they can be related to well-accepted definitions. For the purposes of this report, preference will be given to common English words whenever possible, with explanations of what is meant, without relying on new definitions of words or on poorly defined words.

6.2.4. Spatial characteristics of change

The Younger Dryas cold event, introduced above in section 6.2.2, involved prominent cooling around the north Atlantic, weaker cooling around much of the northern hemisphere, and weak warming in the far south, with uncertainty about changes in many places and probably with relatively minor globally averaged effect (reviewed by Alley, 2007). The most commonly cited records of the Younger Dryas are those that show large signals. Informal discussions by many of us with people outside our field indicate that the strong local signals are at least occasionally 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 is the possibility that an event may start somewhere and then require some climatically notable time interval to propagate to other regions. Limited data supported by process understanding suggest that the Younger Dryas cold event began and ended in the north, with a response delayed by decades or longer in the far south and transmitted there through the ocean (Steig and Alley, 2003; Stocker and Johnsen, 2003). The mere act of relating records from different areas then becomes difficult; an understanding of the processes involved is almost certainly required to support the interpretation.

6.3. Issues concerning reconstruction of rates of change from paleoclimatic indicators

In an ideal world, there would be no need for a chapter on rates of change. If climate records were available from all places and all times, with accurate and precise dating, then rate of change would be immediately evident from inspection of those records. However, as suggested in the previous section, such a simple interpretation is seldom possible.

Consider a hypothetical example. A group of tree trunks, bulldozed by a glacier and incorporated into glacial sediments, is now exposed at a coastal site. Many trees were killed at approximately the same time. The patterns of thick and thin rings, dense and less-dense wood, and isotopic variation of the wood layers contain climatic information (e.g., White et al., 1994). The climatic fluctuations that controlled the tree-ring characteristics can be dated precisely relative to each other—this isotopic event occurred seven years after that one. However, the precise age of the start and end of that climate record may not be available.

If much additional wood of various ages is available nearby, and if a large effort is invested, it may be possible to use the patterns of thick and thin rings and other features to match overlapping trees of different ages, and thus to tie the record to still-living trees and provide a continuous record absolutely dated to the nearest year. If this is not possible, but the trees grew within the time span over which radiocarbon can be used, it may be possible to learn the age of the record to within a few decades or centuries, but no better. If the record is older than radiocarbon, and other dating techniques are not available, even larger errors may be attached to estimates of the time interval occupied by the record.

Reconstruction of the climate changes will have associated uncertainties (were the thick and thin rings controlled primarily by temperature changes, or by moisture changes, for example), but once temperatures or rainfall amounts are estimated for each year, calculation of the rate of change from year to year will involve no additional error because each year is accurately identified. However, learning the spatial pattern of climate change may not be possible, because it will not be possible to relate the events recorded by the tree rings to events in records from other places with their own dating difficulties.

Sometimes, however, it is possible to learn the spatial pattern of the climate change, and to learn how the rate of change at a place compared to the rate of change elsewhere. Volcanic eruptions are discrete events, with major eruptions typically short-lived (hours to days often), so that the layer produced by a single eruption in various lake and marine sediments and glaciers is almost exactly the same age in all. If the same pattern of volcanic fallout is found in multiple cores of lake or ocean sediment or ice, then it is possible to compare the rate of change at those different sites. The uncertainties in knowing the time interval between two volcanic layers may be small or large, but whatever the time interval is, it will be the same in all cores containing those two layers.

These and additional considerations motivate the additional discussion of rates of climate change provided here.

6.3.1. Measurement of rates of change in marine records

In arctic and subarctic marine sediments, radiocarbon dating remains the standard technique to obtain well-dated records over the last 40,000 to 50,000 years. Radiocarbon dating is relatively inexpensive, procedures are well-developed, and materials that can be dated usually are more common than for other techniques. Radiocarbon dating is now conventionally calibrated against other techniques such as tree-ring or uranium-series-disequilibrium techniques, which are more accurate but less widely applicable. Continued improvement of this calibration has occurred (e.g. Stuiver et al., 1998; Hughen et al., 2000; 2004). Instrumental improvements have also occurred. In particular, the availability of Accelerator Mass Spectrometer (AMS) radiocarbon analysis allows dating of milligram quantities of foraminifers, mollusks and other biogenic materials—a single seed or tiny shell can be dated, and this analysis of smaller samples than was possible with previous techniques in turn allows finer time resolution in a single core. Taken together, these advances have greatly enhanced our ability to generate well-constrained age models for high-latitude marine sediment cores. In addition, coring systems such as the Calypso corer have been deployed in the Arctic to recover much longer sediment cores (10 to 60 m long). This allows sampling of relatively long time intervals even in sites where sediment has accumulated rapidy. Sites with faster sediment accumulation allow easier "reading" of the history of short-lived events, so higher-resolution paleo-environmental records can now be generated from high-latitude continental-margin and deep-sea sites. Where dates can be obtained from many levels in a core, it is feasible to evaluate centennial and even multidecadal variability from these archives (e.g. Stoner et al., 2007; Ellison et al., 2006).

However, in the Arctic, particularly along eastern margins of oceans where cold Polar and Arctic water masses influence the environment, the production of carbonate that can be dated by radiocarbon techniques is low, and much of the carbonate produced often dissolves after the producing organism dies. In addition, the carbon used in growing the shells is often "old" (that is, the carbon entered the ocean some decades or centuries before being used by the creature in growing its shell; the date obtained is approximately the time when the carbon entered the ocean, and must be corrected for the time interval between the carbon entering the ocean and being incorporated in the shell).

This marine reservoir correction is often more uncertain in the Arctic than elsewhere (e.g. Bjorck et al., 2003), in part because of the strong but time-varying effect of sea ice blocking exchange between atmosphere and ocean; this uncertainty continues to hamper development of highly constrained chronologies. Some important regions, such as near the eastern side of Baffin Island, have received little study since radiocarbon dating by accelerator mass spectrometry was introduced, so the chronology and Holocene climate evolution of this important margin are still poorly known.

As researchers attempt to develop centennial to multidecadal climate records from marine cores, and achieve correlations between records at sub-millennial resolution, the limits of the dating method are often reached, hampering ability to determine whether high-frequency variability is synchronous or asynchronous between sites. Resource limitations generally restrict radiocarbon dating to samples no more frequently than about 500-year intervals. In marine areas with rapid biological production where sufficient biogenic carbonate is available to obtain highly accurate dates, the instrumental error on individual radiocarbon dates may be as small as ± 20 yrs. But, in many arctic archives, it is not possible to obtain enough carbonate material to achieve that accuracy, and many dates are obtained with standard deviations (one sigma) errors of ±80 yrs to a couple of centuries.

A new approach that uses a combination of paleomagnetic secular variation (PSV) records and radiocarbon dating has been shown to improve relative correlation and chronology well above what each of these methods can achieve on its own (Stoner et al., 2007). The Earth's magnetic field varies in strength and direction over time, and affects the magnetization of sediments deposited. Gross features in the field (reversals of direction) have been used for decades in the interpretation of geologic history, but much shorter-lived, smaller features are now being used, and allow correlation among different records by matching the features.

