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

Chapter 6 — Past Extent and Status of the Greenland Ice Sheet

Chapter Lead Author:

Richard B. Alley, Pennsylvania State University, University Park, PA

Contributing Authors:

John T. Andrews, University of Colorado, Boulder, CO

Garry K.C. Clarke, University of British Columbia, Vancouver, British Columbia, CA

Kurt M. Cuffey, University of California, Berkeley, CA

Svend Funder, University of Copenhagen, DK

Shawn J. Marshall, University of Calgary, Alberta, CA

Jerry X. Mitrovica, University of Toronto, Ontario, CA

Daniel R. Muhs, U.S. Geological Survey, Denver, CO

Bette Otto-Bliesner, National Center for Atmospheric Research, Boulder, CO

ABSTRACT

The *Greenland Ice Sheet* is expected to shrink or disappear with warming, a conclusion based on a survey of paleoclimatic and related information. Recent observations show that the *Greenland Ice Sheet* has melted more in years with warmer summers. Mass loss by melting is therefore expected to increase with warming. But whether the ice sheet shrinks or grows, and at what pace, depend also on snowfall and iceberg production. The Arctic is a complicated system. Reconstructions of past climate and ice sheet configuration (the “paleo-record”) are valuable sources of information that complement process-based models. The paleo-record shows that the *Greenland Ice Sheet* consistently lost mass when the climate warmed, and grew when the climate cooled. Such changes have occurred even at times of slow or zero sea-level change, so changing sea level cannot have been the cause of at least some of the ice sheet changes. In contrast, there are no documented major ice-sheet changes that occurred independent of temperature changes. Moreover, snowfall has increased when the climate warmed, but the ice sheet lost mass nonetheless; increased accumulation in the ice sheet’s center has not been sufficient to counteract increased melting and flow near the edges. Most documented forcings of change, and the changes to the ice sheet themselves, spanned periods of several thousand years, but limited data also show rapid response to rapid forcings. In particular, regions near the ice margin have responded within decades. However, major changes of central regions of the ice sheet are thought to require centuries to millennia. The paleo-record does not yet strongly constrain how rapidly a major shrinkage or nearly complete loss of the ice sheet could occur. The evidence suggests nearly total loss may result from warming of more than a few degrees above

mean 20th century values, but this threshold is poorly defined (perhaps as little as 2°C or more than 7°C). Paleoclimatic records are sufficiently sketchy that the ice sheet may have grown temporarily in response to warming, or changes may have been induced by factors other than temperature, without having been recorded.

6.1 The *Greenland Ice Sheet*

6.1.1. Overview

The *Greenland Ice Sheet* (Figure 6.1) contains by far the largest volume of any present-day Northern Hemisphere ice mass. The ice sheet is approximately 1.7 million square kilometers (km^2) in area, extending as much as 2200 km north to south. The maximum ice thickness is 3367 m, its average thickness is 1600 m (Thomas et al., 2001), and its volume is 2.9 million km^3 (Bamber et al., 2001). Some of the bedrock beneath this ice has been depressed below sea level by the weight of the ice, and a little of this bedrock would remain below sea level following removal of the ice and rebound of the bedrock (Bamber et al., 2001). However, most of the ice that rests on bedrock is above sea level and so would contribute to sea-level rise if it were melted: if the entire ice sheet melted, it is estimated that sea-level would rise about 7.3 m (Lemke et al., 2007).

FIGURE 6.1 NEAR HERE

The ice sheet consists primarily of old snow that has been squeezed to ice under the weight of new snow that accumulates every year. Snow accumulation on the upper surface tends to increase ice-sheet size. Ice sheets lose mass primarily by melting in low-elevation regions, and by forming icebergs that break off the ice margins (calving) and drift away to melt elsewhere. Sublimation, snowdrift (Box et al., 2006), and melting or freezing at the bed beneath the ice are minor terms in the budget, although melting beneath floating extensions called ice shelves before icebergs break off may be important (see 6.1.2, below).

Estimates of net snow accumulation on the *Greenland Ice Sheet* have been presented by Hanna et al. (2005) and Box et al. (2006), among others. Hanna et al. (2005) found for 1961–1990 (an interval of moderately stable conditions before more-recent warming) that surface snow accumulation (precipitation minus evaporation) was about 573 gigatons per year (Gt/yr) and that 280 Gt/yr of meltwater left the ice sheet. The difference of 293 Gt/yr is similar to the estimated iceberg calving flux within broad uncertainties (Reeh, 1985; Bigg, 1999; Reeh et al., 1999). (For reference, return of 360 Gt of ice to the ocean would raise global sea level by 1 millimeter (mm); Lemke et al., 2007.) More-recent trends are toward warming temperatures, increasing snowfall, and more rapidly increasing meltwater runoff (Hanna et al., 2005; Box et al., 2006). Large interannual variability causes the statistical significance of many of these trends to be relatively low, but the independent trends exhibit internal consistency (e.g., warming is expected to increase both melting and snowfall, on the basis of modeling experiments and simple physical arguments, and both trends are observed in independent studies (Hanna et al., 2005; Box et al., 2006)).

Increased iceberg calving has also been observed in response to faster flow of many outlet glaciers and shrinkage or loss of ice shelves (see 6.1.2, below, for discussion of the parts of an ice sheet) (e.g., Rignot and Kanagaratnam, 2006; Alley et al. 2005). The Intergovernmental Panel on Climate Change (IPCC; Lemke et al., 2007) found that "Assessment of the data and techniques suggests a mass balance of the *Greenland Ice Sheet* of between +25 and -60 Gt (-0.07 to 0.17 mm) SLE [sea level equivalent] per year from 1961-2003 and -50 to -100 Gt (0.14 to 0.28 mm SLE) per year from 1993-2003, with even larger losses in 2005". Updates are provided by Alley et al. (2007) (Figure 6.2) and by Cazenave (2006). Rapid changes have been occurring in the ice sheet, and in the ability to observe the ice sheet, so additional updates are virtually certain to be produced.

FIGURE 6.2 NEAR HERE

The long-term importance of these trends is uncertain—short-lived oscillation or harbinger of further shrinkage? This uncertainty motivates some of the interest in the history of the ice sheet.

6.1.2 Ice-sheet behavior

Where delivery of snow or ice (typically as snowfall) exceeds removal (typically by meltwater runoff), a pile of ice develops. Such a pile that notably deforms and flows is called a glacier, ice cap, or ice sheet. (For a more comprehensive overview, see Paterson, 1994; Hughes, 1998; Van der Veen, 1999; or Hooke, 2005, among well-known texts.) Use of these terms is often ambiguous. "Glacier" most typically refers to a relatively small mass in which flow is directed down one side of a mountain, whereas "ice cap" refers to a small mass with flow diverging from a central dome or ridge, and “ice sheet” to a very large ice cap of continental or subcontinental scale. A faster moving “jet” of ice flanked by slower flowing parts of an ice sheet or ice cap may be referred to as an ice stream, but also as an outlet glacier or simply glacier (especially if the configuration of the underlying bedrock is important in delineating the faster moving parts), complicating terminology. Thus, the prominent *Jakobshavn Glacier* (Jakobshavn Isbrae, or Jakobshavn ice stream) is part of the ice sheet on *Greenland*, flowing in a deep bedrock trough but with slower-moving ice flanking the faster-moving ice near the surface.

A glacier or ice sheet spreads under its own weight, deforming internally. The deformation rate increases with the cube of the driving stress, which is proportional to the ice thickness and to the surface slope of the ice. Ice may also move by sliding across the interface between the bottom of the ice and what lies beneath it, i.e., its substrate. Ice motion is typically slow or zero where the ice is frozen to the substrate, but is faster where the ice-substrate interface is close to the melting point. Ice motion can also take place through the deformation of subglacial sediments. This mechanism is important only where subglacial sediments are present and thawed. The contribution of these basal processes ranges from essentially zero to almost all of the total ice motion. Except for floating ice shelves (see below in this section), *Greenland*’s ice generally does not exhibit the gross dominance by basal processes seen in some West Antarctic ice streams.

Most glaciers and ice sheets tend toward a steady configuration. Snow accumulation in higher, colder regions supplies mass, which flows to lower, warmer regions where mass is lost by melting and runoff of the meltwater or by calving of icebergs that drift away to melt elsewhere.

Some ice masses tend to an oscillating condition, marked by ice buildup during a period of slow flow, and then a short-lived surge of rapid ice flow; however, under steady climatic conditions, these oscillations repeat with some regularity and without huge changes in the average size across cycles

Accelerations in ice flow, whether as part of a surging cycle, or in response to long-term ice-sheet evolution or climatically forced change, may occur through several mechanisms. These mechanisms include thawing of a formerly frozen bed, increase in meltwater reaching the bed causing increased lubrication (Zwally et al., 2002; Joughin et al., 1996; Parizek and Alley, 2004), and changes in meltwater drainage causing retention of water at the base of the glacier, which increases lubrication (Kamb et al., 1985). Ice-flow slowdown can similarly be induced by reversing these causes.

Recently, attention has been focused on changes in ice shelves. Where ice flows into a bordering water body, icebergs may calve from grounded (non-floating) ice. Alternatively, the flowing ice may remain attached to the glacier or ice sheet as it flows into the ice-marginal body of water. The attached ice floats on the water and calves from the end of the floating extension, which is called an ice shelf. Ice shelves frequently run aground on local high spots in the bed of the water body on which they float. Ice shelves that occupy embayments or fjords may rub against the rocky or icy sides, and friction from this restrains, or “buttresses,” ice flow. Loss of this buttressing through shrinkage or loss of an ice shelf then allows faster flow of the ice feeding the ice shelf (Payne et al., 2004; Dupont and Alley, 2005; 2006).

Although numerous scientific papers have addressed the effects of changing lubrication or loss of ice-shelf buttressing affecting ice flow, comprehensive ice-flow models generally have not incorporated these processes. These comprehensive models also failed to accurately project recent ice-flow accelerations in *Greenland* and in some parts of the Antarctic ice sheet (Alley et al., 2005; Lemke et al., 2007; Bamber et al., 2007). This issue was cited by IPCC (2007), which provided sea-level projections “excluding future rapid dynamical changes in ice flow” (Table SPM3, WG1) and noted that this exclusion prevented “a best estimate or an upper bound for sea level rise” (p. SPM 15). A paleoclimatic perspective can help inform our understanding of these issues.

As noted above in this section, when subjected to a **step forcing** (e.g., a rapid warming that moves temperatures from one sustained level to another), an ice sheet typically responds by evolving to a new steady state (Paterson, 1994). For example, an increase in accumulation rate thickens the ice sheet. The thicker ice sheet discharges mass faster and, if the ice margin does not move as the ice sheet thickens, the ice sheet becomes on average steeper, which also speeds ice discharge. These changes eventually cause the ice sheet to approach a new configuration—a new steady state—that is in balance with the new forcing. For central regions of cold ice sheets, the time required to complete most of the response to a step change in rate of accumulation (i.e., the response time) is proportional to the ice thickness divided by the accumulation rate. These characteristic times are a few thousands of years (millennia) for the modern *Greenland Ice Sheet* and a few times longer for the ice-age ice sheet (e.g., Alley and Whillans, 1984; Cuffey and Clow, 1997).

A change in the position of the ice-margin will steepen or flatten the mean slope of the ice sheet, speeding or slowing flow. The edge of the ice-sheet will respond first. This response, in turn, will cause a wave of adjustment that propagates toward the ice-sheet center. Fast-flowing marginal regions can be affected within years, whereas the full response of the slow-flowing central regions to a step-change at the coast requires a few millennia.

Warmer ice deforms more rapidly than colder ice. In inland regions, ice sheet response to temperature change is somewhat similar to response to accumulation-rate change, with cooling causing slower deformation, which favors thickening hence higher ice flux through the increased thickness (and perhaps with increasing surface slope also speeding flow), re-establishing equilibrium. However, because most of the deformation occurs in deep ice, and a surface-temperature change requires many millennia to penetrate to that deep ice to affect deformation, most of the response is delayed for a few millennia or longer while the temperature change penetrates to the deep layers, and then the response requires a few more millennia. The calculation is not simple, because the motion of the ice carries its temperature along with it. If melting of the upper surface of an ice sheet develops over a region in which the bottom of the ice is frozen to the substrate, thawing of that basal interface may be caused by penetration of surface meltwater to the bed if water-filled crevasses develop at the surface. The actual penetration of the water-filled crevasse is likely to occur in much less than a single year, perhaps in only a few minutes, rather than over centuries to millennia (Alley et al., 2005).

Numerous ice-sheet models (e.g., Huybrechts, 2002) demonstrate the relative insensitivity of inland ice thickness to many environmental parameters. This insensitivity has allowed reasonably accurate ice-sheet reconstructions using computational models that assume perfectly plastic ice behavior and a fixed yield strength (Reeh 1984; the only piece of information needed in these reconstructions of inland-ice configuration is the footprint of the ice sheet; one need not specify accumulation rate hence mass flux, for example). This insensitivity can be understood from basic physics.

As noted above in this section, the stress that drives ice deformation increases linearly with ice thickness and with the surface slope, and the rate of ice deformation increases with the cube of this stress. Velocity from deformation is obtained by integrating the deformation rate through thickness, and ice flux is the depth-averaged velocity multiplied by thickness. Therefore, for ice frozen to the bed, the ice flux increases with the cube of the surface slope and the fifth power of the thickness. (Ice flux in an ice sheet with a thawed bed would retain strong dependence on surface slope and thickness, but with different numerical values.) If the ice-marginal position is fixed (say, because the ice has advanced to the edge of the continental shelf and cannot advance farther across the very deep water), then the typical surface slope of the ice sheet is also proportional to the ice thickness (divided by the fixed half-width), giving an eighth-power dependence of ice flux on inland thickness. Although an eighth-power dependence is not truly perfectly plastic, it does serve to greatly limit inland-thickness changes—doubling the inland thickness would increase ice flux 256-fold. Because of this insensitivity of the inland thickness to many controlling parameters, changes in ice-sheet volume are controlled more by changes in the areal extent of the ice sheet than by changes in the thickness in central regions (Reeh, 1984; Paterson, 1994).

Such simple mechanistic scalings of ice sheet behaviors can be useful in a pragmatic sense, and they have been used to interpret ice-sheet behavior in the past. However, in modern usage, our physical understanding of ice sheet behaviors is implemented in fully coupled three-dimensional (or reduced-dimensional) ice-dynamical models (e.g., Huybrechts, 2002; Parizek and Alley, 2004; Clarke et al., 2005), which help researchers assimilate and understand relevant data.

6.2 Paleoclimatic Indicators Bearing on Ice-Sheet History

The basis for paleoclimatic reconstruction is discussed in Cronin (1999) and Bradley (1999), among other sources. Here, additional attention is focused on those indicators that help in reconstruction of the history of the ice sheet. Marine indicators are discussed first, followed by terrestrial archives.

6.2.1 Marine Indicators

As discussed in section 6.3 below, the *Greenland Ice Sheet* has at many times in the past been more extensive than it is now, and much of that extension occupied regions that now are below sea level. Furthermore, iceberg-rafted debris and meltwater from the ice sheet can leave records in marine settings related to the extent of the ice sheet and its flux of ice. Marine sediments also preserve important indicators of temperature and of other conditions that may have affected the ice sheet.

Research cruises to the marine shelf and slope margins of west and east *Greenland* dedicated to understanding changes over the times most relevant to the *Greenland Ice Sheet's* history have been undertaken only in the last ten to twenty years. Initially, attention was focused along the *east Greenland shelf* (Marienfeld, 1992b; Mienert et al., 1992; Dowdeswell et al., 1994a), but in the last few years several cruises have extended to the west *Greenland* margin as well (Lloyd, 2006; Moros et al., 2006). Research on adjacent deep-sea basins, such as *Baffin Bay* or *Fram Basin* off North

Greenland, is more complicated because the late Quaternary (less than 450 thousand years old (ka)) sediments contain inputs from several adjacent ice sheets (Dyke et al., 2002; Aksu, 1985; Andrews et al., 1998a: Hiscott et al., 1989). (We use calendar years rather than radiocarbon years unless indicated; conversions include those of Stuiver et al., 1998 and Fairbanks et al., 2005; all ages specified as “ka” or “Ma” are in years before present, where “present” is conventionally taken as the year 1950.) Regardless, only a few geographic areas on the *Greenland* shelf have been investigated. In terms of time, the majority of marine cores from the *Greenland* shelf span the retreat from the last ice age (less than 15 ka). The use of datable volcanic ashes (tephras—a recognizable tephra or ash layer from a single eruption is commonly found throughout broad regions and has the same age in all cores) from Icelandic sources offers the possibility of linking records from around *Greenland* from the time of the layer known as Ash Zone II (about 54 ka) to the present (with appropriate cautions; Jennings et al., 2002a).

The sea-floor around *Greenland* is relatively shallow above “sills” formed during the rifting that opened the modern oceans. Such sills connect *Greenland* to Iceland through *Denmark Strait* and to *Baffin Island* through *Davis Strait*. These 500–600-m-deep sills separate sedimentary records of ice sheet histories into “northern” and “southern” components. Even farther north, sediments shed from north *Greenland* are transported especially into the Fram Basin of the Arctic Ocean (Darby et al., 2002).

The circulation of the ocean around *Greenland* today transports debris-bearing icebergs from the ice sheet. This circulation occurs largely in a clockwise pattern: cold, fresh waters exit the Arctic Ocean through Fram Strait and flow southward along the East Greenland margin as the East Greenland Current (Hopkins, 1991). These waters turn north after rounding the southern tip of *Greenland*. In the vicinity of *Denmark Strait*, warmer water from the Atlantic (modified Atlantic Water from the Irminger Current) turns and flows parallel to the East Greenland Current. This surface current is called the West Greenland Current once it has rounded the southern tip of *Greenland*. On the *East Greenland shelf*, this modified Atlantic Water becomes an “intermediate-depth” water mass (reaching to the deeper parts of the continental shelf, but not to the depths of the ocean beyond the continental shelf), which moves along the deeper topographic troughs on the continental shelf and penetrates into the margins of the calving *Kangerdlugssuaq* ice stream (Jennings and Weiner, 1994; Syvitski et al., 1996). Baffin Bay contains three water masses: Arctic Water in the upper 100–300 meters (m) in all areas, West Greenland Intermediate Water (modified Atlantic Water) between 300–800 m, and Deep Baffin Bay Water throughout the Bay at depths greater than 1200 m (Tang et al., 2004).

Some of the interest in the *Greenland Ice Sheet* is linked to the possibility that meltwater could greatly influence the formation of deep water in the North Atlantic. Furthermore, changes in deep-water formation in the past are linked to climate changes that affected the ice sheet (e.g., Alley, 2007). The major deep-water flow is directed southward through and south of *Denmark Strait* (McCave and Tucholke, 1986). The sediment deposit known as the *Eirik Drift* off southwest *Greenland* is a product of this flow (Stoner et al., 1995). Convection in the *Labrador Sea* forms an upper component of this North Atlantic Deep Water.

Evidence from marine cores and seismic data has been used to reconstruct variations in the *Greenland Ice Sheet* during the last glacial cycle (and, occasionally, into older times). Four types of evidence apply: (1) ice-rafted debris and indications of changes in sediment sources; (2) glacial deposition onto trough-mouth fans; (3) stable-isotope and biotic data that indicate intervals when meltwater was released from the ice sheet; and (4) geophysical data that indicate sea-floor erosion and deposition. Each is discussed briefly in section 6.2.1, below.

6.2.1a Ice-rafted debris and its provenance

Coarse-grained rock material (such as sand and pebbles) cannot be carried far from a continent by wind or current, so the presence of such material in marine cores is of great interest. Small amounts might be delivered in tree roots or attached to uprooted kelp holdfasts (Gilbert, 1990; Smith and Bayliss-Smith, 1998), and rarely a meteorite might be identified, but large quantities of coarse rock material found far from land indicate transport in ice, and so this material is called ice-rafted debris (IRD). Both sea ice and icebergs can carry coarse material, complicating interpretations. However, iceberg-rafted debris usually includes some number of grains larger than 2 mm in size and consistent with the grain-size distribution of glacially transported materials, whereas the sediment entrained in sea ice is typically finer (Lisitzin, 2002). In order to link the *Greenland Ice Sheet* with ice-rafted debris described in marine cores, we must be able to link that debris to specific bedrock sites (i.e., identify its provenance or site of origin). However, such studies are only in their infancy. Proxies for sediment source include radiogenic isotopes (such as εNd; Grousset et al., 2001; Farmer et al., 2003), biomarkers that can be linked to different outcrops of dolomite (Parnell et al., 2007), magnetic properties of sediment (Stoner et al., 1995), and quantitative mineralogical assessment of sediment composition (Andrews, 2008).

6.2.1b Trough mouth fans

The sediments in trough-mouth fans contain histories of sediment sources that may include ice sheets. Sediment is commonly transferred across the continental shelf along large troughs that form major depositional features called trough-mouth fans (TMF) where the troughs widen and flatten at the continental rise (Vorren and Laberg, 1997; O'Cofaigh et al., 2003). Along the East Greenland margin, trough-mouth fans exist off *Scoresby Sund* (Dowdeswell et al., 1997), the Kangerdlugssuaq Trough (Stein, 1996), and the Angamassalik Trough (St. John and Krissek, 2002). Along the west *Greenland* margin, the most conspicuous such fan is a massive body off *Disko Bay* associated with erosion by *Jakobshavn Glacier* and other outlet glaciers in that region. During periods when the ice sheet reached the shelf break, glacial sediments were shed downslope as debris flows (producing coarse, poorly sorted deposits containing large grains in a fine-grained matrix), whereas periods when the ice sheet was well back from the shelf break are marked by sediments containing materials typical of open-marine environments, such as shells of foraminifers, and typical terrestrial materials including ice-rafted debris.

6.2.1c Foraminifers and stable-isotopic ratios of shells

Foraminifers—mostly marine, single-celled planktonic animals, commonly with chalky shells—are widely distributed in sediments, and shells of surface-dwelling (planktic) and bottom-dwelling (benthic) species are commonly found. The particular species present and the chemical and isotopic characteristics of the chalky shells reflect environmental conditions. Variations in the ratios of the stable isotopes of oxygen, ^{18}O to

^{16}O (**δ^{18}O**) are especially widely used. These ratios respond to changes in the global ice volume. Water containing the lighter isotope (^{16}O) evaporates from the ocean more readily, and ice sheets are ultimately composed of that evaporated water, so during times when the ice sheets are larger, the ocean is isotopically heavier. This effect is well known, and it can be corrected for with considerable confidence if the age of a sample is known. Temperature also affects δ^{18}O; warmer air temperatures favor incorporation of the lighter isotope into the shell. Near ice sheets, the abrupt appearance of light isotopes is most commonly associated with meltwater that delivered isotopically light and fresh water (Jones and Keigwin, 1988; Andrews et al., 1994). Around the *Greenland Ice Sheet*, most such records are from near-surface planktic foraminifers of the species *N. pachyderma* sinistral (Fillon and Duplessy, 1980; van Kreveld et al., 2000; Hagen and Hald, 2002), although there are some data from benthic foraminifers (Andrews et al., 1998a; Jennings et al., 2006).

6.2.1d Seismic and geophysical data

Several major shelf troughs and trough-mouth fans have been studied by seismic investigations. Most are high-resolution studies of the sediments nearest the sea floor (seismostratigraphy; O'Cofaigh et al., 2003), although some data on deeper strata are available (airgun profiles; Stein, 1996; Wilken and Mienert, 2006). Sonar reveals the shape of the upper surface of the sediment, and features such as the tracks left by drifting icebergs that plowed through the sediment (Dowdeswell et al., 1994b; Dowdeswell et al., 1996; Syvitski et al., 2001) and the streamlining of the sediment surface caused by glaciation.

