



Contents lists available at ScienceDirect

Quaternary Science Reviews

journal homepage: www.elsevier.com/locate/quascirev

History of the Greenland Ice Sheet: Paleoclimatic insights

Richard B. Alley^{a,*}, J.T. Andrews^b, J. Brigham-Grette^c, G.K.C. Clarke^d, K.M. Cuffey^e,
J.J. Fitzpatrick^f, S. Funder^g, S.J. Marshall^h, G.H. Miller^b, J.X. Mitrovicaⁱ, D.R. Muhs^f,
B.L. Otto-Bliesner^j, L. Polyak^k, J.W.C. White^b

^a Department of Geosciences and Earth and Environmental Systems Institute, Pennsylvania State University, University Park, PA 16802, USA^b Institute of Arctic and Alpine Research, Department of Geological Sciences, University of Colorado, Boulder, CO 80309-0450, USA^c Department of Geosciences, University of Massachusetts, Amherst, MA 01003, USA^d Department of Earth and Ocean Sciences, University of British Columbia, 6339 Stores Road, Vancouver, BC, Canada V6T 1Z4^e Department of Geography, University of California, Berkeley, CA 94720, USA^f Earth Surface Processes, MS-980, U.S. Geological Survey, Denver, CO 80225, USA^g Geological Museum, University of Copenhagen, Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark^h Department of Geography, University of Calgary, 2500 University Drive N.W., Calgary, Alberta, Canada T2N 1N4ⁱ Earth and Planetary Sciences, Harvard University, 20 Oxford Street, Cambridge, MA 02138, USA^j Climate Change Research, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307, USA^k Byrd Polar Research Center, The Ohio State University, 108 Scott Hall, 1090 Carmack Road, Columbus, OH 43210-1002, USA

ARTICLE INFO

Article history:

Received 1 May 2009

Received in revised form

4 January 2010

Accepted 3 February 2010

ABSTRACT

Paleoclimatic records show that the Greenland Ice Sheet consistently has lost mass in response to warming, and grown in response to cooling. 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 and ice-sheet responses 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 paleoclimatic 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 ice-sheet 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.

© 2010 Elsevier Ltd. All rights reserved.

1. The Greenland Ice Sheet

1.1. Overview

The Greenland Ice Sheet (Fig. 1) is approximately 1.7 million km² in area, extending as much as 2200 km north to south. The maximum ice thickness is 3367 m, the average thickness is 1600 m (Thomas et al., 2001), and the volume is 2.9 million km³ (Bamber et al., 2001). The ice has depressed some bedrock below sea level, and a little would remain below sea level following ice removal and

bedrock rebound (Bamber et al., 2001). However, most of the ice rests on bedrock above sea level and would contribute to the globally averaged sea-level rise of 7.3 m if melted completely (Lemke et al., 2007).

The ice sheet is mainly old snow squeezed to ice under the weight of new snow. Mass loss is primarily by melting in low-elevation regions, and by calving icebergs that drift away to melt elsewhere. Sublimation, snowdrift (Box et al., 2006), and melting or freezing at the bed are minor, although melting beneath floating ice shelves before icebergs break off may be locally important (see Section 1.2).

For net snow accumulation on the Greenland Ice Sheet, Hanna et al. (2005) (also see Box et al. (2006), among others) found for

* Corresponding author. Tel.: +1 814 863 1700.

E-mail address: rba6@psu.edu (R.B. Alley).

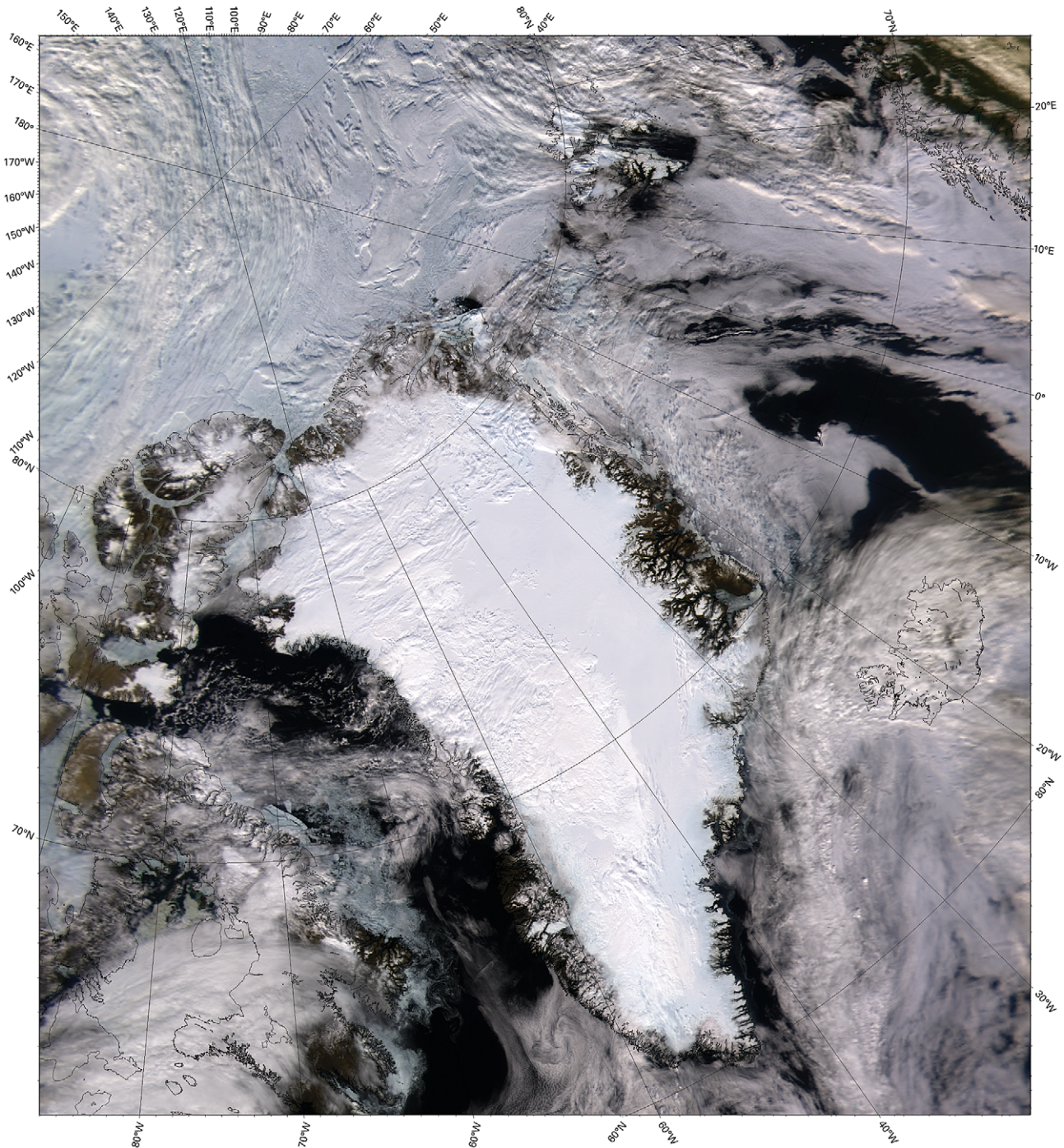


Fig. 1. Satellite image (SeaWiFS) of the Greenland Ice Sheet and surroundings, from July 15, 2000 (<http://www.gsfc.nasa.gov/gsf/earth/pictures/earthpic.htm>).

1961–1990 (an interval of moderately stable conditions before more-recent warming) that surface snow accumulation (precipitation minus evaporation) was $\sim 573 \text{ Gt a}^{-1}$, and that 280 Gt a^{-1} of meltwater left the ice sheet. The difference of 293 Gt a^{-1} is similar to the estimated iceberg calving flux at that time 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 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; Mernild et al., 2008). 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., increasing melting and snowfall are observed, the expected response to warming; Hanna et al., 2005; Box et al., 2006).

Increased iceberg calving from ice-shelf shrinkage and outlet-glacier acceleration has also been observed (e.g., Alley et al., 2005a; Rignot and Kanagaratnam, 2006). The Intergovernmental Panel on Climate Change (IPCC, 2007; 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) sea-level equivalent per year from 1961 to 2003 and -50 to -100 Gt (0.14 – 0.28 mm SLE) per year from 1993 to 2003, with even larger losses in 2005.” Updates include Alley et al. (2007b)

adjustment that propagates toward the ice-sheet center. Fast-flowing marginal regions can be affected within years, but slow-flowing central regions require a few millennia (e.g., Alley et al., 1987; Nick et al., 2009).

Warmer ice deforms more rapidly. In inland regions, ice sheet response to temperature change resembles response to accumulation-rate change: cooling slows deformation, thickening and perhaps steepening the ice sheet to increase ice flux and re-establish equilibrium. However, the few-millennial response is delayed by a few millennia while a surface-temperature change penetrates to the deep ice where most deformation occurs. The calculation is not simple, because moving ice carries its temperature along. Onset of surface-melt penetration to the bed through crevasses might thaw a frozen bed much more rapidly, perhaps in only a few minutes (Alley et al., 2005b; Das et al., 2008).

2. Paleoclimatic indicators bearing on ice-sheet history

Here, marine indicators of ice-sheet change are discussed, followed by terrestrial archives. For a broader overview, see, e.g., Cronin (1999) or Bradley (1999).

2.1. Marine indicators

Paleoceanographic cruises to the Greenland shelf focused on events since ice-sheet formation are mainly limited to the last two decades. Initial work in east Greenland (Marienfeld, 1992b; Mienert et al., 1992; Dowdeswell et al., 1994a) has been extended to west Greenland (Lloyd, 2006; Moros et al., 2006). Few areas have been investigated, and adjacent deep-sea basins are unclear about Greenland history because the late Quaternary sediments contain inputs from several adjacent ice sheets (Aksu, 1985; Andrews et al., 1998a; Hiscott et al., 1989; Dyke et al., 2002). Most marine cores from the Greenland shelf span the retreat from the last ice age (less than 15 ka). (For ages within the range of radiocarbon, we use calibrated or calendar years before present; Stuiver et al., 1998; Fairbanks et al., 2005.) Datable volcanic ashes from Icelandic sources offer the possibility of correlating records around Greenland from the time of 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 500–600-m-deep sills formed during the rifting that opened the modern oceans. These sills connect Greenland to Iceland through Denmark Strait and to Baffin Island through Davis Strait, and separate sedimentary records of ice-sheet history 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 ocean circulation is primarily clockwise around Greenland: 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. On the East Greenland shelf, this modified Atlantic Water is 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, 1996; Syvitski et al., 1996). Baffin Bay contains three water masses: Arctic Water in the upper 100–300 m (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 North Atlantic deep-water formation. Also, past changes in deep-water formation are linked to climate changes that affected the ice sheet (e.g., Alley, 2007). The major deep-water flow runs southward through Denmark Strait (McCave and Tucholke, 1986). The Eirik Drift sediment deposit off southwest Greenland was produced by this flow (Stoner et al., 1995). Convection in the Labrador Sea forms an upper component of this North Atlantic Deep Water.

Marine indicators that especially bear on the history of the Greenland Ice Sheet, discussed next, include: (1) flux and source of ice-rafted debris; (2) glacial deposition on trough-mouth fans; (3) stable-isotopic and biotic data indicating times of ice-sheet meltwater release; and (4) geophysical data indicating sea-floor erosion or deposition.

Except in rare circumstances (Gilbert, 1990; Smith and Bayliss-Smith, 1998), abundant coarse rock material far from land is ice-rafted debris (IRD), carried by sea ice or icebergs. Icebergs usually carry coarser debris (especially >2 mm) than sea ice (Lisitzin, 2002). IRD characteristics can be used to identify sources (e.g., neodymium isotopes (Grousset et al., 2001; Farmer et al., 2003), biomarkers (Parnell et al., 2007), magnetic properties (Stoner et al., 1995), and mineralogy (Andrews, 2008)); however, little such work has been done for Greenland.

Major ice-sheet outlet glaciers advance across the continental shelf in prominent troughs to shed debris flows downslope, forming trough-mouth fans (TMF) where the troughs widen and flatten at the continental rise (Vorren and Laberg, 1997; O’Cofaigh et al., 2003), with sufficiently rapid and continuous sedimentation to provide high-time-resolution records. Open-marine sediments accumulate on TMF when the ice margin is retreated. Prominent trough-mouth fans exist in east Greenland (Scoresby Sund, Dowdeswell et al., 1997; the Kangerdlugssuaq Trough, Stein, 1996; Angamassalik Trough, St. John and Krissek, 2002) and west Greenland (Disko Bay, from Jakobshavn and neighboring glaciers).

In most places, foraminiferal $\delta^{18}\text{O}$ shifts to “heavier” (more positive) values in response to cooling or ice-sheet growth. However, near ice sheets the abrupt appearance of light isotopes is most commonly caused by meltwater (Jones and Keigwin, 1988; Andrews et al., 1994). Around Greenland, measurements are normally made on shells of the near-surface planktic foraminifera *Neogloboquadrina pachyderma* sinistral (Fillon and Duplessy, 1980; van Kreveland et al., 2000; Hagen and Hald, 2002), although some data are available from benthic species (Andrews et al., 1998a; Jennings et al., 2006).

High-resolution seismic and sonar studies (Stein, 1996; O’Cofaigh et al., 2003; Wilken and Mienert, 2006) reveal 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.

2.2. Terrestrial indicators

Although terrestrial records are typically more discontinuous in space and time than are marine records because net erosion is dominant on land whereas net deposition is dominant in most marine settings, useful terrestrial records are available.

2.2.1. Geomorphic indicators

Land-surface characteristics, and especially moraines, provide useful paleoclimatic information (e.g., Sugden and John, 1976). Moraines mark either the maximum ice advance or a still-stand during retreat. Normally, ice readvance destroys older moraines,

although remnants of moraines overrun by a subsequent advance are occasionally preserved and identifiable, especially if the readvancing ice was frozen to its bed.

Radiocarbon (carbon-14) dating of carbonaceous material in a moraine gives a maximum age for ice advance; such dating in lakes behind moraines gives minimum ages for retreat. Increasingly, moraines are dated by measurement of beryllium-10 or other isotopes produced in boulders by cosmic rays (e.g., Gosse and Phillips, 2001). Because cosmic rays penetrate only ~1 m into rock, boulders quarried from beneath the ice following erosion of >~1 m, 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, inheritance of nuclides, and other factors may be nontrivial but potentially reveal further information (e.g., Applegate et al., 2008). Other moraine-dating techniques include historical records, the size of lichen colonies (e.g., Locke et al., 1979; Geirsdottir et al., 2000), soil development, and breakdown of rocks (clast weathering).

Surfaces striated and polished by glacial action, or boulders away from moraines, can also be dated by cosmogenic isotopes. 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.

The highest elevation to which a mountain-glacier moraine extends is close to the equilibrium-line altitude when the moraine formed (snow accumulation caused flow into the glacier above that elevation). Also, glaciers make identifiable landforms, especially if thawed at the base and thus sliding freely, so contrasts in landform appearance can reveal the limits of glaciation (or of wet-based glaciation).

Glaciers respond to many environmental factors, but the balance between snow accumulation and melting is usually the major control. The equilibrium vapor pressure (the ability of warmer air to hold more moisture) increases ~7% °C⁻¹, whereas the increase in melting is ~35% (±10%) °C⁻¹ if melting balances snow accumulation (e.g., Oerlemans, 1994, 2001; Denton et al., 2005). Thus, glacier extent can usually be used as a proxy for summer temperature (duration and warmth of the melt season).

2.2.2. Biological indicators and related features

Biotic indicators are useful in paleoclimatic reconstruction (e.g., Schofield et al., 2007), with important records from lake sediments, with their continuous sedimentation and rich ecosystems (e.g., Björck et al., 2002; Andresen et al., 2004; Ljung and Björck, 2004). Pollen (e.g., Ljung and Björck, 2004; Schofield et al., 2007), microfossils, and macrofossils (such as chironomids, also called midge flies (Brodersen and Bennike, 2003)) are valuable. The isotopic composition of shells or inorganic precipitates records some combination of temperature and water isotopic composition. Physical or physical–biological aspects of lake sediments (e.g., loss on ignition, controlled primarily by the relative abundance of organic matter) are also related to climate. And, types of shells in raised marine deposits (e.g., Dyke et al., 1996; see Section 2.2.3) provide climatic data.

2.2.3. Glacial-isostatic adjustment and relative sea-level indicators near the ice sheet

Shoreline processes produce recognizable features, which may occur above or below modern sea level due to land-elevation change (by geological or glacial-isostatic processes) or ocean-volume changes (especially from ice-sheet growth or decay).

Glacial-isostatic adjustment involves both immediate elastic and slow viscous effects. For example, instantaneous formation and

multi-millennial persistence of an ice sheet would cause nearly instantaneous elastic sinking of the land 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 with a peak thickness of 3–4 km, produced a crustal depression of ~1 km. (For comparison, that ice sheet contained enough water to make a uniform layer only ~70 m thick across the world oceans.)

Outside the depressed region covered by ice, land is gradually pushed up into a peripheral bulge. Subsequent ice melt allows the central depressed region to rebound over millennia or longer. Thus, at sites in Hudson Bay sea level continues to fall on the order of 10 mm a⁻¹ despite disappearance of most of the Laurentide Ice Sheet ~8000 years ago. Also, the loss of ice cover allows the peripheral bulge to subside, causing sea-level rise there (e.g., continuing along the east coast of the United States although at slower rates than rebound in the central part of the former ice sheet). Still farther from the former ice sheet in the so-called “far-field,” sea-level change is dominated by addition or subtraction of water from the ocean. However, in periods of stable ice cover (e.g., during much of the present interglacial), sea level continues to change due to the ongoing gravitational and deformational effects of glacial-isostatic adjustment. Such glacial-isostatic adjustment is responsible for a sea-level fall in parts of the equatorial Pacific of about 3 m during the last 5000 years, exposing corals and shoreline features of this age (Mitrovica and Peltier, 1991; Dickinson, 2001; Mitrovica and Milne, 2002; see Section 2.2.4).

Near-field relative sea-level changes have commonly been used to constrain models of ice-sheet geometry, particularly since the Last Glacial Maximum (which peaked at about 24 ka) (see Lambeck et al., 1998; Tarasov and Peltier, 2002, 2003; Fleming and Lambeck, 2004; Peltier, 2004; discussed in Section 3). Data include 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. Observations on the transition of modern lakes from formerly marine conditions are also useful (e.g., Bennike et al., 2002), as are now-subsea but initially terrestrial archaeological sites (see also Weidick, 1996; Kuijpers et al., 1999).

All glacial-isostatic adjustment studies include uncertainty from the poorly known 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. Also, Greenland studies are complicated by signals from at least two other sources: (1) 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) addition of meltwater from contemporaneous melting (or, in times of glaciation, growth) of all other global ice reservoirs. Thus, estimated volume and extent of the Laurentide Ice Sheet, and the volume of more-distant ice masses, are required to interpret sea-level data from Greenland.