This technique was applied 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 paleomagnetic secular-variation records, along with numerous radiocarbon datings, allowed matching of these cores at the centennial scale (Stoner et al., 2007). In addition, the study has supported development of a well-dated Holocene paleomagnetic secular-variation record for this region (**Fig. 6.1**), which can be used to aid in the dating of nearby lacustrine cores and for synchronization of marine and terrestrial records. Traditionally, volcanic layers such as the Saksunarvatn tephra have been used as time markers for correlation, but these can be used only at the times of major eruptions and not between, whereas the new magnetic technique is continuous. The technique was tested by its ability to independently achieve the same correlations as the volcanic layer, and functioned very well.

As noted above, tephra layers are an important source of chronological control in Arctic marine sediments. Explosive volcanic eruptions from Icelandic and Alaskan volcanoes have resulted in widespread deposition of geochemically distinct tephra layers, each of which marks a unique time. Where the geochemistry of these events is documented, they provide isochrones that can be used to date and synchronize paleoclimate archives (e.g. marine, lacustrine and ice-cores), and to evaluate leads and lags in the climate system. Where radiocarbon dates can be obtained at the same depth in a core as tephra layers, deviations of calibrated ages from the known age of a tephra can be used to determine the marine-reservoir age at that location and time (Eiriksson et al., 2004; Kristjansdottir, 2005; Jennings et al., 2006). An example is the Vedde Ash, a widely dispersed explosive Icelandic tephra that provides a 12,000-year-old constant-time horizon (an isochron) during the Younger Dryas cold period, 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). As many as 22 tephra layers have been identified in Holocene marine cores off north Iceland (Kristjansdottir et al., 2007). Eiriksson et al. (2004) recovered 10 known-age tephra layers of Holocene age. Some of the Icelandic tephras have wide geographic distributions either because these were very large explosive eruptions or because tephra particles were transported on sea ice, whereas nearer to their source, the tephra layers are more numerous and locally distributed. Transport on sea ice may spread the deposition time of a layer to months or years, but the layer will still remain a very short-interval time marker.

6.3.2 Measurement of rates of change in terrestrial records

Terrestrial archives across the Arctic have been tapped to evaluate changes in the climate system in prehistoric times, with particular emphasis on changes in summer temperature, although moisture balance has been addressed in some studies. With sufficient age control, environmental proxies extracted from these archives can be used to evaluate rates of change. Archives that accumulate sediment in a regular and continuous pattern have the highest potential for reconstructing rates of change. The most promising archives are lake sediments and tree rings, both of which add material incrementally over time. Long-lived trees reach only to the fringes of the Arctic, so most reconstructions rely on climate proxies preserved in the sediments that accumulate in lake basins. Trees do extend into 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 rarely exceed 400 years duration (Overpeck et al., 1997). The steady accumulation of calcium carbonate precipitates in caves may also provide a continuous paleoenvironmental record (Lauritzen and Lundberg, 2004), although these archives are relatively rare in the Arctic. This overview focuses on how well we can reconstruct times of rapid change in terrestrial sediment archives from the Arctic, focusing on changes that occurred on timescales of decades to centuries during the past 150,000 years or so, the Late Quaternary.

Much of the terrestrial Arctic was covered by continental ice sheets during the last glacial maximum (until ~15 ka ago), and large areas outside the ice sheet margins were too cold for lake sediment to accumulate. Consequently, most lake records span the time since deglaciation, typically the past 10,000 to 15,000 years. In a few arctic regions, longer, continuous lacustrine records more than 100,000 years long have been recovered, and these rare records provide essential information about past environments and about rates of change in the more distant past (e.g. (Brigham-Grette et al., 2007; Brubaker et al., 2005; Hu et al., 2006; Lozhkin and Anderson, 1995). 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 (Briner et al., 2007; Miller et al., 1999).

The lack, or at least rarity, of terrestrial archives that span the last glaciation hampers our ability to evaluate how rapid, high-magnitude changes seen in ice-core records (Dansgaard – Oeschger, or D-O events) and marine sediment cores (Heinrich, or H- events) are manifest in the terrestrial arctic environment.

6.3.2a Climate indicators and ages.

Deciphering rates of change from lake sediment, or any geological archive, requires a reliable environmental proxy, and a secure geochronology.

<u>Climate/Environmental Proxies</u>: Most high-latitude biological proxies record peak or average summer air temperatures. The most commonly employed paleoenvironmental proxies are biological remains, particularly pollen grains and the siliceous cell walls (frustules) of microscopic, unicellular algae called diatoms, which preserve well and occur in great abundance in lake sediment. In a summary of 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. Diatom assemblages primarily reflect changes in water chemistry, which also carries a strong environmental signal. More recently, biological proxies have expanded to include larval head capsules of non-biting midges (chironomids) that are well preserved in lake

sediment. The distribution of the larval stages of chironomid taxa exhibit a strong summer temperature dependence in the modern environment (Walker et al., 1997), which allows fossil assemblages to be interpreted in terms of past summer temperatures.

In addition to biological proxies that provide information about past environmental conditions, there is also a wide range of physical and geochemical tracers that provide information about past environments. Biogenic silica (mostly produced by diatoms) and organic carbon (mostly derived from the decay of aquatic organisms), and the isotopes of carbon and nitrogen in the organic carbon residues, can be readily measured on small volumes of sediment, allowing the generation of closely spaced data, a key requirement to detect rapid environmental change. Some lakes have sufficiently high levels of calcium and carbonate ions that calcium carbonate precipitates in the sediment column. The isotopes of carbon and oxygen extracted from calcium carbonate deposits in lake sediment offer proxies of past temperatures and precipitation, and have been used to reconstruct times of rapid climate change at high latitudes (e.g., Hu et al., 1999b).

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

<u>Dating lake sediment</u>: 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. Developing a secure depth-age relationship generally falls into one of three categories: 1) direct dating, 2) identification of key stratigraphic markers dated independently at other sites, and 3) dating by correlation to an established record elsewhere. Much similarity exists between the techniques applied in lakes and in marine environments, although with some differences.

Direct dating: 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 archives dating from the past 15,000 years and sometimes beyond, although conditions endemic to the Arctic and described next often prevent application of the technique back to the limit achieved elsewhere of about 40,000 to 50,000 years. The primary challenge to accuracy of radiocarbon dates in Arctic lakes is the low primary productivity of both terrestrial and aquatic vegetation across most of the Arctic, coupled with the low rate of organic-matter decomposition on land. These two factors work together so that dissolved organic carbon incorporated in lake sediment contains a significant proportion of material that grew on land, was stored on land for long times, and was then washed into the lake. The carbon in this terrestrial in-wash is much older than the sediment in which it is deposited, producing dissolved-organic-carbon ages that are anomalously old by centuries to millennia (Wolfe et al., 2005). Dissolved organic carbon includes many different compounds including humic acids; these acids tend to have the lowest reservoir ages among the compounds, and so are most often targeted when no other options are available.