6.2.2 Terrestrial Indicators

Land-based records, like their marine equivalents, can reveal the history of changes in areal extent of ice and of the climate conditions that existed around the ice sheet. Terrestrial records are typically more discontinuous in space and time than are marine records, because net erosion (which removes sediments containing climatic records) is dominant on land whereas net deposition is dominant in most marine settings. Nonetheless, useful records of many time intervals have been assembled from terrestrial indicators. Here, common indicators are briefly described. This treatment is representative rather than comprehensive. Furthermore, the great wealth of indicators, and the interwoven nature of their interpretation, preclude any simple subdivision.

6.2.2a Geomorphic indicators

The land surface itself records the action of ice and thus provides information on ice-sheet history. Glacial deposits known as moraines are especially instructive, but others are also important.

Moraines are composed of sediment deposited around glaciers from material carried on, in, or under the moving ice (e.g., Sugden and John, 1976). A preserved moraine may mark either the maximum extent reached by ice during some advance or a still-stand during retreat. Normally, older moraines are destroyed by ice readvance, although remnants of moraines overrun by a subsequent advance are occasionally preserved and identifiable, especially if the ice that readvanced was frozen to its bed and thus nearly or completely stationary where the ice met the moraine. Because most older moraines are reworked by subsequent advances, most existing moraines record only the time of the most recent glacial maximum and pauses or subsidiary readvances during retreat.

The limiting ages of moraines can be estimated from radiocarbon (carbon-14) dating of carbon-bearing materials incorporated into a moraine (the moraine must be younger than those materials) or deposited in lakes that formed on or behind moraines following ice retreat (the moraine must be older than those materials). Increasingly, moraines are dated by measurement of beryllium-10 or other isotopes produced in boulders by cosmic rays (e.g., Gosse and Phillips, 2001). Cosmic rays penetrate only about 1 m in rock. Thus, boulders that are quarried from beneath the ice following erosion of about 1 m or more of overlying material, or large boulders that fell onto the ice and rolled over during transport, typically start with no cosmogenic nuclides in their upper surfaces but accumulate those nuclides proportional to exposure time. Corrections for loss of nuclides by boulder erosion, for inheritance of nuclides from before deposition, and other factors may be nontrivial but potentially reveal further information. Additional techniques of dating can sometimes be used, including historical records and the increase with time of the size of lichen colonies (e.g., Locke et al., 1979; Geirsdottir et al., 2000), soil development, and breakdown of rocks (clast weathering).

Related information on glacial behavior and ages is also available from the land surface. For ages of events, a boulder need not be in a moraine to be dated using cosmogenic isotopes, and surfaces striated and polished by glacial action can be dated similarly. Glacial retreat often reveals wood or other organic material that died when it was overrun during an advance and that can also be dated using radiocarbon techniques.

In moraines produced by small glaciers, the highest elevation to which a moraine extends is commonly close to the equilibrium-line altitude at the time when the moraine formed. (The equilibrium-line altitude is the altitude above which net snow accumulation occurred and below which mass loss occurred—mass moved into the glacier above that elevation and out below that elevation, controlling the deposition of rock material.) Glaciation produces identifiable landforms, especially if the ice was thawed at the base and thus slid freely across its substrate, so contrasts in the appearance of landforms can be used to map the limits of glaciation (or of wet-based glaciation) where moraines are not available.

Glaciers respond to many environmental factors, but for most glaciers the balance between snow accumulation and melting is the major control on glacier size. Furthermore, with notable exceptions, melting is usually affected more by temperature than is accumulation. The equilibrium vapor pressure (the ability of warmer air to hold more moisture) increases roughly 7% per °C. For a variety of glaciers that balance snow accumulation by melting, the increase in melting is approximately 35% (±10%) per °C (e.g., Oerlemans, 1994; 2001; Denton et al., 2005). Thus, glacier extent can usually be used as a proxy for temperature (duration and warmth of the melt-season), primarily summertime temperature.

6.2.2b Biological indicators and related features

Living things are sensitive to climate. The species found in a tropical rain forest differ from those found on the tundra. By comparing modern species from different places that have different climates, or by looking at changes in species at one place for the short interval of the instrumental record, the relation with climate can be estimated. Assuming that this relation has not changed with time, longer records of climate then can be estimated from occurrence of different species in older sediments (e.g., Schofield et al., 2007). These climate records then can be tied, to some degree, to the state of the ice sheet.

Lake sediments are especially valuable as sources of biotic indicators, because sedimentation (and thus the record) is continuous and the ecosystems in and around lakes tend to be rich (e.g., Bjorck et al., 2002; Ljung and Bjorck, 2004; Andresen et al., 2004). Pollen (e.g., Ljung and Bjorck, 2004; Schofield et al., 2007), microfossils, and macrofossils (such as chironomids, also called midge flies (Brodersen and Bennike, 2003)) are all used to great advantage in reconstructing past climates. The isotopic composition of shells or of inorganic precipitates in lakes records some combination of temperature and of the isotopic composition of the water. Physical aspects of lake sediments, including those linked to biological processes (e.g., loss on ignition, which primarily measures the relative abundance of organic matter in the sediment) are also related to climate. In places where the weight of the ice previously depressed the land below sea level and subsequent rebound raised the land back above sea level and formed lakes (see 6.2.2c, below), the time of onset of lacustrine conditions and the modern height of the lake together provide key information on ice-sheet history (e.g., Bennike et al., 2002).

Raised marine deposits in *Greenland* and surroundings provide an additional and important source of biological indicators of climate change. Many marine deposits now reside above sea level, because of the interplay of changing sea level, geological processes of uplift and subsidence, and isostatic response (ice-sheet growth depressing the land and subsequent ice-sheet shrinkage allowing rebound, with a lagged response; see 6.2.2c, below). Biological materials within those deposits, and especially shells, can be dated by radiocarbon or uranium-thorium techniques (see 6.2.2d, below). Those dates then help fill in the history of relative sea level that can be used to infer ice-sheet loading histories and to reconstruct climates on the basis of the species present (e.g., Dyke et al., 1996).

6.2.2c Glacial isostatic adjustment and relative sea-level indicators near the ice sheet

Within the geological literature, sea level is generally defined as the elevation of the sea surface relative to some adjacent geological feature. (This convention contrasts with the concept of an absolute sea level whose position (the sea surface) is measured relative to some absolute datum, such as the center of Earth.) This definition of sea level is consistent with geological markers of past sea-level change (such as ancient shorelines, shells, and driftwood), which reflect changes in the absolute height of either the sea surface or the geological feature (i.e., an ancient shoreline can be exposed because the surface of the ocean dropped, or land uplifted, or a net combination of land and ocean height changes). During the time periods considered in this report, the dominant processes responsible for such changes, at least on a global scale, have been the mass transfer between ice reservoirs and oceans associated with the ice-age cycles, and the deformational response of Earth to this transfer of mass. This deformational response is formally termed **glacial isostatic adjustment**.

The growth and shrinkage of ice have generally been sufficiently slow that glacial isostatic adjustment of the solid Earth is characterized by both immediate **elastic** and slow viscous (i.e., flow) effects. As an example, if a large ice sheet were to form instantly and then persist for more than a few thousand years, the land would respond by nearly instantaneous elastic sinking, followed by slow subsidence toward isostatic equilibrium as deep, hot rock moved outward from beneath the ice sheet. Roughly speaking, the final depression would be about 30% of the thickness of the ice. Thus the ancient *Laurentide Ice Sheet*, which covered most of Canada and the northeastern United States and whose peak thickness was 3–4 km, produced a crustal depression of about 1 km. (For comparison, that ice sheet contained enough water to make a layer about 70 m thick across the world oceans, much less than the local deformation beneath the ice.) Outside the depressed region covered by ice, land is gradually pushed upward to form a peripheral bulge. As the ice subsequently melts, the central region of depression rebounds, and relative sea level will fall for thousands of years beyond the end of the melting phase. For example, at sites in Hudson Bay, sea-level continues to fall on the order of 1 centimeter per year (cm/yr) despite the disappearance of most of the *Laurentide Ice Sheet* some 8000 years ago. Moreover, the loss of ice cover allows the peripheral bulge to subside, leading to a sea-level rise in such areas (e.g., along the east coast of the United States) that also continues to the present (but involving slower rates of change than for the regions that were beneath the central part of the former ice sheet). As one considers sites farther away from the high-latitude ice cover, in the so-called “far field,” the sea-level change is dominated during deglaciation by the addition of meltwater into the global oceans. However, in periods of stable ice cover, for example during much of the present interglacial, changes in sea level continue as a consequence of the ongoing gravitational and deformational effects of glacial isostatic adjustment. As an example, glacial isostatic adjustment is responsible for a fall in sea level in parts of the equatorial Pacific of about 3 m during the last 5,000 years and for the associated exposure of corals and ancient shoreline features of this age (Mitrovica and Peltier, 1991; Mitrovica and Milne, 2002; Dickinson, 2001). We will return to this point in section 6.2.2d, below.

Nearby (near-field) relative sea-level changes, where the term "relative" denotes the height of an ancient marker relative to the present-day level of the sea, have commonly been used to constrain models of the geometry of ice complexes, particularly since the Last Glacial Maximum (about 24 ka) (e.g., Lambeck et al., 1998; Peltier, 2004). Fleming and Lambeck (2004) compared a set of about 600 relative sea-level data points from sites in *Greenland*; all but the southeast coast and the west coast near *Melville Bugt* (Bay) were represented. Numerical models of glacial isotatic adjustment constrained the history of the *Greenland Ice Sheet* after the Last Glacial Maximum. The Fleming and Lambeck (2004) data set comprised primarily fossil mollusk shells that lived at or below the sea surface but that now are exposed above sea level; because of the unknown depth at which the mollusks lived, they provide a limiting value on sea level. However, Fleming and Lambeck (2004) also included observations on the transition of modern lakes from formerly marine conditions, and constraints associated with the present (sub-sea) location of initially terrestrial archaeological sites (see also Weidick, 1996; Kuijpers et al., 1999). Tarasov and Peltier (2002, 2003) analyzed their own compilation of local sea-level records by coupling glacial isostatic adjustment and climatological models; from this information they inferred ice history into the last interglacial.

Like all glacial isostatic adjustment models, these studies are hampered by uncertainty about the viscoelastic structure of Earth (Mitrovica, 1996), which is generally prescribed by the thickness of the elastic plate and the radial profile of viscosity within the underlying mantle, and this uncertainty has implications for the robustness of the inferred ice history. In addition, the analysis of sea-level records in *Greenland* is complicated by signals from at least two other distant sources: (1) the adjustment of the peripheral bulge associated with the (de)glaciation of the larger North American *Laurentide Ice Sheet*, because this bulge extends into *Greenland* (e.g., Fleming and Lambeck, 2004); and (2) the net addition of meltwater from contemporaneous melting (or, in times of glaciation, growth) of all other global ice reservoirs. Therefore, some constraints on the volume and extent of the *Laurentide Ice Sheet*, and the volume of more-distant ice sheets and glaciers, are required for the analysis of sea-level data from *Greenland*.

6.2.2d Far-field indicators of relative sea-level high-stands

Past changes in the volume of the *Greenland Ice Sheet* are recorded in far-field sea level. All other sources of sea-level change, as well as the change due to glacial isostatic adjustment, are also recorded in far-field sea-level records, so a single history of sea level provides information related to ice-volume change (and to other factors such as thermal expansion and contraction of ocean water) but no information on the relative contribution of individual sources.

The record of past sea level can be reconstructed in many ways. An especially powerful method of reconstruction uses the record of marine deposits or emergent coral reefs that are now found above sea level on geologically relatively stable coasts and islands (that is, in regions not markedly affected by processes linked to plate tectonics). Such records are literally high-water marks (or “bathtub rings”) of past high sea levels. Coastal landforms and deposits provide powerful and independent records of sea-level history compared with the often-cited deep-sea oxygen-isotope record of glacial and interglacial periods. For recording sea-level history, coastal landforms have two advantages as compared with the deep-sea oxygen-isotope record: (1) if corals are present, they can be dated directly; and (2) estimates of ancient sea level may—depending on the geological setting—be possible.

Coastal landforms record high stands of the sea when coral-reefs grew as fast as sea level rose (upper panel in Figure 6.3) or when a stable sea-level high stand eroded marine terraces into bedrock (lower panel in Figure 6.3). Thus, emergent marine deposits, either reefs or terraces, on geologically active, rising coastlines record interglacial periods (Figure 6.4). On a geologically stable or slowly sinking coast, reefs will emerge only from sea-level stands that were higher than at present (Figure 6.4). Past sea levels can thus be determined from stable coastlines, or even rising coastlines if one can make reasoned models of uplift rates. Geologic records of high sea-level stands on geologically relatively stable coasts are especially useful. Although valuable geologic records are found on rising coasts, estimates of past sea level derived from such coasts depend on assumptions about the rate of tectonic uplift, and therefore they embody more uncertainty.

FIGURE 6.3 NEAR HERE

FIGURE 6.4 NEAR HERE

The direct dating of emergent marine deposits is possible because uranium (U) is dissolved in ocean water but thorium (Th) and protactinium (Pa) are not. Certain marine organisms, particularly corals, co-precipitate U directly from seawater during growth. All three of the naturally occurring isotopes of uranium—^{238}U and ^{235}U (both primordial parents) and ^{234}U (a decay product of ^{238}U)—are therefore incorporated into living corals. ^{238}U decays to ^{234}U, which in turn decays to ^{230}Th. The parent isotope ^{235}U decays to ^{231}Pa. Thus, activity ratios of ^{230}Th/^{234}U, ^{238}U/^{234}U, and ^{231}Pa/^{235}U can provide three independent clocks for dating the same fossil coral (e.g., Edwards et al., 1997). Since the 1980s, most workers have employed thermal ionization mass spectrometry (TIMS) to measure U-series nuclides; this method has increased precision, requires much smaller samples, and can extend the useful time period for dating back to at least about 500,000 years.

The coastlines where the most reliable records of past high sea levels can be found are in the tropics and subtropics, where ocean temperatures are warm enough that coral-reefs grow. Within this broad equatorial region, the ideal coastlines for studies of past high sea levels are those that are distant from boundaries of tectonic plates. Such coastlines lie near geologically relatively quiescent continental margins or as islands well within the interiors of large tectonic plates. Even in such locations, however, interpreting past sea levels can include much uncertainty. We highlight two major reasons for this uncertainty.

First, many islands well within the crustal tectonic plate that underlies the Pacific

Ocean, for example, are part of hot-spot volcanic chains. (A major source of internal heat, called a hot spot, leads to a volcano on the overriding tectonic plate; as the plate drifts laterally, the slower-moving hot spot becomes positioned below a different part of the plate, and a new volcano is formed as the previously active volcano becomes extinct. Eventually, a chain of volcanoes is produced, such as the Hawaiian-Emperor seamount chain.) As a volcano grows in elevation, its weight isostatically depresses the land it sits on in the same way that the weight of an ice sheet does, and the cold upper elastic layer of the Earth flexes to form a broad ring-shaped ridge around the low caused by the volcano. Oahu, in the Hawaiian Island chain, is a good example of an island that is apparently experiencing slow uplift, and an associated local sea-level fall, due to volcanic loading on the “Big Island” of Hawaii (Muhs and Szabo, 1994).

Second, the existence of a sea-level highstand of a given age in a stable geologic setting does not necessarily imply that ice volumes were lower at that time relative to the present day, even if the highstand is dated to a previous interglacial. As discussed above, glacial isostatic adjustment, because it involves slow viscous flow of rock, produces global-scale changes in sea-level even during periods when ice volumes are stable. As an example, for the last 5,000 years (long after the end of the last glacial interval), ocean water has moved away from the equatorial regions and toward the former Pleistocene ice complexes to fill the voids left by the subsidence of the peripheral bulge regions produced by the ice sheets. As a result, sea level has fallen (and continues to fall) about 0.5 mm/yr in those far-field equatorial regions (Mitrovica and Peltier, 1991; Mitrovica and Milne, 2002). This process, known as equatorial ocean siphoning, has developed so-called 3-meter beaches and exposed coral reefs that have been dated to the end of the last deglaciation and that are endemic to the equatorial Pacific (e.g., Dickinson, 2001). Thus, the interpretation of such apparent highstands requires correction for glacial isostatic adjustments such that the residual record reflects true changes in ice volume.

6.2.2e Geodetic indicators

Geodetic data are yielding both local and regional constraints on recent changes in the mass of ice-sheets. As an example, land-based measurements of changes in gravity and crustal motions, estimated by using the global positioning system (GPS), are being used to monitor deformation (associated with changes in the distribution of mass) at the periphery of the *Greenland Ice Sheet* (e.g., Kahn et al., 2007). A drawback of these techniques is that few sites have been monitored because of the difficulty of establishing high-quality GPS sites. In contrast, data from the Gravity Recovery and Climate Experiment (GRACE) satellite mission are revealing trends in gravity across the polar ice sheets (at a spatial resolution of about 400 km) from which estimates of both regional and integrated mass flux are being obtained (e.g., Velicogna and Wahr, 2006). A general problem in all attempts to infer recent ice sheet balance, whether from land-based or satellite gravity, GPS, or even altimeter measurements of ice height (e.g., Johannessen et al., 2005; Thomas et al., 2006), is that a measurements must be corrected for the continuing influence of glacial isostatic adjustments. As discussed above (section 6.2.2c), this correction involves uncertainty associated with both the ice sheet history and the viscoelastic structure of Earth.

Accurate glacial isostatic adjustment corrections are also central to regional estimates of ice-sheet mass balance. For the last century global sea-level change has been inferred principally by analyzing records from widely distributed tide gauges (simple sea-level monitoring devices). Most residual rates (those corrected for glacial isostatic adjustment) of tide gauges yield an average 20th century sea-level rise in the range 1.5–2.0 mm/yr (Douglas, 1997) (for additional information on recent trends in sea level, see Solomon et al., 2007).

Furthermore, geographic trends in the residual rates may constrain the sources of the meltwater. In particular, Mitrovica et al. (2001) and Plag and Juttner (2001) have demonstrated that the rapid melting of different ice sheets will have substantially different signatures, or fingerprints, in the spatial pattern of sea-level change. These patterns are linked to the gravitational effects of the lost ice (sea level is raised near an ice sheet because of the gravitational attraction of the ice mass for the adjacent ocean water) and to the elastic (as opposed to viscoelastic) deformation of Earth driven by the rapid unloading. Some ambiguity in determining the source of meltwater arises because of uncertainty in both the original correction for glacial isostatic adjustment and in the correction for the poorly known signature of ocean thermal expansion, as well as from the non-uniform distribution of tide gauge sites.

Other geodetic indicators related to Earth's rotational state also constrain estimates of recent changes in the mass of ice-sheets (Munk, 2002; Mitrovica et al., 2006). Earth's rotation is affected by any redistribution of mass on or inside the planet. Transfer of mass from the poles to the equator slows the planet's rotation (like a spinning ice skater extending her arms to slow her rotation). Moreover, any transfer of mass that is not symmetric about the poles causes "wobble," or true polar wander (TPW) (that is, the position of the north rotation pole moves relative to the surface of the planet). True polar wander for the last century has been estimated using both astronomical and satellite geodetic data. In contrast, changes in the rotation rate (or, as geodesists say, length of day), have been determined for the last few decades by using satellite measurements and for the last few millennia by using observations of eclipses recorded by ancient cultures. Specifically, the timing of ancient eclipses recorded by these cultures differs from the timing one would expect by simply projecting the Earth-Moon-Sun system back in time using the modern rotation rate of Earth. The discrepancy indicates a gradual slowing of Earth's rate of rotation (Munk, 2002). The difference in the rotation-rate history during the last few millennia (after correcting for slowing of Earth's rotation associated with the "drag" of the tides) as compared with the rotation rate of last few decades provides a measure of any anomalous recent melting of polar ice reservoirs. (This difference does not uniquely constrain the individual sources of the meltwater because all sources will be about equally efficient, for a given mass loss rate, at driving these changes in rotation.) True polar wander, after correction for glacial isostatic adjustment, serves as an important complement to this rotation-rate analysis because it does give some information about the source of the meltwater. As an example, melting from the Antarctic, because it is located at the pole, generates very little true polar wander, whereas melting from the *Greenland Ice Sheet*, whose center of mass lies about 15 degrees off Earth's rotation axis, is capable of driving substantial true polar wander (Munk, 2002; Mitrovica et al., 2006).

6.2.2f Ice cores

Ice cores preserve information about many climatic variables that affected the ice sheet, and about how the ice sheet responded to changes in those variables.

Temperature histories derived from ice cores are especially accurate. Several indicators are used, as described next, such as the isotopic ratios of accumulated snow, ice-sheet temperature profiles (using borehole thermometry), and various techniques based on use of gas-isotopic indicators. Agreement among these different indicators increases confidence in the results.

Let us first consider isotopic ratios of the oxygen and hydrogen in accumulated snow (e.g., Jouzel et al., 1997). The ocean contains both normal and “heavy” water: roughly one molecule in 500 incorporates at least one extra neutron in the nucleus of an oxygen or hydrogen atom. The lighter molecules evaporate more easily, and the heavier molecules condense (and thus precipitates) more easily.. As water that evaporated from the ocean is carried by an air mass inland over an ice sheet, the heavy molecules preferentially rain or snow out. The colder the air mass, the more vapor is removed, the more depleted of the heavy molecules is the remaining vapor, and the lighter the isotopic ratios in the next rain or snow. Hence, the isotopic composition of precipitation is linked to temperature of the air mass and, over polar ice sheets, the temperature of the air mass is typically linked to the surface temperature. Oxygen- and hydrogen-isotope ratios are both studied, and they help locate the source of precipitation, track the changing isotopic composition of the moving air mass (“path effects”), and indicate the ice-sheet temperature as well. Because site temperature is most important for this review, one species is sufficient. Results will be discussed here as $\delta^{18}O$, the difference between the ^{18}O:^{16}O ratio of a sample and of standard mean ocean water, normalized by the ratio of the standard and expressed not as percent but as per mil (‰) (percent is parts per hundred, and per mil is parts per thousand).

Although linked to site temperature, $\delta^{18}O$ can be affected by many factors (Jouzel et al., 1997; Alley and Cuffey, 2001), such as change in the ratio of summertime to wintertime precipitation. Hence, additional means of determining past temperatures are required. One of the most reliable is based on the physical temperature of the ice. Just as a frozen turkey takes a long time in a hot oven to warm in the middle, intermediate depths of the central *Greenland Ice Sheet* are colder than ice above or below. Surface ice temperatures equilibrate with air temperature, and basal ice receives some warmth from Earth's heat flow, but the center of the ice sheet has not finished warming from the ice-age cold. If ice flow is understood well at a site, the modern profile of the physical temperature of the ice with increasing depth provides a low-time-resolution history of the surface temperature with increasing time. Joint interpretation of the isotopic ratios and temperatures measured in boreholes (Cuffey et al., 1995; Cuffey and Clow, 1997), or independent interpretation of the borehole temperatures and then comparison with the isotopic ratios (Dahl-Jensen et al., 1998), helps to outline the history of surface air temperature. Furthermore, the relation between isotopic ratio and temperature (α ‰ per °C) becomes a useful paleoclimatic indicator, and changes in this ratio α with time can be used to test hypotheses about the overall changes in seasonality of snowfall and other factors.

The isotopic composition of gases trapped in bubbles in the ice sheet provides an additional indicator of temperature. New-fallen snow contains many interconnected air spaces. Snow turns to ice without melting in central regions of cold ice sheets through solid-state mechanisms that operate more rapidly under higher temperature or higher pressure. Snow in an ice sheet usually transforms to ice within the top few tens of meters.