2.2.4. Far-field indicators of relative sea-level high-stands

Far-field sea-level indicators record the combined history of glacial-isostatic adjustment and ocean volume, including changes in the Greenland Ice Sheet. Marine deposits or emergent coral reefs now found above sea level on geologically relatively stable coasts and islands are especially important, providing high-water marks (or “bathtub rings”) of past high sea levels. For recording sea-level history, coastal landforms have two advantages as compared with the oft-cited deep-sea oxygen-isotope record: (1) if corals are

present, they can be dated directly; and (2) the tie to sea level is more direct, with no complication from changing temperature.

Coastal landforms record high stands of the sea when coral reefs grew as fast as sea level rose (upper panel in Fig. 3) or when a stable sea-level high stand eroded marine terraces into bedrock (lower panel in Fig. 3). Thus, emergent reefs or terraces on geologically rising coastlines record interglacial periods (Fig. 4). On a geologically stable or slowly sinking coast, emergent deposits record only those sea-level stands higher than present (Fig. 4). Past sea levels can thus be determined from stable coastlines, or even rising coastlines if one can make reasoned estimates of uplift rates.

The direct dating of corals is possible because uranium (U) is dissolved in ocean water but thorium (Th) and protactinium (Pa) are almost absent. 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 $^{230}\text{Th}/^{234}\text{U}$, $^{238}\text{U}/^{234}\text{U}$, and $^{231}\text{Pa}/^{235}\text{U}$ can provide three independent clocks for dating the same fossil coral (e.g., Edwards et al., 1997; also Thompson and Goldstein, 2005). Since the 1980s, most workers have employed thermal ionization mass spectrometry (TIMS) to measure U-series nuclides; compared to older techniques, this method yields higher precision from smaller samples, extending useful dating back to at least 500,000 years.

The best records come from geologically quiescent sites far from tectonic-plate boundaries, in tropical and subtropical regions where coral reefs grow. Even in such locations, however,

interpreting past sea levels can include much uncertainty, especially from hot-spot effects and glacial-isostatic adjustment.

First, many islands well inside a tectonic plate are in hot-spot volcanic chains such as the Hawaiian-Emperor seamount chain. Like an ice sheet, a growing volcano isostatically depresses the land beneath, forming a broad ring-shaped high 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, as discussed above, glacial-isostatic adjustment produces global-scale changes in sea level even during periods when ice volumes are stable. As an example, for the last 5000 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 a^{-1} 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).

2.2.5. Geodetic indicators

Many geodetic indicators provide useful information on recent and older ice-sheet change, if sufficient attention is paid to glacial-isostatic adjustment. These include GPS data on elevation changes of rock near the ice sheet (e.g., Kahn et al., 2007), and satellite data

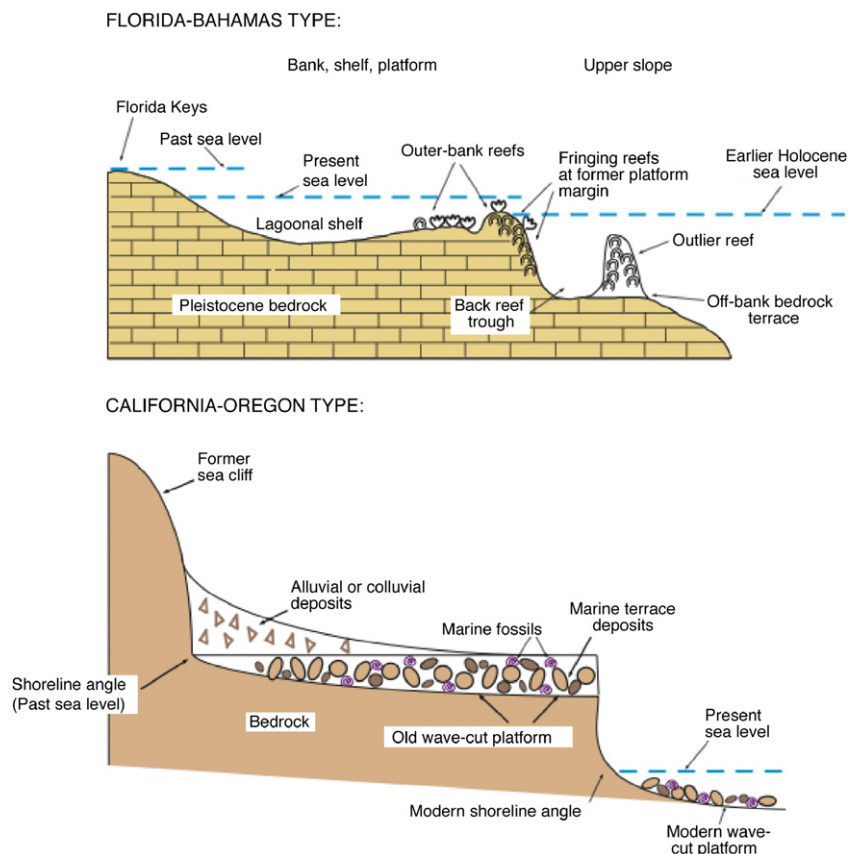


Fig. 3. Cross-sections showing idealized geomorphic and stratigraphic expression of coastal landforms and deposits found on low-wave-energy carbonate coasts of Florida and the Bahamas (upper) and high-wave-energy rocky coasts of Oregon and California (lower). (Vertical elevations are greatly exaggerated.)

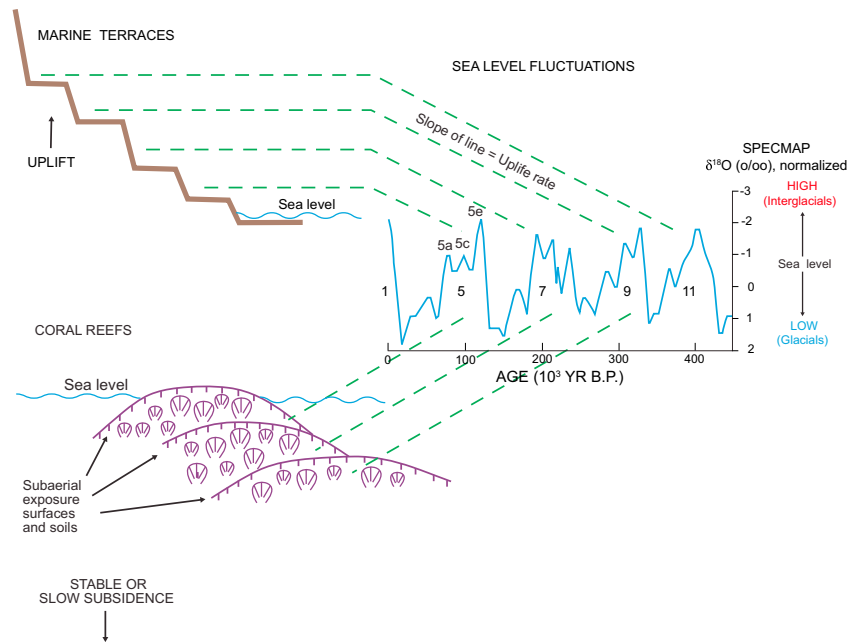


Fig. 4. Relations of oxygen-isotope records in foraminifers of deep-sea sediments to emergent reef or wave-cut terraces on an uplifting coastline (upper) and a tectonically stable or slowly subsiding coastline (lower). Emergent marine deposits record interglacial periods. Oxygen-isotope data shown are from the SPECMAP record (Imbrie et al., 1984). Redrawn from Muhs et al. (2004).

on changing regional mass (Gravity Recovery and Climate Experiment (GRACE) satellite mission; ~400 km resolution; e.g., Velicogna and Wahr, 2006), and altimeter measurements of ice height (e.g., Johannessen et al., 2005; Thomas et al., 2006).

Similarly, tide gauges reveal sea-level change after correction for local effects and glacial-isostatic adjustment (e.g., Douglas, 1997). Deviations from the global-average sea-level trend (~1.5–2 mm a⁻¹ rise in the 20th century) may help constrain meltwater sources. Mitrovica et al. (2001) and Plag and Juttner (2001) showed that rapid melting of different ice sheets will have substantially different signatures, or fingerprints, in the spatial pattern of sea-level change (also see Mitrovica et al., 2009). 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.

Earth's rotation is affected by any redistribution of mass on or in the planet (Munk, 2002; Mitrovica et al., 2006). Mass transfer from poles to equator slows the planet's rotation (like a spinning ice skater extending arms to slow rotation), and any transfer of mass that is not symmetric about the poles causes "wobble," or true polar wander (the position of the rotation pole moves relative to the planet's surface). True polar wander for the last century has been estimated using both astronomical and satellite geodetic data, while changes in the rotation rate (or, as geodesists say, length of day) have been determined for the last few decades from satellite measurements and for the last few millennia from observations of eclipses recorded by ancient cultures. The timing of ancient eclipses differs from the expected timing from simply projecting the Earth–Moon–Sun system back in time using the modern rotation rate of Earth, indicating a gradual slowing of Earth's rotation (Munk, 2002). After correcting for slowing of

Earth's rotation associated with the "drag" of the tides, this rotation-rate change 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 at driving these changes in rotation.) True polar wander, after correction for glacial-isostatic adjustment, gives some information about the meltwater source; in particular, melting centers progressively further from the pole will be more efficient at perturbing its location. (Munk, 2002; Mitrovica et al., 2006).

2.2.6. Ice cores

Ice cores provide broad information on climate forcing and response. Temperature histories are especially accurate, and agreement among several indicators increases confidence in the results.

A widely used indicator is δ¹⁸O, the difference between the ¹⁸O:¹⁶O ratio of a sample and of standard mean ocean water, normalized by the ratio of the standard and expressed as per mil (‰); hydrogen isotopic ratios offer additional information (e.g., Jouzel et al., 1997). Preferential condensation of the heavier species causes them to be progressively depleted in air masses, and thus in precipitation, with cooling. Although linked to site temperature, δ¹⁸O also is affected by the seasonal distribution of precipitation and other factors (Jouzel et al., 1997; Alley and Cuffey, 2001), requiring additional paleothermometers.

The temperature of the ice gives one of the most reliable. Just as the center of a turkey cooks slowly in a warm oven, intermediate depths of the central Greenland Ice Sheet have not yet finished warming from the ice age. When ice flow is understood well, this provides a low-time-resolution history of the surface temperature, which can be extracted through 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). This calibrates the relation between δ¹⁸O and temperature (α‰ °C⁻¹) as a new paleoclimatic

indicator, sensitive to changes in seasonality of snowfall and other factors.

The isotopic composition of trapped gases also records temperature. Snow is converted to permeable firn and then impermeable bubbly ice over a few tens of meters by solid-state processes that are faster under higher temperature or higher pressure. Only diffusion mixes the gases deeper than the wind-mixed upper few meters. Gravity slightly enriches heavier species in the air trapped in bubbles, proportional to the thickness of the air column in which diffusion dominates (Sowers et al., 1992).

If a sudden temperature change occurs at the surface, about 100 years are required for most of the temperature change at the depth of bubble trapping. However, a temperature gradient across gases in diffusive equilibrium slightly separates them by thermal fractionation, with the heavier gases moving toward the cold end (Severinghaus et al., 1998). 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, continuing until temperature change in the deep firn removes the temperature gradient over 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 determination of the temperature difference that existed vertically in the firn at the time of bubble trapping, and of 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 surface 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.

Past ice-accumulation rates can be obtained from the measured thickness of annual layers corrected for ice-flow thinning (e.g., Alley et al., 1993; Cuffey and Clow, 1997). Or, the firn thickness can be estimated from gas–isotope fractionation or the number density of bubbles (Spencer et al., 2006), and combined with temperature estimates to constrain accumulation rates. Aerosols of all types are added to the ice sheet with falling snow and between snowfalls; knowledge of the accumulation rate (hence dilution of the aerosols) allows estimation of time-histories of atmospheric loading (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 Nishiizumi, 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 may be estimated by counting annual layers (e.g., Alley et al., 1993; Andersen et al., 2006) or by correlation with other records (Blunier and Brook, 2001). Atmospheric composition “fingerprinting” of samples from Greenland ice cores

matched to longer Antarctic records (Suwa et al., 2006) showed that old ice exists in central Greenland (Chappellaz et al., 1997; Suwa et al., 2006) at depths where flow processes have mixed the layers (Alley et al., 1997). Where layers are continuous and unmixed, other features in ice cores such as chemically distinctive ash from particular volcanic eruptions can be correlated with independently dated records (e.g., Zielinski et al., 1994; Finkel and Nishiizumi, 1997). Flow models also can aid in dating.

The elevation history of ice sheets is indicated by the total gas content of the ice (Raynaud et al., 1997). The density at pore close-off is nearly constant, with a small and fairly well known correction for climatic conditions. Because air pressure varies with elevation, total moles of trapped gas per kilogram of ice decrease with increasing ice-sheet thickness. Fluctuations in total gas content linked to changing layering in the firn or other issues probably limit accuracy to ~500 m (Raynaud et al., 1997).

More information on ice-sheet changes comes from the current distribution of isochronous surfaces. An explosive volcanic eruption deposits an acidic layer of one age on the ice sheet, which can be mapped after burial using radar (Whillans, 1976; Jacobel and Welch, 2005) and dated at ice-core sites (e.g., Eisen et al., 2004). The modeled modern distribution of isochronous surfaces (and of other properties such as temperature) for hypothesized climate (primarily accumulation rate affecting burial and temperature) and ice-sheet flow (primarily changes in surface elevation and extent; e.g., Clarke et al., 2005) can be compared to measured data to estimate optimal climate histories.

3. History of the Greenland Ice Sheet

3.1. Ice-sheet onset and early fluctuations

Before 65 Ma, dinosaurs lived on a high-CO₂ world that was warm to high latitudes; mean-annual temperature exceeded 14 °C at 71°N based on occurrence of crocodile-like champsosaurs (Tarduno et al., 1998; also see Markwick, 1998; Vandermark et al., 2007). The ocean surface near the North Pole warmed from ~18 °C to a peak of ~23 °C during the short-lived Paleocene–Eocene Thermal Maximum about 55 Ma (Sluijs et al., 2006). 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 quickly; sand and coarser materials in an Arctic Ocean core from ~46 Ma (Moran et al., 2006; St. John, 2008) probably indicate ice rafting from glaciers. A core from ~75°N latitude in the Norwegian–Greenland Sea off East Greenland contains ice-rafted debris with at least some likely from glaciers rather than sea ice, from ~38 to 30 Ma (late Eocene into Oligocene time). Certain characteristics of this debris point to an East Greenland source and exclude Svalbard, the next-nearest landmass (Eldrett et al., 2007). This ice-rafted debris may represent isolated mountain glaciers or more-extensive ice-sheet cover. (This interval, until ~16 Ma, is not well-represented in the central-Arctic core of Moran et al. (2006) owing to erosion or little deposition.)

Ice-rafted debris, interpreted as representing iceberg as well as sea-ice transport, was actively delivered to the open Arctic Ocean at 16 Ma, and volumes increased ~14 Ma and again ~3.2 Ma (Moran et al., 2006; 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 ~7.3 Ma, based on ice-rafted debris near Greenland in the Irminger Basin. From its geographical pattern, the increase in ice-rafted debris ~3.2 Ma probably had sources in Greenland, Scandinavia, and the North

American landmass, although fingerprinting the debris to particular source rocks (e.g., Hemming et al., 2002) has not been done. Also, no direct evidence shows whether this debris was supplied to the ocean by an extensive ice sheet or by coastal-mountain glaciers in the absence of ice from Greenland's central lowlands. Despite lack of conclusive evidence, Greenland seems to have supported at least some glaciation since at least 38 Ma; glaciation left more records after ~14 Ma (middle Miocene). Thus, as Earth cooled from the "hothouse" conditions of the Cretaceous, ice sheets began to form on Greenland.

Following establishment of ice on Greenland, a notable warm interval ~2.4 Ma is recorded by the Kap København Formation of North Greenland (Funder et al., 2001). This 100-m-thick unit of sand, silt, and clay was deposited primarily in shallow-marine conditions, and contains fossil biota recording a switch from Arctic to subarctic to boreal assemblages in ~20,000 years or less. Funder et al. (2001) postulated complete deglaciation of Greenland at this time, primarily based on the great summer warmth at this far-northern site, although no comprehensive record of the whole ice sheet is available.

3.2. The most recent million years

The broad outline of ice-age cycling, based on $\delta^{18}\text{O}$ of benthic foraminifera, is plotted in Fig. 4 with the warm marine isotope stages numbered. This section focuses on those prior to MIS 5e.

3.2.1. Far-field sea-level indications

High sea levels are expected during interglacials (MIS 5, 7, 9, and 11; Fig. 4) with the amplitudes of the various high stands linked to the long-term mass balance of ice sheets including the Greenland Ice Sheet. Fragmentary and poorly dated deposits suggest higher-than-present sea level during MIS 11, ~400 ka, when orbital geometry was similar to the configuration during the current interglacial (Berger and Loutre, 1991).

Hearty et al. (1999) proposed that marine deposits found in a cave about 21 m above modern sea level on the tectonically stable island of Bermuda date to MIS 11. Coral pebbles in these deposits yielded U-series ages >500 ka, but Hearty et al. (1999) interpreted the deposits to date to about 400 ka based primarily on an overlying deposit that dates to about 400 ka.

A marine deposit from the "Anvilian marine transgression", possibly from MIS 11, occurs at altitudes up to 22 m along the tectonically stable Seward Peninsula and Arctic Ocean coast of Alaska (Kaufman et al., 1991), landward of Pelukian (MIS 5, ~74–130 ka) marine deposits (age assignments to marine isotope stages differ in different usage, and are provided here for reference, not definition). Amino-acid ratios in mollusks (Kaufman and Brigham-Grette, 1993) show that the Anvilian deposit is older than the Pelukian deposits but younger than deposits thought to be of Pliocene age (~1.8–5.3 Ma). Kaufman et al. (1991) reported that basaltic lava, with an average age based on several analyses of 470 ± 190 ka, overlies deposits of the Nome River glaciation, which in turn overlie Anvilian marine deposits. Kaufman et al. (1991) thus proposed that the Anvilian marine transgression dates to about 400 ka and correlates with MIS 11.

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). These results taken together, if accurate, imply that all of the Greenland Ice Sheet (Willerslev et al., 2007; see Section 3.2.2), all of the West Antarctic ice sheet, and part of the East Antarctic ice sheet 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)).

The MIS 9 (about 303–331 ka) high stand is poorly constrained but not documented to be significantly above modern. Coral reefs of this age are known from tectonically rising Barbados (Bender et al., 1979). Stirling et al. (2001) reported that well-preserved, well-dated (U-series dates between about 334 ± 4 and 293 ± 5 ka) fringing reefs occur on tectonically stable Henderson Island in the southeastern Pacific Ocean as high as ~29 m above sea level; however, slow uplift (<0.1 m/1000 a) due to volcanic loading from growth of nearby Pitcairn Island complicates interpretation, and a correction for maximum uplift rate would put the MIS 9 sea-level estimate below present sea level. Multer et al. (2002) reported U-series ages of ~370 ka for a coral (*Montastrea annularis*) from a fossil reef drilled at 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), and correction following Gallup et al. (1994) yields an age closer to 300–340 ka, suggesting that 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 fewer on tectonically relatively stable coasts. Exceptions are a pair of U-series ages of ~200 ka from coral-bearing marine deposits ~2 m above sea level on Bermuda (Muhs et al., 2002), and a single coral age of 235 ± 4 ka from a near-surface *M. annularis* coral in quarry spoil piles on Long Key in the Florida Keys (Muhs et al., 2004). The latter showed higher-than-modern initial $^{234}\text{U}/^{238}\text{U}$, indicating a true age closer to ~220–230 ka using the Gallup et al. (1994) correction scheme. These sparse data suggest that sea level stood close to its present level during MIS 7.