The large and variable reservoir age of dissolved organic carbon has led most researchers to avoid it for dating, and instead they concentrate on sufficiently large, identifiable organic remains such as seeds, shells, leaves, or other materials, typically called macrofossils. Macrofossils of things living on land, such as land plants, almost always yield accurate radiocarbon ages because the carbon in the plant was fully and recently exchanged (equilibrated) with the atmosphere. Similarly, aquatic plants are equilibrated with the carbon in the lake water, which for most lakes is equilibrated with the atmosphere. However, some lakes contain sufficient calcium carbonate, which typically contains old carbon not equilibrated with the atmosphere, such that the ^{14}C activity of the lake water is not in equilibrium with the atmosphere, a fundamental assumption for accurate radiocarbon dating. In these settings, known as hard-water lakes, marcofossils of terrestrial origin are targeted for dating. In lakes without this hard-water effect, either terrestrial or aquatic macrofossils may be targeted. Although macrofossils have been shown to be more reliable than bulk-carbon dates in Arctic lakes, in many instances terrestrial macrofossils washed into lake basins are derived from stored reservoirs in the landscape, and have radiocarbon ages significantly (hundreds of years) older than the deposition of the enclosing sediments.

For young sediment (20th Century), the best dating methods are ^{210}Pb (age range of about 100 to 150 years) and identification of the atmospheric nuclear testing spike of the early 1960s, usually either with peak abundances of ^{137}Cs, ^{239,240}Pu or ^{241}Am. These methods usually provide high-precision age control for sediments deposited within the past century.

Some lakes preserve annual laminations, due to strong seasonality in either biological or physical parameters. If laminations can be shown to be annual, chronologies can be derived by simply counting the number of annual laminations, or varves (Francus et al., 2002; Hughen et al., 1996; Snowball et al., 2002).

For Late Quaternary sediments beyond the range of radiocarbon dating, available dating methods include optically-stimulated luminescence (OSL) dating, amino acid racemization (AAR) dating, cosmogenic radionuclide (CRN) dating, uranium-series disequilibrium (U-series) dating, and for volcanic sediment, potassium-argon or argon-argon (K-Ar or $^{40/39}$Ar) dating (see, e.g., Cronin, 1999 or Bradley, 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 changes in measured proxies occurred can be derived within reasonable uncertainties. U-series dating has stringent depositional-system requirements that must be met to be applicable. For the terrestrial realm, calcium carbonate accumulations precipitated in a regular fashion in caves (flowstones, stalagmites, stalactites) offer the optimal materials. In these instances, high-precision ages can be derived for the entire Late Quaternary time period.

Stratigraphic markers: As noted in the previous subsection, 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. These tephra layers provide time-synchronous marker horizons that offer key tools to constrain the geochronology of lacustrine sediment records. The tephra layers can also serve to precisely synchronize records derived from lacustrine, marine and ice-sheet archives, thereby allowing a better assessment of leads and lags in the climate system and the phasing of abrupt changes identified in different archives. Most tephras have diagnostic geochemical signatures that allow them to be securely identified as to source, and with modest age constraints, to the actual eruptive event, which may be well dated in source-proximal regions, and such tephras then become dating tools in a technique known as tephrachronology.

As indicated in section 6.3.1, systematic centennial to millennial changes in the Earth's magnetic field (paleomagnetic secular variation) (**Fig. 6.1**) have been used to correlate between high-latitude lacustrine sedimentary archives, and between marine and lacustrine records in the same region (Snowball et al., 2007; Stoner et al., 2007). Lacustrine records of paleomagnetic secular variation calibrated with varved sediments have been used for dating in Scandinavia (Ojala and Tiljander, 2003; Saarinen, 1999; Snowball and Sandgren, 2004)]. Recent work on marine sediments suggests that paleomagnetic secular variation can provide a viable means of marine/terrestrial correlation.

"Wiggle matching": In some instances, very-high-resolution down-core analytical profiles from sedimentary archives with only moderate age constraints can be conclusively correlated to a well dated high-resolution 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 remains a potential tool to provide age control for Arctic lake sediment records.

6.3.2b Potential for reconstructing rates of environmental change in the terrestrial Arctic

A goal of paleoclimate research is to understand rapid changes on human timescales of decades to centuries. The major challenges to meet this goal for the Arctic include uncertainties in the timescales of terrestrial archives and in the interpretation of the various environmental proxies. Although uncertainties are widespread in both aspects, neither presents a fundamental impediment to the primary goal, quantifying rates of change.

Precision vs Accuracy: Many Arctic lake archives are dated with high precision, but with greater uncertainty in their accuracy. One can say, for example, that a particular climate change recorded in a section of core occurred over a 500-year interval with little uncertainty, but the exact age of the start and end of that 500-year interval are much less certain. This is due to systematic errors in the proportion of old carbon incorporated into the humic acid fraction of the dissolved organic carbon used to date the lake sediment. Although this fraction, or “reservoir age”, varies through the Holocene, changes in the reservoir age occur relatively slowly.

Figure 6.2 shows a segment of a sediment core from the Eastern Canadian Arctic, for which six humic acid dates define an age-depth relation with an uncertainty of only ±65 years, but the humic acid ages are systematically 500 to 600 years too old. In this situation, calculating rates of change over decades to centuries can be done with confidence, although determining whether a rapid change at this site is correlative with a rapid change elsewhere is much less certain due to the large uncertainty in the accuracy of the humic acid dates.

Figure 6.3 similarly provides an example of rapid change in an environmental proxy in an Arctic lake sediment core, for which the rate of change can be estimated with certainty, but the timing of the change is less certain.

6.3.3 Measurement of rates of change in ice-core records

Ice-core records have figured especially prominently in the discussion of rates of change, over the time interval for which such records are available. One special advantage of ice cores is that they collect climate indicators from many different regions. In central Greenland, for example, the dust trapped in ice cores has been isotopically and chemically fingerprinted, and comes from central Asia (Biscaye et al., 1997), the methane has widespread sources including 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 allows one to learn whether climate changes in widespread regions occurred at the same time or different times, with much better time resolution than is available from comparison of individual records with 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-isotopic composition of the ice. (The calibration of this template to actual temperature is achieved in various ways, as discussed in chapter 7, but the major changes in the isotopic ratios correlate to major changes in temperature with very high confidence, as discussed there.) However, due to post depositional processes including diffusion in firn and ice (Johnsen, 1977; Whillans and Grootes, 1985; Cuffey and Steig, 1998; Johnsen and others, 2000), the resolution of the isotope records does decrease with increasing age and depth, initially due to processes in the porous firn, and later due to enhanced 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 over 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 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 and others 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 for isotopic ratios, so confident modeling of the chemical diffusion is not possible and the degree of smoothing is not as well quantified as one would like. Persistence of relatively sharp steps 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 using mainly annual cycles of dust (Meese and others 1997), and the NGRIP core using annual layers in different ionic constituents together with the visible dust layers (cloudy bands; **Fig. 6.4**) back to 42 ka ago (Andersen and others 2006, Svensson and others 2006). **Figure 6.4** shows the visible cloudy bands in a 72 ka old section of the NGRIP core. The cloudy bands are generally assumed to be due to tiny gas bubbles that form on dust particles when the core is being brought to surface. During storage of core in the laboratory, these bands fade somewhat. However, the very sharp nature of the bands when recovered suggests that diffusive smoothing has not been important, and that high-time-resolution data are preserved.