The intermediate material is called firn, and the transformation is complete when bubbles are isolated so that the air spaces are no longer interconnected to the surface. Wind moving over the ice sheet typically mixes gases in the pore spaces of the firn only in the uppermost few meters or less. Diffusion mixes the gases deeper than this. Gases are slightly separated by gravity (Sowers et al., 1992), with the air trapped in bubbles slightly isotopically heavier than in the free atmosphere, proportional to the thickness of the air column in which diffusion dominates.

If a sudden temperature change occurs at the surface, the resulting temperature change of the firn beneath requires typically about 100 years to penetrate to the depth of bubble trapping. However, when a temperature gradient is applied across gases in diffusive equilibrium, the gases are separated by thermal fractionation as well as by gravity, with the heavier gases moved thermally to the colder end (Severinghaus et al., 1998). Equilibrium of gases is obtained in a few years, far faster than the time for heat flow to remove the temperature gradient across the firn. Within a few years after an abrupt temperature change at the surface, newly forming bubbles will begin to trap air with very slight (but easily measured) anomalies in gas-isotope compositions, and this trapping of slightly anomalous air will continue for a century or so. Because different gases have different sensitivities to temperature gradients and to gravity, measuring isotopic ratios of several gases (such as argon and nitrogen) allows researchers to determine the temperature difference that existed vertically in the firn at the time of bubble trapping and to determine the thickness of firn in which wind was not mixing the gas (Severinghaus et al., 1998). If the surface temperature changed very quickly, the magnitude of the temperature difference across the firn will peak at the magnitude of the surface-temperature change; for a slower change, the temperature difference across the firn will always be less than the total temperature change at the surface. If the climate was relatively steady before an abrupt temperature change, such that the depth-density profile of the firn came into balance with the temperature and the accumulation rate, and if the accumulation rate is known independently (see below), then the number of years or amount of ice between the gas-phase and ice-phase indications of abrupt change provides information on the mean temperature before the abrupt change (Severinghaus et al., 1998). With so many independent thermometers, highly confident paleothermometry is possible.

Ice cores can provide information on climatic indicators other than temperature. Past ice-accumulation rates are most readily obtained by measuring the thickness of annual layers in ice cores corrected for ice-flow thinning (e.g., Alley et al., 1993). In other methods, the thickness of firn can be approximated by measurements of gas-isotope fractionation or of the number and density of bubbles (Spencer et al., 2006); these measurements combined with temperature estimates constrain accumulation rates as well. Aerosols (very small liquid and solid particles) of all types fall with snow and during intervals when snow is not falling, and are incorporated into the ice sheet; with knowledge of the accumulation rate (hence dilution of the aerosols), time histories of atmospheric loading of those aerosols can be estimated (e.g., Alley et al., 1995a). Dust and volcanic fallout (e.g., Zielinski et al., 1994) help constrain the cooling effects of aerosols (particles) blocking the Sun. Cosmogenic isotopes (beryllium-10 is most commonly measured) reflect cosmic-ray bombardment of the atmosphere, which is modulated by the strength of Earth's magnetic field and by solar activity (e.g., Finkel and

Nizhiizumi, 1997). The observed correlation in paleoclimatic records between indicators of climate and indicators of solar activity (Stuiver et al., 1997; Muscheler et al., 2005; Bard and Frank, 2006)—and the lack of correlation with indicators of magnetic-field strength (Finkel and Nishiizumi, 1997; Muscheler et al., 2005)—help researchers understand climate changes.

Ages in ice cores are most commonly estimated by counting annual layers (e.g., Alley et al., 1993; Andersen et al., 2006) and by correlation with other records (Blunier and Brook, 2001). Several indicators of atmospheric composition from *Greenland* ice cores that were matched with similar (but longer) records from Antarctica (Suwa et al., 2006) showed that old ice exists in central *Greenland* (Suwa et al., 2006; Chappellaz et al., 1997) at depths where flow processes have mixed the layers (Alley et al., 1997). In regions of continuous and unmixed layers, other features in ice cores, such as chemically distinctive ash from particular volcanic eruptions, can be correlated with independently dated records (e.g., Finkel and Nishiizumi, 1997; Zielinski et al. 1994). Flow models also can be used to aid in dating.

The past elevation of ice-sheets is indicated by the total gas content of the ice (Raynaud et al., 1997) at a given depth and age. As noted above in this section, bubbles are pinched off (pore close-off) from interconnected air spaces in the firn a few tens of meters down. The density of the ice at this pore close-off is nearly constant, with a small and fairly well known correction for climatic conditions. Because air pressure varies with elevation and elevation varies with ice thickness, the total number of trapped molecules of gas per unit volume of ice is correlated with ice-sheet thickness. Small elevation changes cannot be detected (because of additional fluctuations in total gas content that are likely linked to changing layering in the firn that affects trapped bubbles), but elevation changes of greater than 500 m are detectable with confidence (Raynaud et al., 1997).

Additional information on ice-sheet changes comes from the current distribution of isochronous surfaces (surfaces that have the same age throughout) in the ice sheet. An explosive volcanic eruption will deposit an acidic ash layer of a single age on the surface of the ice sheet, and that layer can be identified after burial by using radar (Whillans, 1976). Ages of reflectors can be determined at ice-core sites (e.g., Eisen et al., 2004), and the layers can then be mapped throughout broad areas (Jacobel and Welch, 2005). A model can be used to predict the current distribution of isochronous surfaces (as well as some other properties, such as temperature) for any hypothesis that combines the history of climatic forcing (primarily accumulation rate affecting burial and temperature) and ice-sheet flow (primarily changes in surface elevation and extent) (e.g., Clarke et al., 2005). Optimal histories can be estimated in this way.

6.3 History of the *Greenland Ice Sheet*

6.3.1 Ice-Sheet Onset and Early Fluctuations

Prior to 65 million years ago (Ma), dinosaurs lived on a high-CO_2, warm world that usually lacked permanent ice at sea level. The high latitudes were warm; Tarduno et al. (1998) provided a minimum estimate of the mean-annual temperature during this time of over 14°C at 71°N based on occurrence of crocodile-like champsosaurs (also see Vandermark et al., 2007; Markwick, 1998). Sluijs et al. (2006) showed that the ocean surface warmed near the North Pole from about 18°C to peak temperatures of 23°C during the short-lived Paleocene-Eocene Thermal Maximum about 55 Ma. Such warm temperatures preclude permanent ice near sea level and, indeed, no evidence of such ice has been found (Moran et al., 2006).

Cooling following the Paleocene-Eocene Thermal Maximum may have allowed ice to reach sea level fairly quickly; sand and coarser materials found in a core from the Arctic Ocean sea floor and dated at about 46 Ma (Moran et al. 2006; St. John, 2008) are most easily (but not with absolute certainty) interpreted as indicating ice rafting linked to glaciers. Ice-rafted debris likely traceable at least in part to glaciers rather than to sea ice is found in a core recovered from about 75°N latitude in the *Norwegian-Greenland Sea* off East Greenland; the core is dated between about 38 and 30 Ma (late Eocene into Oligocene time). Certain characteristics of this debris point to an East Greenland source and exclude *Svalbard*, the next-nearest land mass (Eldrett et al., 2007). It is not known whether this ice-rafted debris represents isolated mountain glaciers or more-extensive ice-sheet cover.

The central Arctic Ocean sediment core of Moran et al. (2006) shows a highly condensed record that suggests erosion or little deposition across this interval of ice rafting off *Greenland* studied by Eldrett et al. (2007; see previous paragraph) and until about 16 Ma. Ice-rafted debris, interpreted as representing iceberg as well as sea-ice transport, was actively delivered to the open-ocean site studied by Moran et al. (2006) at 16 Ma, and volumes increased about 14 Ma and again about 3.2 Ma (also see Shackleton et al., 1984; Thiede et al., 1998; Kleiven et al., 2002). St. John and Krissek (2002) suggested onset of sea-level glaciation in southeastern *Greenland* at about 7.3 Ma, on the basis of ice-rafted debris near *Greenland* in the *Irminger Basin*. Because of its geographical pattern, the increase in ice-rafted debris about 3.2 Ma is thought to have had sources in *Greenland*, Scandinavia, and the North American landmass (*Laurentide Ice Sheet*). However, tying the debris to particular source rocks (e.g., Hemming et al., 2002) has not been possible. Additionally, no direct evidence shows whether this debris was supplied to the ocean by an extensive ice sheet or by vigorous glaciers that drained coastal mountains in the absence of ice from *Greenland*'s central lowlands. Despite the lack of conclusive evidence, *Greenland* seems to have supported at least some glaciation since at least 38 Ma; glaciation left more records after about 14 Ma (middle Miocene). Thus, as Earth cooled from the "hothouse" conditions extant during the time of dinosaurs, ice sheets began to form on *Greenland*.

Following the establishment of ice in *Greenland*, a notable warm interval about 2.4 million years (m.y.) ago is recorded by the *Kap København* Formation of North *Greenland* (Funder et al., 2001). This formation is a 100-m-thick unit of sand, silt, and clay deposited primarily in shallow marine conditions. Fossil biota in the deposit switch from Arctic to subarctic to boreal assemblages during the depositional interval. The unit was deposited rapidly, perhaps in 20,000 years or less. Funder et al. (2001) postulated complete deglaciation of *Greenland* at this time, primarily on the basis of the great summertime warmth indicated at this far-northern site, although clearly there is no comprehensive record of the whole ice sheet.

6.3.2 The Most Recent Million Years

Fragmented records on land combined with lack of unequivocal indicators in the ocean complicate ice-sheet reconstructions. Nonetheless, many additional indications of ice-sheet change are available between the time of the *Kap København* Formation and the most recent 100,000 years. Locally, ice expanded during colder times and ice retreated during warmer times, but data provide no comprehensive overviews of the ice sheet. This section (6.3.2) summarizes data especially from marine isotope stage (MIS) 11 (about 440 ka; see chapter 3.5 on Chronology) to MIS 5 (about 130 ka), although dating uncertainties allow the possibility that some of the samples are older than MIS 11, and detailed consideration of MIS 5 is deferred to subsequent sections.

Glacial-interglacial cycles have been studied by examining the oxygen isotope composition of foraminifers in deep-sea cores, and we how have a fairly detailed picture of how glacial ice has expanded and retreated during the past 2 m.y. or so (the Quaternary period). Figure 6.4 shows the four most recent glacial-interglacial cycles: peaks represent interglacial periods (relatively high sea levels) and troughs represent glacial periods (relatively low sea levels). Glacial periods in the oxygen isotope record are called “stages” and are numbered back in time with even numbers; interglacial stages are numbered back in time with odd numbers. Thus, the present interglacial is marine isotope stage (MIS) 1 and the preceding glacial period is MIS 2.

FIGURE 6.4 NEAR HERE

6.3.2a Far-field sea-level indications

In the absence of clear and well-dated records proximal to the *Greenland Ice Sheet*, records of global sea level that may be related to changes on *Greenland* are of interest. If we consider only the past few glacial cycles, it is most likely that sea level was as high as or higher than present during previous interglacial times (MIS 5, 7, 9, and 11; Figure 6.4). Under the assumption that any glacial-isostatic-adjustment contributions to these relative highstands of sea level were small, and thus that highstands of sea level were primarily related to changes in ice volume, the amplitudes of the various highstands of sea level provide a measure of the long-term mass balance of the *Greenland Ice Sheet* and other contemporaneous ice masses.

Far from the *Greenland Ice Sheet*, some fragmentary and poorly dated deposits suggest a higher-than-present sea-level stand during MIS 11, about 400 ka. Sea-level history of MIS 11 [about 362–420 ka] (as noted in section 3.5, Chronology, age assignments to marine isotope stages may differ in different usages; both age ranges and marine isotope stage names are given here for information, not as definitions) is of particular interest to paleoclimatologists because the Earth-Sun orbital geometry during that interglacial epoch is similar to the configuration during the current interglacial (Berger and Loutre, 1991).

Hearty et al. (1999) proposed that marine deposits found in a cave on the tectonically stable island of Bermuda date to the MIS 11 interglacial epoch. These marine deposits are about 21 m above modern sea level, and they contain coral pebbles that have been dated by U-series techniques. Hearty et al. (1999) interpreted the deposits to date to about 400 ka, although the coral pebbles were dated older than 500 ka. The authors' interpretation is based primarily on an overlying deposit that dates to about 400 ka. Although the deposit appears to record an old sea stand markedly higher than present, the chronology is still uncertain.

An Alaskan marine deposit is also found at altitudes of up to 22 m (Kaufman et al., 1991), similar to altitudes of the cave deposit on Bermuda. The deposit, representing what has been called the “Anvilian marine transgression,” extends along the Seward Peninsula and Arctic Ocean coast of Alaska. This part of Alaska is tectonically stable. It is landward of Pelukian (MIS 5 (about 74–130 ka)) marine deposits. Amino-acid ratios in mollusks (Kaufman and Brigham-Grette, 1993) show that the Anvilian deposit is easily distinguishable from last-interglacial (locally called Pelukian) deposits, but it is younger than deposits thought to be of Pliocene age (about 1.8–5.3 Ma). Kaufman et al. (1991) reported that basaltic lava overlies deposits of the Nome River glaciation, which in turn overlie Anvilian marine deposits. An average of several analyses on the lava yields an age of 470 ± 190 ka. Within the broad limits permitted by this age, and using reasonable rates of changes in the amino-acid ratios of marine mollusks, Kaufman et al. (1991) proposed that the Anvilian marine transgression dates to about 400 ka and correlates with MIS 11.

Other far-field evidence supports the concept that during MIS 11 sea level was higher than at present. Oxygen-isotope and faunal data from the Cariaco Basin off Venezuela provide independent evidence of a higher-than-present sea level during MIS 11 (Poore and Dowsett, 2001). If the Bermudan cave deposits and the Anvilian marine deposits of Alaska prove to be genuine manifestations of a ~400 ka-old high sea stand, the implication for climate history is that all of the *Greenland Ice Sheet* (Willerslev et al., 2007; see section 6.3.2b, below), all of the West Antarctic ice sheet, and part of the East Antarctic ice sheet would have disappeared at this time (these being generally accepted as the most vulnerable ice masses); preservation of the *Greenland Ice Sheet* would require much more loss from the East Antarctic ice sheet, which is widely considered to be relatively stable (e.g., Huybrechts and de Wolde, 1999).

Until recently, no reliably dated emergent marine deposits from MIS 9 [about 303–331 ka] had been found on tectonically stable coasts, although coral reefs of this age have been recognized for some time on the tectonically rising island of Barbados (Bender et al., 1979). Stirling et al. (2001) reported that well-preserved fringing reefs are found on Henderson Island in the southeastern Pacific Ocean. Reef elevations on this tectonically stable island are as high as about 29 m above sea level, and U-series dates between about 334 ± 4 and 293 ± 5 ka correlate with MIS 9. Despite the good preservation of the corals and the reefs they are found in, and the reliable U-series ages, it is uncertain how high sea level was at this time. Although Henderson Island is geologically stable, it is experiencing slow uplift (less than 0.1 m/1,000 yr) due to volcanic loading by the emplacement of nearby Pitcairn Island. A correction for maximum uplift rate, therefore, could put the MIS 9 ancient level estimate below present sea level. Multer et al. (2002) reported U-series ages of about 370 ka for a coral (*Montastrea annularis*) from a fossil reef drilled at a locality called Pleasant Point in Florida Bay. This coral showed clear evidence of open-system conditions (i.e., it was not completely chemically isolated from its surroundings since formation, a requirement for the measured age to be accurate), and the age is probably closer to 300–340 ka, if we use the correction scheme of Gallup et al. (1994). If so, the age suggests that during MIS 9, sea level was close to but not much above the present level.

As with MIS 9, several MIS 7 (about 190–241 ka) reef or terrace records have been found on tectonically rising coasts (Bender et al., 1979; Gallup et al., 1994; Edwards et al., 1997), but far fewer have been found on tectonically relatively stable coasts. However, two recent reports show evidence of MIS 7 sea-level high stands on tectonically stable islands. One is a pair of U-series ages of about 200 ka from coral-bearing marine deposits about 2 m above sea level on Bermuda (Muhs et al., 2002). The other is a single coral age from the Florida Keys (Muhs et al., 2004). They collected samples of near-surface *Montastrea annularis* corals in quarry spoil piles on Long Key. Analysis of a single sample shows an apparent age of 235 ± 4 ka. The higher-than-modern initial $^{234}U/^{238}U$ value indicates a probable bias to an older age by about 7 ka; thus, the true age may be closer to about 220–230 ka, if we again use the Gallup et al. (1994) correction scheme. If valid, these data suggest that sea level may have stood close to its present level during the interglacial period MIS 7. Much more study is needed to confirm these preliminary ages, however.

Taken together, these data point to MIS 11 as a time in which sea level likely was notably higher than at present, although the data are sufficiently sparse that stronger conclusions are not warranted. If so, melting of *Greenland* ice seems likely, mostly on the basis of elimination: *Greenland* meltwater is thought to be able to supply much of the sea-level rise needed to explain the observations, and the alternative—extracting an additional 7 m of sea-level rise through melting in East Antarctica—is not considered as likely. Marine isotope stages 9 and 7 seem to have had sea levels similar to modern ones.

6.3.2b Ice-sheet indications

The cold MIS 6 ice age (about 130–188 ka) may have produced the most extensive ice in *Greenland* (Wilken and Meinert, 2006). Recently described glacial deposits in east Greenland support this view (Adrielsson and Alexanderson, 2005), although more-extensive, older deposits are known locally (Funder et al., 2004). Funder et al. (1998) reconstructed thick ice (greater than1000 m) during MIS 6 in areas of *Jameson Land* (east Greenland) that now are ice-free. However, no confident ice-sheet-wide reconstructions based on paleoclimatic data are available for MIS 6 ice.

Both northwest and east Greenland preserve widespread marine deposits from early in the MIS 5 interglacial (the interglacial previous to the present one) (about 74–130 ka), and particularly from the warmest subdivision of MIS 5, called MIS 5e (about 123 ka). Depression of the land from the weight of MIS 6 ice allowed incursion of seawater as ice melted during the transition to MIS 5e. The resulting deposits were not reworked by the subsequent incursion of seawater during the transition from the most recent glaciation (MIS 2, which peaked about 24 ka or slightly more recently) to the modern interglacial (MIS 1, less than 11 ka). Thus, seawater moved farther inland during the transition from MIS 6 (glacial) to MIS 5 (interglacial) than during the transition from MIS 2 (most recent glacial) to MIS 1 (current interglacial).

Several hypotheses can explain this difference. Perhaps most simply, there may have been more ice on *Greenland* causing greater isostatic depression during MIS 6 than during MIS 2. However, if some or all of the older deposits survived being overridden by cold-based ice of MIS 2, additional possibilities exist. Because isostatic uplift occurs while ice is thinning but before the ice margin melts enough to allow incursion of seawater, perhaps the MIS 6 ice melted faster and allowed incursion of seawater over more-depressed land than was true for MIS 2 ice. Additionally, at the time during MIS 6 that ice in *Greenland* receded and thus allowed incursion of seawater, global sea level might have been higher than during the corresponding part of MIS 2 (perhaps because of relatively earlier melting of MIS 6 ice on North America or elsewhere beyond *Greenland*). More-detailed modeling of glacial isostatic adjustment will be required to test these hypotheses. Nonetheless, the leading hypothesis seems to be that ice was more extensive in MIS 6 than in MIS 2.

A particularly interesting new result comes from analysis of materials found in ice cores from the deepest part of the ice sheet. Willerslev et al. (2007) attempted to amplify DNA in three samples: (1) silty ice at the base of the *Greenland Ice Sheet* from the *Dye-3* drill site (on the southern dome of the ice sheet) and the *GRIP* drill site (at the crest of the main dome of the ice sheet), (2) “clean” ice just above the silty ice of these sites, and (3) the *Kap København* formation. The *Kap København*, clean-ice, and *GRIP* silty samples did not yield identifiable quantities of DNA (probably indicating post-depositional changes for *Kap København* perhaps during room-temperature storage following collection, and showing that long-distance transport is not important for supplying large quantities of DNA to the ice of the central part of the sheet). However, it was possible to prepare extensive materials from the *Dye 3* silty ice. These materials indicate a northern boreal forest, compared to the tundra environment that exists in coastal sites at the same latitude and lower elevation today. . The taxa indicate mean July temperatures then above 10°C and minimum winter temperatures above –17°C at an elevation of about 1 km above sea level (allowing for isostatic rebound following ice melting). Dating of this warm, reduced-ice time is uncertain, but a tentative age of 450–800 ka is probably consistent with the indications of high sea level in MIS 11.

Nishiizumi et al. (1996) reported on radioactive cosmogenic isotopes in rock core collected from beneath the ice at the *GISP2* site (central Greenland, 28 km west of the *GRIP* site at the *Greenland* summit). Joint analysis of beryllium-10 and aluminum-26 indicated a few-millennia-long interval of exposure to cosmic rays (hence ice cover of thickness less than 1 m or so) about 500 ± 200 ka. This information is consistent with, and thus provides further support for, the DNA results of Willerslev et al. (2007). This work was presented at a scientific meeting and in an abstract but not in a refereed scientific journal, and thus it is subject to lower confidence than is other evidence discussed in this report.

No long, continuous climate records from *Greenland* itself are available for the time interval occupied by the boreal forest at *Dye-3* reported by Willerslev et al. (2007). Marine-sediment records from around the North Atlantic point toward MIS 11, at about 440 ka, as the most likely time of anomalous warmth. Owing to orbital forcing factors (reviewed in Droxler et al., 2003), this interglacial seems to have been anomalously long compared with those before and after. As discussed above, indications of sea level above modern level exist for this interval (Kindler and Hearty, 2000), but much uncertainty remains (see Rohling et al., 1998; Droxler et al., 2003). Records of sea-surface-temperature in the North Atlantic indicate that MIS 11 temperatures were similar to those from the current interglacial (Holocene) within 1°–2°C; slightly cooler, similar, or slightly warmer conditions have all been reported (e.g., Bauch et al., 2000; de Abreu et al. 2005; Helmke et al., 2003; McManus et al., 1999, Kandiano and Bauch, 2003). The longer of these records show no other anomalously warm times within the age interval most consistent with the Willerslev et al. (2007) dates. (Notice, however, that during MIS 5e locally higher temperatures are indicated in *Greenland* than are indicated in the far-field sea-surface temperatures. Thus, the absence of warm temperatures far from the ice sheet does not guarantee the absence of warm temperatures close to the ice sheet; see 6.3.3, below.) The independent indications of high global sea level during MIS 11, as discussed above in section 6.3.2a, and of major *Greenland Ice Sheet* shrinkage or loss at that time, are mutually consistent.

The *Greenland Ice Sheet* is thought to complete most of its response to a step forcing in climate within a few millennia (e.g., Alley and Whillans, 1984; Cuffey and Clow, 1997). Thus, any of the interglacials during the last 420,000 years was long enough for the ice sheet to have completed most of its response to the end-of-ice-age forcings (although smaller forcings during the interglacials may have precluded a completely steady state). Thus, it is not obvious how a longer-yet-not-warmer interglacial, as suggested by MIS 11 indicators in the North Atlantic away from *Greenland*, would have caused notable or even complete loss of the *Greenland Ice Sheet*, although this result cannot be ruled out completely. Many possible interpretations remain: greater *Greenland* warming in MIS 11 than indicated by marine records from well beyond the ice sheet, large age error in the Willerslev et al. (2007) estimates, great warmth at Dye-3 yet a reduced but persistent *Greenland Ice Sheet* nearby, and others. One possible interpretation is that the threshold for notable shrinkage or loss of *Greenland* ice is just 1°–2°C above the temperature reached during MIS 5e, thus falling within the error bounds of the data.