Taken together, these data point to MIS 11 as a time when sea level likely was notably higher than now, although the data are sufficiently sparse that stronger conclusions are not warranted. If so, melting of most or all Greenland's ice seems likely, mostly on the basis of elimination: Greenland meltwater is 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.

3.2.2. Ice-sheet indications

The cold MIS 6 ice age (about 130–188 ka) may have produced the most extensive ice in Greenland (Wilken and Mienert, 2006; also see Roberts et al., 2009). 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 than 1000 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 (~74–130 ka), and particularly from MIS 5e (~123 ka), showing that 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). More Greenland ice causing greater isostatic depression during MIS 6 than during MIS 2 would explain these data. 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 for MIS 2 ice. Also, when the MIS 6 Greenland ice receded to allow seawater incursion, 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, although excess MIS 6 ice may be the simplest explanation.

Ice-core data from central Greenland suggest a major ice retreat, perhaps during stage 11. Willerslev et al. (2007) attempted to amplify DNA in: (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 Dye-3 silty ice yielded much identifiable DNA, which was lacking from the others. This absence at Kap København may indicate post-depositional changes, perhaps during room-temperature storage following collection. The lack of DNA at GRIP shows that there is no important wind-blown source, which suggests that the Dye-3 material has a relatively local source. Dye-3 DNA indicates a northern boreal forest, with mean July temperatures 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), compared to the tundra environment that exists in coastal sites at the same latitude and lower elevation today. 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 (so far, only in an abstract) on cosmogenic isotopes in rock core collected from beneath 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 of exposure to cosmic rays (hence ice cover <1 m or so) $\sim 500 \pm 200$ ka, consistent with the results of Willerslev et al. (2007).

No long, continuous climate records from Greenland itself are available from this time. 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 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., McManus et al., 1999; Bauch et al., 2000; Helmke et al., 2003; Kandiano and Bauch, 2003; De Abreu et al., 2005). 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 in the far-field sea-surface temperatures, so absence of warm temperatures far from the ice sheet does not guarantee absence closer to the ice sheet; see Section 3.3.) The independent indications of high global sea level during MIS 11, as discussed above in Section 3.2.1, 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), so the last few interglacials were 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. 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. A consistent 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 extensive ice coverage and large ice ages began, while temperatures surrounding Greenland were not grossly higher than recently. The rate of mass loss is unconstrained; the long interglacial at MIS 11 allows the possibility of very slow 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.

3.3. Marine isotope stage 5e

3.3.1. Far-field sea-level indications

Widespread coral-reef and marine deposits now above sea level from tectonically stable coasts including Australia, the Bahamas, Bermuda and the Florida Keys document a sea-level high stand during MIS 5e (Muhs, 2002).

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, with most coral growth 128–121 ka (Stirling et al., 1995, 1998). Because the corals grew at some unknown depth below sea level, 4 m is a minimum for the last-interglacial sea-level rise, presuming no additional isostatic corrections.

The islands of the Bahamas are tectonically stable, although perhaps slowly subsiding owing to carbonate loading. Well-preserved fossil reefs, many with corals in growth position, occur up to 5 m above sea level (Chen et al., 1991). On San Salvador Island, reef ages range from 130.3 ± 1.3 to 119.9 ± 1.4 ka. Many fossil reefs contain the coral *Acropora palmata*, which almost always lives within the upper 5 m of the water column (Goreau, 1959), providing a fairly precise constraint on the former water depth.

Tectonically stable Bermuda (Section 3.2.1) lacks MIS 5e fossil reefs, but numerous coral-bearing marine deposits fringe the island, with ages from ~ 119 ka to ~ 113 ka (Muhs et al., 2002). These deposits are 2–3 m above present sea level, although overlying wind-blown sand prevents precise location of the former shoreline.

For the tectonically stable Florida Keys, Fruijtier et al. (2000) reported ages for corals from Windley Key, Upper Matecumbe Key, and Key Largo that, when corrected for high initial $^{234}\text{U}/^{238}\text{U}$ 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 1–2 m above sea level, and on Key Largo 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, local 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, *M. 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, during MIS 5e local sea level was at least 6 m higher than modern-day sea level (Figs. 5 and 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.

The above estimates generally presume that glacial-isostatic adjustment has not notably affected the sites at the key times. Kopp et al. (2009) recently analyzed a global compilation of sea-level markers from MIS 5e using a Bayesian statistical approach that accounts for both the noisy and sparse nature of the database and the geographically variable signature of glacial-isostatic adjustment. In particular, they derived a covariance between local (i.e., site-specific) and globally averaged sea level (GSL) by running many hundreds of ice-age sea-level predictions, and their Bayesian formulation yielded posterior probability functions for GSL and the 1000-year average GSL rate through the last-interglacial. They concluded with 95%, 67% and 33% probability that GSL exceeded 6.6 m, 8.0 m and 9.4 m, respectively, sometime during MIS 5e. The same probability thresholds were reached for millennial average GSL rates of 5.6 m ka⁻¹, 7.4 m ka⁻¹ and 9.2 m ka⁻¹, respectively. They also argued that it was 95% likely that both the Antarctic and Greenland Ice Sheets lost at least 2.5 m of equivalent sea level during MIS 5e (though not necessarily at the same time), though they cautioned that their adoption of Gaussian priors could not incorporate hard bounds on the maximum ice sheet mass loss. In any event, the “very likely” 6.6 m GSL high stand indicates

significant melting of both the Greenland and Antarctic ice sheets (also see Overpeck et al., 2006).

3.3.2. Conditions in Greenland

Paleoclimate data show strong warmth peaking ~130 ka on and around Greenland during MIS 5e. As summarized by CAPE (2006), terrestrial data show peak summertime temperatures ~4 °C above recent in northwest Greenland and ~5 °C above recent in east Greenland (and thus 2–4 °C above the mid-Holocene warmth [~6 ka]; Funder et al., 1998, and see below), with near-shore marine conditions 2–3 °C above recent in east Greenland. Climate-model simulations (Otto-Bliesner et al., 2006) forced by the strong summertime increase of sunshine (insolation) in MIS 5e as compared to now show close agreement, with local summertime warming maxima around Greenland of 4–5 °C in those north-western 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 °C were simulated.

The sea-level record in East Greenland (in and near 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. If correct, this suggests rapid response of the Greenland Ice Sheet to climate forcing.

3.3.3. Ice-sheet changes

The MIS 5e Greenland Ice Sheet covered a smaller area than now, but by how much is not known with certainty. The most

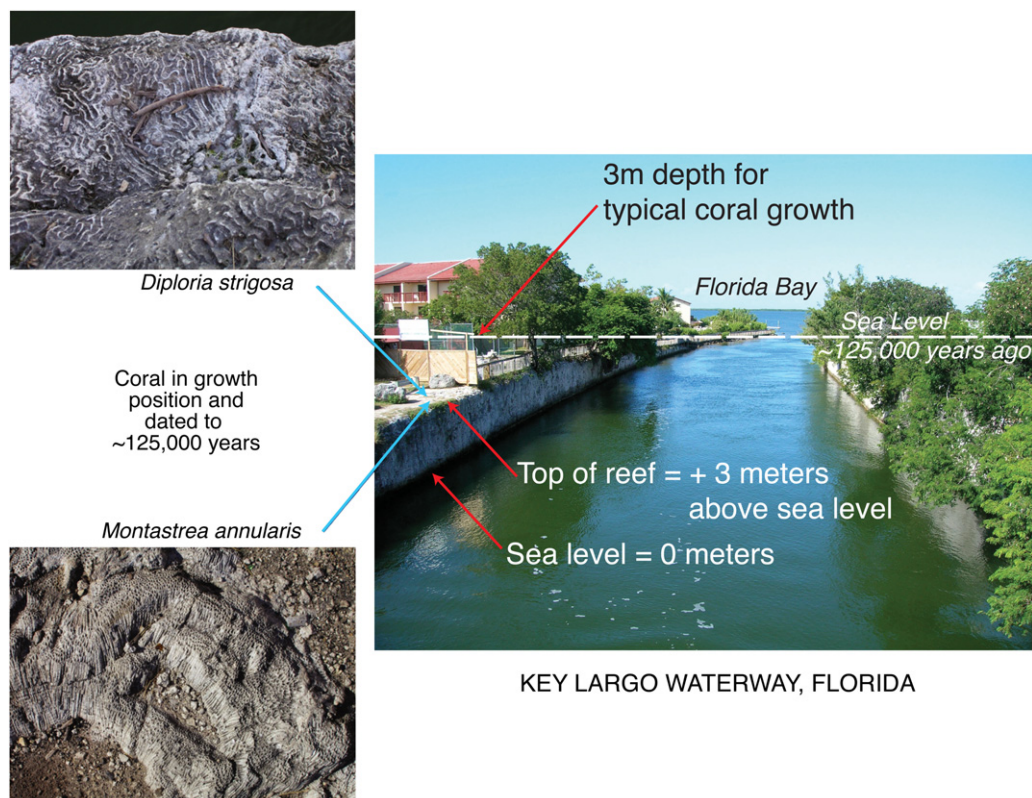


Fig. 5. Photographs of last-interglacial (MIS 5e) reef and corals on Key Largo, Florida, their elevations, probable water depths, and estimated paleo-sea level. Photographs by D.R. Muhs.

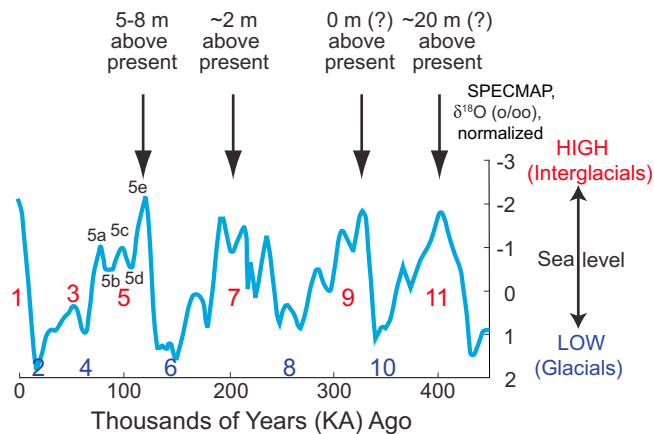


Fig. 6. Oxygen-isotope data from the SPECMAP record (Imbrie et al., 1984), with indications of sea-level stands for different interglacials, assuming minimal glacial-isostatic adjustments to the observed reef elevations. Numbers identify Marine Isotope Stages (MIS) 1–11.

compelling evidence is the absence of pre-MIS 5e ice in the ice cores from south, northwest, and east Greenland (Dye-3, Camp Century, and Renland, respectively). In 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), most likely MIS 5e. Moreover, the isotopic 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, but 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, with older ice beneath each rock, but this seems highly improbable.)

At Dye-3 this basal ice layer has $\delta^{18}\text{O} = -23\text{‰}$, compared to a value of -30‰ for modern snowfall in the source region up-flow from Dye-3. At Camp Century, a value of $\delta^{18}\text{O} = -25\text{‰}$ is reported for basal ice versus -31.5‰ in the source region (see Table 2 of Koerner, 1989). These $\sim 7\text{‰}$ differences 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 regrowing when these MIS 5e ices were deposited. (Note that Carlson et al. (2008) also indicate greater melting of the southern ice dome of Greenland during MIS 5e than during MIS 1, based on marine-sediment records from Eirik Drift.)

In combination, the absence of pre-MIS 5e ice, and anomalously low-elevation sources of the basal ice, indicate considerable retreat of the Greenland margin 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, especially at Dye-3. Unfortunately the picture is cloudy – not unlike the basal ice itself, which has a small amount of silt and sand dispersed through it. 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 firn densification in a cold, dry environment. This clean ice has $\delta^{18}\text{O} = -30.5\text{‰}$. The bottom 4 m of the silty ice is radically different; it has $\delta^{18}\text{O} = -23\text{‰}$, 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 (formed by meltwater refreezing 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 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 among rock particles, by grain-boundary diffusion, and possibly by other processes. Unfortunately, there is no way to distinguish rigorously how much of 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 3.2.2, Willerslev et al. (2007) found that organic material 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.)

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, providing a way to rapidly re-establish the ice sheet in deglaciated regions and to preserve soil or permafrost materials as the ice re-grows. During cooling, sliding advances the ice margin more rapidly than would be possible if the ice were frozen to the bed. Furthermore, the sliding will transport 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 3.5.2), 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, re-glaciation of a site such as Dye-3 might 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 normal, cold-environment, ice-sheet ice from MIS 5e. 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 that 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)) first addressed this question using the information from the central Greenland cores as constraints, and in particular the oxygen isotopic ratios. 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 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 $\alpha = 0.4\text{‰ per }^{\circ}\text{C}$), the southern dome of Greenland completely melted after a sustained (for at least 2000 years) climate warming (mean annual, but with summer most important) of $\sim 7^{\circ}\text{C}$ higher than present. In a different scenario (sensitivity $\alpha = 0.67\text{‰ per }^{\circ}\text{C}$), 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 $5\text{--}6^{\circ}\text{C}$) 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 (Kapsner et al., 1995; Cuffey and Clow, 1997), whereas most models assume that snowfall (and hence accumulation) increases with temperature. This lack of correlation suggests that models are over-predicting the extent to which increased snowfall partly balances 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 add observational constraints to optimize the model parameters. For example, the new ability to model passive tracers in ice sheets (Clarke and Marshall, 2002) allows comparison of predicted and observed isotope profiles at ice-core sites. By using these capabilities, Tarasov and Peltier (2003) estimated MIS 5e ice volume constrained by the measured ice-temperature profiles at GRIP and GISP2 and by the $\delta^{18}\text{O}$ 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 and estimate 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 were the seasonal and latitudinal Milankovitch changes in solar insolation at the top of the atmosphere (Berger, 1978), which produced anomalously high summer insolation in the Northern Hemisphere during the first half of MIS 5e (about 130–123 ka) (Otto-Bliesner et al., 2006; Overpeck et al., 2006). In response, atmosphere–ocean general circulation models (AOGCMs) simulate approximately correct sensitivity, reproducing the proxy-derived summer warmth for the Arctic of up to 5 °C, and placing 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, which has a mid-range climate sensitivity compared to similar models; Kiehl and Gent, 2004), the orbitally induced warmth of MIS 5e caused snow and sea-ice loss, causing 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 Arctic sea-ice extent from April into November, allowing 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 summer. Together with simulated precipitation rates, which overall were not greatly different from present, the simulated mass balance of the Greenland Ice Sheet was negative. Then, as now, the surface of the ice sheet melted primarily in the summer.

Temperatures and precipitation produced by the NCAR CCSM model for 130 ka were then used to drive an updated version of the ice-flow model used by Cuffey and Marshall (2000) (which thus also lacked representations of some physical processes that would accelerate ice-sheet response and increase sensitivity to climate change; warming caused by ice-albedo feedback from bedrock during ice retreat was also omitted). The modeled 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 melting during an extended season led to a negative mass balance. Marginal retreat over several millennia lowered the surface, amplifying the negative mass-balance and accelerating 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. Melting increased in response to both direct effects (warmer atmospheric temperatures) and indirect effects (reduction of ice-sheet altitude and size). The model simulated a steep-sided MIS 5e ice sheet in central and northern Greenland (Otto-Bliesner et al., 2006) (Fig. 7).

If the Greenland Ice Sheet's southern dome did not survive the peak interglacial warmth, as suggested by most available data (see above), the model suggests that the Greenland Ice Sheet contributed 1.9–3.0 m of MIS 5e sea-level rise (another 0.3–0.4 m rise was produced by meltwater from ice on Arctic Canada and Iceland) over 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 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. (Note that 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).)

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

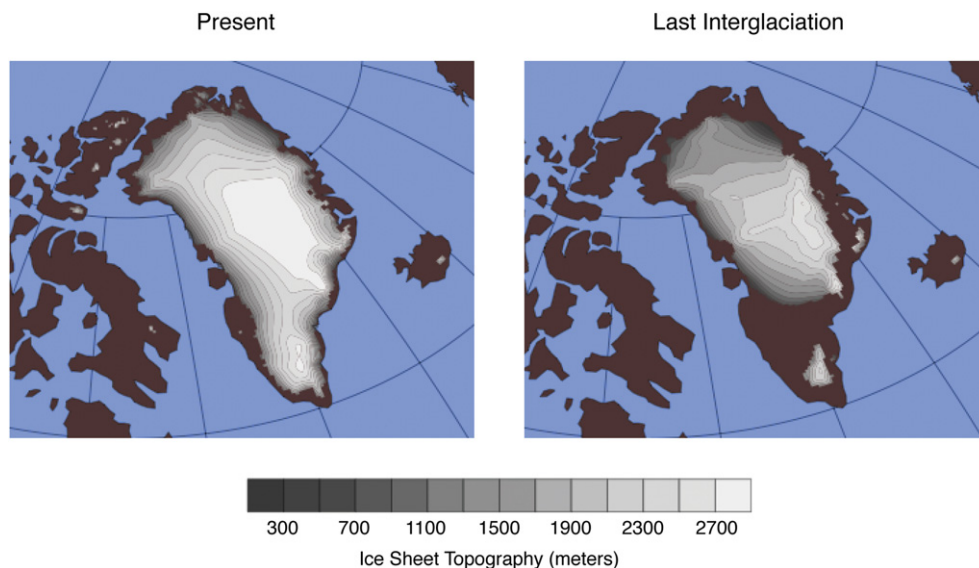


Fig. 7. Modeled configuration of the Greenland Ice Sheet today (left) and in MIS 5e (right), from Otto-Bliesner et al. (2006). [Reprinted with permission of American Association for the Advancement of Science.]

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.

The efforts summarized above suggest that melting of the Greenland Ice Sheet contributed as little as 1–2 m or as much as 4–5 m to sea-level rise during MIS 5e, in response to climatic temperature changes of perhaps 2°–7 °C. The higher numbers for the warming are based on estimates that include the feedbacks from melting of the ice sheet, and the associated sea-level estimates are strongly favored by the statistical/modeling analysis of Kopp et al. (2009). Therefore, central values in the 3–4 m and 3°–4 °C range may be appropriate.

3.4. Post-MIS 5e cooling to the Last Glacial Maximum (LGM, or MIS 2)

3.4.1. 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 (e.g., Cuffey et al., 1995; Jouzel et al., 1997; Dahl-Jensen et al., 1998; Severinghaus et al., 1998; Johnsen et al., 2001).

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 (Fig. 8) from the Eirik Drift (Stoner et al., 1995), rapid sedimentation buried the record from MIS 5e to ~13 m depth. At that site, the $\delta^{18}\text{O}$ change of planktonic foraminiferal shells of ~1.5‰ from MIS 5e to

5d is consistent with cooling as well as ice growth on land, and occurs with a rapid increase in magnetic susceptibility indicating delivery of glacially derived sediments.