6.4 Classes of changes, and their rates

The day-to-night and summer-to-winter changes are typically larger, but have less persistent effect on the climate, than long-lived features such as ice ages. This simple observation suggests that it is wise to separate rates of change based on persistence. As discussed in section 4.2 on forcings, effects from the aging of the sun can be discounted on “short” timescales of 100 million years or less, but many other forcings must be considered. A selection of these is treated in turn below. Over the last ice-age cycle, special reliance is placed on Greenland ice-core records because of their high time resolution and confident paleothermometery. Because Greenland is only a small part of the whole Arctic, this should be borne in mind.

6.4.1 Tectonic time scales

As discussed in section 4.2 on forcings, moving continents and related slow shifts in global biogeochemical cycling, together with evolving life forms, can have profound local and global effects on climate over tens of millions of years. If a continent moves from equator to pole, the climate of that continent will change greatly. In addition, by affecting ocean currents, ability to grow ice sheets, cloud patterns and more, the moving continent may have an effect on global and regional climates as well, although this effect will in general be much more subtle than the effect on the continent's own climate (e.g.. Donnadieu et al., 2006).

Over the last tens of millions of years, the primary direct effect of drifting continents on the Arctic probably has been to affect the degree to which the Arctic Ocean connects to the lower latitudes, by altering the "gateways" between land masses; the Arctic Ocean, primarily surrounded by land masses, has persisted throughout that time (Moran et al. 2006). Much attention has been directed to the possibility that the warmth of the Arctic during certain times, such as the Eocene ~50 Ma ago, was linked to enhanced ocean-heat transport then compared to other, colder times. However, this appears unlikely based on models (e.g., Bice et al., 2000). Furthermore, the data also indicate that this was unlikely; the late-Eocene Arctic Ocean appears to have supported a dense growth of pond weed (Azola), which is understood to grow in brackish waters (those notably fresher than full marine salinity) (Moran et al., 2006). A more-vigorous ocean circulation then would have introduced fully marine waters and would have transported the pond weed away. A great range of studies indicates that higher atmospheric carbon-dioxide concentrations during that earlier time were important in causing the warmth (section 6.2.5, Royer et al., 2007).

The Arctic of ~50 Ma ago appears to have been ice-free, at least near sea level, indicating minimum wintertime temperatures above freezing. Section 7.3.1 includes some indications of temperatures in that time, with perhaps 20°C a useful benchmark for Arctic-wide, average annual temperature. Recent values are closer to -15°C, which would indicate a cooling of roughly 35°C over ~50 million years. The implied rate is then in the neighborhood of 0.7°C/million years or 0.0000007°C/yr. One could pick time intervals over which little or no change occurred, and intervals within the last 50 million years during which the rate of change was somewhat larger; a "tectonic" value of ~1°C/million years or less may be useful.

6.4.2 Orbital time scales

As described in section 4.3 on forcings, features of Earth's orbit cause very small changes in globally averaged incoming solar radiation (insolation), but large changes (more than 10%) in local sunshine, serving primarily to move sunshine from north to south and back or from poles to equator and back, depending on which of the orbital features is considered. The leading interpretation (e.g., Imbrie et al., 1993) is that ice sheets grow, and the world enters an ice age, when reduced summer sunshine at high northern latitudes allows survival of snow without melting; ice sheets melt, and the world exits an ice age, when enhanced summer sunshine at high northern latitudes melts snow there. 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 (feedbacks). Changes in greenhouse-gas concentrations (especially carbon dioxide), in reflection of the sun's energy (ice-albedo feedback, plus some changes in vegetation), and in blocking of the sun by dust are prominent in interpretations, and all appear to be required to explain the size and pattern of the reconstructed changes (Jansen et al., 2007).

The globally averaged change from ice-age to interglacial is typically estimated as 5-6°C (e.g., Jansen et al., 2007). Changes in the Arctic clearly were larger. In central Greenland, typical glacial and interglacial temperatures differed by ~15°C, with a maximum warming from the most-recent ice age of ~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.

In central Greenland, the coldest time of the ice age was about 24 ka ago, although as discussed in chapter 7, some records place the extreme value of the most recent ice age slightly more recently. Kaufman et al. (2004) analyzed the timing of the peak warmth of the Holocene across broad regions of the Arctic; near the melting ice sheet on North America, peak warmth was delayed until most of the ice was gone, whereas peak warmth far from the ice sheet was reached before 8 ka ago, in some regions by a few millennia.

A useful order-of-magnitude estimate may be that the temperature change associated with the end of the ice age was ~15^{o}C in about ~15 ka or about 1^{o}C/ka or 0.001^{o}C/yr, with peak rates perhaps twice that. The ice-age cycle over the last few hundred thousand years is often described as consisting of ~90 ka of cooling followed by ~10 ka of warming, or something similar, implying faster warming than cooling (see **Fig. 7.9**). Thus, rates notably slower than 1-2^{o}C/ka are clearly observed at times.

Kaufman et al. (2004) indicated that the warmest times of the current or Holocene interglacial (MIS 1) in the western-hemisphere part of the Arctic were, for average land, 1.6±0.8^{o}C above mean 20th-century values. The peak warmth occurred before 12 ka ago in western Alaska, and after 3 ka ago in some places near Hudson Bay, with a typical value near 7–8 ka ago. Thus, the orbital signal during the Holocene has been less than or equal to approximately 0.2^{o}C/ka, or 0.0002^{o}C/yr.

6.4.3 Millenial or abrupt climate changes

Exceptional attention has been focused on the abrupt climate changes exhibited in Greenland ice-core records and many other records, from the most recent ice age and beyond (see National Research Council, 2002; Alley et al., 2003; Alley, 2007).

The more recent of these changes had been well-known for decades, from a range of studies primarily in Europe and focused on lake and bog sediments and the moraines left by retreating ice sheets. However, most research was focused on the slower ice-age cycles, which were easier to study in paleoclimatic archives.

The first deep ice core through the Greenland ice sheet, at Camp Century in 1966, produced a $\delta^{18}O$ isotopic profile that showed unexpectedly rapid and strong climatic shifts through the entire last glacial period (Dansgaard et al., 1969; 1971; Johnsen et al., 1972). The fastest observed sharp transitions from cold to warm seemed to have occurred on the timescale of centuries, clearly much faster than Milankovitch timescales.