The data strongly indicate that *Greenland*'s ice was notably reduced, or lost, sometime after ice coverage became extensive and large ice ages began, while temperatures surrounding *Greenland* were not grossly higher than they have been recently. The rate of mass loss within the warm period is unconstrained; the long interglacial at MIS 11 allows the possibility of very slow loss or much faster loss. If the cosmogenic isotopes in the *GISP2* rock core are interpreted at face value, then the time over which ice was absent was only a few millennia.

6.3.3 Marine Isotope Stage 5e

6.3.3a Far-field sea-level indications

Investigators studying sea-level history have paid most attention to sea level during the last interglacial, MIS 5 (about 71–122 ka), and specifically to MIS 5e (about 123 ka). The evidence of past sea level during MIS 5e along tectonically stable coasts is summarized here (Muhs, 2002). Sea-level high stand during MIS 5e is best estimated from coral reef and marine deposits now above sea level at sites in Australia, the Bahamas, Bermuda, and the Florida Keys.

On the coast and islands of tectonically stable Western Australia, emergent coral reefs and marine deposits now 2–4 m above sea level are widespread and well-preserved. U-series ages of the fossil corals at mainland localities and Rottnest Island range from 128 ± 1 to 116 ± 1 ka (Stirling et al., 1995, 1998). The main period of last-interglacial coral growth was a restricted interval from about 128–121 ka (Stirling et al., 1995, 1998). Because the highest corals are about 4 m above sea level at present but grew at some unknown depth below sea level, 4 m is a minimum for the amount of last-interglacial sea-level rise.

The islands of the Bahamas are tectonically stable, although they may be slowly subsiding owing to carbonate loading on the Bahamian platform. Fossil reefs in the

Bahamas are well preserved (Chen et al., 1991), reefs have elevations up to 5 m above sea level, and many corals are in growth position. On San Salvador Island, reef ages range from 130.3 ± 1.3 to 119.9 ± 1.4 ka. The sea level record of the Bahamas is particularly valuable because many reefs contain the coral *Acropora palmata*, a species that almost always lives within the upper 5 m of the water column (Goreau, 1959). Thus, fossil reefs containing this species place a fairly precise constraint on the former water depth.

As discussed above (section 6.3.2a), Bermuda is tectonically stable. Bermuda does not host MIS 5e fossil reefs, but numerous coral-bearing marine deposits fringe the island. A number of U-series ages of corals from Bermuda range from about 119 ka to about 113 ka (Muhs et al., 2002). The deposits are found 2–3 m above present sea level, although overlying wind-blown sand prevents precise estimates of where the former shoreline lay.

The Florida Keys, not far from the Bahamas, are also tectonically stable. Fruijtier et al. (2000) reported ages for corals from Windley Key, Upper Matecumbe Key, and Key Largo that, when corrected for high initial $^{234}U/^{238}U$ values (Gallup et al., 1994), are in the range of 130–121 ka. The last-interglacial MIS 5 reef on Windley Key is 3–5 m above present sea level, on Grassy Key it is 1–2 m above sea level, and on Key Largo it is 3–4 m above modern sea level.

The collective evidence from Australia, Bermuda, the Bahamas, and the Florida Keys shows that sea level was above its present stand during MIS 5e. On the basis of measurements of the reefs themselves, sea level then was at least 4–5 m higher than sea level now. An additional correction should be applied for the water depth at which the various coral species grew. Most coral species found in Bermuda, the Bahamas, and the Florida Keys require water depths of at least a few meters for optimal growth, and many live tens of meters below the ocean surface. For example, *Montastrea annularis*, the most common coral found in MIS 5e reefs of the Florida Keys, has an optimum growth depth of 3–45 m and can live as deep as 80 m (Goreau, 1959). A minimum rise in sea level is calculated thusly: fossil reefs are 3 m above present sea level, and the most conservative estimate of the depth at which they grew is 3 m. Thus, the MIS 5e sea level was at least 6 m higher than modern-day sea level (Figures 6.5, 6.6). A summary of additional sites led Overpeck et al. (2006) to indicate a sea-level rise of 4 m to more than 6 m during MIS 5e.

FIGURE 6.5 NEAR HERE

FIGURE 6.6 NEAR HERE

Existing estimates generally presume that glacial isostatic adjustment have not notably affected the sites at the key times. The data set, and the accuracy of the dates (also see Thompson and Goldstein, 2005) are becoming sufficient to support, in future work, improved corrections for glacial isostatic adjustment.

The implications of a 4 m to more than 6 m sea-level highstand during the last interglacial are as follows: (1) all or most of the *Greenland Ice Sheet* would have melted; or (2) all or most of the West Antarctic ice sheet would have melted; or (3) parts of both would have melted. Both ice sheets may indeed have melted in part, but greater melting is likely from *Greenland* (Overpeck et al., 2006), as described in section 6.3.3c, below.

6.3.3b Conditions in Greenland

Paleoclimate data provide strong evidence for notable warmth on and around *Greenland* during MIS 5e, with peak temperatures occurring ~130 ka. As summarized by CAPE (2006), terrestrial data indicate peak summertime temperatures ~4^{o}C above recent in NW *Greenland* and ~5^{o}C above recent in east *Greenland* (and thus 2–4^{o}C above the mid-Holocene warmth [~6 ka]; Funder et al., 1998, and see below), with near-shore marine conditions 2–3^{o}C above recent in east *Greenland*. Climate-model simulations by Otto-Bliesner et al. (2006) show that the strong summertime increase of sunshine (insolation) in MIS 5e as compared to now caused strong warming, which was amplified by ice-albedo and other feedbacks. Simulated summertime warming around *Greenland* exhibited local maxima of 4-5^{o}C in those northwestern and eastern coastal regions for which terrestrial and shallow-marine summertime data are available and show matching warmings; elsewhere over *Greenland* and surroundings, typical warmings of ~3^{o}C were simulated.

The sea-level record in East Greenland (*Scoresby Sund*) indicates a two-step inundation at the start of MIS 5e. Of the possible interpretations, Funder et al. (1998) favored one in which early deglaciation of the coastal region of *Greenland* preceded much of the melting of non-*Greenland* land ice, so that early coastal flooding after deglaciation of isostatically depressed land was followed by uplift and then by flooding attributable to sea-level rise as that far-field land ice melted. Additional testing of this idea would be very interesting, as it suggests that the *Greenland Ice Sheet* has responded rapidly to climate forcing in the past.

Much of the evidence of climate change in *Greenland* comes from ice-core records. As discussed next, these changes cannot be estimated independent of a discussion of the ice sheet, because of the possibility of thickness change. Hence, the changes in the ice sheet are discussed before additional evidence bearing on forcing and response.

6.3.3c Ice-sheet changes

The *Greenland Ice Sheet* during MIS 5e covered a smaller area than it does now. How much smaller is not known with certainty. The most compelling evidence is the absence of pre-MIS 5e ice in the ice cores from south, northwest, and east Greenland (the locations *Dye-3, Camp Century,* and *Renland* drilling sites, respectively). In all of these cores, the climate record extends through the entire last glacial epoch and then terminates at the bed in a layer of ice deposited in a much warmer climate (Koerner, 1989; Koerner and Fisher, 2002). This basal ice is most likely MIS 5e ice. Moreover, the composition of this ice is not an average of glacial and interglacial values, as would be expected if it were a mixture of ices from earlier cold and warm climates. Instead, the ice composition exclusively indicates a climate considerably warmer than that of the Holocene. (One cannot entirely eliminate the possibility that each core independently bottomed on a rock that had been transported up from the bed, and that older ice lies beneath each rock, but this seems highly improbable.)

At Dye-3, the oxygen isotope composition of this basal ice layer is reported as $\delta^{18}O = –23‰$, which means that it is 23‰ (or 2.3%) lighter than standard mean ocean water. Moreover, a value of $\delta^{18}O = –30‰$ is reported for modern snowfall in the source region (up-flow from the site of Dye-3). At *Camp Century*, a value of $\delta^{18}O = –25‰$ is reported for basal ice; a value of $\delta^{18}O = -31.5‰$ is reported in the source region (see Table 2 of Koerner, 1989). These changes of about 7‰ are much larger than the MIS 5e-to-MIS 1 climatic signal (about 3.3‰, according to the central Greenland cores; see below in this section). Thus, the MIS 5e ice at *Dye-3* and *Camp Century* not only indicates a warmer climate but also a much lower source elevation: the ice sheet was re-growing when these MIS 5e ices were deposited.

In combination, these two observations (absence of pre-MIS 5e ice, and anomalously low-elevation sources of the basal ice) indicate that the Greenland margin had retreated considerably during MIS 5e. Of greatest importance is that retreat of the margin northward past *Dye-3* implies that the southern dome of the ice sheet was nearly or completely gone.

In this context it is useful to understand the genesis of the basal ice layer, and the layer at *Dye-3* in particular. Unfortunately the picture is cloudy—not unlike the basal ice itself, which has a small amount of silt and sand dispersed through it, making it opaque. This silty basal layer is about 25 m thick (Souchez et al., 1998). Overlying it is “clean” (not notably silty) ice that appears to be typical of polar ice sheets. Its total gas content and gas composition indicate that the ice formed by normal densification of firn in a cold, dry environment. The oxygen isotope composition of this clean ice is –30.5‰. The bottom 4 m of the silty ice is radically different; its oxygen isotope value is –23‰, and its gas composition indicates substantial alteration by water. The total gas content of this basal silty ice is about half that of normal cold ice formed from solid-state transformation of firn, the carbon dioxide content is 100 times normal, and the oxygen/nitrogen ratio is less than 20% that of normal cold ice. This basal silty layer may be superimposed ice (ice formed by refreezing of meltwater in snow on a glacier or ice sheet, as Koerner (1989) suggested for the entire silty layer), or it may be non-glacial snowpack, or it may be a remnant of segregation ice in permafrost (permafrost commonly contains relatively “clean” although still impure lenses of ice, called segregation ice).

In any case, the upper 21 m of the silty ice may be explained as a mixture of these two end members (Souchez et al. 1998). As they deform, ice sheets do mix ice layers by small-scale structural folding (e.g., Alley et al., 1995b), by interactions between rock particles, by grain-boundary diffusion, and possibly by other processes. Unfortunately, there is no way to distinguish rigorously how much this ice really is a mixture of these end-member components and how much of it is warm-climate (presumably MIS 5e) normal ice-sheet ice. The difficulty is that the bottom layer is not itself well mixed (its gas composition is highly variable), so a mixing model for the middle layer uses an essentially arbitrary composition for one end member. Souchez et al. (1998) used the composition at the top of the bottom layer for their mixing calculations, but it could just as well be argued that the composition here is determined by exchange with the overlying layer and is not a fixed quantity.

As discussed in section 6.3.2b, above, in a recent study, Willerslev et al. (2007) examined biological molecules in the silty ice from *Dye-3*, including DNA and amino acids. They concluded that organic material contained in that *Dye-3* ice originated in a boreal forest (remnants of diagnostic plants and insects were identified). This environment implies a very much warmer climate than at the present margin in *Greenland* (e.g., July temperatures at 1 km elevation above 10°C), and hence it also suggests a great antiquity for this material; no evidence suggests that MIS 5e in

Greenland was nearly this warm. Indeed, Willerslev et al. (2007) also inferred the age of the organic material and the age of exposure of the rock particles, using several methods. They concluded that a 450–800 ka age is most likely, although uncertainties in all four of their dating techniques prevented a definitive statement. This conclusion suggests that the bottom ice layer (the source of rock material in the overlying mixed layer) is much older than MIS 5e.

This evidence admits of two principal interpretations. One is that this material survived the MIS 5e deglaciation by being contained in permafrost. The second is that the MIS 5e deglaciation did not extend as far north as the Dye-3 site, and that local topography allowed ice to persist, isolated from the large-scale flow. This latter hypothesis (apparently favored by Willerslev et al., 2007) does not explain the several-hundred-thousand-year hiatus within the ice, however, or the purely interglacial composition of the entire basal ice, both of which favor the permafrost interpretation. (Both hypotheses can be modified slightly to allow short-distance ice-flow transport to the Dye-3 site; e.g., Clarke et al., 2005.)

Ice-sheets can also slide at their margins. Sliding near the modern margin of the *Greenland Ice Sheet* (e.g., Joughin et al., 2008a) provides a way to rapidly re-establish the ice sheet in deglaciated regions and to preserve soil or permafrost materials as the ice re-grows, as described next. Marginal regions of the *Greenland Ice Sheet* are thawed at the bottom and slide over the materials beneath (e.g., Joughin et al., 2008a)—on a thin film of water or possibly thicker water or soft sediments. During a time of cooling, sliding advances the ice margin more rapidly than would be possible if the ice were frozen to the bed. Furthermore, the sliding will bring to a given point ice that was deposited elsewhere and at higher elevation; subsequently, that ice may freeze to the bed. As discussed below (section 6.3.5b), widespread evidence shows a notable advance of the ice-sheet margin during the last few millennia. Regions near the ice-sheet margin, and icebergs calving from that margin, now contain ice that was deposited somewhere in the accumulation zone at higher elevation and that slid into position (e.g., Petrenko et al., 2006). Were sliding not present, one might expect that re-glaciation of a site such as *Dye-3* would have required cooling until the site became an accumulation zone, followed by slow buildup of the ice sheet.

In contrast to all the preceding information from south-, northwest-, and east-*Greenland* ice cores, the ice cores from central *Greenland* (the *GISP2* and *GRIP* cores; Suwa et al., 2006) and north-central Greenland (the *NGRIP* core) do contain MIS 5e ice that is normal, cold-environment, ice-sheet ice. Unfortunately, none of these cores contains a complete or continuous MIS 5e chronology. Layering of the *GISP2* and *GRIP* cores is disrupted by ice flow (Alley et al., 1995b) and, in the *NGRIP* core, basal melting has removed the early part of MIS 5e and any older ice (Dahl-Jensen et al., 2003). The central Greenland cores do reveal two important facts: MIS 5e was warmer than MIS 1 (oxygen isotope ratios were 3.3‰ higher than modern ones), and the elevation in the center of the ice sheet was similar to that of the modern ice sheet, although the ice sheet was probably slightly thinner in MIS 5e (within a few hundred meters of elevation, based on the total gas content). Thus, if we consider also evidence from the other cores, the ice sheet shrank substantially under a warm climate, but it persisted in a narrower, steeper form.

What climate conditions were responsible for driving the ice sheet into this configuration? The answer is not clear. None of the paleoclimate proxy information is continuous over time, both precipitation and temperature changes are important, and some factors related to ice flow are poorly constrained. Cuffey and Marshall (2000; also see Marshall and Cuffey, 2000) were the first to address this question using the information from the central Greenland cores as constraints. In particular, Cuffey and Marshall (2000) noted that oxygen isotope ratios were at least 3.3‰ higher during MIS 5e, and they used this value to constrain the climate forcing on an ice sheet model. Because the isotopic composition depends on the elevation of the ice-sheet surface as well as on temperature change at a constant elevation, these analyses generated both climate histories and ice-sheet histories. Results depended critically on the isotopic sensitivity parameter relating isotopic composition to temperature and on the way past accumulation rates are estimated, which have large uncertainties. Furthermore, there was no attempt to model increased flow in response to changes of calving margins, or increased flow in response to production of surface meltwater (see Lemke et al., 2007). Thus, the ice sheet model was conservative; a given climatic temperature change produced a smaller response in the modeled ice sheet than is expected in nature.

In the reconstruction favored by Cuffey and Marshall (isotopic sensitivity α = 0.4‰ per °C), the southern dome of Greenland completely melted after a sustained (for at least 2,000 years) climate warming (mean annual, but with summer most important) of approximately 7°C higher than present. In a different scenario (sensitivity α = 0.67‰ per °C), the southern ice sheet margin did not retreat past Dye-3 after a sustained warming of 3.5°C. Thus an intermediate scenario (sustained warming of 5°–6°C) is required, in this view, to cause the margin to retreat just to Dye-3. Given the conservative representation of ice dynamics in the model, a smaller sustained warming would in fact be sufficient to accomplish such a retreat. How much smaller is not known, but it could be quite small. Outflow of ice can increase by a factor of two in response to modest changes in air and ocean temperatures at the calving margins (see Lemke et al., 2007).

Mass balance depends on numerous variables that are not modeled, introducing much uncertainty. Examples of these variables are storm-scale weather controls on the warmest periods within summers, similar controls on annual snowfall, and increased warming due to exposure of dark ground as the ice sheet retreats. In contrast to the under-representation of ice dynamics, however, no major observations show that the models are fundamentally in error with respect to surface mass-balance forcings.

A hint of a serious error is, however, provided by the record of accumulation rate from central *Greenland*. During the past about 11,000 years (MIS 1) variations in snow accumulation and in temperature show no consistent correlation (Cuffey and Clow, 1997; Kapsner et al., 1995), whereas most models assume that snowfall (and hence accumulation) will increase with temperature. This lack of correlation suggests that models are over-predicting the extent to which increased snowfall will partly balance increased melting in a warmer climate. If this MIS 1 situation in central Greenland applied to much of the ice sheet in MIS 5e, then models would require less warming to match the reconstructed ice-sheet footprint. Again, the real ice sheet appears to be more vulnerable than the model ones. We refer to this observation as only a “hint” of a problem, however, because snowfall on the center of *Greenland* may not represent snowfall over the whole ice sheet, for which other climatological influences come into play.

The climate forcing for the Cuffey and Marshall (2000) ice dynamics model, like that of most recent models that explore Greenland's glacial history, is driven by a single paleoclimate record, the isotope-based surface temperature at the Summit ice core sites. From this information, temperature and precipitation fields are derived and then combined to obtain a mass balance forcing over space and time, which is then applied to the entire ice sheet. This approach can be criticized for eliminating all local-scale climate variability, but few observations would allow such variability to be adequately specified.

Recent efforts to estimate the minimum MIS 5e ice volume for *Greenland* have much in common with the Cuffey and Marshall (2000) approach, but they focus on adding observational constraints that optimize the model parameters. For example, the new ability to model the movement of materials passively entrained in ice sheets (Clarke and Marshall, 2002) now allows the predicted and observed isotope profiles at ice core sites to be compared. By using these capabilities, Tarasov and Peltier (2003) produced new estimates of MIS 5e ice volume that were constrained by the measured ice-temperature profiles at *GRIP* and *GISP2* and by the $\delta^{18}O$ profiles at *GRIP*, *GISP2*, and *NorthGRIP*. Their conservative estimate is that the *Greenland Ice Sheet* contributed enough meltwater to cause a 2.0–5.2 m rise in MIS 5e sea level; the more likely range is 2.7–4.5 m—lower than the 4.0–5.5 m estimate of Cuffey and Marshall (2000).

Ice-core sites closer to the ice sheet margins, such as *Camp Century* and *Dye-3*, better constrain ice extent than do the central Greenland sites (Lhomme et al., 2005). These authors added a tracer transport capability to the model used by Marshall and Cuffey (2000) and attempted to optimize the model fit to the isotope profiles at *GRIP*, *GISP2*, *Dye-3* and *Camp Century*. For now, their estimate of a 3.5–4.5 m maximum MIS

5e sea-level rise attributable to meltwater from the *Greenland Ice Sheet* is the most comprehensive estimate based on this technique (Lhomme et al., 2005).

The discussion just previous rested on interpretation of paleoclimatic data from the central Greenland ice cores to drive a model to match the inferred ice-sheet “footprint” (and sometimes other indicators) and thus learn volume changes in relation to temperature changes. An alternative approach is to use what we know about climate forcings to drive a coupled ocean-atmosphere climate model and then test the output of that model against paleoclimatic data from around the ice sheet. If the model is successful, then the modeled conditions can be used over the ice sheet to drive an ice-sheet model to match the reconstructed ice-sheet footprint. From response to forcing changes we then learn volume changes. This latter approach avoids the difficulty of inferring the “α” parameter relating isotopic composition of ice to temperature, and of assuming a relation between temperature and snow accumulation, although this latter approach obviously raises other issues. The latter approach was used by Otto-Bliesner et al. (2006; also see Overpeck et al., 2006).

The primary forcings of Arctic warmth during MIS 5e are the seasonal and latitudinal changes in solar insolation at the top of the atmosphere associated with periodic, cyclical changes in Earth’s orbit (Berger, 1978). Earth’s orbit varies in its obliquity (the inclination of Earth’s spin axis to the orbital plane, which peaked at about 130 ka), eccentricity (the out-of-roundness of Earth’s elliptical orbit around the Sun), and precession (the timing of closest approach to the Sun on the elliptical orbit relative to hemispheric seasons). The net effect of these factors was anomalously high summer insolation in the Northern Hemisphere during the first half of this interglacial (about 130–

123 ka) (Otto-Bliesner et al., 2006; Overpeck et al., 2006). Atmosphere-Ocean General Circulation Models of the climate (AOGCMs) have used the MIS 5e seasonal and latitudinal insolation changes to calculate both the seasonal temperatures and precipitation of the atmosphere, as well as changes to sea ice and ocean temperatures. These models simulate approximately correct sensitivity to the MIS 5e orbital forcing. They reproduce the proxy-derived summer warmth for the Arctic of up to 5°C, and they place the largest warming over northern Greenland, northeast Canada, and Siberia (CAPE, 2006; Jansen et al., 2007).

In one of the models that has been extensively analyzed, the NCAR CCSM (National Center for Atmospheric Research Community Climate System Model), the orbitally induced warmth of MIS 5e caused loss of snow and sea ice, which in turn caused positive albedo feedbacks that reduced reflection of sunlight (Otto-Bliesner et al., 2006). The insolation anomalies increased sea-ice melting early in the northern spring and summer seasons, and reduced the extent of Arctic sea ice from April into November. The simulated reduced summer sea ice allowed the North Atlantic to warm, particularly along coastal regions of the Arctic and the surrounding waters of *Greenland*. Feedbacks associated with the reduced sea ice around *Greenland* and decreased snow depths on *Greenland* further warmed *Greenland* during the summer months. In combination with simulated precipitation rates, which overall were not substantially different from present rates, the simulated mass balance of the *Greenland Ice Sheet* resulting from the model was negative**.** Then, as now, the surface of the ice sheet melted primarily in the summer.

The NCAR CCSM model has a mid-range climate sensitivity among comprehensive atmosphere-ocean models; that is, this model generates mid-range warming in response to doubling of CO_2 or other specified forcing (Kiehl and Gent, 2004). Temperatures and precipitation produced by the NCAR CCSM model for 130 ka were then used to drive an ice-flow model. (The model used an updated version of that used by Cuffey and Marshall (2000), and thus it also lacked representations of some physical processes that would accelerate ice-sheet response and increase sensitivity to climate change.) The ice-flow model simulated the likely configuration of the MIS 5e *Greenland Ice Sheet*, for comparison with paleoclimatic data on ice-sheet configuration. In this model, the *Greenland Ice Sheet* proved sensitive to the warmer summer temperatures when melting was taking place. Increased melting outweighed the increase in snowfall. For all but the summit of *Greenland* and isolated coastal sites, increased rates of melting and the extended ablation season led to a negative mass balance in response to the orbitally induced changes in temperature and snowfall. As the simulated ice sheet retreated for several millennia, the loss of ice mass lowered the surface of the *Greenland Ice Sheet*, which amplified the negative mass-balance and accelerated retreat. The *Greenland Ice Sheet* responded to the seasonal orbital forcings because it is particularly sensitive to warming in summer and autumn, rather than in winter when temperatures are too cold for melting. The modeled *Greenland Ice Sheet* melted in response to both direct effects (warmer atmospheric temperatures) and indirect effects (reduction of its altitude and size).