The broad picture indicates:

- a general cooling from MIS 5e (about 123 ka) to MIS 2 (coldest temperatures were at about 24 ka in Greenland; Alley et al., 2002),
- warming to the mid-Holocene/MIS 1 a few millennia ago,
- irregular or bumpy cooling into the Little Ice Age of one to a few centuries ago,
- and then a bumpy warming (see Section 3.5.2).

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.

The cooling from MIS 5e to MIS 2, and then warming into MIS 1 (the Holocene), were punctuated by shorter-lived “millennial” events in which central Greenland warmed abruptly – ~10 °C in a few years to decades – cooled gradually, then cooled more abruptly, gradually warmed slightly, and then repeated the sequence (Fig. 9) (also see Alley, 2007). 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). Marine-sediment cores from around the North Atlantic and beyond show temperature histories closely tied to those 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

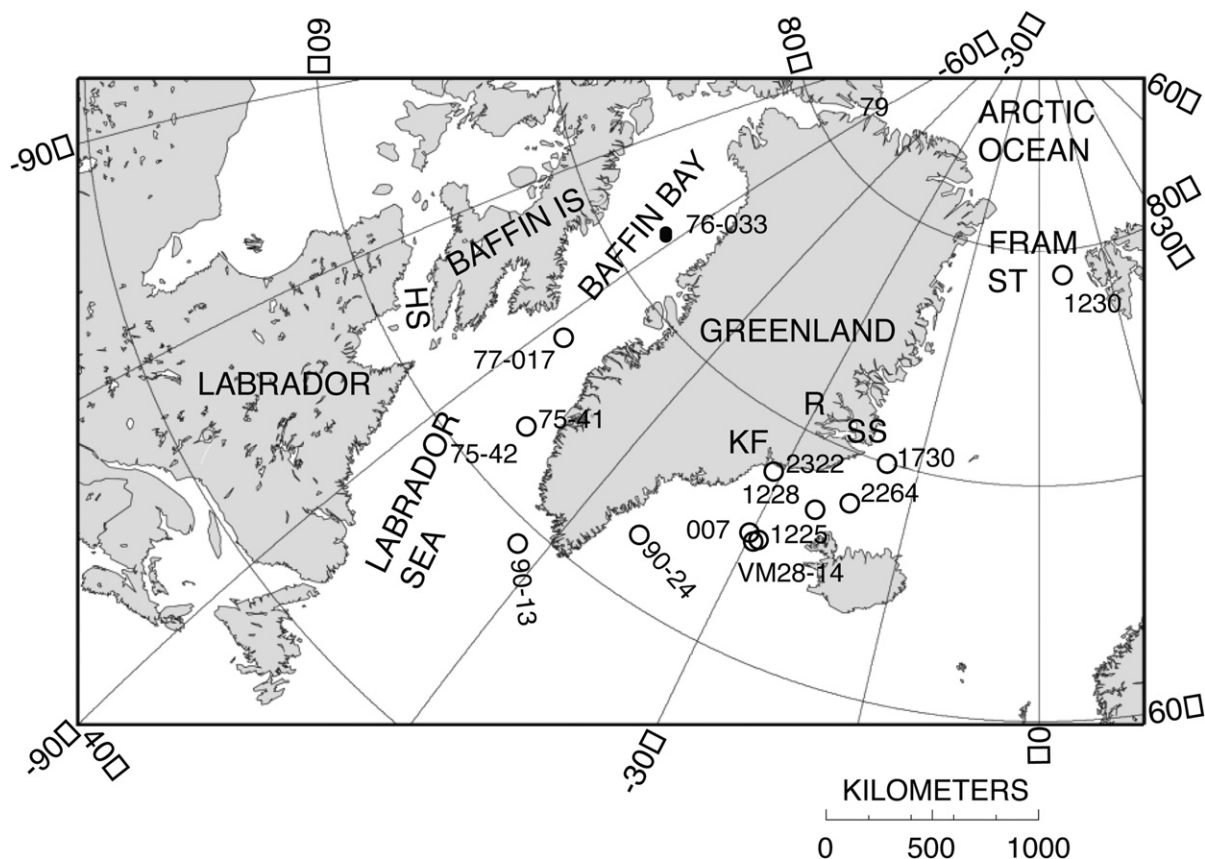


Fig. 8. Location map with core locations discussed in text. Full core identities are as follows: 79 = LSSL2001-079; 75-41 and -42 = HU75-4, -42; 77-017 = HU77-017; 76-033 = HU76-033; 90-013 = HU90-013; 1230 = PS1230; 2264 = PS2264; 1225 and 1228 = JM96-1225, -1228; 007 = HU93-007; 2322 = MD99-2322; 90-24 = SU90-24. HS = Hudson Strait, source for major Heinrich events; R = location of the Renland Ice Cap.

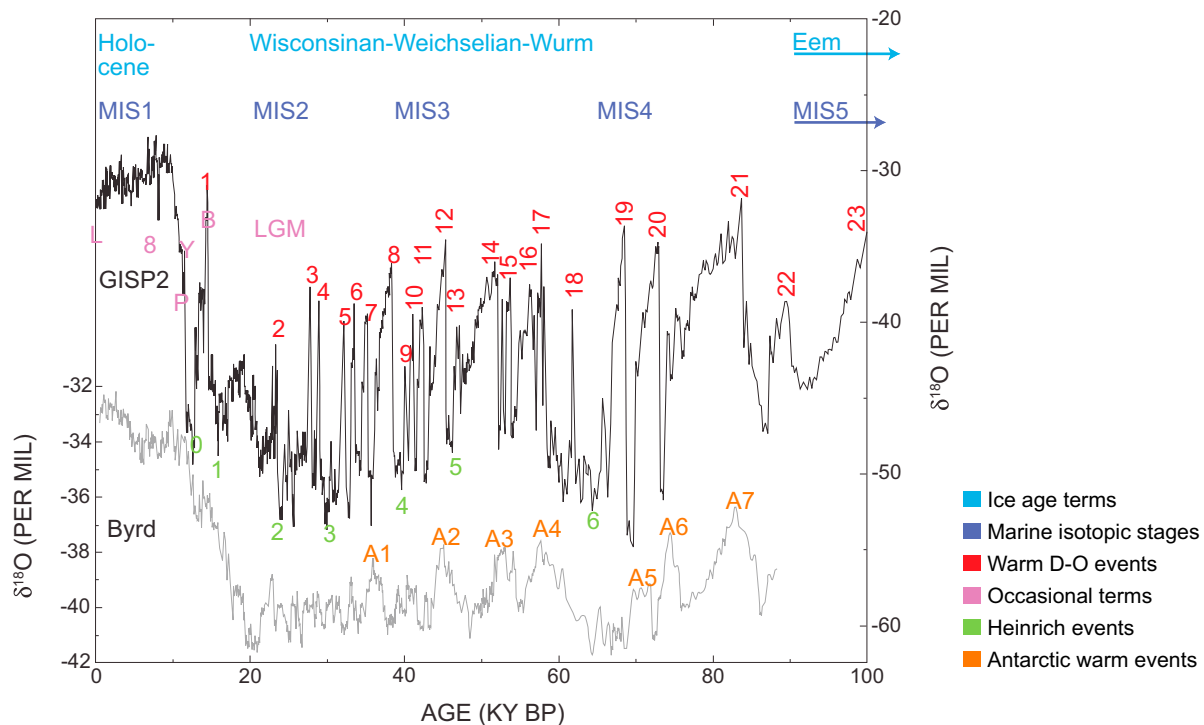


Fig. 9. Ice-isotopic records ($\delta^{18}\text{O}$, a proxy for temperature, with less-negative values indicating warmer conditions) from GISP2, Greenland (Grootes and Stuiver, 1997) (scale on right) and Byrd Station, Antarctica (scale on left), as synchronized by Blunier and Brook (2001), with various climate-event terminology indicated. Ice age terms are shown in light blue (top); the classical Eemian/Sangamonian is slightly older than shown here, as is the peak of marine isotope stage (MIS, shown in dark blue) 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 = 8 k 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). [Reprinted, with permission, from the *Annual Review of Earth and Planetary Sciences*, Volume 35 ©2007 by Annual Reviews www.annualreviews.org].

a modified seesaw in far-southern regions (Fig. 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). Fig. 10, panel A compares near-surface data from a transect of cores (Andrews, 2008) to $\delta^{18}\text{O}$ data from the Renland ice core, just inland from Scoresby Sund (Johnsen et al., 1992a, 2001) (Fig. 8). The complexity observed in this comparison likely arises because of the rich nature of the marine indicators. As noted in Section 2.1, 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). Cooling caused by freshwater-induced reduction in the formation of deep water might cause 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 panel A of Fig. 10 probably 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.

The slower tens-of-millennia cycling of the climate records is well explained by features of Earth's orbit and 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), with colder intervals having more-extensive wintertime sea ice (Denton et al., 2005), in turn coupled to changes in deep-water formation in the North Atlantic and thus

to the “conveyor-belt” circulation (e.g., Broecker, 1995; Alley, 2007). (Note that while the qualitative nature of these events is well-established, 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 (Fig. 9). The material in these debris layers 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 they also contain other materials from almost all glaciated regions around the North Atlantic (Hemming, 2004).

3.4.2. Ice-sheet changes

With certain qualifications, the Greenland Ice Sheet during this time 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 also had a larger ice volume. This result emerges both from limited data on ice-core total gas content (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 1.2, ice-sheet marginal retreat tends to thin central regions, whereas marginal advance tends to thicken central regions. Also, because ice

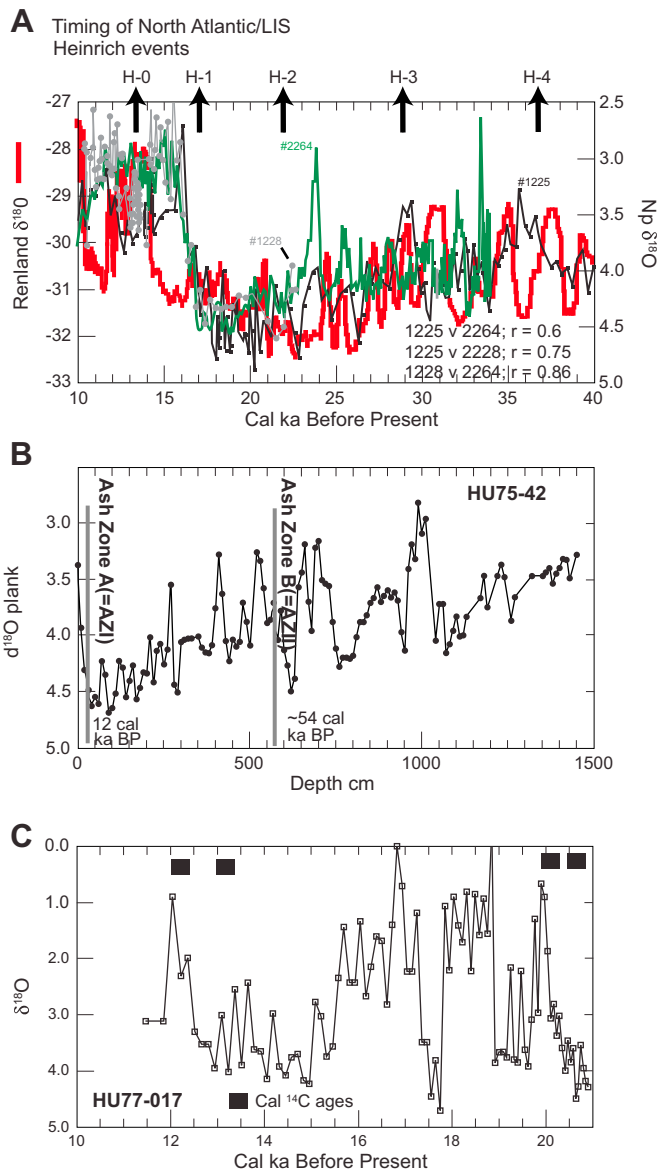


Fig. 10. A) Variations in near-surface plankton $\delta^{18}\text{O}$ values from a series of sediment cores north to south of Denmark Strait (see Fig. 6, 8), namely: PS2264, JM96-1225 and 1228 plotted with the $\delta^{18}\text{O}$ from the Renland Ice Cap. (Correlation coefficients between sediment cores shown in lower right); B) $\delta^{18}\text{O}$ variations in cores HU75-42 (NW Labrador Sea); C) $\delta^{18}\text{O}$ variations in cores HU77-017 from north of the Davis Strait.

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, east Greenland (Funder et al., 1998; also see Hakansson et al., 2009), which show:

- ice advanced during the MIS 5d and 5b coolings without fully filling the Scoresby Sund fjord,
- ice retreated 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),
- ice advanced to the mouth of Scoresby Sund, probably during MIS 4,
- and ice remained there into MIS 2, building the extensive fjord-mouth moraine.

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 Sund mouth 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, while debris-flow activity on the northern portion predates MIS5, the southern portion received deposits fairly recently, suggesting ice advance 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 from ~31–19 ka (Andrews et al., 1997, 1998a; Jennings et al., 2002a). Widespread geomorphic evidence shows that ice reached the shelf break around south Greenland (Funder et al., 2004; Weidick et al., 2004).

In the Thule region of northwestern Greenland, the data are consistent with advances in colder MIS 5d, 5b, 4 and especially MIS 2, retreats in warmer 5c and 5a, possibly in MIS 3, and surely in MIS 1 (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 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 the Ellesmere Island Inuitian 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 known (Blake et al., 1996; Kelly et al., 1999). A recent reconstruction (Funder et al., 2004) suggested that the merged Greenland and North American ice spread to the northeast and southwest along what is now Nares Strait, but with ice shelves rather than grounded ice extending toward the Arctic Ocean and Baffin Bay. The lack of a high marine limit just south of Smith 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 exists between slow 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, preventing strong conclusions.

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.

Looking first at ice extent, the generally accepted picture has been 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 offshore of 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), and Roberts et al. (2009); also see Rinterknecht et al. (2009), suggest that the MIS 2 ice in southwestern Greenland reached the Hellefisk moraines based on reconstructed

thickness near the modern coast. On the east coast, the evidence noted above from the mouth of Scoresby Sund and the trough-mouth fan opens the possibility of more-extensive ice there than is indicated by the generally accepted picture, extending to the mid-shelf or the shelf edge, and this is supported by the new observations on the northeast shelf by Evans et al. (2009). Similarly, the work of Blake et al. (1996) in Greenland's far northwest may indicate that wet-based ice reached the shelf edge. 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 help.

Glaciological understanding indicates that ice sheets almost always respond to climatic or other environmental forcings (such as sufficiently large sea-level change; Alley et al., 2007a). Exceptions include the inability of grounded ice to advance beyond the continental-shelf edge, and relative climatic insensitivity during the advance stage of the tidewater-glacier cycle (Meier and Post, 1987). Thus, 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, 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 extending back through MIS 3 and even into MIS 4 (from Baffin Bay, the Eirik Drift off south-western Greenland, the Irminger and Blossville Basins (e.g., cores SU90-24 and PS2264, Fig. 8), and the Denmark Strait) (Fig. 8) show wide variations in $\delta^{18}\text{O}$ of near-surface planktic foraminifers during MIS 3. These were first documented by Fillon and Duplessy (1980) in cores HU75-041 and -042 from south of Davis Strait (Figs. 8, and 10 panel B), before confident ice-core documentation of large millennial oscillations (Dansgaard–Oeschger or D-O events; Johnsen et al., 1992b; Dansgaard et al., 1993). In addition, Fillon and Duplessy (1980) found “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 (Fig. 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, also found large oscillations in planktonic foraminiferal $\delta^{18}\text{O}$ (Elliot et al., 1998; Hagen, 1999; van Kreveland et al., 2000; Hagen and Hald, 2002) likely recording climate forcing and ice-sheet response (see Section 3.4.1 and Fig. 10A).

Cores from the Scoresby Sund and Kangerdlugssuaq trough mouth fans also have distinct layers rich in ice-rafted debris (Stein et al., 1996; Andrews et al., 1998a; Nam and Stein, 1999). Such layers in cores HU93030-007 and MD99-2260 from the Kangerdlugssuaq trough-mouth fan (Dunhill, 2005) (Fig. 8) are approximately coeval with Heinrich events 3 and 2 (Fig. 9). Similar layers on the Scoresby Sund trough-mouth fan (Stein et al., 1996), although not as well dated, are at least approximately coeval with the Heinrich events. The new results from Verplanck et al. (2009) also implicate at least the southern Greenland Ice Sheet in such fluctuations.

Several reports have invoked the small (Hubbard et al., 2006) Icelandic Ice Sheet as a major contributor to North Atlantic sediment (Bond and Lotti, 1995; Elliot et al., 1998; Grousset et al., 2001), but Farmer et al. (2003) and Andrews (2008) argued instead for a greater role for the eastern Greenland Ice Sheet. Andrews (2008) argued that data from Iceland and Denmark Strait precluded an

Icelandic contribution for Heinrich event 3 (also see Verplanck et al., 2009).

The history of the western margin of the Greenland Ice Sheet during MIS 3 is partially obscured by lack of well-dated records from the trough-mouth fan off Disko Island, and by intermingling of Greenland- and Laurentide-sourced materials in Baffin Bay. 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) (Fig. 11).

Core PS1230 from Fram Strait, which records sediment export by ice sheets around the Arctic Ocean (Darby et al., 2002), shows ice-rafted debris intervals associated with major contributions from north Greenland ~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.

Interpretation of IRD changes continues to be difficult, because IRD 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 causes. The very large changes in sediment fluxes from the 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, but those fluctuations cannot be demonstrated uniquely.

3.5. Ice-Sheet retreat from the Last Glacial Maximum (MIS 2)

3.5.1. Climatic history and forcing

The coldest conditions recorded in Greenland ice cores since MIS 6 were reached about 24 ka (Fig. 9; also see Alley et al., 2002), the approximate age of minimum local mid-summer sunshine and Heinrich event H2. The Denmark Strait sediment-core suite (Figs. 8 and 10A) plus additional cores (VM28-14 and HU93030-007) have a slightly younger $\delta^{18}\text{O}$ extremum (~18–20 ka), with values of 4.6‰ indicating cold, salty waters.

The “orbital” warming signal in ice-core and other records is fairly weak until ~19 ka (Alley et al., 2002). The prominent, very rapid onset of warmth ~14.7 ka (the Bølling interstadial) followed more than 1/3 of the total deglacial warming; that pre-14.7 ka orbital warming was interrupted by Heinrich event H1. Bølling warmth was followed by general cooling (punctuated by the short-lived but prominent Older Dryas and Inter-Allerød cold periods), before faster cooling into the Younger Dryas ~12.8 ka. Gradual warming through the Younger Dryas led to a step warming at the end of the Younger Dryas ~11.5 ka. A ramp warming to above recent values by ~9 ka, punctuated by the short-lived cold event of the Preboreal Oscillation ~11.2–11.4 ka (Björck et al., 1997; Geirsdóttir et al., 1997; Hald and Hagen, 1998; Fisher et al., 2002; Andrews and Dunhill, 2004; van der Plicht et al., 2004; Kobashi et al., 2008), was followed by the short-lived cold event ~8.3–8.2 ka (the “8 k event”; e.g., Alley and Ágústsdóttir, 2005).