These results did not stimulate much additional research immediately; the record occurred close to the glacier bed, and it may be that many people suspected that the records had been altered by ice-flow processes. There were, however, data from quite different archives pointing to the same possibility of large and rapid climate change. For example, the Grand Pile pollen profile (Woillard, 1978; Woillard, 1979) showed that the last interglacial (MIS 5) ended rapidly during an interval estimated at 150±75 yrs, comparable to the Camp Century findings. The Grand Pile pollen data also pointed to many sharp warming events during the last ice age.

The next deep core in Greenland at the Dye-3 radar station was drilled by the US, Danish, and Swiss members of the Greenland Ice Sheet Program (Dansgaard et al., 1982). The violent climatic changes, as Willi Dansgaard termed them, matched the often-discarded Camp Century results. The cause for these strong climatic oscillations had already been hinted at by Ruddiman and Glover (1975) and Ruddiman and McIntyre (1981), who studied the oceanic evidence for the large climatic oscillations involving the strong warming into the Bolling, cooling into the Younger Dryas, and warming into the Preboreal. They assigned the cause for these strong climatic anomalies to thermohaline circulation changes combined with strong zonal winds partly driving the surface currents in the north Atlantic, resulting in drastic north-south shifts of the polar front. In the light of the ice core data, the oscillations around the Younger Dryas were part of a long row of similar events, which Dansgaard et al. (1984) and Oeschger et al. (1984) likewise 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.

The results of the Dye 3 core went a long way toward settling the issue of the existence of abrupt climate change. Further results from year-by-year ice sampling during the Younger Dryas warming from this same core pushed the definition of abrupt from the century time scale to the decadal and nearly annual scale (Dansgaard et al., 1989), with Alley et al. (1993) opening the possibility that much of an abrupt change was completed in a single year for at least one climatic variable (snow accumulation at the GISP2 site).

In addition to the GISP2, GRIP and DYE-3 cores, ice core evidence has been strengthened by new deep ice cores at Siple Dome in West Antarctica and North-GRIP in northern Greenland. New high-resolution measurement techniques have provided subannual resolution for several parameters, and these data have been used for the North-GRIP core to provide absolute dating, the GICC05 chronology, back to 60 ka ago (Svensson et al., 2005; Rasmussen et al., 2006; Vinther et al., 2006). The GISP2 and GRIP ice cores have also been synchronized with the North-GRIP core through MIS2 (Rasmussen et al., 2006; in press).

The temperature shifts into the warm intervals in the millennial climate changes, which are called interstadials (Johnsen et al., 1992; Dansgaard et al., 1993), have been found to vary from 10 to 16^{o}C based on borehole thermometry (Cuffey et al., 1995; Johnsen et al., 1995; Jouzel et al., 1997) and 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).

The North-GRIP core, the most recent of the Greenland deep cores and the one on which the most effort was expended in counting annual layers, shows that typically the rapid warmings into interstadials are recorded as increases in only 20 years in the 20-year averages of isotopic values during MIS 2 and MIS 3, indicating temperature changes of 0.5°C/yr or faster.

In the Holocene period, the approximately 160-year long cold event about 8.2 ka ago, with 4 to 5°C cooling (Leuenberger et al., 1999), began in less than 20 years, and perhaps much less. The cooling is believed to have been caused by the emptying of Lake Agassiz (reviewed by Alley and Agustsdottir, 2005) and the fast transitions found bear witness to the dynamic nature of the North Atlantic circulation in jumping to a new mode.

The Younger Dryas and the 8.2 ka cold event (section 7.3.5a) are well-known in Europe and into Arctic regions, but appear to have been much weaker or absent in other Arctic regions (see reviews by Alley and Agustsdottir, 2005 and Alley, 2007; note that strong signals of these events are found in some but not all lower-latitude regions). The signal of the Younger Dryas did extend across the Arctic to Alaska (see Peteet, 1995a; 1995b; Hajdas et al., 1998). Lake sediment records from the Eastern Canadian Arctic contain evidence for both excursions (Miller et al., 2005).

The 8.2 ka event is recorded at two sites as a notable glacier readvance of cirque glaciers and outlet glaciers of local ice caps at 8,200 ± 100 years (Miller et al., 2005). In some lakes not dominated by runoff of meltwater from glaciers, a reduction in primary productivity is apparent at the same time. These records suggest colder summers during the event without a dramatic reduction in precipitation, producing positive mass balances and glacier re-advances. For most local glaciers, this was the last important readvance before they receded behind their Little Ice Age margins. Organic carbon accumulation in a West-Greenland lake sediment record suggests a decrease in biotic productivity synchronous with the negative $\delta^{18}O$ excursion in the GRIP ice core (Willemse and Törnqvist, 1999).

Few Arctic lakes contain records that extend through the Younger Dryas time. But despite the strong signal indicative of rapid, dramatic Younger Dryas cooling in

Greenland ice cores, there are no definitive records documenting or refuting accompanying glacier expansion or cold around the edge of the Greenland ice sheet (Bjorck et al., 2002; Funder and Hansen, 1996) (discussed in Chapter 7), Svalbard (Svendson and Mangerud, 1992), or Arctic Canada (Miller et al., 2005). This is consistent with the joint observations that the events primarily occurred in wintertime whereas most paleoclimatic indicators are more sensitive to summertime conditions, and that the events primarily occurred in the north Atlantic and surroundings with reduced amplitude away from the north Atlantic (Denton et al., 2005; Alley, 2007; also see Bjorck et al., 2002). This means in turn that the rate of climate change associated with these events, although truly spectacular in the north Atlantic, was much smaller elsewhere (poorly constrained, but perhaps only one-tenth as large in many parts of the Arctic, and with a region of zero temperature change somewhere on the planet separating the northern regions of cooling from the southern regions of weak warming). The globally averaged signal in temperature change was weak, although in some regions very strong changes seem to have occurred in rainfall (e.g., Cai et al., 2008).

6.4.4 Higher-frequency events especially in the Holocene

The Holocene record, although showing greatly muted fluctuations in temperature compared to earlier times, is not entirely without variations. As noted above, there is a slow variation during the Holocene linked to orbital forcing and decay of the great ice sheets. Riding on the back of this are oscillations of roughly 1^{o}C or less, at various temporal spacings. Great effort has been invested in determining what is signal versus noise in these records, because the signals are so small, and issues of whether events are broadly synchronous or not become important.