The simulated MIS 5e *Greenland Ice Sheet* was a steep-sided ice sheet in central and northern *Greenland* (Otto-Bliesner et al., 2006) (Figure 6.7). The model did not incorporate feedbacks associated with the exposure of bedrock as the ice sheet retreated, potential meltwater-driven or ice-shelf-driven ice-dynamical processes, or time-evolving orbital forcing, so the model was probably less sensitive and more slowly responsive to warming than the real ice sheet, as noted just above. The lateral extent of the modeled minimal *Greenland Ice Sheet* was constrained by ice core data (see above). If the *Greenland Ice Sheet*'s southern dome did not survive the peak interglacial warmth, as suggested by those data (Koerner and Fisher, 2002; Lhomme et al., 2005), then the model suggests that the *Greenland Ice Sheet* contributed enough meltwater to account for 1.9–3.0 m of sea-level rise (another 0.3–0.4 m rise was produced by meltwater from ice on Arctic Canada and Iceland) for several millennia during the last interglacial. The evolution through time of the *Greenland Ice Sheet*'s retreat and the linked rate at which sea level rose cannot be constrained by paleoclimatic observational data or current ice-sheet models. Furthermore, because the ice-sheet model was forced by conditions appropriate for 130 ka rather than being forced by more realistic, slowly time-varying conditions, the details of the modeled time-evolution of the *Greenland Ice Sheet* are not expected to exactly match reality. Sensitivity studies that set melting of the *Greenland Ice Sheet* at a more rapid rate than suggested by the ice-sheet model indicate that the meltwater added to the North Atlantic was not sufficient to induce oceanic and other climate changes that would have inhibited melting of the *Greenland Ice Sheet* (Otto-Bliesner et al., 2006).

FIGURE 6.7 NEAR HERE

The atmosphere-ocean modeling driven by known forcings produces reconstructions that match many data from around *Greenland* and the Arctic. The earlier work of Cuffey and Marshall (2000) had found that a very warm and snowy MIS 5e, or a more modest warming with less increase in snowfall, could be consistent with the data, and the atmosphere-ocean model favors the more modest temperature change. (The results of the different approaches, although broadly compatible, do not agree in detail, however.) The Otto-Bliesner et al. (2006) modeling leads to a somewhat smaller sea-level rise from melting of the *Greenland Ice Sheet* than does the earlier work of Cuffey and Marshall (2000). A temperature rise of 3°–4°C and a sea-level rise of 3–4 m may be consistent with the data, with notable uncertainties.

Considering all of the efforts summarized above, as little as 1–2 m or as much as 4–5 m of ice may have been removed from the *Greenland Ice Sheet* during MIS 5e, in response to climatic temperature changes of perhaps 2°–7°C. At least the higher numbers for the warming are based on estimates that include the feedbacks from melting of the ice sheet. Central values in the 3–4 m and 3°–4°C range may be appropriate.

6.3.4 Post-MIS 5e Cooling to the Last Glacial Maximum (LGM, or MIS 2)

6.3.4a Climate forcing

Both climate and ice-sheet reconstructions become more confident for times younger than MIS 5e. The climatic records derived from ice cores are especially good. The *Greenland* ice cores, primarily from the *GRIP*, *NGRIP*, and *GISP2* cores but also from *Camp Century*, *Dye-3*, and *Renland* cores, provide what are probably the most reliable paleoclimatic records of any sites on Earth (e.g., Cuffey et al., 1995; Dahl-Jensen et al., 1998; Johnsen et al., 2001; Jouzel et al., 1997; Severinghaus et al., 1998).

The paleoclimate information derived from near-field marine records is less robust. Because sediment accumulated rapidly in depositional centers adjacent to glaciated margins, relatively few cores span all of the last 130,000 years. In core HU90-013 (Figure 6.8) from the Eirik Drift (Stoner et al., 1995), rapid sedimentation buried the sediments from MIS 5e to about 13 m depth. At that site, the $\delta^{18}O$ of planktonic foraminiferal shells changes markedly from MIS 5e to 5d.The change, of close to 1.5‰, is consistent with cooling as well as ice growth on land, and it is associated with a rapid increase in magnetic susceptibility that indicates delivery of glacially derived sediments.

FIGURE 6.8 NEAR HERE

The broad picture, which is based on ice-core, far-field and near-field marine records, and more, indicates the following for climatic conditions most relevant to the *Greenland Ice Sheet*:

- a general cooling from MIS 5e (about 123 ka) to MIS 2 (coldest temperatures were at about 24 ka; Alley et al., 2002),
- warming to the mid-Holocene/MIS 1 a few millennia ago,
- cooling into the Little Ice Age of one to a few centuries ago,
- and then a bumpy warming (see section 6.3.5b, below).

The cooling trend from MIS 5e involved temperature minima in MIS 5d, 5b, and 4 before reaching the coldest of these minima in MIS 2, with maxima in MIS 5c, 5a, and 3.

Throughout the cooling from MIS 5e to MIS 2, and the subsequent warming into MIS 1 (the Holocene), shorter-lived “millennial” events occurred. During these events, central *Greenland* warmed abruptly—roughly 10°C in a few years to decades—cooled gradually, then cooled more abruptly, gradually warmed slightly, and then repeated the sequence (Figure 6.9) (also see Alley, 1998). The abrupt coolings were usually spaced about 1500 years apart, although longer intervals are often observed (e.g., Alley et al., 2001; Braun et al., 2005).

FIGURE 6.9 NEAR HERE

Marine sediment cores from around the North Atlantic and beyond show temperature histories closely tied to those recorded in *Greenland* (Bond et al., 1993). Indeed, the *Greenland* ice cores appear to have recorded quite clearly the template for millennial climate oscillations around much of the planet (although that template requires a modified seesaw in far-southern regions (Figure 6.9) (Stocker and Johnsen, 2003)).

Closer to the ice sheet, marine cores display strong oscillations that correlate in time with that template, but with more complexity in the response (Andrews, 2008). Figure 6.10, panel A shows data from a transect of cores (Andrews, 2008) and compares the marine near-surface isotopic variations with $\delta^{18}O$ data from the *Renland* ice core, just inland from *Scoresby Sund* (Johnsen et al., 1992a; 2001) (Figure 6.8). The complexity observed in this comparison likely arises because of the rich nature of the marine indicators. As noted in section 6.2.1c, above, the oxygen isotope composition of surface-dwelling foraminiferal shells becomes lighter when the temperature increases and also when meltwater supply is increased to the system (or meltwater removal is reduced). If cooling is caused by freshwater-induced reduction in the formation of deep water, then one may observe either heavier or lighter isotopic ratios, depending on whether the core primarily reflects the temperature change or the freshwater change. Some of the signals in Figure 6.10, panel A likely involve delivery of additional meltwater (which could have had various sources, such as melting of icebergs) to the vicinity of the core during colder times.

FIGURE 6.10 NEAR HERE

The slower tens-of-millennia cycling of the climate records is well explained by features of Earth's orbit and by associated influences of Earth-system response to the orbital features (especially changes in atmospheric CO_2 and other greenhouse gases, ice-albedo feedbacks, and effects of changing dust loading), and strongly modulated by the response of the large ice sheets (e.g., Broecker, 1995). The faster changes are rather clearly linked to switches in the behavior of the North Atlantic (e.g., Alley, 2007): colder intervals mark times of more-extensive wintertime sea ice, and warmer intervals mark times of lesser sea ice (Denton et al., 2005). These links are in turn coupled to changes in deep-water formation in the North Atlantic and thus to "conveyor-belt" circulation (e.g., Broecker, 1995; Alley, 2007). (Note that a fully quantitative mechanistic understanding of forcing and response of these faster changes is still being developed; e.g., Stastna and Peltier, 2007.)

Of particular interest relative to the ice sheets is the observation that iceberg-rafted debris is much more abundant throughout the North Atlantic during some cold intervals, called Heinrich events (Figure 6.9). The material in this debris is largely tied to sources in Hudson Bay and Hudson Strait at the mouth of Hudson Bay, and thus to the

North American *Laurentide Ice Sheet*, but it also contains other materials from almost everywhere around the North Atlantic (Hemming, 2004).

6.3.4b Ice-sheet changes

With certain qualifications, the behavior of the *Greenland Ice Sheet* during this interval was closely tied to the climate: the ice sheet expanded with cooling and retreated with warming. Records are generally inadequate to assess response to millennial changes, and dating is typically sufficiently uncertain that lead-or-lag relations cannot be determined with high confidence, but colder temperatures were accompanied by more-extensive ice.

Furthermore, with some uncertainty, the larger footprint of the *Greenland Ice Sheet* during colder times corresponded with a larger ice volume. This conclusion emerges both from limited data on total gas content of ice cores (Raynaud et al., 1997) indicating small changes in thickness, and from physical understanding of the ice-flow response to changing temperature, accumulation rate, ice-sheet extent, and other changes in the ice. As described in section 6.1.2, above, the retreat of ice-sheet margins tends to thin central regions, whereas the advance of margins tends to thicken central regions. Moreover, because ice thickness in central regions is relatively insensitive to changes in accumulation rate (or other factors), marginal changes largely dominate the ice-volume changes.

The best records of ice-sheet response during the cooling into MIS 2 are probably those from the *Scoresby Sund* region of east *Greenland* (Funder et al., 1998). These records indicate

- ice advances during the coolings of MIS 5d and 5b that did not fully fill the *Scoresby Sund* fjord,
- retreats during the relatively warmer MIS 5c and 5a (although 5c and 5a were colder than MIS 5e or MIS 1; e.g., Bennike and Bocher, 1994),
- advance to the mouth of *Scoresby Sund*, probably during MIS 4,
- and remaining there into MIS 2, building the extensive moraine at the mouth of the Sund.

Whether ice advanced beyond the mouth of the Sund during this interval remains unclear. Most reconstructions place the ice edge very close to the mouth (e.g., Dowdeswell et al., 1994a; Mangerud and Funder, 1994). However, the recent work of Hakansson et al. (2007) indicates wet-based ice on the south side of the mouth of the Sund at a site that is 250 m above modern sea level at the Last Glacial Maximum (MIS 2). Such a position almost certainly requires ice advance past the mouth. Seismic studies and cores on the *Scoresby Sund* trough-mouth fan offshore indicate that, on the southern portion of the fan, debris flows have been deposited fairly recently, whereas on the northern portion this activity pre-dates MIS 5 (O'Cofaigh et al., 2003). It is not clear how such debris flow activity occurred unless the ice had advanced well onto the shelf (O'Cofaigh et al., 2003).

To the south of *Scoresby Sund*, at *Kangerdlugssuaq*, ice extended to the edge of the continental shelf during about 31–19 ka (Andrews et al., 1997, 1998a; Jennings et al., 2002a). These data, combined with widespread geomorphic evidence that ice reached the shelf break around south *Greenland*, are then the primary evidence for extensive ice cover of this age in southern *Greenland* (Funder et al., 2004; Weidick et al., 2004).

In the Thule region of northwestern *Greenland*, the data are consistent both with the broad climate picture (the MIS 5e to MIS 2 sequence) and with ice-sheet response as in *Scoresby Sund* (advances in colder MIS 5d, 5b, 4 (about 59–73 ka) and especially MIS 2, retreats in warmer 5c and 5a, possibly in MIS 3 (about 24–59 ka), and surely in MIS 1; see Figure 6.6 for general chronology) (Kelly et al., 1999). However, the dating is not secure enough to insist on much beyond the warmth of MIS 5e (marked by retreated ice), the cold of MIS 2 (marked by notably expanded ice), and the ice's subsequent retreat.

The extent of ice at the glacial maximum also remains in doubt in the northwestern part of the *Greenland Ice Sheet*. The submarine moraines at the edge of the continental shelf are poorly dated. Ice from *Greenland* did merge with that from *Ellesmere Island*, thus joining the great *Greenland Ice Sheet* with the Innuitian sector of the North American *Laurentide Ice Sheet* (England, 1999; Dyke et al., 2002). However, whether ice advanced to the edge of the continental shelf in widespread regions to the north and south of the merger zone is poorly understood (Blake et al., 1996; Kelly et al., 1999). A recent reconstruction (Funder et al., 2004) favors advance of grounded ice to the shelf edge in the northwest, merging with North American ice, and with the merged ice spreading to the northeast and southwest along what is now *Nares Strait* to feed ice shelves extending toward the Arctic Ocean and *Baffin Bay*. The lack of a high marine limit just south of *Smith Sund* (Sound) in the northwest is prominent in that interpretation—more-extensive ice would have pushed the land down more and allowed the ocean to advance farther inland following deglaciation, and then subsequent isostatic uplift would have raised the marine deposits higher. But, a trade-off does exist between slow retreat and small retreat in controlling the marine limit. This trade-off has been explored by some workers (e.g., Huybrechts, 2002; Tarasov and Peltier, 2002), but the relative sea-level data are not as sensitive to the earlier part (about 24 ka) as to the later, and so strong conclusions are not available.

Thus, the broad picture of ice advance in cooling conditions and ice retreat in warming conditions is quite clear. Remaining issues include the extent of advance onto the continental shelf (and if it was limited, why), and the rates and times of response.

We will look first at ice extent. The generally accepted picture has been one of expansion to the edge of the continental shelf in the south, much more limited expansion in the north, and a transition somewhere between *Kangerdlugssuaq* and *Scoresby Sund* on the east coast (Dowdeswell et al., 1996). On the west coast, the moraines that typically lie 30–50 km beyond the modern coastline (and even farther along troughs) are usually identified with MIS 2. The shelf-edge moraines (usually called Hellefisk moraines and usually roughly twice as far from the modern coastline as the presumably MIS 2 moraines) are usually identified with MIS 6, although few solid dates are available (Funder and Larsen, 1989). On the east coast, the evidence from the mouth of *Scoresby Sund* and the trough-mouth fan, noted above in this section, opens the possibility of more-extensive ice there than is indicated by the generally accepted picture; ice may have extended to the mid-shelf or the shelf edge. Similarly, the work of Blake et al. (1996) in *Greenland*'s far northwest may indicate that ice reached the shelf edge. The indications of Blake et al. (1996) are geomorphically consistent with wet-based ice. The increasing realization that cold-based ice is sometimes extensive yet geomorphically inactive (e.g., England, 1999) further complicates interpretations. No evidence overturns the conventional view of expansion to the shelf-edge in the south, expansion to merge with

North American ice in the northwest, and expansion onto the continental shelf but not to the shelf-edge elsewhere. Thus, this interpretation is probably favored, but additional data would clearly be of interest.

Glaciological understanding indicates that ice sheets almost always respond to climatic or other environmental forcings (such as sufficiently large sea-level change). The most prominent exception may be advance to the edge of the continental shelf under conditions that would allow further advance if a huge topographic step in the sea floor were not present. (Similarly, ice may not respond to relatively small climate changes, such as during the advance stage of the tidewater-glacier cycle (Meier and Post, 1987)). If this assessment is accurate, and if the *Greenland Ice Sheet* at the time of the Last Glacial Maximum terminated somewhere on the continental shelf rather than at the shelf edge around part of the coastline, then glaciological understanding indicates that the ice sheet should have responded to short-lived climate changes.

The near-field marine record is consistent with such fluctuations, as discussed next. However, owing to the complexity of the controls on the paleoclimatic indicators, unambiguous interpretations are not possible.

Several marine sediment cores extend back through MIS 3 and even into MIS 4 (the cores were obtained from *Baffin Bay*, the *Eirik Drift* off southwestern *Greenland*, the *Irminger* and Blosseville Basins (e.g., cores SU90-24 & PS2264, Figure 6.8), and from the Denmark Strait) (Figure 6.8). In many of those cores, the $\delta^{18}O$ of near-surface planktic foraminifers varies widely during MIS 3. These variations were initially documented by Fillon and Duplessy (1980) in cores HU75-041 and -042 from south of *Davis Strait* (Figures 6.8 and 6.10, panel B), and this documentation preceded the recognition of large millennial oscillations (Dansgaard-Oeschger or D-O events; Johnsen at al., 1992b, Dansgaard et al., 1993) in the *Greenland* ice core records. In addition, Fillon and Duplessy (1980) also contributed information on the down-core numbers of volcanic-ash (tephra) shards in these two cores. These authors identified “Ash Zone B” in core HU75-042, which is correlated with the North Atlantic Ash Zone II, for which the current best-estimate age is about 54 ka (Figure 6.10B; it is associated with the end of interstadial 15 as identified by Dansgaard et al., 1993). Subsequent work, especially north and south of *Denmark Strait*, has also shown large oscillations in planktonic foraminiferal $\delta^{18}O$ (Elliott et al., 1998; Hagen, 1999; van Kreveld et al., 2000; Hagen and Hald, 2002). As noted in section 6.3.4a, above, and shown in Figure 6.10A, the transect of cores appears to show both climate forcing and ice-sheet response in the millennial oscillations, although strong conclusions are not possible.

Cores from the *Scoresby Sund* and *Kangerdlugssuaq* trough mouth fans, two of the major outlets of the eastern *Greenland Ice Sheet*, also have distinct layers that are rich in ice-rafted debris (Stein et al., 1996; Andrews et al., 1998a; Nam and Stein, 1999). Cores HU93030-007 and MD99-2260 from the *Kangerdlugssuaq* trough-mouth fan (Dunhill, 2005) (Figure 6.8) consist of alternating layers with more and less ice-rafted debris that overlie a massive debris flow. Material above the debris flow is dated about 35 ka. The debris-rich layers have radiocarbon dates that are approximately coeval with Heinrich events 3 and 2 (Figure 6.9). On the *Scoresby Sund* trough-mouth fan, Stein et al (1996) also recorded intervals rich in ice-rafted debris that they quantified by counting the number of clasts greater than 2 mm as observed on X-rays. Although these cores are not as well dated as many from sites south of the Scotland-Greenland Ridge, they do indicate that such debris was delivered to the fan in pulses that may be approximately coeval with the North Atlantic Heinrich events.

Although several reports have invoked the Iceland Ice Sheet as a major contributor to North Atlantic sediment (Bond and Lotti, 1995; Elliot et al., 1998; Grousset et al., 2001), Farmer et al. (2003) and Andrews (2008) have questioned this assertion. They argue that the eastern *Greenland Ice Sheet* has been an ignored source of ice-rafted debris in the eastern North Atlantic south of the Scotland-Greenland Ridge. In particular, Andrews (2008) argued that the data from *Iceland* and *Denmark Strait* precluded any Icelandic contribution for Heinrich event 3. As noted by Huddard et al (2006), the area of the Iceland Ice Sheet during the Last Glacial Maximum was only 200,000 km^2 with an annual loss of ~600 km^3, and only ~150 km^3 of this loss was associated with calving. This is less than one-half the estimated calving rate of the present day *Greenland Ice Sheet* (Reeh, 1985).

The marine evidence from the western margin of the *Greenland Ice Sheet* for fluctuations of the ice sheet during MIS 3 is confounded by two facts: there are no published chronologies from the trough-mouth fan off *Disko Island*, and the stratigraphic record from *Baffin Bay* consists of glacially derived sediments from the *Greenland Ice Sheet* and from the *Laurentide Ice Sheet* including its Innuitian section (Dyke et al., 2002). Evidence for major ice-sheet events during MIS 3 is abundant, as is seen throughout *Baffin Bay* in layers rich in carbonate clasts transported from adjacent continental rocks (Aksu, 1985; Andrews et al., 1998b; Parnell et al., 2007) (Figure 6.11).

FIGURE 6.11 NEAR HERE

Core PS1230 from Fram Strait, which records the export of sediments from ice sheets around the Arctic Ocean (Darby et al., 2002), shows ice-rafted debris intervals associated with major contributions from north *Greenland* about 32, 23, and 17 ka. These debris intervals correspond closely in timing with ice-rafted debris events from the Arctic margins of the *Laurentide Ice Sheet.*

The fact that ice-rafted debris does not directly indicate ice-sheet behavior presents a continuing difficulty. Iceberg rafting of debris at an offshore site may increase owing to several possible factors: faster flow of ice from an adjacent ice sheet; flow of ice containing more clasts; loss of an ice shelf (most ice shelves experience basal melting, tending to remove debris in the ice, so ice-shelf loss would allow calving of bergs bearing more debris); cooling of ocean waters that allows icebergs—and their debris—to reach a site, loss of extensive coastal sea ice that allows icebergs to reach sites more rapidly (Reeh, 2004), alterations in currents or winds that control iceberg drift tracks, or other changes. The very large changes in volume of incoming sediment from the North American *Laurentide Ice Sheet* during Heinrich events (Hemming, 2004) are generally interpreted to be true indicators of ice-dynamical changes (e.g., Alley and MacAyeal, 1994), but even that is debated (e.g., Hulbe et al., 2004). Thus, the marine-sediment record is consistent with *Greenland* fluctuations in concert with millennial variability during the cooling into MIS 2. Moreover, trained observers have interpreted the records as indicating millennial oscillations of the *Greenland Ice Sheet* in concert with climate, but those fluctuations cannot be demonstrated uniquely.

6.3.5 Ice-Sheet Retreat from the Last Glacial Maximum (MIS 2)

6.3.5a Climatic history and forcing

As shown in Figure 6.9 (also see Alley et al., 2002), the coldest conditions recorded in *Greenland* ice cores since MIS 6 were reached about 24 ka, which corresponds closely in time with the minimum in local midsummer sunshine and with Heinrich Event H2. The suite of sediment cores from *Denmark Strait* (Figures 6.8 and 6.10A) plus data from other sediment cores (VM28-14 and HU93030-007) indicate that the most extreme values indicating Last Glacial Maximum in $\delta^{18}O$ of marine foraminifera occurred ~18–20 ka (slightly younger than the Last Glacial Maximum values in the ice cores) with values of 4.6‰ indicating cold, salty waters.

The "orbital" warming signal in ice-core records and other climate records is fairly weak until perhaps 19 ka or so (Alley et al., 2002). The very rapid onset of warmth about 14.7 ka (the Bølling interstadial) is quite prominent. However, more than a third of the total deglacial warming was achieved before that abrupt step, and that pre-14.7 ka orbital warming was interrupted by Heinrich event H1. Bølling warmth was followed by general cooling (punctuated by two prominent but short-lived cold events, usually called the Older Dryas and the Inter-Allerød cold period), before faster cooling led into the Younger Dryas about 12.8 ka. Gradual warming then occurred through the Younger Dryas, followed by a step warming at the end of the Younger Dryas about 11.5 ka. This abrupt warming was followed by ramp warming to above recent values by 9 ka or so, punctuated by the short-lived cold event of the Preboreal Oscillation about 11.2–11.4 ka (Bjorck et al., 1997; Geirsdottir et al., 1997; Hald and Hagen, 1998; Fisher et al., 2002; Andrews and Dunhill, 2004; van der Plicht et al., 2004; Kobashi et al., in press), and followed by the short-lived cold event about 8.3–8.2 ka (the “8k event”; e.g., Alley and Agustsdottir, 2005).

The cold times of Heinrich events H2, H1, the Younger Dryas, the 8k event, and probably other short-lived cold events including the Preboreal Oscillation are linked to greatly expanded wintertime sea ice in response to decreases in near-surface salinity and to the strength of the overturning circulation in the North Atlantic (see review by Alley, 2007). The cooling associated with these oceanic changes probably affected summers in and around *Greenland* (but see Bjorck et al., 2002 and Jennings et al., 2002a), but the changes were largest in wintertime (Denton et al., 2005).

Peak MIS 1/Holocene summertime warmth before and after the 8.2-ka event was, for roughly millennial averages, ~1.3^{o}C above late Holocene values in central *Greenland*, based on frequency of occurrence of melt layers in the *GISP2* ice core (Alley and Anandakrishnan, 1995), with mean-annual changes slightly larger although still smaller than ~2^{o}C (and with correspondingly larger wintertime changes); other indicators are consistent with this interpretation (Alley et al., 1999). Indicators from around *Greenland* similarly show mid-Holocene warmth, although with different sites often showing peak warmth at slightly different times (Funder and Fredskild, 1989). Peak Holocene warmth was followed by cooling (with oscillations) into the Little Ice Age. The ice-core data indicate that the century- to few-century-long anomalous cold of the Little Ice Age was ~1^{o}C or slightly more (Johnsen, 1977; Alley and Koci, 1990; Cuffey et al., 1994).