The cold times of Heinrich events H2, H1, the Younger Dryas, the 8 k event, and probably other short-lived cold events including the Preboreal Oscillation are linked to greatly expanded wintertime sea ice in response to decreased near-surface salinity and strength of

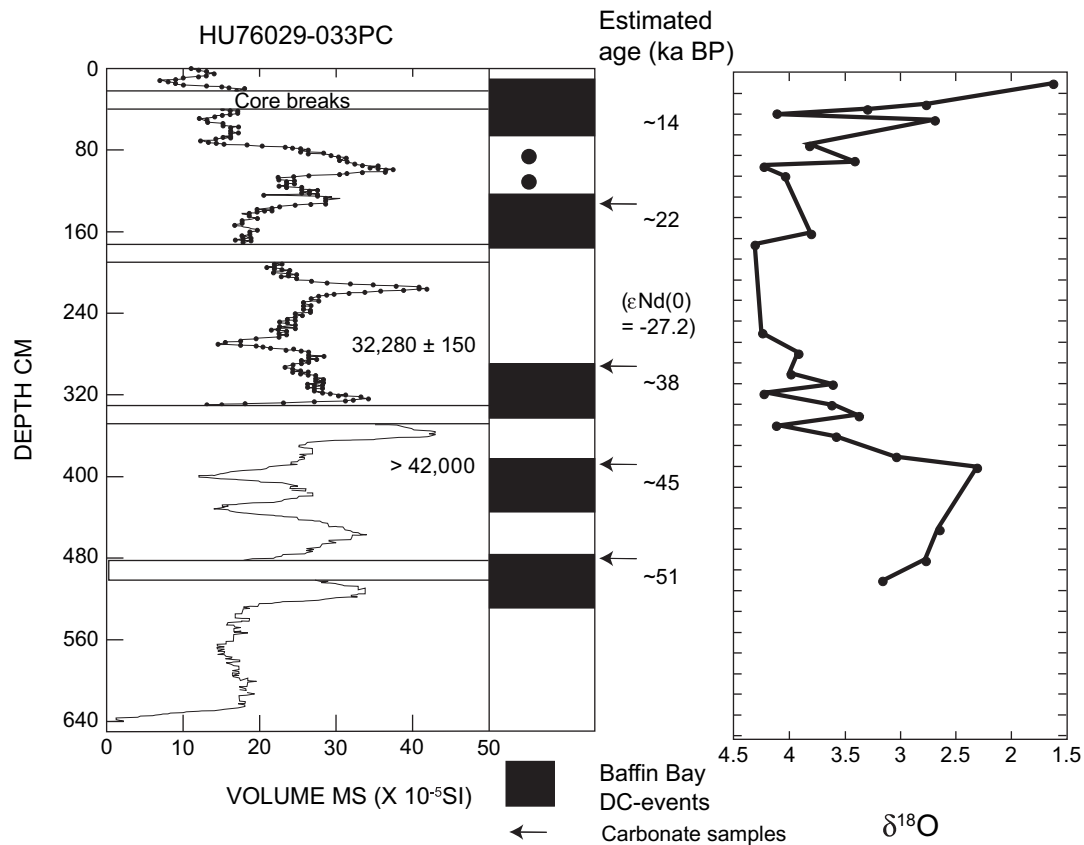


Fig. 11. Variations in detrital carbonate in core HU76-033 from Baffin Bay showing down-core variations in magnetic susceptibility and $\delta^{18}\text{O}$.

the North Atlantic overturning circulation (e.g., Alley, 2007). The cooling in and near Greenland with these oceanic changes was largest in winter (Denton et al., 2005) but probably affected summer (cf. Björck et al., 2002 and Jennings et al., 2002a).

Peak MIS 1/Holocene summertime warmth before and after the 8 k event was, for roughly millennial averages, $\sim 1.3^\circ\text{C}$ 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 $\sim 2^\circ\text{C}$ (and with correspondingly larger wintertime changes); other indicators are consistent with this interpretation (Alley et al., 1999; also see Vinther et al., 2009). 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 $\sim 1^\circ\text{C}$ or slightly more (Johnsen, 1977; Alley and Koci, 1990; Cuffey et al., 1994).

3.5.2. Ice-sheet changes

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

The time of onset of retreat from the Last Glacial Maximum is poorly known because most of the evidence is now below sea level. Funder et al. (1998) suggested that Scoresby Sund ice was most extended ~ 24 – 19 ka, with retreat beginning when many ice

masses began notable contribution to sea-level rise (e.g., Peltier and Fairbanks, 2006). Roberts et al. (2009) suggested that ice-sheet thinning in the southwest began ~ 21 ka and was clearly occurring after ~ 18 ka.

Extensive deglaciation that left clear records is typically more recent. A core from Hall Basin (core 79, Fig. 8), the northernmost of the basins between northwest Greenland and Ellesmere Island, has foraminifera dating to ~ 16.2 ka, showing retreat of the land ice flowing to the Arctic Ocean by this time (Mudie et al., 2006). At Sermilik Fjord in southwest Greenland, retreat from the shelf preceded ~ 16 ka (Funder, 1989c). The ice margin 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 Younger Dryas climatic anomaly) (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 (Fig. 12).

Along most of the Greenland coast, radiocarbon dates much older than the end of the Younger Dryas are rare, likely because the Greenland Ice Sheet had not yet retreated. Radiocarbon dates become common near the end of the Younger Dryas and especially during the Preboreal interval, and 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

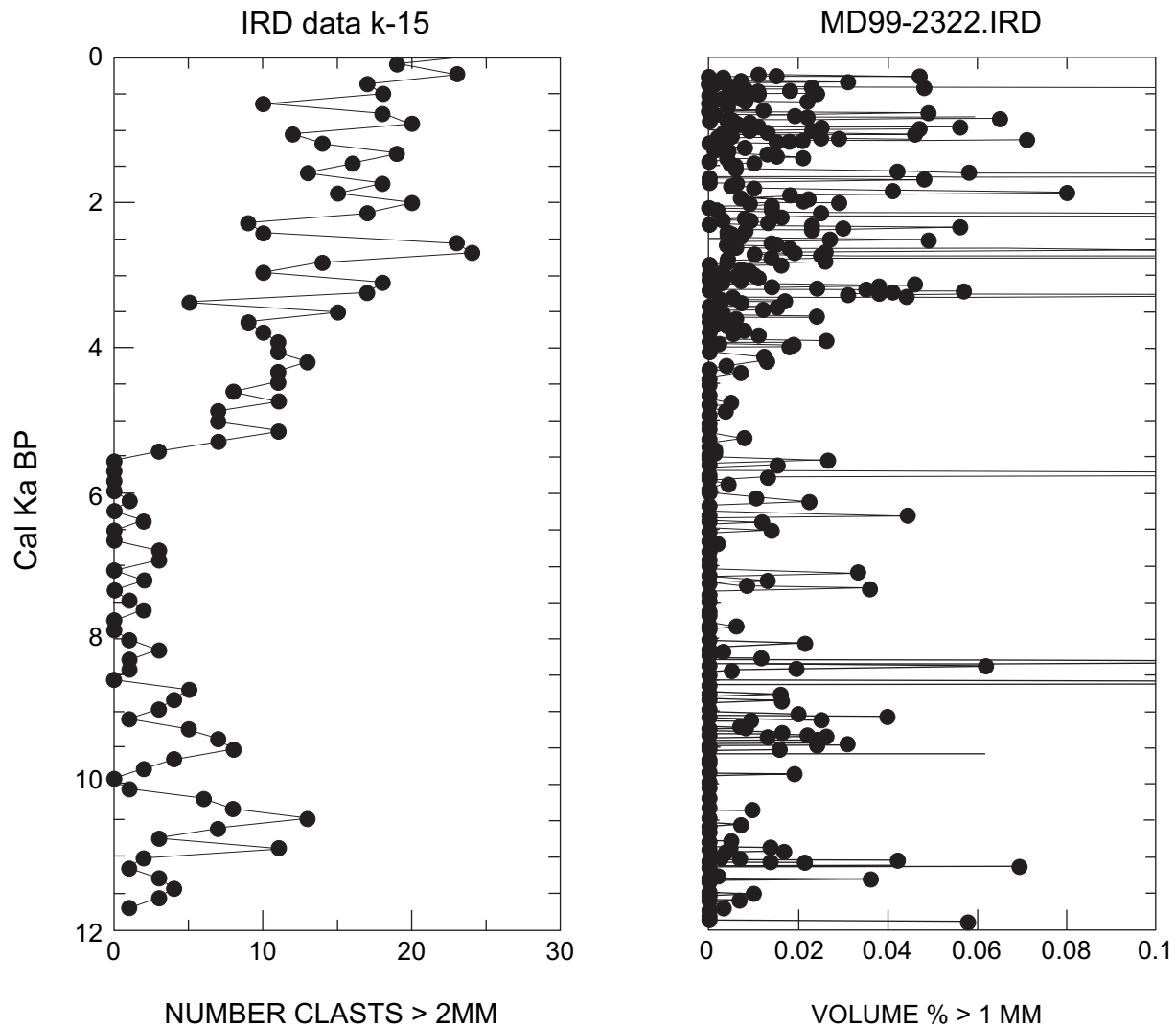


Fig. 12. Inputs of ice-rafted debris in two cores off east Greenland (see Fig. 8). The data for K-15 is based on counts of clasts >2 mm counted in 2-cm thick depth increments from X-radiographs, whereas the data from MD9999-2322 (Jennings et al., 2006) are the log of weights of the >1 mm sediment fraction.

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, differentiation of the Preboreal Oscillation from the longer Preboreal generally is not possible, and linking particular dates to the Preboreal versus the Younger Dryas often is difficult.

Given the prominence of the end of the Younger Dryas cold event in ice-core records ($\sim 10^\circ\text{C}$ warming in about 10 years; Severinghaus et al., 1998), lack of confidently dated moraines abandoned in response to the warming may seem surprising. Part of the difficulty is solved by the Denton et al. (2005) result that most of the warming was in winter. Björck 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.

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 reflects the rising mid-summer insolation (a function of Earth's orbit). The Younger Dryas was long enough for coastal-mountain glaciers to reflect both the cooling into the event and the warming during the event before the terminal step, and the ice-sheet margin probably behaved similarly (as discussed in Section 3.4.2, and below). As shown by Vacco et al. (2009), this climate history would produce moraine sets 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 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. Funder et al. (1998) suggested that the last resurgence of glaciers in the Scoresby Sund region of east Greenland, known as the Milne Land Stade, was correlated with the Preboreal Oscillation, although a Younger Dryas age for at least some of the moraines could not be excluded (Funder et al., 1998; Denton et al., 2005). The additional work of Kelly et al. (2008) and Hall et al. (2008) (also see Miller, 2008) is perhaps most easily interpreted as indicating a more-extensive early Younger Dryas advance followed by a less-extensive Preboreal Oscillation advance. Although data and modeling remain sufficiently sketchy that strong conclusions do not seem warranted, 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 ~ 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 was interrupted by intervals of moraine formation, especially ~ 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 8 k cold event (Long et al., 2006). The onset of late Holocene readvance is not well-dated. Funder (1989c) suggested ~ 3 ka for west Greenland, the approximate time when relative sea-level fall from isostatic rebound switched to begin relative sea-level rise of ~ 5 m, probably in response to depression of the land by the advancing ice load (also see Long et al., 2009). Similar considerations place the onset of readvance somewhat earlier in the south, where relative sea-level fall switched to relative rise of ~ 10 m beginning ~ 8 –6 ka (Sparrenbom et al., 2006a,b).

The late Holocene advance culminated in different areas at different times, especially in the mid-1700s, 1850–1890, and ~ 1920 (Weidick et al., 2004), followed by retreat. (For an extensive review of changes in small glaciers detached from the main ice sheet, see Kelly and Lowell (2009). Behavior has been broadly similar to the main ice sheet, although small glaciers complete their response to climate changes more rapidly.) Evidence of relative sea-level changes is consistent with this deglacial history (Funder, 1989d; Tarasov and Peltier, 2002, 2003; Fleming and Lambeck, 2004). Fleming and Lambeck (2004) used an iterative technique to reconstruct the ice-sheet volume over time to match relative sea-level curves based on indicators including flights of raised beaches on many coasts of Greenland, and as much as 160 m above modern sea level in West Greenland. They obtained a Last Glacial Maximum ice-sheet volume $\sim 42\%$ larger than modern (3.1 m of additional sea-level equivalent, compared with the modern 7.3 m; interestingly, Huybrechts (2002) obtained a model-based estimate of 3.1 m of excess ice at the Last Glacial Maximum, while Simpson et al. (2009) obtained 4.1 m). 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 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 more than 1500 m 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, introducing additional uncertainty.

In the ICE5G model, Peltier (2004) (with 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, providing 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). The previous ICE4G reconstruction had an excess ice volume during the

Last Glacial Maximum relative to the present of 6 m; this is 2.8 m in ICE5G. 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 modern. In a similar, more-recent modeling effort, Simpson et al. (2009) estimated an excess ice volume at glacial maximum of 4.1 m of sea-level equivalent, increasing to 4.6 m at 16.5 ka before shrinking to 0.17 m smaller than present 5–4 ka, and then re-growing. Carlson et al. (2008) presented marine-sediment evidence that the Greenland Ice Sheet in south Greenland retreated essentially synchronously with regional warming.

The 20th century warmed from the Little Ice Age to about 1930, sustained this 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,b,c), and retreat and mass loss are now widespread (e.g., Alley et al., 2005a). Study of declassified satellite images shows that at least Helheim Glacier in the southeast of Greenland 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 (Kahn et al., 2007). It is also consistent with GRACE satellite gravity observations, which indicate a mean mass loss in the period from April, 2002 to April, 2006 equivalent to 0.5 mm a^{-1} of globally uniform sea-level rise (Velicogna and Wahr, 2006).

As discussed in Section 2.2.5, geodetic measurements of perturbations in Earth's rotational state also help constrain estimates of 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 that 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, finding that an anomalous 20th century contribution of as much as about 1 mm a^{-1} of sea-level rise is consistent with the data; partitioning of this value into signals from melting of mountain glaciers, Antarctic ice, and the Greenland Ice Sheet is non-unique. 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 a^{-1} of sea-level rise.

4. Discussion

Glaciers and ice sheets are controlled by many climatic factors and by internal dynamics. Attribution of a given ice-sheet change to a particular cause is generally difficult, and 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 grew with reduced precipitation (snowfall) and shrank when snowfall increased, so precipitation changes cannot have controlled ice-sheet behavior. However, the data do not preclude the possibility that 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 and more-widespread temperatures are closely correlated. 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.

It appears unlikely that something else controlled both the temperature and the ice sheet. The only physically plausible control is sea level, with warming causing ice melting beyond Greenland, and the resulting sea-level rise forcing retreat of the Greenland Ice Sheet by floating marginal regions. However, data show times when this explanation is not sufficient. As described in Section 3.2.2, Greenland deglaciation seems to have led global sea-level rise at MIS 6. Ice expanded with cooling from MIS 5e to MIS 5d from a reduced state that 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 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. (2007a), 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., 2007a). 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, many of the changes in the ice sheet clearly 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 1.2. Summertime warming increases melt-water 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 (e.g., Holland et al., 2008), reducing friction at the ice-water interface and so increasing 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” here in

discussing ice-sheet forcing. In detail, ocean and air temperature will not correlate exactly, so additional studies may quantify the relative importance of changes in ocean and air temperatures.

Most of the forcings of past ice-sheet behavior considered here were applied slowly. Orbital changes in sunshine, greenhouse-gas forcing, and sea level have all occurred 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, records are available with sufficiently high-time-resolution and sufficiently accurate dating 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.

Many of the observations are summarized in Fig. 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. 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., 2007a). 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 Fig. 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 ice-sheet shrinkage.

5. Synopsis

Paleoclimatic data show that the Greenland Ice Sheet has changed greatly with time. From physical understanding, 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 values (the last few millennia). Snow accumulation rate is inversely related to ice-sheet volume (less ice when snowfall is higher), and so 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

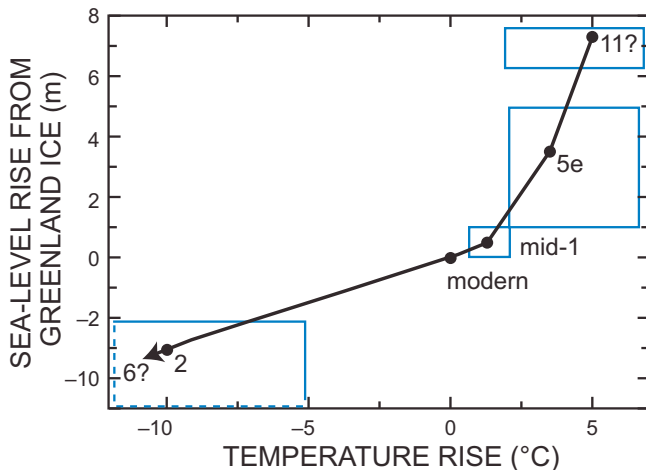


Fig. 13. A best-guess representation of the dependence of the volume of the Greenland Ice Sheet on temperature. Large uncertainties should be understood, and any ice-volume changes in response to sea-level changes correlated with temperature changes are included (although, as discussed in the text, temperature 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 $\sim 6^\circ\text{C}$ 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 $\sim 15^\circ\text{C}$ (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 $\sim 10^\circ\text{C}$ 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 the lower end somewhat arbitrarily set as 1 m. 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 $\sim 0.5\text{ m}$ 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 sections 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 Section 3.3.3. 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.

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.

Acknowledgements

RBA acknowledges partial support from the US National Science Foundation under grants 0531211 and 0424589. GHM acknowledges partial support from the US National Science Foundation under grants ARC0714074 and ATM0318479. LP acknowledges partial support from the US National Science Foundation under grants ARC0612473 and 0806999. JWCW acknowledges partial support from the US National Science Foundation under grants 0806387, 053759, and 059512. BLO acknowledges support from the US National Science Foundation through its sponsorship of the National Center for Atmospheric Research.