A few rather simple conclusions can be stated with some confidence. Ice-core records from Greenland show the forcing and response of individual volcanic eruptions. A large explosive eruption causes a cooling of roughly 1^{o}C in Greenland, with the cooling and then warming each lasting roughly 1 year (Grootes and Stuiver, 1997; Stuiver et al., 1997), although with a “tail” of cooling that lasts longer. Thus, the temperature changes associated with volcanic eruptions are strong, 1°C/year, 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 older changes; e.g., Braun et al., 2005) is of considerable interest. Ice-core records are prominent in reconstruction of solar forcing (e.g., Muscheler et al., 2007; Bard 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 solar role in the oscillations from the so-called Medieval Warm Period through the Little Ice Age and the subsequent warming to recent conditions. The reader is especially referred to Hegerl et al. (2007). In Greenland, the Little Ice Age/Medieval Warm Period oscillation had an amplitude of roughly 1°C. Attribution exercises show that much of this 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), but some fraction is probably attributable to solar forcing (Hegerl et al., 2007). Although the time from Medieval Warm Period to Little Ice Age to recent warmth is about 1 millennium, there is sufficient structure in that interval that the changes involved are probably closer to 1°C/century, and with some fraction of that attributable to solar forcing, some to volcanic, and some perhaps to oceanic processes. Because recent studies tend to indicate greater importance for the volcanic influence than the solar forcing (Hegerl et al., 2007), changes of 0.3°C/century may a reasonable estimate of an upper limit for the solar forcing observed, with notable uncertainty. Identification of weak variations of the ice-core isotopic ratios correlative with the sunspot cycles and other inferred solar periodicities similarly indicates a weak solar influence (Stuiver et al., 1997; Grootes and Stuiver, 1997). Whether a weak solar influence acting over millennial time scales is evident in poorly quantified paleoclimatic indicators (Bond et al., 2001) remains a hotly debated topic, but the ability to explain the Medieval Warm Period-Little Ice Age oscillation without appeal to such a periodicity, and the evidently very small role of any solar forcing in those events, largely exclude a major role for such millennial oscillations in the Holocene.

The warming from the Little Ice Age extends into the instrumental record, generally consistent with the considerations above. In the reconstruction of Delworth and Knutson (2000), the Arctic sections show warming of roughly 1°C in the first half of the 20th century (and with peak warming rates of twice that average), likely arising from some combination of volcanic, solar, and human (McConnell et al., 2007) forcing perhaps with some oceanic role, followed by weak cooling and then a similar warming in the latter 20th century (roughly 1°C per 30 years) primarily attributable to human forcing with little and perhaps opposing natural forcing (Hegerl et al., 2007).

As noted in section 4.2 on forcings (see above; also see Bard and Delaguye, 2008), the lack of correlation between climatic indicators and indicators of past magnetic-field strength, or between climatic indicators and indicators of infall rate of extraterrestrial materials, means that any role of these possible forcings must be minor if not truly zero.

6.5 Summary

The discussion in the previous session produced estimates for peak rates of climate change associated with different causes. These are plotted in a summary fashion in **Figure 6.5**. As one goes to longer times, the total size of changes increases, but the rate of change decreases. Such behavior is unsurprising; a sprinter changes position very rapidly but does not sustain the rate, so that over a few hours the marathon-runner covers more ground. To illustrate this, regression lines were added through the tectonic, ice-age, volcano, volcanoes and solar points; abrupt climate changes and human-caused changes were omitted from this regression because of difficulty in estimating an Arctic-wide value.

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

Thus far, the 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 the human influence continues beyond a century and rises above the "mid-range" A1B scenario. There exists no generally accepted way of formally assessing the impacts 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 over all time scales considered. The data suggest the twin hypotheses that the human influence on rate and size of climate change thus far has not stood out strongly from other causes of climate change, but that projected human changes in the future may do so.

The report here relied much more heavily on ice-core data from Greenland than would be 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 chapter. If widely applied, such research could remove the over-reliance on Greenland data.

FIGURE CAPTIONS

Figure 6.1 A comparison of paleomagnetic secular variations (inclination and declination, left) records, tephrochronology (right) and calibrated radiocarbon ages for cores MD99-2269 and -2322 (center) provides a Holocene stratigraphic template for the Denmark Straits region (After Stoner et al., 2007 and Kirstjansdottir et al., 2007). Solid lines denote the occurrence of tephra horizons in core 2269.

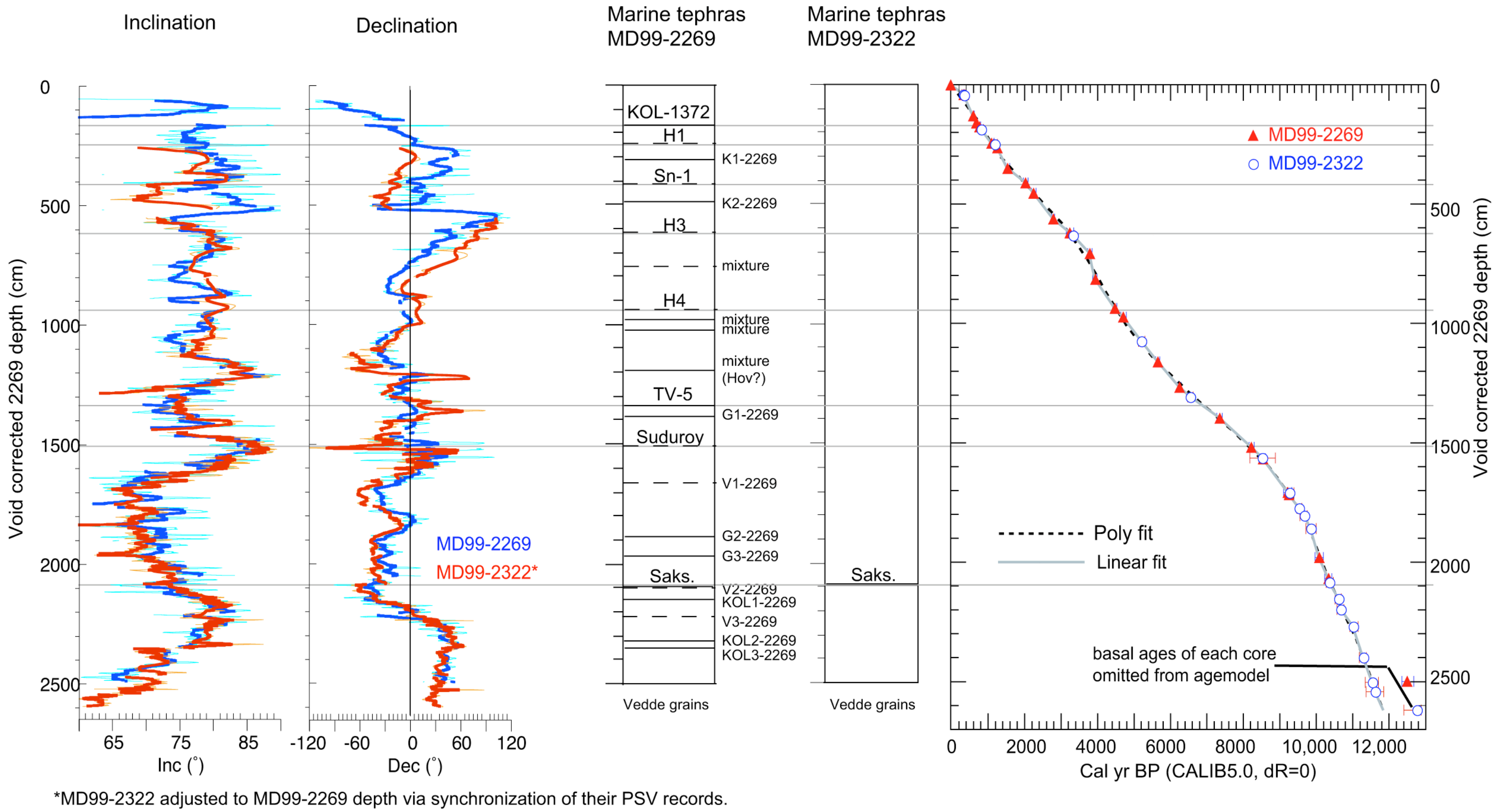
Figure 6.2 Precision vs. accuracy. Six AMS ^{14}C dates on the humic acid (HA) fraction of the total dissolved organic carbon (DOC) extracted from a sediment core from the Eastern Canadian Arctic (blue circles), and one AMS ^{14}C date on macrofossil of aquatic moss from the 75.6 cm, the same depth as one of the HA-DOC dates. Dashed line is the best estimate of the age-depth model for the core. HA-DOC dates taken 1 to 2 cm apart show a systematic downcore trend suggesting the precision is within the uncertainty of the measurements (±40 to ±80 yr), whereas the discrepancy between macrofossil and HA-DOC dates from the same level demonstrates an uncertainty in the accuracy of the HA-DOC ages of nearly 600 years. Data are from [Miller et al., 1999].