6*.3.5b Ice-sheet changes*

The *Greenland Ice Sheet* lost about 40% of its area (Funder et al., 2004) and a notable fraction of its volume (see below; also Elverhoi et al., 1998) after the peak of the last glaciation about 24–19 ka. These losses are much less than those of the warmer Laurentide and Fennoscandian Ice Sheets (essentially complete loss) and much more than those in the colder Antarctic.

The time of onset of retreat from the Last Glacial Maximum is poorly defined because most of the evidence is now below sea level. Funder et al. (1998) suggested that the ice was most extended in the *Scoresby Sund* area from about 24,000 to about 19,000 ka, on the basis of a comparison of marine and terrestrial data. This interval started at the coldest time in *Greenland* ice cores (which corresponds with the millennial Heinrich event H2) and extends to roughly the time when sea-level rise became notable because many ice masses around the world retreated (e.g., Peltier and Fairbanks, 2006).

Extensive deglaciation that left clear records is typically more recent. For example, a core from Hall Basin (core 79, Figure 6.8), the northernmost of a series of basins that lie between northwest *Greenland* and Ellesmere Island, has a date on hand-picked foraminifers of about 16.2 ka. This date implies that the land ice flowing to the Arctic Ocean had retreated by this time (Mudie et al., 2006). At *Sermilik Fjord* in southwest *Greenland*, retreat from the shelf preceded about 16 ka (Funder, 1989c). The ice was at the modern coastline or back into the fjords along much of the coast by approximately Younger Dryas time (13–11.5 ka, but with no implication that this position is directly linked to the climatic anomaly of the Younger Dryas) (Funder, 1989c; Marienfeld, 1992b; Andrews et al., 1996; Jennings et al., 2002b; Lloyd et al., 2005; Jennings et al., 2006). In the Holocene, the marine evidence of ice-rafted debris from the east-central *Greenland* margin (Marienfeld, 1992a; Andrews et al., 1997; Jennings et al., 2002a; Jennings et al., 2006) shows a tripartite record with early debris inputs, a middle-Holocene interval with very little such debris, and a late Holocene (neoglacial) period that spans the last 5–6 ka of steady delivery of such debris (Figure 6.12).

FIGURE 6.12 NEAR HERE

Along most of the *Greenland* coast, radiocarbon dates much older than the end of Younger Dryas time are rare, likely because of persistent cover by the *Greenland Ice Sheet*. Radiocarbon dates become common near the end of the Younger Dryas and especially during the Preboreal interval, and they remain common for all younger ages, indicating deglaciation (Funder, 1989a,b,c). The term “Preboreal” typically refers to the millennium-long interval following the Younger Dryas; the Preboreal Oscillation is a shorter-lived cold event within this interval, but the terminology has sometimes been used loosely in the literature. Owing to uncertainty about the radiocarbon “reservoir” age of the waters in which mollusks lived and other issues, it typically is not possible to assess whether a given date traces to the Preboreal Oscillation or the longer Preboreal. These uncertainties typically preclude linking a particular date with Preboreal or with Younger Dryas.

Given the prominence of the end of the Younger Dryas cold event in ice-core records (it was marked by a temperature increase of about 10°C in about 10 years; Severinghaus et al., 1998), it may seem surprising at first that widespread moraines abandoned in response to that warming have not been identified with confidence. Part of the difficulty is solved by the hypothesis of Denton et al. (2005), who argued that most of the warming occurred in winter. Bjorck et al. (2002) and Jennings et al. (2002a) argued for notable summertime warmth in *Greenland* during the Younger Dryas, but from Denton et al. (2005) and Lie and Paasche (2006), at least some warming or lengthening of the melt season probably occurred at the end of the Younger Dryas. The terminal Younger Dryas warming then would be expected to have affected glacier and ice-sheet behavior.

All ice-core records from *Greenland* show clearly that the temperature drop into the Younger Dryas was followed by a millennium of slow warming before the rapid warming at the end (Johnsen et al., 2001; North Greenland Ice Core Project Members, 2004). The slow warming perhaps reflected rising mid-summer insolation (a function of Earth's orbit) during that time. The Younger Dryas was certainly long enough for coastal mountain glaciers to reflect both the cooling into the event and the warming during the event before the terminal step. The ice-sheet margin probably would have been influenced by these changes as well (as discussed in section 6.3.4b, above, and in this section below). If the ice margin did advance with the cooling into the Younger Dryas, and did retreat during the Younger Dryas and its termination, then moraine sets would be expected from near the start of the Younger Dryas and from the cooling of the Preboreal Oscillation after the Younger Dryas (perhaps with minor moraines marking small events during the latter-Younger Dryas retreat). Because so much of the ice-sheet margin was marine at the start of the Younger Dryas, events of that age would not be recorded well.

Much study has focused on the spectacular late-glacial moraines of the *Scoresby Sund* region of east Greenland (Funder et al., 1998; Denton et al., 2005). Funder et al. (1998) suggested that the last resurgence of glaciers in the region, known as the Milne

Land Stade, was correlated with the Preboreal Oscillation, although a Younger Dryas age for at least some of the moraines, perhaps with both Preboreal Oscillation and Younger Dryas present, cannot be excluded (Funder et al., 1998; Denton et al., 2005). Data and modeling remain sufficiently sketchy that strong conclusions do not seem warranted, but the available results are consistent with rapid response of the ice to forcing, with warming causing retreat.

Retreat of the ice sheet from the coastline passed the position of the modern ice margin about 8 ka and continued well inland, perhaps more than 10 km in west *Greenland* (Funder, 1989c), up to 20 km in north *Greenland* (Funder, 1989b), and perhaps as much as 60 km in parts of south *Greenland* (Tarasov and Peltier, 2002). Reworked marine shells and other organic matter of ages 7–3 ka found on the ice surface and in younger moraines document this retreat (Weidick et al., 1990; Weidick, 1993). In west *Greenland*, the general retreat from the coast was interrupted by intervals during which moraines formed, especially about 9.5–9 ka and 8.3 ka (Funder, 1989c). These moraines are not all of the same age and are not, in general, directly traceable to the short-lived 8k cold event about 8.3–8.2 ka (Long et al., 2006). Timing of the onset of late Holocene readvance is not tightly constrained. Funder (1989c) suggested about 3 ka for west *Greenland*, the approximate time when relative a sea-level fall (from isostatic rebound of the land) switched to begin a relative sea-level rise of about 5 m (perhaps in part a response to depression of the land by the advancing ice load). Similar considerations place the onset of readvance somewhat earlier in the south, where relative sea-level fall switched to relative rise of about 10 m beginning about 8–6 ka (Sparrenbom et al., 2006a; 2006b).

The late Holocene advance culminated in different areas at different times, especially in the mid-1700s, 1850–1890, and near 1920 (Weidick et al., 2004). Since then, ice has retreated from this maximum.

Evidence of relative sea-level changes is consistent with this history (Funder, 1989d; Tarasov and Peltier, 2002; 2003; Fleming and Lambeck, 2004). Flights of raised beaches or other marine indicators are observed on many coasts of *Greenland*, and they lie as much as 160 m above modern sea level in west Greenland.

Fleming and Lambeck (2004) used an iterative technique to reconstruct the ice-sheet volume over time to match relative sea-level curves. They obtained an ice-sheet volume at the time of the Last Glacial Maximum about 42% larger than modern (3.1 m of additional sea-level equivalent in the ice sheet, compared with the modern value of 7.3 m of sea-level equivalent; interestingly, Huybrechts (2002) obtained a model-based estimate of 3.1 m of excess ice at the Last Glacial Maximum). Fleming and Lambeck (2004) estimated that 1.9 m of the 3.1 m of excess ice during the Last Glacial Maximum persisted at the end of the Younger Dryas. In their reconstruction, ice of the Last Glacial Maximum terminated on the continental shelf in most places, but it extended to or near the shelf edge in parts of southern Greenland, northeast Greenland, and in the far northwest where the *Greenland Ice Sheet* coalesced with the Innuitian ice from North America. Ice along much of the modern coastline was more than 500 m thick, and it was more than 1500 m thick in some places. Mid-Holocene retreat of about 40 km behind the present margin before late Holocene advance was also indicated. Rigorous error limits are not available, and modeling of the Last Glacial Maximum did not include the effects of the Holocene retreat behind the modern margin, so additional uncertainty is introduced.

In the ICE5G model, Peltier (2004) (with a Greenland Ice Sheet history based on Tarasov and Peltier, 2002) found that the relative sea-level data were inadequate to constrain Greenland ice-sheet volume accurately. In particular, these constraints provide only a partial history of the ice-sheet footprint and no information on the small—but nonzero—changes inland. Thus, Tarasov and Peltier (2002; 2003) and Peltier (2004) chose to combine ice-sheet and glacial isostatic adjustment modeling with relative-sea-level observations to derive a model of the ice-sheet geometry extending back to the Eemian (MIS 5e, about 125–130 ka). The previous ICE4G reconstruction had been characterized by an excess ice volume during the Last Glacial Maximum, relative to the present, of 6 m; this volume is reduced to 2.8 m in ICE5G. Later shrinkage of the *Greenland Ice Sheet* largely occurred in the last 10 ka in the ICE5G reconstruction, and proceeded to a mid-Holocene (7-6 ka) volume about 0.5 m less than at present, before regrowth to the modern volume.

The 20th century warmed from the Little Ice Age to about 1930, sustained warmth into the 1960s, cooled, and then warmed again since about 1990 (e.g., Box et al., 2006). The earlier warming caused marked ice retreat in many places (e.g., Funder, 1989a; 1989b; 1989c), and retreat and mass loss are now widespread (e.g., Alley et al., 2005). Study of declassified satellite images shows that at least for *Helheim Glacier* in the southeast of Greenland, the ice was in a retreated position in 1965, advanced after that during a short-lived cooling, and has again switched to retreat (Joughin et al., 2008b). This latest phase of retreat is consistent with global positioning system–based inferences of rapid melting in the southeastern sector of the *Greenland Ice Sheet* (Khan et al., 2007).

It is also consistent with GRACE satellite gravity observations, which indicate a mean mass loss in the period April 2002–April 2006 equivalent to 0.5 mm/yr of globally uniform sea-level rise (Velicogna and Wahr, 2006).

As discussed in section 6.2.2e, above, geodetic measurements of perturbations in Earth's rotational state can also help constrain the recent ice-mass balance. Munk (2002) suggested that length-of-day and true-polar-wander data were well fit by a model of ongoing glacial isostatic adjustment, and that this fit precluded a contribution from the *Greenland Ice Sheet* to recent sea-level rise. Mitrovica et al. (2006) reanalyzed the rotation data and applied a new theory of true polar wander induced by glacial isostatic adjustment. They found that an anomalous 20th-century contribution of as much as about 1 mm/yr of sea-level rise is consistent with the data; the partitioning of this value into signals from melting of mountain glaciers, Antarctic ice, and the *Greenland Ice Sheet* is non-unique. Interestingly, Mitrovica et al. (2001) analyzed a set of robust tide-gauge records and found that the geographic trends in the glacial isostatic adjustment–corrected rates suggested a mean 20th century melting of the *Greenland Ice Sheet* equivalent to about 0.4 mm/yr of sea-level rise.

6.4 Discussion

Glaciers and ice sheets are highly complex, and they are controlled by numerous climatic factors and by internal dynamics. Textbooks have been written on the controls, and no complete list is possible. The attribution of a given ice-sheet change to a particular cause is generally difficult, and it requires appropriate modeling and related studies.

It remains, however, that in the suite of observations as a whole, the behavior of the *Greenland Ice Sheet* has been more closely tied to temperature than to anything else. The *Greenland Ice Sheet* shrank with warming and grew with cooling. Because of the generally positive relation between temperature and precipitation (e.g., Alley et al., 1993), the ice sheet has tended to grow with reduced precipitation (snowfall) and to shrink when the atmospheric mass supply increased, so precipitation changes cannot have controlled ice-sheet behavior. However, local or regional events may at times have been controlled by precipitation.

The hothouse world of the dinosaurs and into the Eocene occurred with no evidence of ice reaching sea level in *Greenland*. The long-term cooling that followed is correlated in time with appearance of ice in *Greenland*.

Once ice appeared, paleoclimatic archives record fluctuations that closely match not only local but also widespread records of temperature, because local temperatures correlate closely with more-widespread temperatures. Because any ice-albedo feedback or other feedbacks from the *Greenland Ice Sheet* itself are too weak to have controlled temperatures far beyond *Greenland*, the arrow of causation cannot have run primarily from the ice sheet to the widespread climate.

One must consider whether something controlled both the temperature and the ice sheet, but this possibility appears unlikely. The only physically reasonable control would be sea level, in which warming caused melting of ice beyond *Greenland*, and the resultant sea-level rise forced retreat of the *Greenland Ice Sheet* by floating marginal regions and speeding iceberg calving and ice-flow spreading. However, data point to times when this explanation is not sufficient. There at least is a suggestion at MIS 6 that *Greenland* deglaciation led strong global sea-level rise, as described in section 6.3.2b, above. Ice expanded from MIS 5e to MIS 5d from a reduced ice sheet, which would have had little contact with the sea. Much of the retreat from the MIS 2 maximum took place on land, although fjord glaciers did contact the sea. Ice re-expanded after the mid-Holocene warmth against a baseline of very little change in sea level but in general with slight sea-level rise—opposite to expectations if sea-level controls the ice sheet. Similarly, the advance of Helheim Glacier after the 1960s occurred with a slightly rising global sea level and probably a slightly rising local sea level.

At many other times the ice-sheet size changed in the direction expected from sea-level control as well as from temperature control, because trends in temperature and sea level were broadly correlated. Strictly on the basis of the paleoclimatic record, it is not possible to disentangle the relative effects of sea-level rise and temperature on the ice sheet. However, it is notable that terminal positions of the ice are marked by sedimentary deposits; although erosion in *Greenland* is not nearly as fast as in some mountain belts such as coastal *Alaska*, notable sediment supply to grounding lines continues. And, as shown by Alley et al. (2007), such sedimentation tends to stabilize an ice sheet against the effects of relative rise in sea level. Although a sea-level rise of tens of meters could overcome this stabilizing effect, the ice would need to be nearly unaffected for many millennia by other environmental forcings, such as changing temperature, to allow that much sea-level rise to occur and control the response (Alley et al., 2007). Strong temperature control on the ice sheet is observed for recent events (e.g., Zwally et al., 2002; Thomas et al., 2003; Hanna et al., 2005; Box et al., 2006) and has been modeled (e.g., Huybrechts and de Wolde, 1999; Huybrechts, 2002; Toniazzo et al., 2004; Ridley et al., 2005; Gregory and Huybrechts, 2006).

Thus, it is clear that many of the changes in the ice sheet were forced by temperature. In general, the ice sheet responded oppositely to that expected from changes in precipitation, retreating with increasing precipitation. Events explainable by sea-level forcing but not by temperature change have not been identified. Sea-level forcing might yet prove to have been important during cold times of extensively advanced ice; however, the warm-time evidence of Holocene and MIS 5e changes that cannot be explained by sea-level forcing indicates that temperature control was dominant.

Temperature change may affect ice sheets in many ways, as discussed in section 6.1.2. Warming of summertime conditions increases meltwater production and runoff from the ice-sheet surface, and may increase basal lubrication to speed mass loss by iceberg calving into adjacent seas. Warmer ocean waters (or more-vigorous circulation of those waters) can melt the undersides of ice shelves, which reduces friction at the ice-water interface and so increases flow speed and mass loss by iceberg calving. In general, the paleoclimatic record is not yet able to separate these influences, which leads to the broad use of "temperature" in discussing ice-sheet forcing. In detail, ocean temperature will not exactly correlate with atmospheric temperature, so the possibility may exist that additional studies could quantify the relative importance of changes in ocean and in air temperatures.

Most of the forcings of past ice-sheet behavior considered here have been applied slowly. Orbital changes in sunshine, greenhouse-gas forcing, and sea level have all varied on 10,000-year timescales. Purely on the basis of paleoclimatic evidence, it is generally not possible to separate the ice-volume response to incremental forcing from the continuing response to earlier forcing. In a few cases, sufficiently high time resolution and sufficiently accurate dating are available to attempt this separation for ice-sheet area. At least for the most recent events during the last decades of the 20th century and into the 21st century, ice-marginal changes have tracked forcing, with very little lag. The data on ice-sheet response to earlier rapid forcing, including the Younger Dryas and Preboreal Oscillation, remain sketchy and preclude strong conclusions, but results are consistent with rapid temperature-driven response.

A summary of many of the observations is given in Figure 6.13, which shows changes in ice-sheet volume in response to temperature forcing from an assumed “modern” equilibrium (before the warming of the last decade or two). Error bars cannot be placed with confidence. A discussion of the plotted values and error bars is given in the caption to Figure 6.13. Some of the ice-sheet change may have been caused directly by temperature and some by sea-level effects correlated with temperature; the techniques used cannot separate them (nor do modern models allow complete separation; Alley et al., 2007). However, as discussed above in this section, temperature likely dominated, especially during warmer times when contact with the sea was reduced because of ice-sheet retreat. Again, no rates of change are implied. The large error bars on Figure 6.13 remain disturbing, but general covariation of temperature forcing and sea-level change from *Greenland* is indicated. The decrease in sensitivity to temperature with decreasing temperature also is physically reasonable; if the ice sheet were everywhere cooled to well below the freezing point, then a small warming would not cause melting and the ice sheet would not shrink.

FIGURE 6.13 NEAR HERE

6.5 Synopsis

Paleoclimatic data show that the *Greenland Ice Sheet* has changed greatly with time. Physical understanding indicates that many environmental factors can force changes in the size of an ice-sheet. Comparison of the histories of important forcings and of ice-sheet size implicates cooling as causing ice-sheet growth, warming as causing shrinkage, and sufficiently large warming as causing loss. The evidence for temperature control is clearest for temperatures similar to or warmer than recent temperatures (the last few millennia). Snow accumulation rate is inversely related to ice-sheet volume (less ice when snowfall is higher), and thus the snow-accumulation rate in general is not the leading control on ice-sheet change. Rising sea level tends to float marginal regions of ice sheets and force retreat, so the generally positive relation between sea level and temperature means that typically both reduce the volume of the ice sheet. However, for some small changes during the most recent millennia, marginal fluctuations in the ice sheet have been opposed to those expected from local relative sea-level forcing but in the direction expected from temperature forcing. These fluctuations, plus the tendency of ice-sheet margins to retreat from the ocean during intervals of shrinkage, indicate that sea-level change is not the dominant forcing at least for temperatures similar to or above those of the last few millennia. High-time-resolution histories of ice-sheet volume are not available, but the limited paleoclimatic data consistently show that short-term and long-term responses to temperature change are in the same direction. The best estimate from paleoclimatic data is thus that warming will shrink the *Greenland Ice Sheet*, and that warming of a few degrees is sufficient to cause ice-sheet loss. Tightly constrained numerical estimates of the threshold warming required for ice-sheet loss are not available, nor are rigorous error bounds, and rate of loss is very poorly constrained. Numerous opportunities exist for additional data collection and analyses that would reduce these uncertainties.

Figure 6.1 Satellite image (SeaWiFS) of the Greenland Ice Sheet and surroundings, from July 15, 2000 (http://www.gsfc.nasa.gov/gsfc/earth/pictures/earthpic.htm).

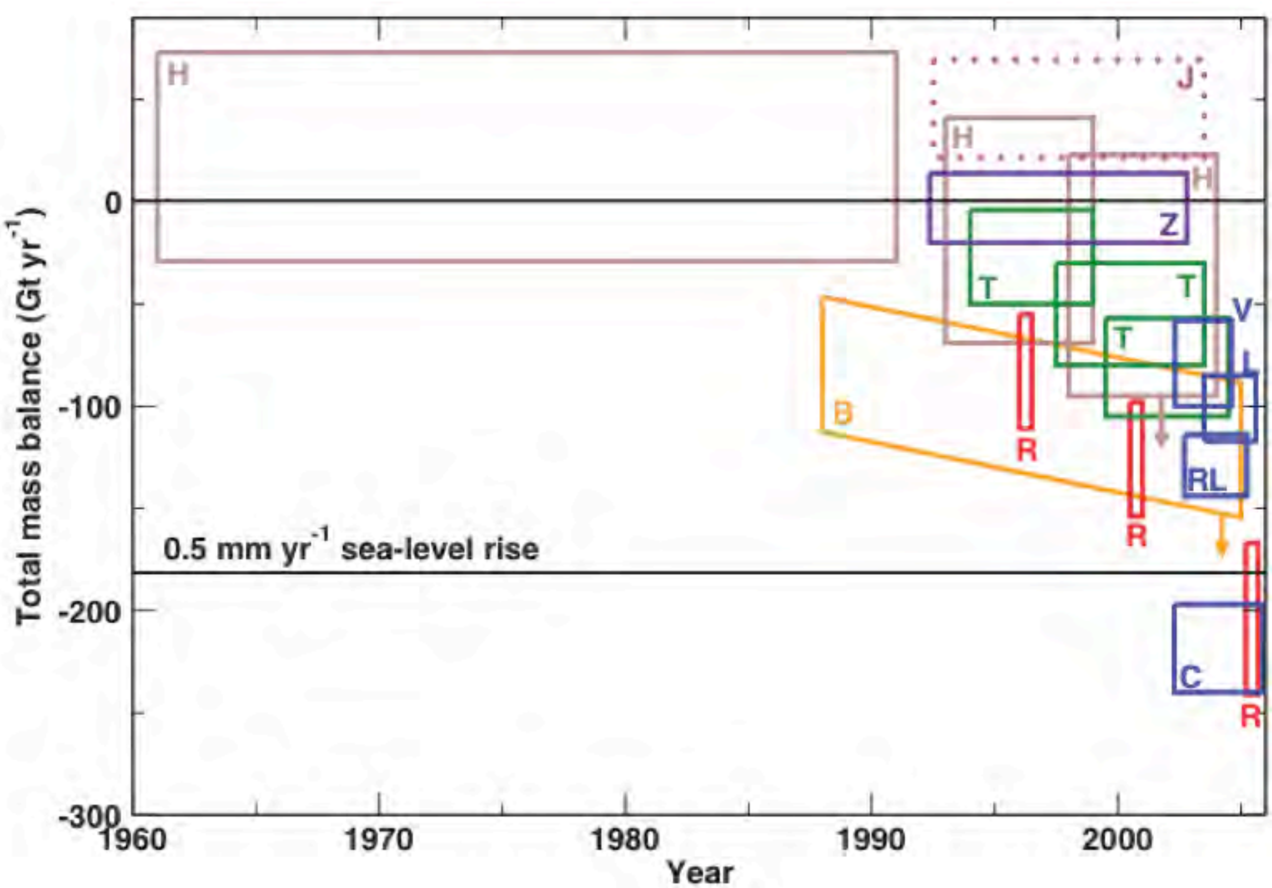
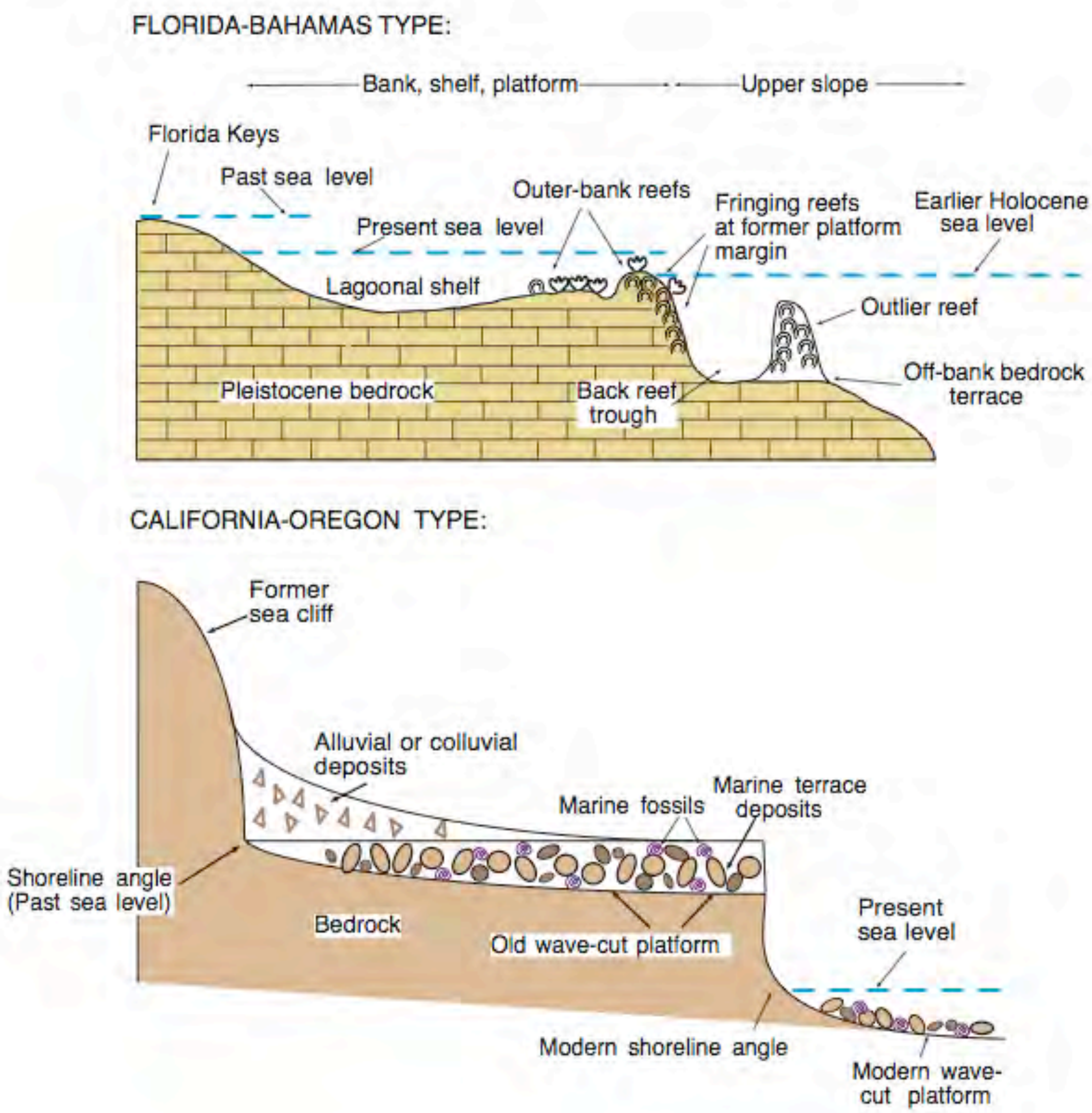