References

- Adrielsson, L., Alexanderson, H., 2005. Interactions between the Greenland Ice Sheet and the Liverpool Land coastal ice cap during the last two glaciation cycles. *Journal of Quaternary Science* 20, 269–283.
- Aksu, A.E., 1985. Climatic and oceanographic changes over the past 400,000 years – evidence from deep-sea cores on Baffin Bay and David Strait. In: Andrews, J.T. (Ed.), *Quaternary Environments – Eastern Canadian Arctic, Baffin Bay and Western Greenland*. Allen and Unwin, Boston, pp. 181–209.
- Alley, R.B., 2007. Wally was right – predictive ability of the North Atlantic “conveyor belt” hypothesis for abrupt climate change. *Annual Review of Earth and Planetary Sciences* 35, 241–272.
- Alley, R.B., Ágústssdóttir, A.M., 2005. The 8 k event: cause and consequences of a major Holocene abrupt climate change. *Quaternary Science Reviews* 24, 1123–1149. doi:10.1016/j.quascirev.2004.12.004.
- Alley, R.B., Anandakrishnan, S., 1995. Variations in melt-layer frequency in the GISP2 ice core – implications for Holocene summer temperatures in central Greenland. *Annals of Glaciology* 21, 64–70.
- Alley, R.B., Cuffey, K.M., 2001. Oxygen- and hydrogen-isotopic ratios of water in precipitation – beyond paleothermometry. In: Valley, J.W., Cole, D. (Eds.), *Stable Isotope Geochemistry. Reviews in Mineralogy and Geochemistry* 43, 527–553.
- Alley, R.B., Koci, B.R., 1990. Recent warming in central Greenland? *Annals of Glaciology* 14, 6–8.
- Alley, R.B., MacAyeal, D.R., 1994. Ice-rafted debris associated with binge/purge oscillations of the Laurentide Ice Sheet. *Paleoceanography* 9 (4), 503–511.
- Alley, R.B., Whillans, I.M., 1984. Response of the East Antarctic ice sheet to sea-level rise. *Journal of Geophysical Research* 89C, 6487–6493.
- Alley, R.B., Blankenship, D.D., Rooney, S.T., Bentley, C.R., 1987. Till beneath ice stream B. 4. A coupled ice-till flow model. *Journal of Geophysical Research* 92B, 8931–8940.
- Alley, R.B., Meese, D.A., Shuman, C.A., Gow, A.J., Taylor, K.C., Grootes, P.M., White, J.W.C., Ram, M., Waddington, E.D., Mayewski, P.A., Zielinski, G.A., 1993. Abrupt increase in snow accumulation at the end of the Younger Dryas event. *Nature* 362, 527–529.
- Alley, R.B., Finkel, R.C., Nishiizumi, K., Anandakrishnan, S., Shuman, C.A., Mershon, G.R., Zielinski, G.A., Mayewski, P.A., 1995a. Changes in continental and sea-salt atmospheric loadings in central Greenland during the most recent deglaciation. *Journal of Glaciology* 41 (139), 503–514.
- Alley, R.B., Gow, A.J., Johnsen, S.J., Kipfstuhl, J., Meese, D.A., Thorsteinsson, Th., 1995b. Comparison of deep ice cores. *Nature* 373, 393–394.
- Alley, R.B., Gow, A.J., Meese, D.A., Fitzpatrick, J.J., Waddington, E.D., Bolzan, J.F., 1997. Grain-scale processes, folding, and stratigraphic disturbance in the GISP2 ice core. *Journal of Geophysical Research* 102 (C12), 26,819–26,830.
- Alley, R.B., Ágústssdóttir, A.M., Fawcett, P.J., 1999. Ice-core evidence of late-Holocene reduction in north Atlantic ocean heat transport. In: Clark, P.U., Webb, R.S., Keigwin, L.D. (Eds.), *Mechanisms of Global Climate Change at Millennial Time Scales. Geophysical Monograph*, vol. 112. American Geophysical Union, Washington, DC, pp. 301–312.
- Alley, R.B., Anandakrishnan, S., Jung, P., 2001. Stochastic resonance in the North Atlantic. *Paleoceanography* 16, 190–198.
- Alley, R.B., Brook, E.J., Anandakrishnan, S., 2002. A northern lead in the orbital band – North–south phasing of ice-age events. *Quaternary Science Reviews* 21, 431–441.
- Alley, R.B., Clark, P.U., Huybrechts, P., Joughin, I., 2005a. Ice-sheet and sea-level changes. *Science* 310, 456–460.
- Alley, R.B., Dupont, T.K., Parizek, B.R., Anandakrishnan, S., 2005b. Access of surface meltwater to beds of sub-freezing glaciers – preliminary insights. *Annals of Glaciology* 40, 8–14.
- Alley, R.B., Anandakrishnan, S., Dupont, T.K., Parizek, B.R., Pollard, D., 2007a. Effect of sedimentation on ice-sheet grounding-line stability. *Science* 315, 1838–1841.
- Alley, R.B., Spencer, M.K., Anandakrishnan, S., 2007b. Ice-sheet mass balance: assessment, attribution and prognosis. *Annals of Glaciology* 46, 1–7.
- Alley, R.B., Fahnestock, M., Joughin, I., 2008. Climate change: understanding glacier flow in changing times (perspective). *Science* 322, 1061–1062.
- Allison, I., Alley, R.B., Fricker, H.A., Thomas, R.H., Warner, R.C., 2009. Ice sheet mass balance and sea level. *Antarctic Science* 21, 413–426.

- Andersen, K.K., Svensson, A., Rasmussen, S.O., Steffensen, J.P., Johnsen, S.J., Bigler, M., Röthlisberger, R., Ruth, U., Siggaard-Andersen, M.-L., Dahl-Jensen, D., Vinther, B.M., Clausen, H.B., 2006. The Greenland ice core chronology 2005, 15–42 ka. Part 1: constructing the time scale. *Quaternary Science Reviews* 25 (23–24), 3246–3257.
- Andresen, C.S., Björck, S., Bennike, O., Bond, G., 2004. Holocene climate changes in southern Greenland – evidence from lake sediments. *Journal of Quaternary Science* 19, 783–795.
- Andrews, J.T., 2008. The role of the Iceland Ice Sheet in sediment delivery to the North Atlantic during the late Quaternary – how important was it? Evidence from the area of Denmark Strait. *Journal of Quaternary Science* 23, 3–20.
- Andrews, J.T., Dunhill, G., 2004. Early to mid-Holocene Atlantic water influx and deglacial meltwater events, Beaufort Sea slope, Arctic Ocean. *Quaternary Research* 61, 14–21.
- Andrews, J.T., Erlenkeuser, H., Tedesco, K., Aksu, A., Jull, A.J.T., 1994. Late Quaternary (Stage 2 and 3) meltwater and Heinrich events, NW Labrador Sea. *Quaternary Research* 41, 26–34.
- Andrews, J.T., Jennings, A.E., Cooper, T., Williams, K.M., Mienert, J., 1996. Late Quaternary sedimentation along a fjord to shelf (trough) transect, East Greenland (ca. 68°N). In: Andrews, J.T., Austin, W., Bergsten, H., Jennings, A.E. (Eds.), *Late Quaternary Paleogeography of North Atlantic Margins*. Geological Society, London, pp. 153–166.
- Andrews, J.T., Smith, L.M., Preston, R., Cooper, T., Jennings, A.E., 1997. Spatial and temporal patterns of iceberg rafting (IRD) along the East Greenland margin, ca. 68 N, over the last 14 cal. ka. *Journal of Quaternary Science* 12, 1–13.
- Andrews, J.T., Cooper, T.A., Jennings, A.E., Stein, A.B., Erlenkeuser, H., 1998a. Late Quaternary iceberg-rafted detritus events on the Denmark Strait/Southeast Greenland continental slope (about 65° N) – related to North Atlantic Heinrich events? *Marine Geology* 149, 211–228.
- Andrews, J.T., Kirby, M.E., Aksu, A., Barber, D.C., Meese, D., 1998b. Late Quaternary detrital carbonate (DC-) events in Baffin Bay (67°–74° N) – do they correlate with and contribute to Heinrich Events in the North Atlantic? *Quaternary Science Reviews* 17, 1125–1137.
- Applegate, P.J., Lowell, T.V., Alley, R.B., 2008. Comment on “Absence of cooling in New Zealand and the adjacent ocean during the Younger Dryas chronozone”. *Science* 320, 746d. doi:10.1126/science.1152098.
- Bamber, J.L., Layberry, R.L., Gogineni, S.P., 2001. A new ice thickness and bed data set for the Greenland Ice Sheet 1 – measurement, data reduction, and errors. *Journal of Geophysical Research* 106, 33,773–33,780.
- Bamber, J.L., Alley, R.B., Joughin, I., 2007. Rapid response of modern day ice sheets to external forcing. *Earth and Planetary Science Letters* 257, 1–13.
- Bard, E., Frank, M., 2006. Climate change and solar variability – what's new under the sun? *Earth and Planetary Science Letters* 248, 1–14.
- Bauch, H.A., Erlenkeuser, H., Helmke, J.P., Struck, U., 2000. A paleoclimatic evaluation of marine oxygen isotope stage 11 in the high-northern Atlantic (Nordic seas). *Global and Planetary Change* 24, 27–39.
- Baur, O., Kuhn, M., Featherstone, W.E., 2009. GRACE-derived ice-mass variations over Greenland by accounting for leakage effects. *Journal of Geophysical Research-Solid Earth* 114 Art. No. B06407.
- Bender, M.L., Fairbanks, R.G., Taylor, F.W., Matthews, R.K., Goddard, J.G., Broecker, W.S., 1979. Uranium-series dating of the Pleistocene reef tracts of Barbados, West Indies. *Geological Society of America Bulletin Part 1* 90, 577–594.
- Bennike, O., Bocher, J., 1994. Land biotas of the last interglacial–glacial cycle on Jameson Land, East Greenland. *Boreas* 23, 479–487.
- Bennike, O., Björck, S., Lambeck, K., 2002. Estimates of South Greenland late-glacial ice limits from a new relative sea level curve. *Earth and Planetary Science Letters* 197, 171–186.
- Berger, A.L., 1978. Long-term variations of caloric insolation resulting from the earth's orbital elements. *Quaternary Research* 9, 139–167.
- Berger, A., Loutre, M.F., 1991. Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10, 297–317.
- Bigg, G.R., 1999. An estimate of the flux of iceberg calving from Greenland. Arctic, Antarctic, and Alpine Research 31, 174–178.
- Björck, S., Rundgren, M., Ingolfsson, O., Funder, S., 1997. The Preboreal oscillation around the Nordic Seas – terrestrial and lacustrine responses. *Journal of Quaternary Science* 12, 455–465.
- Björck, S., Bennike, O., Rose, P., Andresen, C.S., Bohncke, S., Kaas, E., Conley, D., 2002. Anomalously mild Younger Dryas summer conditions in southern Greenland. *Geology* 30, 427–430.
- Blake Jr., W., Jackson, H.R., Currie, C.G., 1996. Seafloor evidence for glaciation, northernmost Baffin Bay. *Bulletin of the Geological Survey of Denmark* 43, 157–168.
- Blunier, T., Brook, E.J., 2001. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291, 109–112.
- Bond, G., Broecker, W., Johnson, S., McManus, J., Labeyrie, L., Jouzel, J., Bonani, G., 1993. Correlation between climate records from North Atlantic sediments and Greenland ice. *Nature* 365, 507–508.
- Bond, G.C., Lott, R., 1995. Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation. *Science* 267, 1005–1009.
- Box, J.E., Bromwich, D.H., Veenhuis, B.A., Bai, L.-S., Stroeve, J.C., Rogers, J.C., Steffen, K., Haran, T., Wang, S.-H., 2006. Greenland ice-sheet surface mass balance variability (1988–2004) from calibrated Polar MM5 output. *Journal of Climate* 19 (12), 2783–2800.
- Bradley, R.S., 1999. *Paleoclimatology – Reconstructing Climate of the Quaternary*. Academic Press, San Diego, 613 pp.
- Braun, H., Christ, L.M., Rahmstorf, S., Ganopolski, A., Mangini, A., Kubatzki, C., Roth, K., Kromer, B., 2005. Possible solar origin of the 1,470-year glacial climate cycle demonstrated in a coupled model. *Nature* 438, 208–211.
- Brodersen, K.P., Bennike, O., 2003. Interglacial Chironomidae (Diptera) from Thule, Northwest Greenland – matching modern analogues to fossil assemblages. *Boreas* 32, 560–565.
- Broecker, W.S., 1995. *The Glacial World According to Wally*. Eldigio Press, Palisades, NY, 89 pp.
- CircumArctic PaleoEnvironments (CAPE) – Last Interglacial Project Members:, Anderson, P., Bennike, O., Bigelow, N., Brigham-Grette, J., Duvall, M., Edwards, M., Frechette, B., Funder, S., Johnsen, S., Knies, J., Koerner, R., Lozhkin, A., Marshall, S., Matthiessen, J., Macdonald, G., Miller, G., Montoya, M., Muhs, D., Otto-Bliesner, B., Overpeck, J., Reeh, N., Sejrup, H.P., Spielhagen, R., Turner, C., Velichko, A., 2006. Last Interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25, 1383–1400.
- Carlson, A.E., Stoner, J.S., Donnelly, J.P., Hillaire-Marcel, C., 2008. Response of the southern Greenland Ice Sheet during the last two deglaciations. *Geology* 36, 359–362.
- Cazenave, A., 2006. How fast are the ice sheets melting? *Science* 314, 1250–1252.
- Cazenave, A., Dominh, K., Guinehut, S., Berthier, E., Llovel, W., Ramillien, G., Ablain, M., Larnicol, G., 2009. Sea level budget over 2003–2008: a reevaluation from GRACE space gravimetry, satellite altimetry and Argo. *Global and Planetary Change* 65, 83–88.
- Chappellaz, J., Brook, E., Blunier, T., Malaize, B., 1997. CH₄ and δ¹⁸O of O₂ records from Antarctic and Greenland ice – a clue for stratigraphic disturbance in the bottom part of the Greenland Ice Core Project and the Greenland Ice Sheet Project 2 ice cores. *Journal of Geophysical Research* 102 (C12), 26,547–26,557.
- Chen, J.H., Curran, H.A., White, B., Wasserburg, G.J., 1991. Precise chronology of the last interglacial period – ²³⁴U–²³⁰Th data from fossil coral reefs in the Bahamas. *Geological Society of America Bulletin* 103, 82–97.
- Chen, J.L., Wilson, C.R., Tapley, B.D., 2006. Satellite gravity measurements confirm accelerated melting of Greenland ice sheet. *Science* 313 (5795), 1958–1960.
- Clarke, G.K.C., Marshall, S.J., 2002. Isotopic balance of the Greenland Ice Sheet – modeled concentrations of water isotopes from 30,000 BP to present. *Quaternary Science Reviews* 21, 419–430.
- Clarke, G.K.C., Lhomme, N., Marshall, S.J., 2005. Tracer transport in the Greenland Ice Sheet – three-dimensional isotopic stratigraphy. *Quaternary Science Reviews* 24, 155–171.
- Cronin, T.M., 1999. *Principles of Paleoclimatology*. Columbia University Press, New York, 560 pp.
- Cuffey, K.M., Brook, E.J., 2000. Ice sheets and the ice-core record of climate change. In: Jacobson, M.C., Charlson, R.J., Rodhe, H., Orians, G.H. (Eds.), *Earth System Science – From Biogeochemical Cycles to Global Change*. Academic Press, Burlington, MA, pp. 459–497.
- Cuffey, K.M., Clow, G.D., 1997. Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition. *Journal of Geophysical Research* 102 (C12), 26,383–26,396.
- Cuffey, K.M., Marshall, S.J., 2000. Substantial contribution to sea-level rise during the last interglacial from the Greenland Ice Sheet. *Nature* 404, 591–594.
- Cuffey, K.M., Alley, R.B., Grootes, P.M., Bolzan, J.F., Anandakrishnan, S., 1994. Calibration of the δ¹⁸O isotopic paleothermometer for central Greenland, using borehole temperatures. *Journal of Glaciology* 40 (135), 341–349.
- Cuffey, K.M., Clow, G.D., Alley, R.B., Stuiver, M., Waddington, E.D., Saltus, R.W., 1995. Large Arctic temperature change at the glacial–Holocene transition. *Science* 270, 455–458.
- Dahl-Jensen, D., Mosegaard, K., Gundestrup, N., Clow, G.D., Johnsen, S.J., Hansen, A.W., Balling, N., 1998. Past temperatures directly from the Greenland Ice Sheet. *Science* 282, 268–271.
- Dahl-Jensen, D., Gundestrup, N., Gogineni, S.P., Miller, H., 2003. Basal melt at NorthGRIP modeled from borehole, ice-core and radio-echo sounded observations. *Annals of Glaciology* 37, 207–212.
- Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdóttir, A.E., Jouzel, J., Bond, G., 1993. Evidence for general insatibility of past climate from a 250-kyr ice-core record. *Nature* 364, 218–220.
- Darby, D.A., Bischof, J., Spielhagen, R.F., Marshall, S.A., Herman, S.W., 2002. Arctic ice export events and their potential impact on global climate during the late Pleistocene. *Paleoceanography* 17, 15–1–15–17. doi:10.1029/2001PA000639.
- Das, S.B., Joughin, I., Behn, M.D., Howat, I.M., King, M.A., Lizarralde, D., Bhatia, M.P., 2008. Fracture propagation to the base of the Greenland Ice Sheet during supraglacial lake drainage. *Science* 320, 778–781.
- De Abreu, C., Abrantes, F.F., Shackleton, N.J., Tzedakis, P.C., McManus, J.F., Oppo, D.W., Hall, M.A., 2005. Ocean climate variability in the eastern North Atlantic during interglacial marine isotope stage 11 – a partial analogue to the Holocene? *Paleoceanography* 20 Art. No. PA3009.
- Denton, G.H., Alley, R.B., Comer, G.C., Broecker, W.S., 2005. The role of seasonality in abrupt climate change. *Quaternary Science Reviews* 24, 1159–1182.
- Dickinson, W.R., 2001. Paleoshoreline record of relative Holocene sea levels on Pacific Islands. *Earth-Science Reviews* 55, 191–234.
- Douglas, B.C., 1997. Global sea rise: a re-determination. *Surveys in Geophysics* 18 (2–3), 279–292.
- Dowdeswell, J.A., Uenzelmann-Neben, G., Whittington, R.J., Marienfeld, P., 1994a. The Late Quaternary sedimentary record in Scoresby Sund, East Greenland. *Boreas* 23, 294–310.