Figure 6.3 Down-core changes in organic carbon (measured as loss-on-ignition (LOI)) in a lake sediment core from the Eastern Canadian Arctic. The dramatic increase in organic carbon at the base of the record, from ~2 to >20%, occurred 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].

Figure 6.4. A linescan image showing the NGRIP annual cloudy bands from 2528.35 to 2530.0 m depth. The age is about 72 ka BP corresponding to the Greenland Inter Stadial 19. The annual layer thickness is close to 1.5 cm. (Svensson et al., 2005)

Figure 6.5. Summary of estimated peak rates of change, and sizes of changes, associated with various classes of causes. Error bars are not provided because of difficulty of quantifying, but high precision is not implied. Note that both panels have logarithmic

scales on both axes (log-log plots) to allow the huge range of behavior to be shown in a single figure. The natural changes over the Little Ice Age-Medieval Warm Period 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; 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. Overall, the intent is that these numbers are representative of the Arctic, but Greenland ice-core data have been especially prominent in determinations, with the instrumental record included in assessing the human effects (see Delworth and Knutson, 2000 and Hegerl et al., 2007). One cannot exclude the possibility that the "human" contribution is overestimated and that natural fluctuations contributed to the late-20th-century change, but one also cannot exclude the possibility that the "human" contribution was larger than shown here with natural variability offsetting some of the change; the ability of climate models to explain widespread changes in climate based primarily on human forcing, and the evidence that there is little natural forcing over the latter 20th century (Hegerl et al., 2007), motivate plotting 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 Figure 10.7 of Meehl et al. (2007; also see Chapman and Walsh, 2007). This is shown as the square labeled A1B, with a different symbol chosen to show the fundamental difference of this scenario-based projection from data-based interpretations for the other results on the figure. Human changes clearly could be smaller or larger than shown as A1B, and may continue to possibly much larger values further in the future, with no guarantee that human disturbance will end before the end of the 21st century, as plotted here. The regression lines are through tectonic, ice-age, solar, volcano, and volcanoes, and are included solely to guide the eye and not to imply mechanisms.



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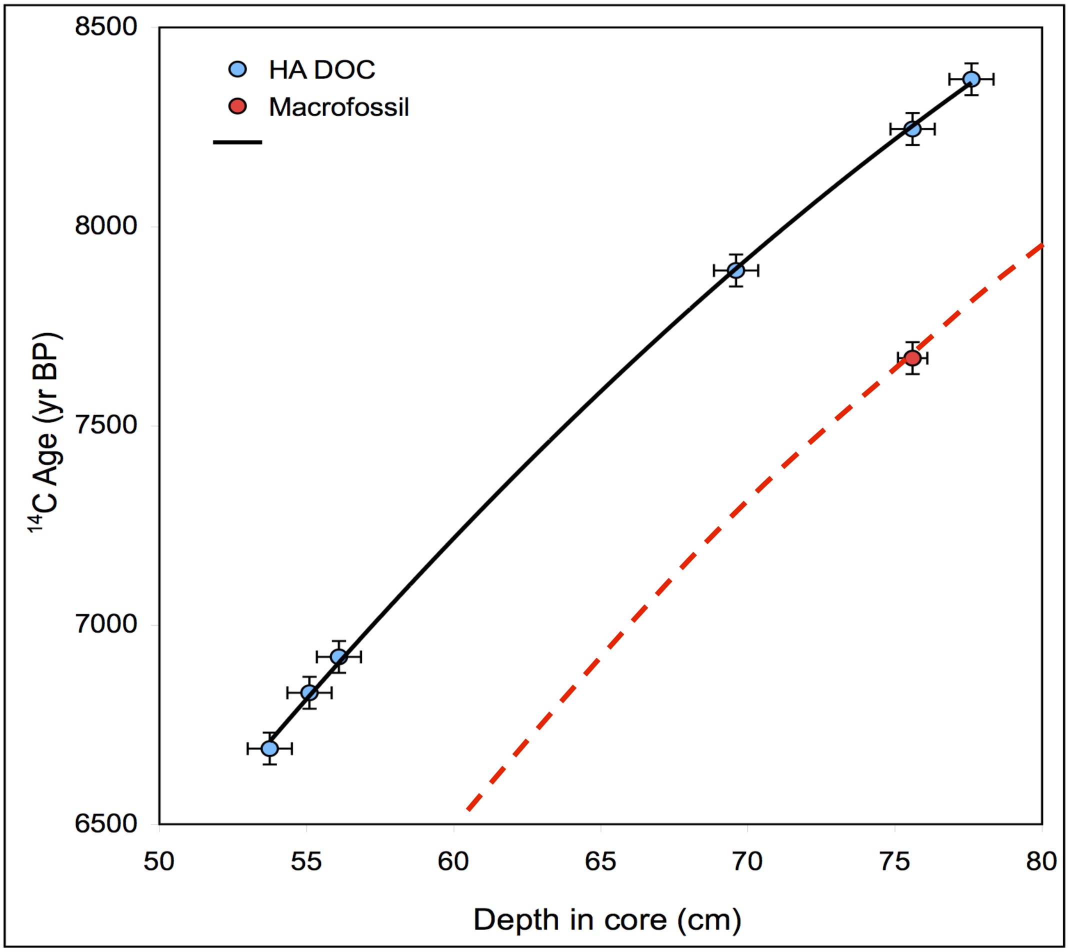