Figure 6.2 Recently published estimates of the mass balance of the Greenland Ice Sheet through time (modified from Alley et al., 2007). A Total Mass Balance of 0 indicates neither growth nor shrinkage, and -180 Gt yr-1 indicates ice-sheet shrinkage contributing to sea-level rise of 0.5 mm/yr, as indicated. Each box extends from the beginning to the end of the time interval covered by the estimate, with the upper and lower lines indicating the uncertainties in the estimates. A given color is associated with a particular technique, and the different letters identify different studies. Two estimates have arrows attached, because those authors indicated that the change is probably larger than shown. The dotted box in the upper right is a frequently-cited study that applies only to the central part of the ice sheet, which is thickening, and misses the faster thinning in the margins.



2064 **Figure 6.3** Cross-sections showing idealized geomorphic and stratigraphic expression of
2065 coastal landforms and deposits found on low-wave-energy carbonate coasts of Florida
2066 and the Bahamas (upper) and high-wave-energy rocky coasts of Oregon and California
2067 (lower). (Vertical elevations are greatly exaggerated.)

2068

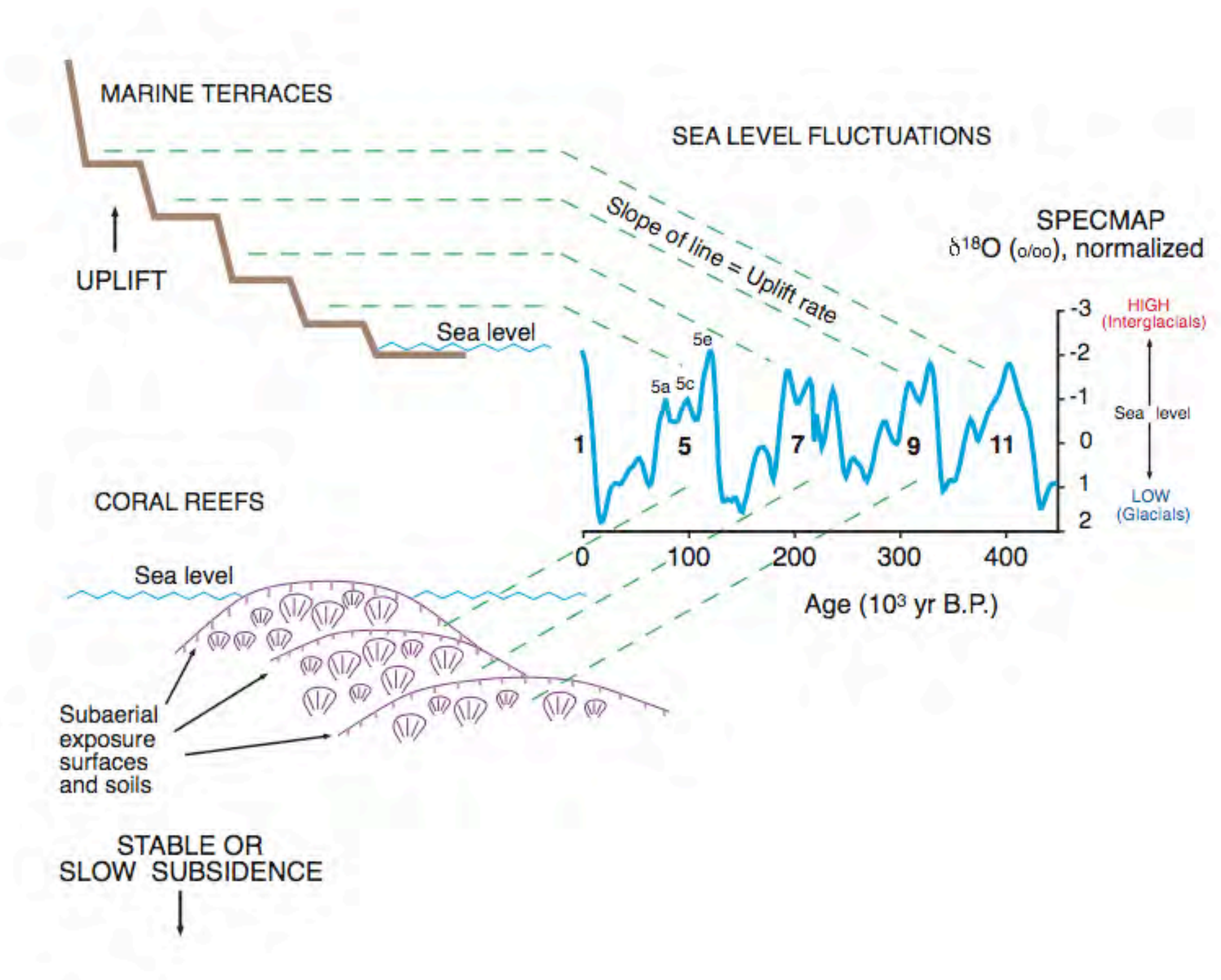


Figure 6.4 Relations of oxygen isotope records in foraminifers of deep-sea sediments to emergent reef or wave-cut terraces on an uplifting coastline (upper) and a tectonically stable or slowly subsiding coastline (lower). Emergent marine deposits record interglacial periods. Oxygen isotope data shown are from the SPECMAP record (Imbrie et al., 1984). Redrawn from Muhs et al. (2004).

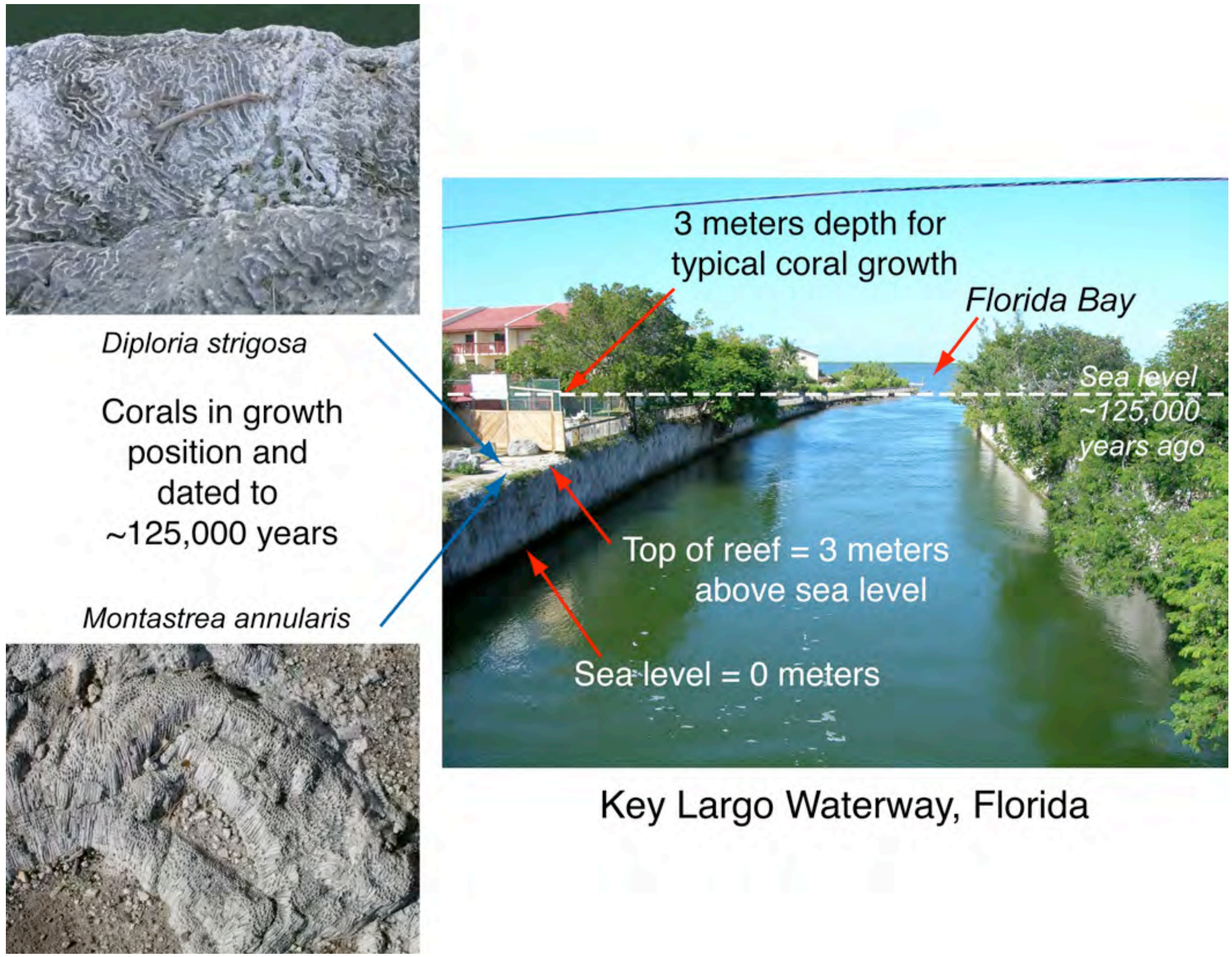


Figure 6.5 Photographs of last-interglacial (MIS 5e) reef and corals on Key Largo, Florida, their elevations, probable water depths, and estimated paleo-sea level. Photographs by D.R. Muhs.

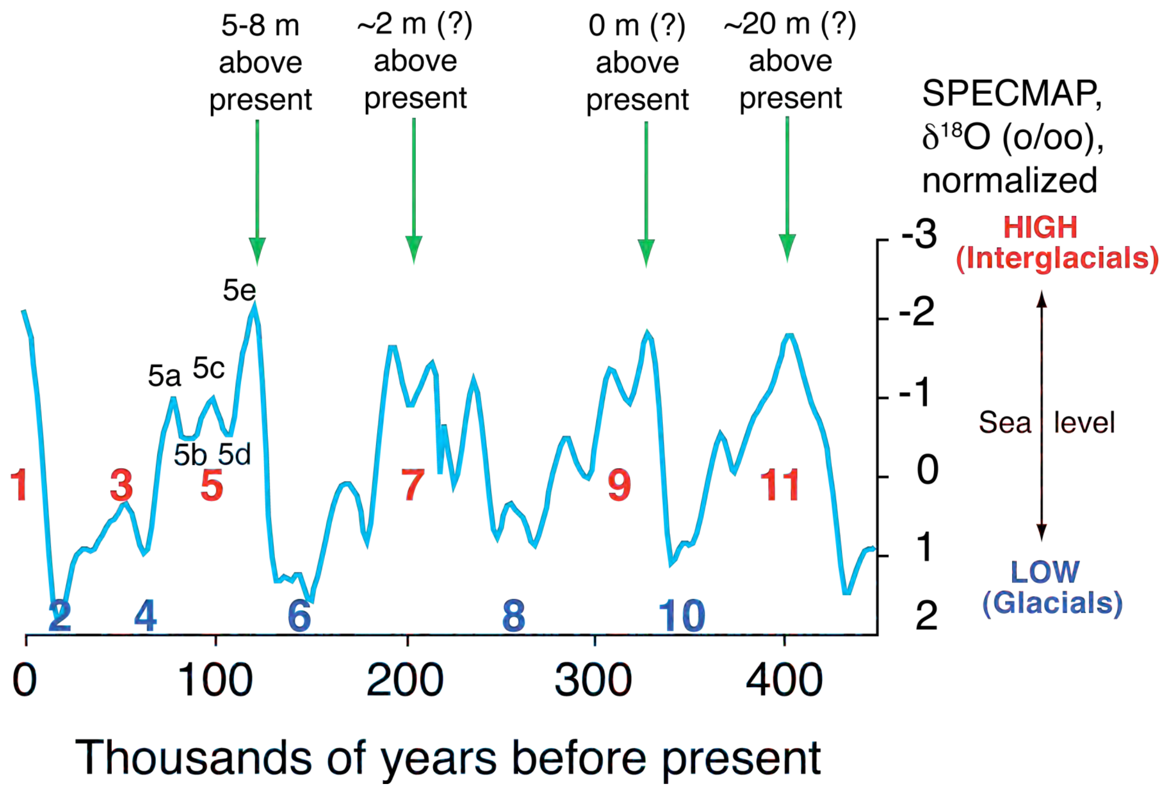


Figure 6.6 Oxygen isotope data from the SPECMAP record (Imbrie et al., 1984), with indications of sea-level stands for different interglacials, assuming minimal glacial isostatic adjustments to the observed reef elevations. Numbers identify Marine Isotope Stages (MIS) 1 through 11.

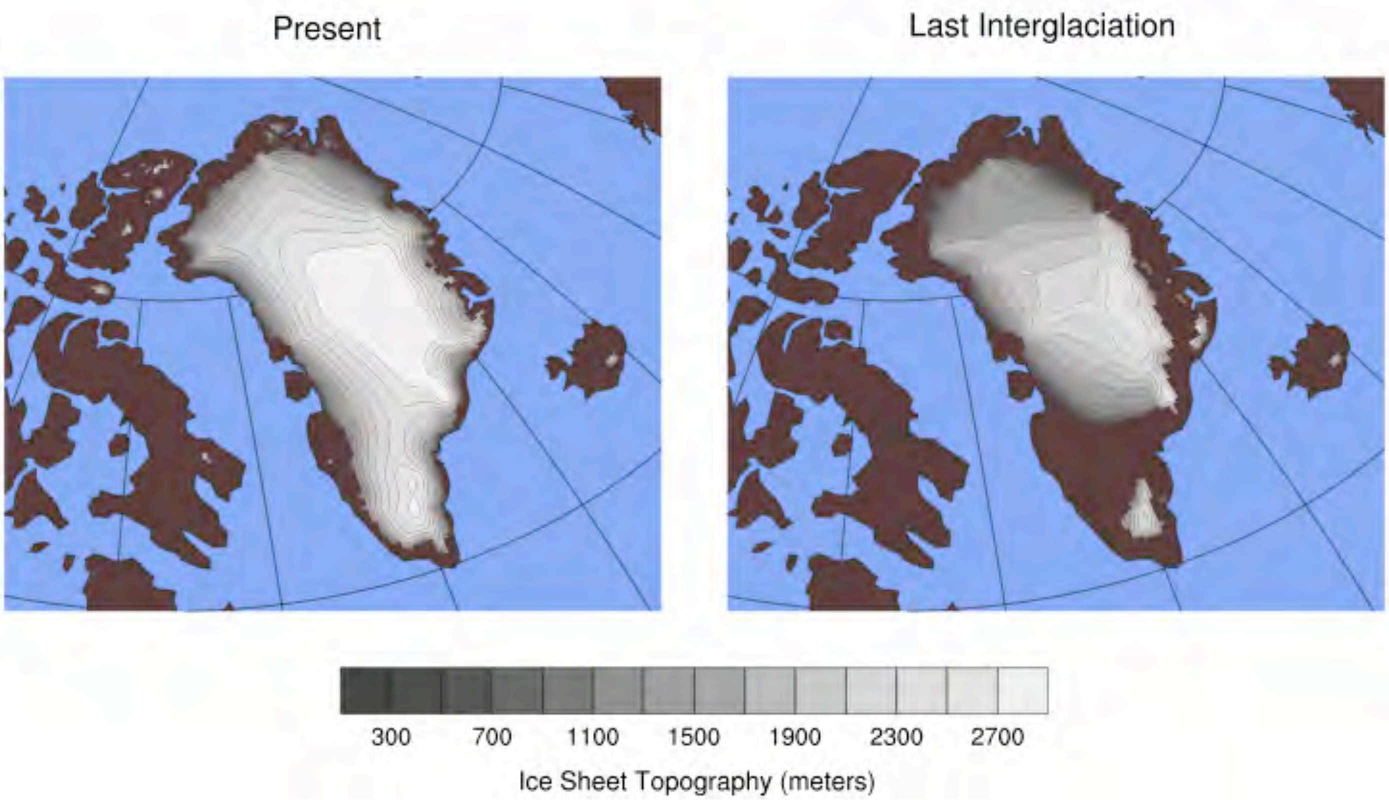


Figure 6.7 Modeled configuration of the Greenland Ice Sheet today (left) and in MIS 5e (right), from Otto-Bliesner et al. (2006).

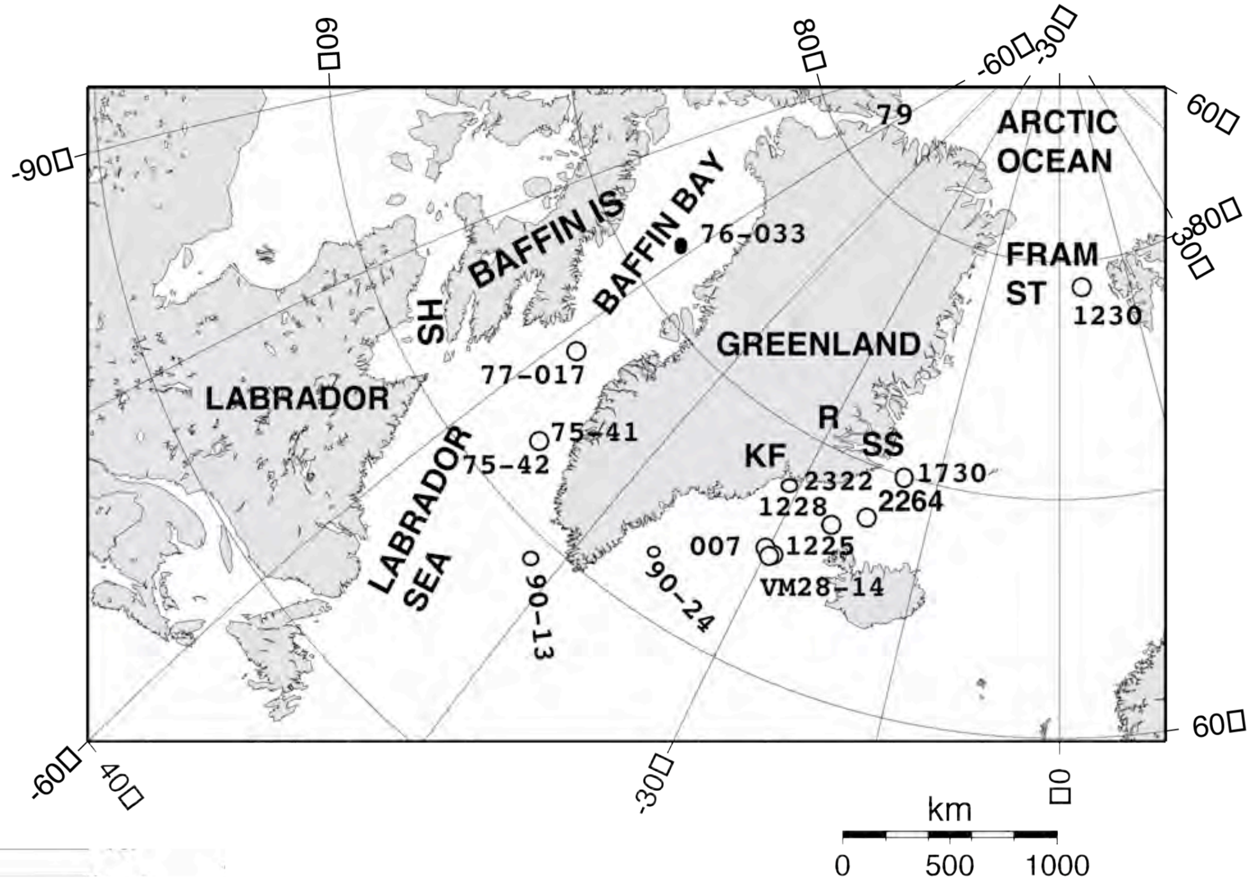
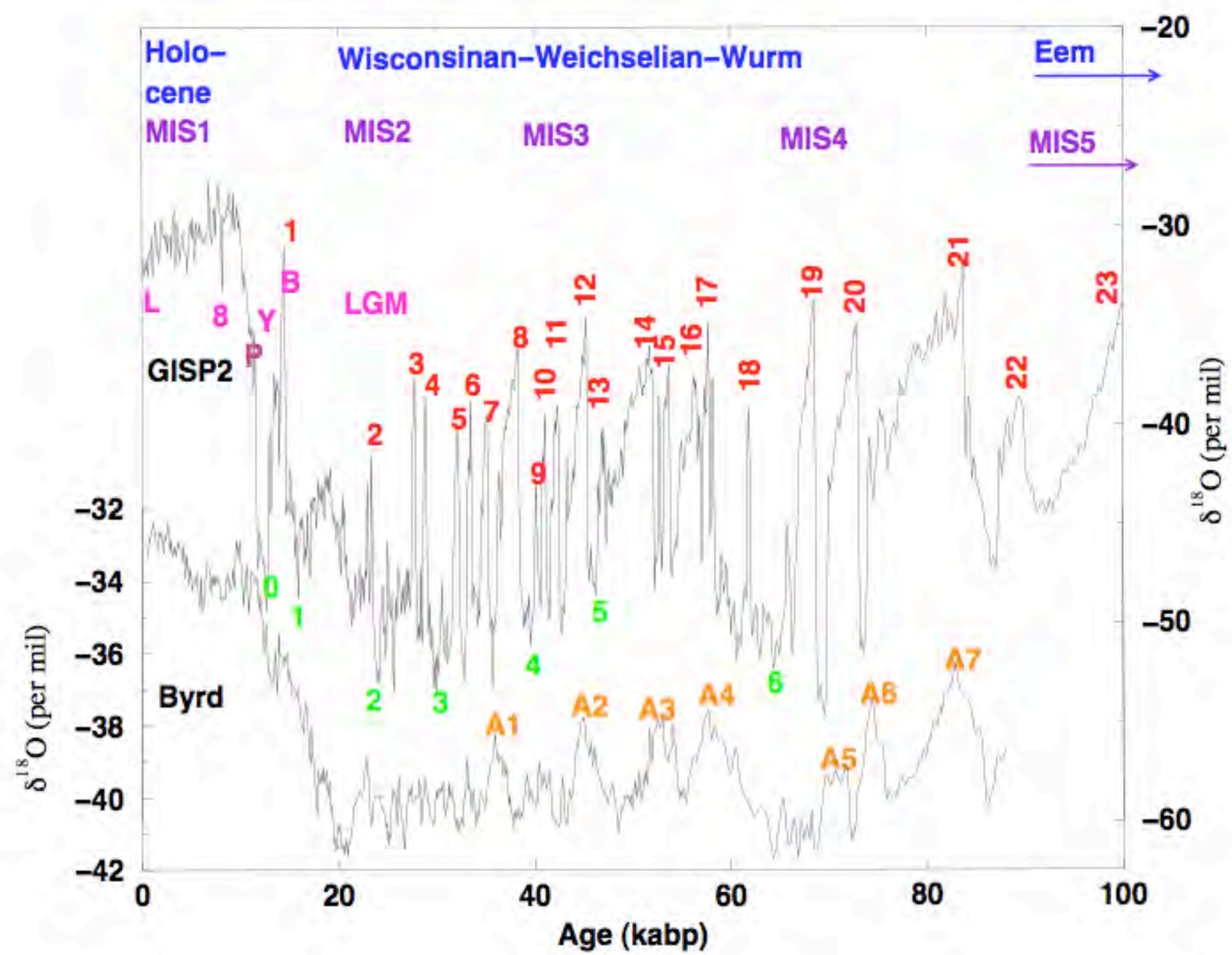


Figure 6.8 Location map with core locations discussed in the text. Full core identities are as follows: 79=LSSLL2001-079; 75-41 and-42=HU75-4,-42; 77-017=HU77-017; 76-033=HU76-033; 90-013=HU90-013; 1230=PS1230; 2264=PS2264; 1225 and 1228=JM96-1225,-1228; 007=HU93-007; 2322= MD99-2322; 90-24=SU90-24. HS=Hudson Strait, source for major Heinrich events; R = location of the Renland Ice Cap.