- Dowdeswell, J.A., Whittington, R.J., Marienfeld, P., 1994b. The origin of massive diamicton facies by iceberg rafting and scouring, Scoresby Sund, East Greenland. *Sedimentology* 41, 21–35.
- Dowdeswell, J.A., Kenyon, N.H., Elverhoi, A., Laberg, J.S., Hollender, F.-J., Mienert, J., Siegert, M.J., 1996. Large-scale sedimentation on the glacier-influenced Polar North Atlantic margins – long-range side-scan sonar evidence. *Geophysical Research Letters* 23, 3535–3538.
- Dowdeswell, J.A., Kenyon, N.H., Laberg, J.S., 1997. The glacier-influenced Scoresby Sund Fan, east Greenland continental margin – evidence from GLORIA and 3.5 kHz records. *Marine Geology* 143, 207–221.
- Droxler, A.W., Alley, R.B., Howard, W.R., Poore, R.Z., Burckle, L.H., 2003. Unique and exceptionally long interglacial marine isotope stage 11 – window into Earth future climate. In: Droxler, A.W., Poore, R.Z., Burckle, L.H. (Eds.), *Earth's Climate and Orbital Eccentricity – The Marine Isotope Stage 11 Question*. Geophysical Monograph, vol. 137. American Geophysical Union, pp. 1–14.
- Dunhill, G., 2005. Iceland and Greenland margins – a comparison of depositional processes under different glaciological and oceanographic settings. PhD dissertation, Geological Sciences, University of Colorado, Boulder.
- Dupont, T.K., Alley, R.B., 2005. Assessment of the importance of ice-shelf buttressing to ice-sheet flow. *Geophysical Research Letters* 32, L04503. doi:10.1029/2004GL020224.
- Dupont, T.K., Alley, R.B., 2006. Role of small ice shelves in sea-level rise. *Geophysical Research Letters* 33, L09503.
- Dyke, A.S., Dale, J.E., McNeely, R.N., 1996. Marine molluscs as indicators of environmental change in glaciated North America and Greenland during the last 18,000 years. *Geographie Physique et Quaternaire* 50, 125–184.
- Dyke, A.S., Andrews, J.T., Clark, P.U., England, J.H., Miller, G.H., Shaw, J., Veilleux, J.J., 2002. The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21, 9–31.
- Edwards, R.L., Cheng, H., Murrell, M.T., Goldstein, S.J., 1997. Protactinium-231 dating of carbonates by thermal ionization mass spectrometry – implications for Quaternary climate change. *Science* 276, 782–786.
- Eisen, O., Nixdorf, U., Wilhelms, F., Miller, H., 2004. Age estimates of isochronous reflection horizons by combining ice core, survey, and synthetic radar data. *Journal of Geophysical Research* 109 (B4), B04106.
- Eldrett, J.S., Harding, I.C., Wilson, P.A., Butler, E., Roberts, A.P., 2007. Continental ice in Greenland during the Eocene and Oligocene. *Nature* 446, 176–179.
- Elliot, M., Labeyrie, L., Bond, G., Cortijo, E., Turon, J.-L., Tiseray, N., Duplessy, J.-C., 1998. Millennial-scale iceberg discharges in the Irminger Basin during the last glacial period – relationship with the Heinrich events and environmental settings. *Paleoceanography* 13, 433–446.
- Elverhoi, A., Dowdeswell, J.A., Funder, S., Mangerud, J., Stein, R., 1998. Glacial and oceanic history of the polar North Atlantic margins – an overview. *Quaternary Science Reviews* 17, 1–10.
- England, J., 1999. Coalescent Greenland and Innuitian ice during the Last Glacial Maximum – revising the Quaternary of the Canadian High Arctic. *Quaternary Science Reviews* 18, 421–456.
- Evans, J., Cofaigh, C.O., Dowdeswell, J.A., Wadhams, P., 2009. Marine geophysical evidence for former expansion and flow of the Greenland Ice Sheet across the north-east Greenland continental shelf. *Journal of Quaternary Science* 24, 279–293.
- Fairbanks, R.G., Mortlock, R.A., Tzu-Chien, C., Cao, L., Kaplan, A., Guilderson, T.P., Fairbanks, T.W., Bloom, A.L., Grootes, P.M., Nadeau, M.-J., 2005. Radiocarbon calibration curve spanning 0 to 50,000 years BP based on paired $^{230}\text{Th}/^{234}\text{U}$ and ^{14}C dates on pristine corals. *Quaternary Science Reviews* 24, 1781–1796.
- Farmer, G.L., Barber, D.C., Andrews, J.T., 2003. Provenance of Late Quaternary ice-proximal sediments in the North Atlantic – Nd, Sr and Pd isotopic evidence. *Earth and Planetary Science Letters* 209, 227–243.
- Fillon, R.H., Duplessy, J.C., 1980. Labrador Sea bio-, tephro-oxygen isotopic stratigraphy and Late Quaternary paleoceanographic trends. *Canadian Journal of Earth Sciences* 17, 831–854.
- Finkel, R.C., Nishiizumi, K., 1997. Beryllium 10 concentrations in the Greenland Ice Sheet Project 2 ice core from 3–40 ka. *Journal of Geophysical Research* 102 (C12), 26,699–26,706.
- Fisher, T.G., Smith, D.G., Andrews, J.T., 2002. Preboreal oscillation caused by a glacial Lake Agassiz flood. *Quaternary Science Reviews* 21, 873–878.
- Fleming, K., Lambeck, K., 2004. Constraints on the Greenland Ice Sheet since the Last Glacial Maximum from sea-level observations and glacial-rebound models. *Quaternary Science Reviews* 23, 1053–1077.
- Frujtier, C., Elliot, T., Schlager, W., 2000. Mass-spectrometric ^{234}U – ^{230}Th ages from the Key Largo Formation, Florida Keys, United States – constraints on diagenetic age disturbance. *Geological Society of America Bulletin* 112, 267–277.
- Funder, S., 1989a. Quaternary geology of east Greenland. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1, 839 pp.; also Geological Society of America, the *Geology of North America K-1*, pp. 756–763.
- Funder, S., 1989b. Quaternary geology of north Greenland. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1, 839 pp.; also Geological Society of America, the *Geology of North America K-1*, pp. 763–769.
- Funder, S., 1989c. Quaternary geology of west Greenland. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1, 839 pp.; also Geological Society of America, the *Geology of North America K-1*, pp. 749–756.
- Funder, S., 1989d. Sea level history. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1, 839 pp.; also Geological Society of America, the *Geology of North America K-1*, pp. 772–774.
- Funder, S., Fredskild, B., 1989. Paleofaunas and floras. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1; also Geological Society of America, the *Geology of North America K-1*, pp. 775–783.
- Funder, S., Larsen, H.C., 1989. Quaternary geology of the shelves adjacent to Greenland. In: Fulton, R.J. (Ed.), *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada (Chapter 13), *Geology of Canada*, no. 1; also Geological Society of America, the *Geology of North America K-1*, pp. 769–772.
- Funder, S., Hjort, C., Landvik, J.Y., Nam, S.I., Reeh, N., Stein, R., 1998. History of a stable ice margin East Greenland during the Middle and Upper Pleistocene. *Quaternary Science Reviews* 17, 77–123.
- Funder, S., Bennike, O., Böcher, J., Israelson, C., Petersen, K.S., Simonarson, L.A., 2001. Late Pliocene Greenland – the Kap København formation in North Greenland. *Bulletin of the Geological Society of Denmark* 48, 117–134.
- Funder, S., Jennings, A.E., Kelly, M.J., 2004. Middle and late Quaternary glacial limits in Greenland. In: Ehlers, J., Gibbard, P.L. (Eds.), *Quaternary Glaciations – Extent and Chronology*, Part II. Elsevier, New York, pp. 425–430.
- Gallup, C.D., Edwards, R.L., Johnson, R.G., 1994. The timing of high sea levels over the past 200,000 years. *Science* 263, 796–800.
- Geirsdottir, A., Hardardottir, H., Eiríksson, J., 1997. The depositional history of the Younger Dryas – Preboreal Budi moraines in south-central Iceland. *Arctic and Alpine Research* 29, 13–23.
- Geirsdottir, A., Hardardottir, J., Andrews, J.T., 2000. Late Holocene terrestrial geology of Miki and I.C. Jacobsen Fjords, East Greenland. *The Holocene* 10, 125–134.
- Gilbert, R., 1990. Rafting in glacial marine environments. In: Dowdeswell, J.A., Scourse, J.D. (Eds.), *Glacimarine Environments – Processes and Sediments*. Geological Society, London, pp. 105–120.
- Goreau, T.F., 1959. The ecology of Jamaican coral reefs I – species composition and zonation. *Ecology* 40 (1), 67–90.
- Gosse, J.C., Phillips, F.M., 2001. Terrestrial in situ cosmogenic nuclides – theory and application. *Quaternary Science Reviews* 20, 1475–1560.
- Gregory, J.M., Huybrechts, P., 2006. Ice-sheet contributions to future sea-level change. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences* 364 (1844), 1709–1731.
- Grootes, P.M., Stuiver, M., 1997. Oxygen 18/16 variability in Greenland snow and ice with 10^{-3} to 10^5 -year time resolution. *Journal of Geophysical Research* 102C, 26,455–26,470.
- Grousset, F.E., Cortijo, E., Huon, S., Herve, L., Richter, T., Burdloff, D., Duprat, J., Weber, O., 2001. Zooming in on Heinrich layers. *Paleoceanography* 16, 240–259.
- Hagen, S., 1999. North Atlantic paleoceanography and climate history during the last ~70 cal. ka years. PhD dissertation, Department of Geology, University of Tromsø, Norway, 110 pp.
- Hagen, S., Hald, M., 2002. Variation in surface and deep water circulation in the Denmark Strait, North Atlantic, during marine isotope stages 3 and 2. *Paleoceanography* 17, 13–11–13–16. doi:10.1029/2001PA000632.
- Hakansson, L., Briner, J., Alexanderson, H., Aldahan, A., Possnert, G., 2007. ^{10}Be ages from central east Greenland constrain the extent of the Greenland Ice Sheet during the Last Glacial Maximum. *Quaternary Science Reviews* 26 (19–21), 2316–2321. doi:10.1016/j.quascirev.2007.08.001.
- Hakansson, L., Alexanderson, H., Hjort, C., Moller, P., Briner, J.P., Aldahan, A., Possnert, G., 2009. Late Pleistocene glacial history of Jameson Land, central East Greenland, derived from cosmogenic Be-10 and Al-26 exposure dating. *Boreas* 38, 244–260.
- Hald, M., Hagen, S., 1998. Early Preboreal cooling in the Nordic seas region triggered by meltwater. *Geology* 26, 615–618.
- Hall, B.L., Baroni, C., Denton, G.H., Kelly, M.A., Lowell, T.V., 2008. Relative sea-level change, Kjøve Land, Scoresby Sund, East Greenland: implications for seasonality in Younger Dryas time. *Quaternary Science Reviews* 27, 605–618.
- Hanna, E., Huybrechts, P., Janssens, I., Cappelen, J., Steffens, K., Stephens, A., 2005. Runoff and mass balance of the Greenland Ice Sheet – 1958–2003. *Journal of Geophysical Research* 110, D13108. doi:10.1029/2004JD005641.
- Hearty, P.J., Kindler, P., Cheng, H., Edwards, R.L., 1999. A +20 m middle Pleistocene sea-level highstand (Bermuda and the Bahamas) due to partial collapse of Antarctic ice. *Geology* 27, 375–378.
- Helmke, J.P., Bauch, H.A., Erlenkeuser, H., 2003. Development of glacial and interglacial conditions in the Nordic seas between 1.5 and 0.35 Ma. *Quaternary Science Reviews* 22, 1717–1728.
- Hemming, S.R., 2004. Heinrich events – massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Reviews of Geophysics* 42 Art. No. RG1005.
- Hemming, S.R., Vorren, T.O., Kleman, J., 2002. Provinciality of ice rafting in the North Atlantic – application of Ar-40/Ar-39 dating of individual ice rafted hornblende grains. *Quaternary International* 95, 75–85.
- Hiscott, R.N., Aksu, A.E., Nielsen, O.B., 1989. Provenance and dispersal patterns, Pliocene–Pleistocene section at site 645, Baffin Bay. In: Stewart, S.K. (Ed.), *Proceedings, ODP. Scientific Results*, vol. 105. Ocean Drilling Program, College Station, Texas, pp. 31–52.
- Holland, D.M., Thomas, R.H., de Young, B., Ribergaard, M.H., Lyberth, B., 2008. Acceleration of Jakobshavn Isbra triggered by warm subsurface ocean waters. *Nature Geoscience* 1, 659–664.

- Hooke, R. Le B., 2005. *Principles of Glacier Mechanics*. Cambridge University Press, Cambridge, 429 pp.
- Hopkins, T.S., 1991. The GIN Sea – a synthesis of its physical oceanography and literature review 1972–1985. *Earth Science Reviews* 30, 175–318.
- Hubbard, A., Sugden, D., Dugmore, A., Norddahl, H., Petersson, H.G., 2006. A modelling insight into the Icelandic Last Glacial Maximum ice sheet. *Quaternary Science Reviews* 25, 2283–2296.
- Hughes, T.J., 1998. *Ice Sheets*. Oxford University Press, New York, 343 pp.
- Hulbe, C.L., MacAyeal, D.R., Denton, G.H., Kleman, J., Lowell, T.V., 2004. Catastrophic ice shelf breakup as the source of Heinrich event icebergs. *Paleoceanography* 19 Art. No. PA1004.
- Huybrechts, P., 2002. Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles. *Quaternary Science Reviews* 21, 203–231.
- Huybrechts, P., de Wolde, J., 1999. The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming. *Journal of Climate* 12, 2169–2188.
- Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L., Shackleton, N.J., 1984. The orbital theory of Pleistocene climate: support from a revised chronology of the marine $\delta^{18}\text{O}$ record. In: Berger, A., Imbrie, J., Hays, H., Kukla, G., Saltzman, B. (Eds.), *Milankovitch and Climate: Understanding the Response to Astronomical Forcing*, Proceedings of the NATO Advanced Research Workshop Held 30 November–4 December, 1982 in Palisades, NY. D. Reidel Publishing, Dordrecht, p. 269.
- Intergovernmental Panel on Climate Change (IPCC), 2007. Summary for policy-makers. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), *Climate Change 2007—The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge and New York, 18 pp.
- Jacobel, R.W., Welch, B.C., 2005. A time marker at 17.5 kyr BP detected throughout West Antarctica. *Annals of Glaciology* 41, 47–51.
- Jansen, E., Overpeck, J., Briffa, K.R., Duplessy, J.-C., Joos, F., Masson-Delmotte, V., Olago, D., Otto-Bliesner, B., Peltier, W.R., Rahmstorf, S., Ramesh, R., Raynaud, D., Rind, D., Solomina, O., Villalba, R., Zhang, D., 2007. Palaeoclimate. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), *Climate Change 2007—The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge, United Kingdom and New York, pp. 434–497.
- Jennings, A.E., Weiner, N.J., 1996. Environmental change in eastern Greenland during the last 1300 years: evidence from foraminifera and lithofacies in Nansen Fjord, 68 degrees N. *Holocene* 6, 179–191.
- Jennings, A.E., Grönvold, K., Hilberman, R., Smith, M., Hald, M., 2002a. High-resolution study of Icelandic tephra in the Kangerlussuaq Trough, Southeast Greenland, during the last deglaciation. *Journal of Quaternary Science* 7, 747–757.
- Jennings, A.E., Knudsen, K.L., Hald, M., Hansen, C.V., Andrews, J.T., 2002b. A mid-Holocene shift in Arctic sea ice variability on the East Greenland shelf. *The Holocene* 12, 49–58.
- Jennings, A.E., Hald, M., Smith, M., Andrews, J.T., 2006. Freshwater forcing from the Greenland Ice Sheet during the Younger Dryas: evidence from southeastern Greenland shelf cores. *Quaternary Science Reviews* 25, 282–298.
- Johannessen, O.M., Khvorostovsky, K., Miles, M.W., Bobylev, L.P., 2005. Recent ice-sheet growth in the interior of Greenland. *Science* 310, 1013–1016.
- Johnsen, S.J., 1977. Stable isotope homogenization of polar firn and ice. In: *Proceedings of a Symposium on Isotopes and Impurities in Snow and Ice*, I.U.G. G. XVI, General Assembly, Grenoble Aug.–Sept. 1975, vol. 118. IAHS-AISH Publ., Washington D.C., pp. 210–219.
- Johnsen, S., Clausen, H.B., Dansgaard, W., Gundestrup, N.S., Hansson, M., Johnsson, P., Steffensen, J.P., Sveinbjörnsdóttir, A.E., 1992a. A “deep” ice core from East Greenland. *Meddelelser om Grønland Geoscience* 29, 3–22.
- Johnsen, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B., Steffensen, J.P., 1992b. Irregular glacial interstadials recorded in a new Greenland ice core. *Nature* 359, 311–313.
- Johnsen, S.J., Dahl-Jensen, D., Gundestrup, N., Steffensen, J.P., Clausen, H.B., Miller, H., Masson-Delmotte, V., Sveinbjörnsdóttir, A.E., White, J., 2001. Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. *Journal of Quaternary Science* 16, 299–307.
- Jones, G.A., Keigwin, L.D., 1988. Evidence from the Fram Strait (78°N) for early deglaciation. *Nature* 336, 56–59.
- Joughin, I., Tulaczyk, S., Fahnestock, M., Kwok, R., 1996. A mini-surge on the Ryder Glacier, Greenland, observed by satellite radar interferometry. *Science* 274, 228–230.
- Joughin, I., Das, S.B., King, M.A., Smith, B.E., Howat, I.M., Moon, T., 2008a. Seasonal speedup along the western flank of the Greenland ice sheet. *Science* 320 (5877), 781–783.
- Joughin, I., Howat, I., Alley, R.B., Ekstrom, G., Fahnestock, M., Moon, T., Nettles, M., Truffer, M., Tsai, V.C., 2008b. Ice front variation and tidewater behavior on Helheim and Kangerdlugssuaq Glaciers, Greenland. *Journal of Geophysical Research* 113, F01004. doi:10.1029/2007JF00837.
- Jouzel, J., Alley, R.B., Cuffey, K.M., Dansgaard, W., Grootes, P., Hoffmann, G., Johnsen, S.J., Koster, R.D., Peel, D., Shuman, C.A., Stievenard, M., Stuiver, M., White, J., 1997. Validity of the temperature reconstruction from water isotopes in ice cores. *Journal of Geophysical Research* 102 (C12), 26,471–26,487.
- Kahn, S.A., Wahr, J., Stearns, L.A., Hamilton, G.S., van Dam, T., Larson, K.M., Francis, O., 2007. Elastic uplift in southeast Greenland due to rapid ice mass loss. *Geophysical Research Letters* 34, L21701. doi:10.1029/2007GL031468.
- Kamb, B., Raymond, C.F., Harrison, W.D., Engelhardt, H., Echelmeyer, K.A., Humphrey, N., Brugman, M.M., Pfeffer, T., 1985. Glacier surge mechanism: 1982–1983 surge of variegated glacier, Alaska. *Science* 227, 469–479.
- Kandiano, E.S., Bauch, H.A., 2003. Surface ocean temperatures in the north-east Atlantic during the last 500,000 years – evidence from foraminiferal census data. *Terra Nova* 15, 265–271.
- Kapsner, W.R., Alley, R.B., Shuman, C.A., Anandakrishnan, S., Grootes, P.M., 1995. Dominant control of atmospheric circulation on snow accumulation in central Greenland. *Nature* 37, 52–54.
- Kaufman, D.S., Brigham-Grette, J., 1993. Aminostratigraphic correlations and paleotemperature implications, Pliocene–Pleistocene high sea level deposits, northwestern Alaska. *Quaternary Science Reviews* 12, 21–33.
- Kaufman, D.S., Walter, R.C., Brigham-Grette, J., Hopkins, D.M., 1991. Middle Pleistocene age of the Nome River glaciation, northwestern Alaska. *Quaternary Research* 36, 277–293.
- Kelly, M.A., Lowell, T.V., 2009. Fluctuations of local glaciers in Greenland during latest Pleistocene and Holocene time. *Quaternary Science Reviews* 28, 2088–2106. doi:10.1016/j.quascirev.2008.12.008.
- Kelly, M., Funder, S., Houmark-Nielsen, M., Knudsen, K.L., Kronborg, C., Landvik, J., Sorby, L., 1999. Quaternary glacial and marine environmental history of northwest Greenland – a review and reappraisal. *Quaternary Science Reviews* 18, 373–392.
- Kelly, M.A., Lowell, T.V., Hall, B.L., Schaefer, J.M., Finkel, R.C., Goehring, B.M., Alley, R.B., Denton, G.H., 2008. A Be-10 chronology of lateglacial and Holocene mountain glaciation in the Scoresby Sund region, east Greenland: implications for seasonality during lateglacial time. *Quaternary Science Reviews* 27, 2273–2282.
- Kiehl, J.T., Gent, P.R., 2004. The community climate system model, version two. *Journal of Climate* 17, 3666–3682.
- Kindler, P., Hearty, P.J., 2000. Elevated marine terraces from Eleuthera (Bahamas) and Bermuda: sedimentological, petrographic and geochronological evidence for important deglaciation events during the middle Pleistocene. *Global and Planetary Change* 24, 41–58.
- Kleiven, H.F., Jansen, E., Fronval, T., Smith, T.M., 2002. Intensification of Northern Hemisphere glaciations in the circum Atlantic region (3.5–2.4 Ma) – ice-rafted detritus evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 184, 213–223.
- Kobashi, T., Severinghaus, J.P., Barnola, J.-M., 2008. $4 \pm 1.5^\circ\text{C}$ abrupt warming 11,270 years ago identified from trapped air in Greenland ice. *Earth and Planetary Science Letters* 268, 397–407.
- Koerner, R.M., 1989. Ice core evidence for extensive melting of the Greenland Ice Sheet in the last interglacial. *Science* 244, 964–968.
- Koerner, R.M., Fisher, D.A., 2002. Ice-core evidence for widespread Arctic glacier retreat in the last interglacial and the early Holocene. *Annals of Glaciology* 35, 19–24.
- Kopp, R.E., Simons, F.J., Mitrovica, J.X., Maloof, A.C., Oppenheimer, M., 2009. Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462, 863–868.
- Kuijpers, A., Abrahamsen, N., Hoffmann, G., Hühnerbach, V., Konradi, P., Kunzendorf, H., Mikkelsen, N., Thiede, J., Weinrebe, W., scientific party of RV Poseidon, surveyors of the Royal Danish Administration for Navigation, and Hydrology, 1999. Climate change and the Viking-age fjord environment of the Eastern Settlement, South Greenland. *Geology of Greenland Survey Bulletin* 183, 61–67.
- Lambeck, K., Smither, C., Johnston, P., 1998. Sea-level change, glacial rebound and mantle viscosity for northern Europe. *Geophysical Journal International* 134, 102–144.
- Lemke, P., Ren, J., Alley, R.B., Allison, I., Carrasco, J., Flato, G., Fujii, Y., Kaser, G., Mote, P., Thomas, R.H., Zhang, T., 2007. Observations: changes in snow, ice and frozen ground. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge and New York, 996 pp.
- Lhomme, N., Clarke, G.K.C., Marshall, S.J., 2005. Tracer transport in the Greenland Ice Sheet – constraints on ice cores and glacial history. *Quaternary Science Reviews* 24, 173–194.
- Lie, O., Paasche, O., 2006. How extreme was northern hemisphere seasonality during the Younger Dryas? *Quaternary Science Reviews* 25, 404–407. doi:10.1016/j.quascirev.2005.11.003.
- Lisitzin, A.P., 2002. *Sea-Ice and Iceberg Sedimentation in the Ocean – Recent and Past*. Springer-Verlag, Berlin, 563 pp.
- Ljung, K., Björck, S., 2004. A lacustrine record of the Pleistocene–Holocene boundary in southernmost Greenland. *GFF (Geological Society of Sweden)* 126, 273–278. Part 3.
- Lloyd, J.M., 2006. Late Holocene environmental change in Disko Bugt, west Greenland – interaction between climate, ocean circulation and Jakobshavn Isbrae. *Boreas* 35, 35–49.
- Lloyd, J.M., Park, L.A., Kuijpers, A., Moros, M., 2005. Early Holocene palaeoceanography and deglacial chronology of Disko Bugt, West Greenland. *Quaternary Science Reviews* 24, 1741–1755.