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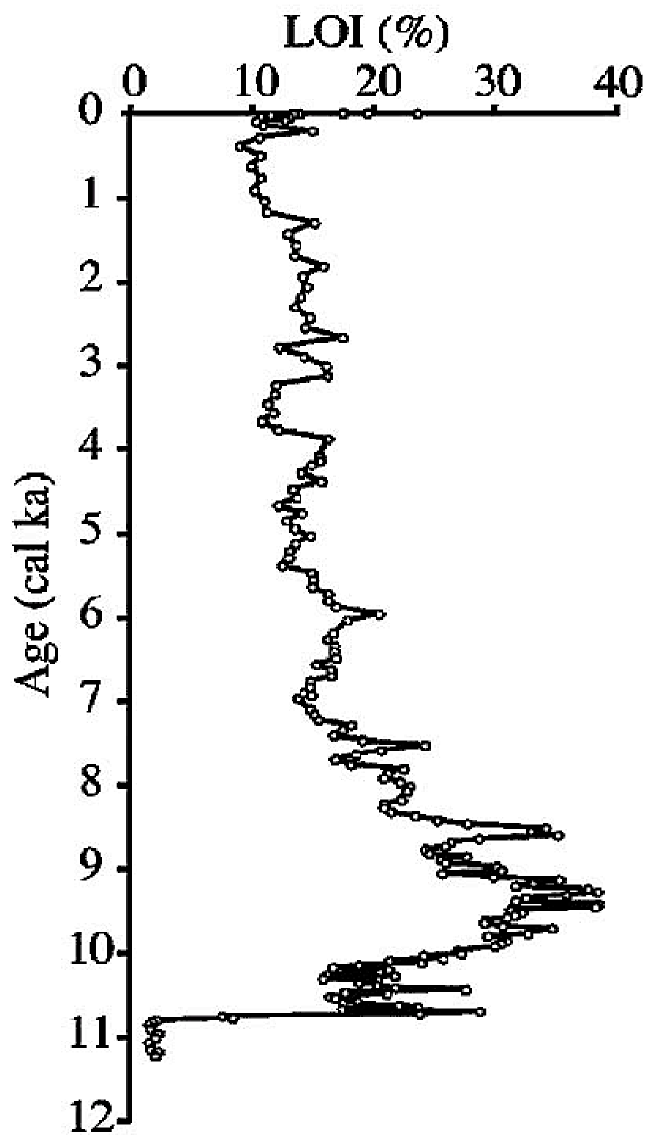

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Figure 6.4. A linescan image showing the NGRIP annual cloudy bands from 2528.35 to 2530.0 m depth. The age is about 72 ka BP corresponding to the Greenland Inter Stadial 19. The annual layer thickness is close to 1.5 cm. (Svensson et al., 2005)

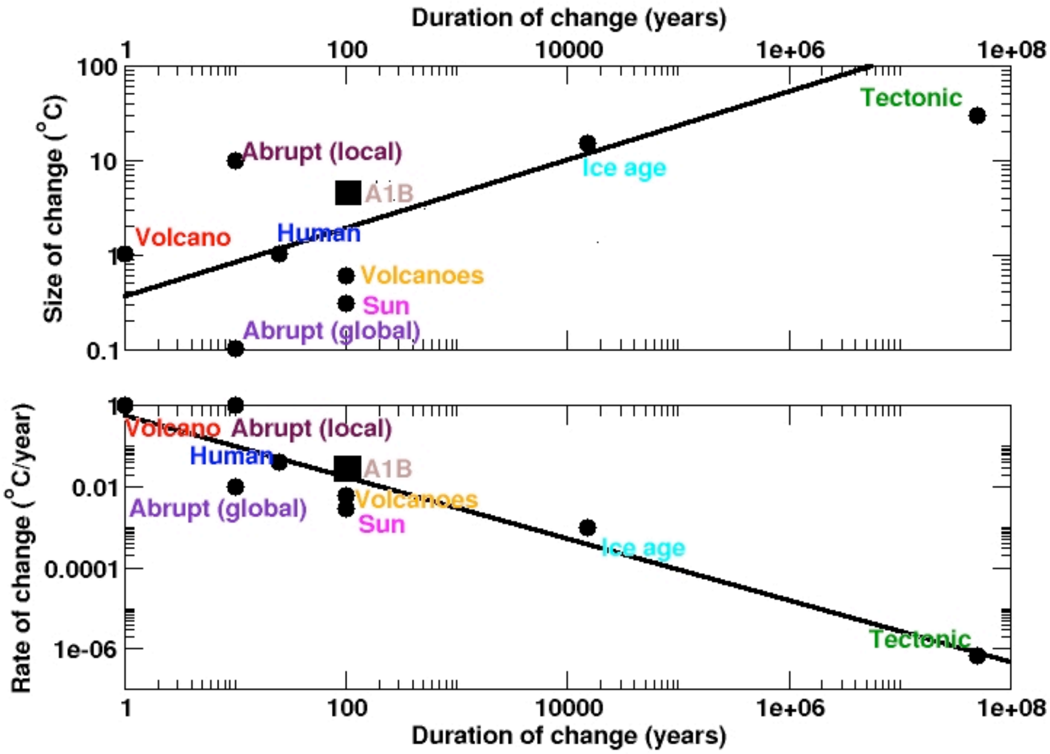

Figure 6.5. Summary of estimated peak rates of change, and sizes of changes, associated with various classes of causes. Error bars are not provided because of difficulty of quantifying, but high precision is not implied. Note that both panels have logarithmic scales on both axes (log-log plots) to allow the huge range of behavior to be shown in a single figure. The natural changes over the Little Ice Age-Medieval Warm Period 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; 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. Overall, the intent is that these numbers are representative of the Arctic, but Greenland ice-core data have been especially prominent in determinations, with the instrumental record included in assessing the human effects (see Delworth and Knutson, 2000 and Hegerl et al., 2007). One cannot exclude the possibility that the "human" contribution is overestimated and that natural fluctuations contributed to the late-20th-century change, but one also cannot exclude the possibility that the "human" contribution was larger than shown here with natural variability offsetting some of the change; the ability of climate models to explain widespread changes in climate based primarily on human forcing, and the evidence that there is little natural forcing over the latter 20th century (Hegerl et al., 2007), motivate plotting 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 Figure 10.7 of Meehl et al. (2007; also see Chapman and Walsh, 2007). This is shown as the square labeled A1B, with a different symbol chosen to show the fundamental difference of this scenario-based projection from data-based interpretations for the other results on the figure. Human changes clearly could be smaller or larger than shown as A1B, and may continue to possibly much larger values further in the future, with no guarantee that human disturbance will end before the end of the 21st century, as plotted here. The regression lines are through tectonic, ice-age, solar, volcano, and volcanoes, and are included solely to guide the eye and not to imply mechanisms.

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