Holo–
cene
Wisconsinan–Weichselian–Wurm
Eem
MIS1
MIS2
MIS3
MIS4
MIS5
L
8
Y
P
B
1
LGM
GISP2
Byrd
1
2
3
4
5
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1
2
3
4
5
6
A1
A2
A3
A4
A5
A6
A7
δ 18O (per mil)
δ 18O (per mil)
Age (kabp)
0
20
40
60
80
100
−20
−30
−40
−50
−60
−32
−34
−36
−38
−40
−42

(

Figure 6.9 Ice-isotopic records ($\delta^{18}O$, a proxy for temperature, with less-negative values indicating warmer conditions) from GISP2, *Greenland* (Grootes and Stuiver, 1997) (scale on right) and Byrd Station, Antarctica (scale on left), as synchronized by Blunier and Brook (2001), with various climate-event terminology indicated. Ice age terms are shown in blue (top); the classical Eemian/Sangamonian is slightly older than shown here, as is the peak of marine isotope stage (MIS, shown in purple) 5, known as 5e. Referring specifically to the GISP2 curve, the warm Dansgaard-Oeschger events or stadial events, as numbered by Dansgaard et al. (1993), are indicated in red; Dansgaard-Oeschger event 24 is older than shown here. Occasional terms (L = Little Ice Age, 8 = 8k event, P=Preboreal Oscillation (PBO), Y = Younger Dryas, B = Bølling-Allerød, and LGM = Last Glacial Maximum) are shown in pink. Heinrich events are numbered in green just below the GISP2 isotopic curve, as placed by Bond et al. (1993). The Antarctic warm events A1–A7, as identified by Blunier and Brook (2001), are indicated for the Byrd record. Modified from Alley (2007).

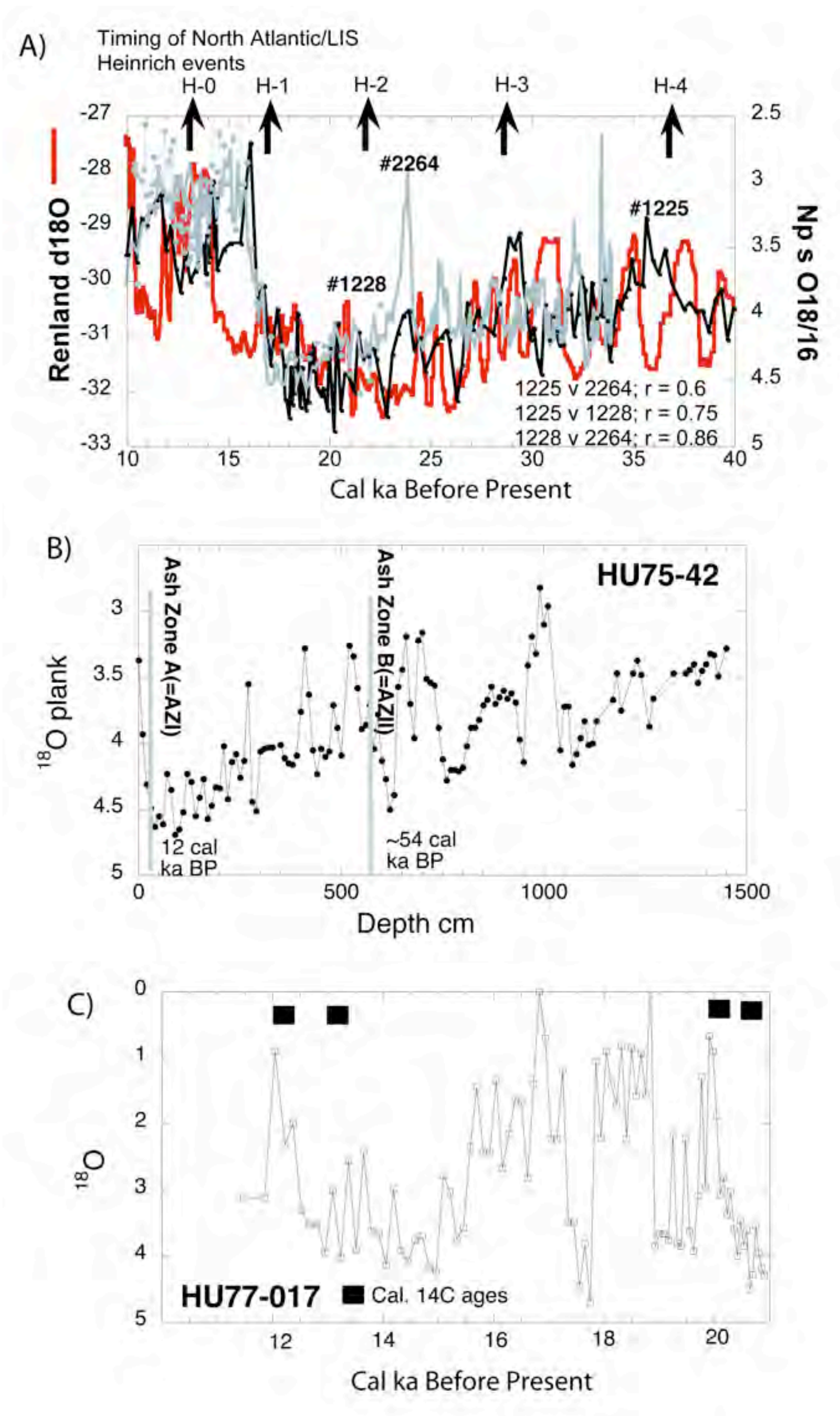

Figure 6.10 A) Variations in δ^{18}O from a series of cores north to south of Denmark Strait (see Fig. 6.8), namely: PS2264, JM96-1225 and 1228 plotted against the δ^{18}O from the Renland Ice Cap. B) δ18O variations in cores HU75-42 (NW Labrador Sea). C) Stable oxygen variations in cores HU77-017 from north of the Davis Strait.

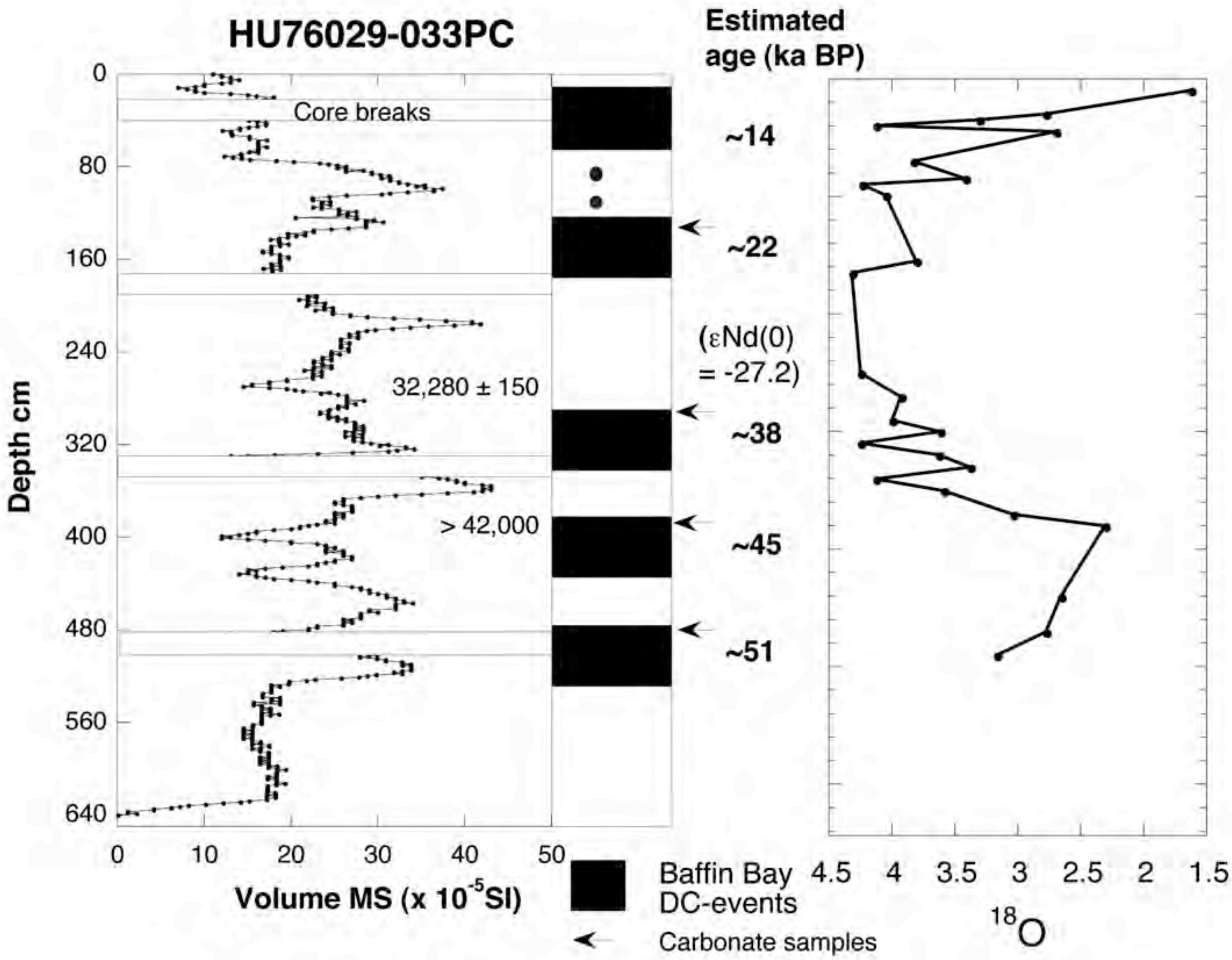

Figure 6.11 Variations in detrital carbonate (pieces of old rock) in core HU76-033 from Baffin Bay (Figure 6.8) showing down-core variations in magnetic susceptibility and $\delta^{18}O$.

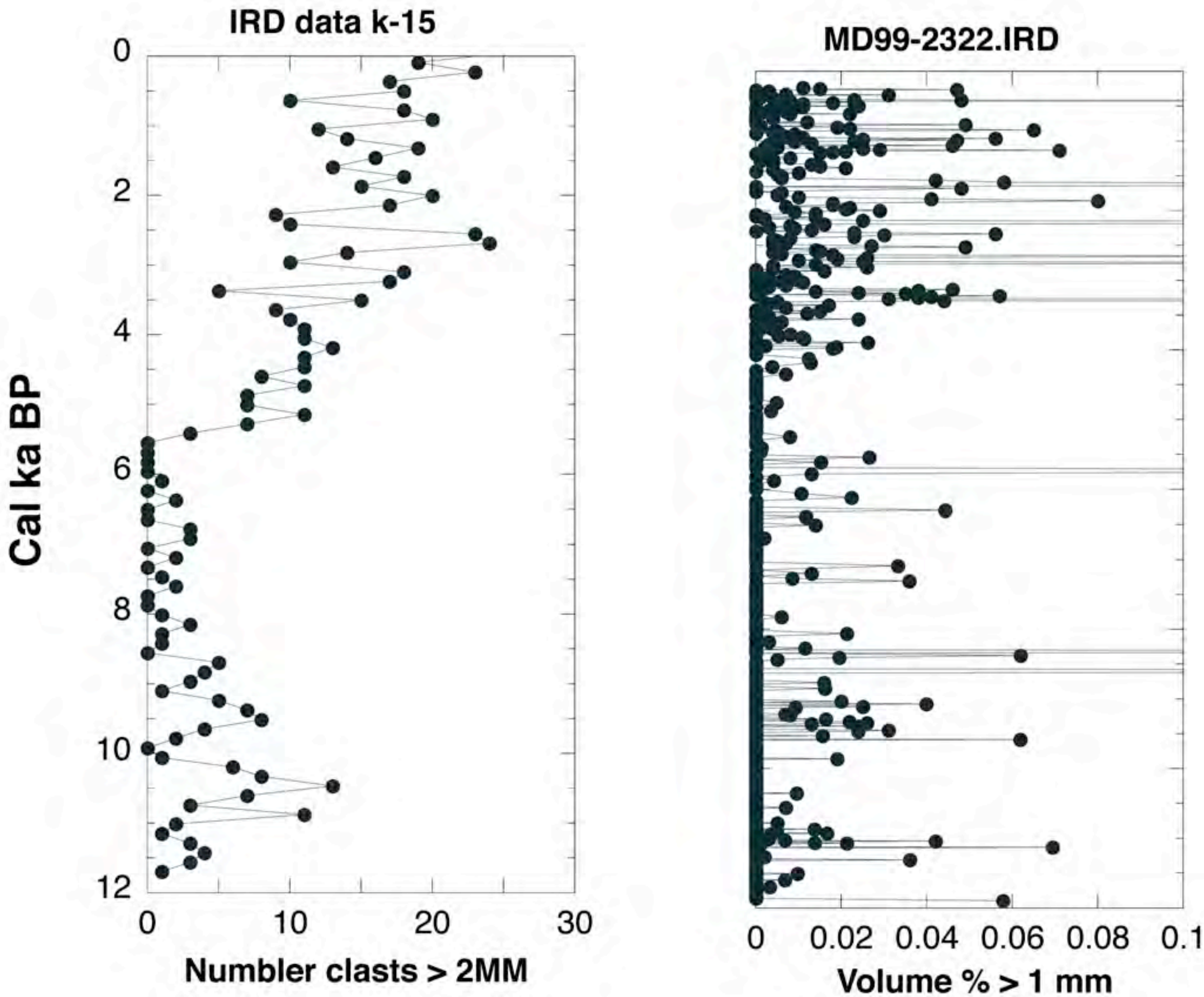

Figure 6.12 Holocene ice-rafted debris concentrations from MD99-2322 off Kangerdlugssuaq Fjord, east *Greenland* (Figure 6.8) showing log values of the percent of sediment > 1 mm and the weight % of quartz in the < 2mm sediment fraction.

1

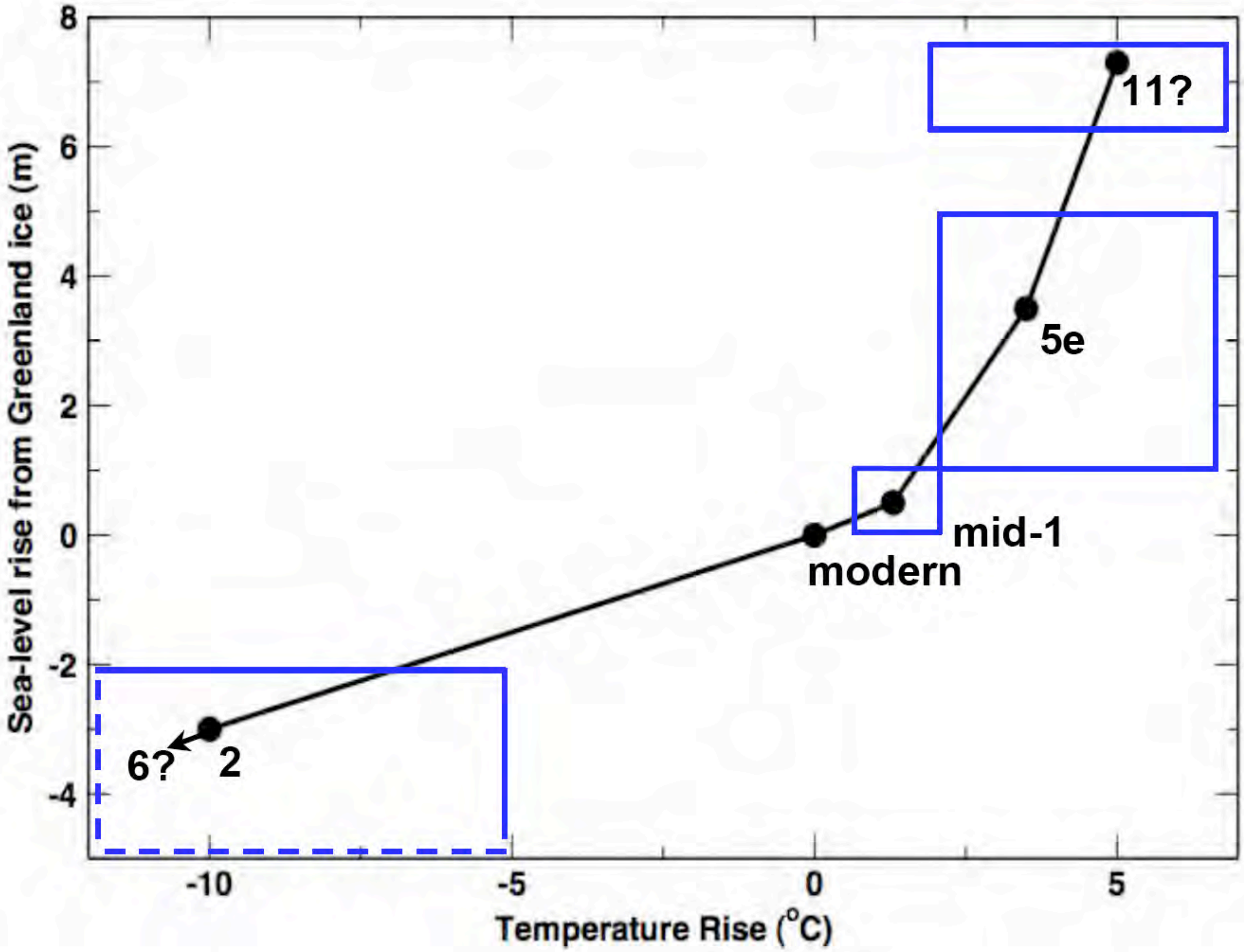
8
6
4
2
0
-2
-4
Sea-level rise from Greenland ice (m)
-10
-5
0
5
Temperature Rise (°C)
11?
5e
mid-1
modern
2
6?

Figure 6.13 A best-guess representation of the dependence of the volume of the Greenland Ice Sheet on temperature. Large uncertainties should be understood, and any ice-volume changes in response to sea-level changes correlated with temperature changes are included (although, as discussed in the text, temperature changes probably dominated forcing, especially at warmer temperatures when the reduced ice sheet had less contact with the sea). Recent values of temperature and ice volume (perhaps appropriate for 1960 or so) are assigned 0,0. The Last Glacial Maximum was probably ~6°C colder than modern for global average (e.g., Cuffey and Brook, 2000; data and results summarized in Jansen et al., 2007). Cooling in central *Greenland* was ~15°C (with peak cooling somewhat more; Cuffey et al., 1995). Some of the central-*Greenland* cooling was probably linked to strengthening of the temperature inversion that lowers near-surface temperatures relative to the free troposphere (Cuffey et al., 1995). A cooling of ~10°C is thus plotted. The ice-volume-change estimates of Peltier (2004; ICE5G) and Fleming and Lambeck (2004) are used, with the upper end of the uncertainty taken to be the ICE4G estimate (see Peltier, 2004), and somewhat arbitrarily set as 1 m on the lower side. The arrow indicates that the ice sheet in MIS 6 was more likely than not slightly larger than in MIS 2, and that some (although inconsistent) evidence of slightly colder temperatures is available (e.g., Bauch et al., 2000). The mid-Holocene result from ICE5G (Peltier, 2004) of an ice sheet smaller than modern by ~0.5 m of sea-level equivalent is plotted; the error bars reflect the high confidence that the mid-Holocene ice sheet was smaller than modern, with similar uncertainty assumed for the other side. Mid-Holocene temperature is taken from the Alley and Anandakrishnan (1995) summertime melt-layer history of central *Greenland*, with their 0.5°C uncertainty on the lower side, and a wider uncertainty on the upper side to include larger changes from other indicators (which are probably weighted by wintertime changes that have less effect on ice-sheet mass balance, and so are not used for the best estimate; Alley et al., 1999). As discussed in 6.3.3b and c, MIS 5e (the Eemian) is plotted with a warming of 3.5°C and a sea-level rise of 3.5 m. The uncertainties on sea-level change come from the range of data-constrained models discussed in 6.3.3c. The temperature uncertainties reflect the results of Cuffey and Marshall (2000) on the high side, and the lower values simulated over *Greenland* by Otto-Bliesner et al. (2006). Loss of the full ice sheet is also plotted, to reflect the warmer conditions that may date to MIS 11 if not earlier, and perhaps also to the

Pliocene times of the Kap København Formation. Very large warming is indicated by the paleoclimatic data from *Greenland*, but much of that warming probably was a feedback from loss of the ice sheet itself (Otto-Bliesner et al., 2006). Data from around the North Atlantic for MIS 11 and other interglacials do not show significantly higher temperatures than during MIS 5e, allowing the possibility that sustaining MIS 5e levels for a longer time led to loss of the ice sheet. Slight additional warming is indicated here, within the error bounds of the other records, based on assessment that MIS 5e was sufficiently long for much of the ice-sheet response to have been completed, so that additional warmth was required to cause additional retreat. The volume of ice possibly persisting in highlands even after loss of central regions of the ice sheet is poorly quantified; 1 m is indicated.

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