- Locke, W.W.I., Andrews, J.T., Webber, P.J., 1979. A manual for lichenometry. British Geomorphological Research Group Technical Bulletin 26 47pp.
- Long, A.J., Roberts, D.H., Dawson, S., 2006. Early Holocene history of the west Greenland Ice Sheet and the GH-8.2 event. *Quaternary Science Reviews* 25, 904–922.
- Long, A.J., Woodroffe, S.A., Dawson, S., Roberts, D.H., Bryant, C.L., 2009. Late Holocene relative sea level rise and the Neoglacial history of the Greenland ice sheet. *Journal of Quaternary Science* 24, 345–359.
- Luthcke, S.B., Zwally, H.J., Abdalati, W., Rowlands, D.D., Ray, R.D., Nerem, R.S., Lemoine, F.G., McCarthy, J.J., Chinn, D.S., 2006. Recent Greenland ice mass loss by drainage system from satellite gravity observations. *Science* 314, 1286–1289. doi:10.1126/science.1130776.
- Mangerud, J., Funder, S., 1994. The interglacial–glacial record at the mouth of Scoresby Sund, East Greenland. *Boreas* 23, 349–358.
- Mariénfeld, P., 1992a. Recent sedimentary processes in Scoresby Sund, East Greenland. *Boreas* 21, 169–186.
- Mariénfeld, P., 1992b. Postglacial sedimentary history of Scoresby Sund, East Greenland. *Polarforschung* 60, 181–195.
- Marshall, S.J., Cuffey, K.M., 2000. Peregrinations of the Greenland Ice Sheet divide in the last glacial cycle – implications for central Greenland ice cores. *Earth and Planetary Science Letters* 179, 73–90.
- Markwick, P.J., 1998. Fossil crocodilians as indicators of Late Cretaceous and Cenozoic climates – implications for using palaeontological data in reconstructing palaeoclimate. *Palaeogeography, Palaeoclimatology, Palaeoecology* 137, 205–271.
- McCave, I.N., Tucholke, B.E., 1986. Deep current-controlled sedimentation in the western North Atlantic. In: Vogt, P.R., Tucholke, B.E. (Eds.), *The Geology of North America: The Western North Atlantic Region*. Geological Society of America, pp. 451–468.
- McManus, J.F., Oppo, D.W., Cullen, J.L., 1999. A 0.5-million-year record of millennial-scale climate variability in the North Atlantic. *Science* 283, 971–975.
- Meier, M.F., Post, A., 1987. Fast tidewater glaciers. *Journal of Geophysical Research* 92 (B9), 9051–9058.
- Mernild, S.H., Liston, G.E., Hiemstra, C.A., Steffen, K., 2008. Surface melt area and water balance modeling on the Greenland Ice Sheet 1995–2005. *Journal of Hydrometeorology* 9, 1191–1211.
- Mienert, J., Andrews, J.T., Milliman, J.D., 1992. The East Greenland continental margin (65°N) since the last deglaciation – changes in sea floor properties and ocean circulation. *Marine Geology* 106, 217–238.
- Miller, G.H., 2008. Greenland's elusive Younger Dryas. *Quaternary Science Reviews* 27, 2271–2272.
- Mitrova, J.X., 1996. Haskell [1935] revisited. *Journal of Geophysical Research* 101, 555–569.
- Mitrova, J.X., Milne, G.A., 2002. On the origin of Late Holocene highstands within equatorial ocean basins. *Quaternary Science Reviews* 21, 2179–2190.
- Mitrova, J.X., Peltier, W.R., 1991. On post-glacial geoid relaxation over the equatorial oceans. *Journal of Geophysical Research* 96, 20,053–20,071.
- Mitrova, J.X., Tamisiea, M.E., Davis, J.L., Milne, G.A., 2001. Recent mass balance of polar ice sheets inferred from patterns of global sea-level change. *Nature* 409, 1026–1029.
- Mitrova, J.X., Wahr, J., Matsuyama, I., Paulson, A., Tamisiea, M.E., 2006. Reanalysis of ancient eclipse, astronomic and geodetic data – a possible route to resolving the enigma of global sea-level rise. *Earth and Planetary Science Letters* 243, 390–399.
- Mitrova, J.X., Gomez, N., Clark, P.U., 2009. The sea-level fingerprint of West Antarctic Collapse. *Science* 323, 753.
- Moran, K., Backman, J., Brinkhuis, H., Clemens, S.C., Cronin, T., Dickens, G.R., Eynaud, F., Gattacceca, J., Jakobsson, M., Jordan, R.W., Kaminski, M., King, J., Koç, N., Krylov, A., Martinez, N., Matthiessen, J., McInroy, D., Moore, T.C., Onodera, J., O'Regan, A.M., Pälike, H., Rea, B., Rio, D., Sakamoto, T., Smith, D.C., Stein, R., St. John, K., Suto, I., Suzuki, N., Takahashi, K., Watanabe, M., Yamamoto, M., Farrell, J., Frank, M., Kubik, P., Jokat, W., Kristoffersen, Y., 2006. The Cenozoic palaeoenvironment of the Arctic Ocean. *Nature* 441, 601–605.
- Moros, M., Jensen, K.G., Kuijpers, A., 2006. Mid- to late-Holocene hydrological variability and climatic variability in Disko Bugt, central West Greenland. *The Holocene* 16, 357–367.
- Mudie, P.J., Rochon, A., Prins, M.A., Soenarjo, D., Troelstra, S.R., Levac, E., Scott, D.B., Roncaglia, L., Kuijpers, A., 2006. Late Pleistocene–Holocene marine geology of Nares Strait region – Palaeoceanography for foraminifera and dinoflagellate cysts, sedimentology and stable isotopes. *Polarforschung* 74 (1/3), 169–183. doi:10.1013/epic.29931.d001.
- Muhs, D.R., 2002. Evidence for the timing and duration of the last interglacial period from high-precision uranium-series ages of corals on tectonically stable coastlines. *Quaternary Research* 58, 36–40.
- Muhs, D.R., Szabo, B.J., 1994. New uranium-series ages of the Waimanalo Limestone, Oahu, Hawaii – implications for sea level during the last interglacial period. *Marine Geology* 118, 315–326.
- Muhs, D.R., Simmons, K.R., Steinke, B., 2002. Timing and warmth of the last interglacial period – new U-series evidence from Hawaii and Bermuda and a new fossil compilation for North America. *Quaternary Science Reviews* 21, 1355–1383.
- Muhs, D.R., Wehmiller, J.F., Simmons, K.R., York, L.L., 2004. Quaternary sea level history of the United States. In: Gillespie, A.R., Porter, S.C., Atwater, B. F. (Eds.), *The Quaternary Period in the United States*. Elsevier, Amsterdam, pp. 147–183.
- Multer, H.G., Gischler, E., Lundberg, J., Simmons, K.R., Shinn, E.A., 2002. Key Largo Limestone revisited – Pleistocene shelf-edge facies, Florida Keys, USA. *Facies* 46, 229–272.
- Munk, W., 2002. Twentieth century sea level – an enigma. *Proceedings of the National Academy of Sciences of the United States of America* 99, 6550–6555.
- Muscheler, R., Beer, R., Kubik, P.W., Synal, H.A., 2005. Geomagnetic field intensity during the last 60,000 years based on Be-10 and Cl-36 from the Summit ice cores and C-14. *Quaternary Science Reviews* 24, 1849–1860.
- Nam, S.I., Stein, R., 1999. Late Quaternary variations in sediment accumulation rates and their paleoenvironment implications: a case study from the East Greenland continental margin. *GeoResearch Forum* 5, 223–240.
- Nick, M.F., Veli, A., Howat, I.M., Joughin, I., 2009. Large-scale changes in Greenland outlet glacier dynamics triggered at the terminus. *Nature Geoscience* 2, 110–114.
- Nishiizumi, K., Finkel, R.C., Pongani, K.V., Graf, T., Kohl, C.P., Marti, K., 1996. In situ produced cosmogenic nuclides in GISP2 rock core from Greenland Summit (Abstract, Fall Meeting 1996). *Eos, Transactions of the American Geophysical Union* 77 (46), F428. Abstract OS41B-10.
- North Greenland Ice Core Project Members, 2004. High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431, 147–151.
- O'Cofaigh, C., Taylor, J., Dowdeswell, J.A., Pudsey, C.J., 2003. Palaeo-ice streams, trough mouth fans and high-latitude continental slope sedimentation. *Boreas* 32, 37–55.
- Oerlemans, J., 1994. Quantifying global warming from the retreat of glaciers. *Science* 264, 243–245.
- Oerlemans, J., 2001. *Glaciers and Climatic Change*. A.A. Balkema Publishers, Lisse, 148 pp.
- Otto-Bliesner, B.L., Overpeck, J.T., Marshall, S.J., Miller, G., 2006. Simulating Arctic climate warmth and icefield retreat in the Last Interglaciation. *Science* 311, 1751–1753.
- Overpeck, J.T., Otto-Bliesner, B.L., Miller, G.H., Muhs, D.R., Alley, R.B., Kiehl, J.T., 2006. Paleoclimatic evidence for future ice-sheet instability and rapid sea-level rise. *Science* 311, 1747–1750.
- Parizek, B.R., Alley, R.B., 2004. Implications of increased Greenland surface melt under global-warming scenarios: ice-sheet simulations. *Quaternary Science Reviews* 23, 1013–1027.
- Parnell, J., Bowden, S., Taylor, C., Andrews, J.T., 2007. Biomarker determination as a provenance tool for detrital carbonate events (Heinrich events?) – fingerprinting Quaternary glacial sources in Baffin Bay. *Earth and Planetary Science Letters* 257, 71–82.
- Paterson, W.S.B., 1994. *The Physics of Glaciers*, third ed. Pergamon, Oxford, 480 pp.
- Payne, A.J., Veli, A., Shepherd, A.P., Wingham, D.J., Rignot, E., 2004. Recent dramatic thinning of largest West Antarctic ice stream triggered by oceans. *Geophysical Research Letters* 31, L23401. doi:10.1029/2004GL021284.
- Peltier, W.R., 2004. Global glacial isostasy and the surface of the ice-age Earth – the ICE-5 G (VM2) Model and GRACE. *Annual Review of Earth and Planetary Sciences* 32, 111–149. doi:10.1146/annurev.earth.32.082503.144359.
- Peltier, W.R., Fairbanks, R.G., 2006. Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* 25, 3322–3337.
- Petrenko, V.V., Severinghaus, J.P., Brook, E.J., Reeh, N., Schaefer, H., 2006. Gas records from the West Greenland ice margin covering the Last Glacial Termination: a horizontal ice core. *Quaternary Science Reviews* 25, 865–875.
- Pfeffer, W.T., Harper, J.T., O'Neil, S., 2008. Kinematic constraints on glacier contributions to 21st-century sea-level rise. *Science* 321, 1340–1343.
- Plag, H.-P., Juttner, H.-U., 2001. Inversion of global tide gauge data for present-day ice load changes. In: Yamanouchi, T. (Ed.), *Proceedings of the International Symposium on Environmental Research in the Arctic and Fifth Ny-Alesund Science Seminar. Memoirs of the National Institute of Polar Research (Japan)* 54, 301–317.
- Poore, R.Z., Dowsett, H.J., 2001. Pleistocene reduction of polar ice caps – evidence from Cariaco Basin marine sediments. *Geology* 29, 1–74.
- Ramillien, G., Lombard, A., Cazenave, A., Ivins, E.R., Llubes, M., Remy, F., Biancale, R., 2006. Interannual variations of the mass balance of the Antarctica and Greenland ice sheets from GRACE. *Global and Planetary Change* 53, 198–208.
- Raynaud, D., Chappellaz, J., Ritz, C., Martinerie, P., 1997. Air content along the Greenland Ice Core Project core – a record of surface climatic parameters and elevation in central Greenland. *Journal of Geophysical Research* 102 (C12), 26,607–26,613.
- Reeh, N., 1984. Reconstruction of the glacial ice covers of Greenland and the Canadian Arctic islands by 3-dimensional, perfectly plastic ice-sheet modeling. *Annals of Glaciology* 5, 115–121.
- Reeh, N., 1985. Greenland ice-sheet mass balance and sea-level change. In: *Glaciers, Ice Sheets, and Sea Level – Effect of a CO₂ Induced Climatic Change*. National Academy Press, Washington D.C., pp. 155–171.
- Reeh, N., 2004. Holocene climate and fjord glaciations in Northeast Greenland – implications for IRD deposition in the North Atlantic. *Sedimentary Geology* 165, 333–342.
- Reeh, N., Mayer, C., Miller, H., Thomsen, H.H., Weidick, A., 1999. Present and past climate control on fjord glaciations in Greenland – implications for IRD-deposition in the sea. *Geophysical Research Letters* 26, 1039–1042.
- Ridley, J.K., Huybrechts, P., Gregory, J.M., Lowe, J.A., 2005. Elimination of the Greenland Ice Sheet in a high CO₂ climate. *Journal of Climate* 18, 3409–3427.

- Rignot, E., Kanagaratnam, P., 2006. Changes in the velocity structure of the Greenland Ice Sheet. *Science* 311, 986–990.
- Rignot, E., Box, J.E., Burgess, E., Hanna, E., 2008. Mass balance of the Greenland ice sheet from 1958 to 2007. *Geophysical Research Letters* 35, L20502. doi:10.1029/2008GL035417.
- Rinterknecht, V., Gorokhov, Y., Schaefer, J., Caffee, M., 2009. Preliminary Be-10 chronology for the last deglaciation of the western margin of the Greenland Ice Sheet. *Journal of Quaternary Science* 24, 270–278.
- Roberts, D.H., Long, A.J., Schnable, C., Davies, B.J., Xu, S., Simpson, M.J.R., Huybrechts, P., Sowers, T., Brook, E.J., Alley, R.B., Bender, M.L., 1998. Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391, 141–146.
- Rohling, E.J., Fenton, M., Jorissen, F.J., Bertrand, P., Ganssen, G., Caulet, J.P., 1998. Magnitudes of sea-level lowstands of the past 500,000 years. *Nature* 394, 162–165.
- Schofield, J.E., Edwards, K.J., McMullen, J.A., 2007. Modern pollen–vegetation relationships in subarctic southern Greenland and the interpretation of fossil pollen data from the Norse landnam. *Journal of Biogeography* 34, 473–488.
- Severinghaus, J.P., Sowers, T., Brook, E.J., Alley, R.B., Bender, M.L., 1998. Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391, 141–146.
- Shackleton, N.J., Backman, J., Zimmerman, H., Kent, D.V., Hall, M.A., Roberts, D.G., Schnitker, D., Baldauf, J.G., Desprairies, A., Homrighausen, R., Huddleston, P., Keene, J.B., Kaltenback, A.J., Krumsiek, K.A.O., Morton, A.C., Murray, J.W., Westberg-Smith, J., 1984. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North-Atlantic region. *Nature* 307, 620–623.
- Shepherd, A., Wingham, D., 2007. Recent sea-level contributions of the Antarctic and Greenland ice sheets. *Science* 315, 1529–1532.
- Simpson, M.J.R., Milne, G.A., Huybrechts, P., Long, A.J., 2009. Calibrating a glaciological model of the Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level and ice extent. *Quaternary Science Reviews* 28, 1631–1657.
- Sluijs, A., Schouten, S., Pagani, M., Wolter, M., Brinkhuis, H., Damste, J.S.S., Dickens, G.R., Huber, M., Reichert, G.J., Stein, R., Matthiessen, J., Lourens, L.J., Pedentchouk, N., Backman, J., Moran, K., 2006. Subtropical arctic ocean temperatures during the Palaeocene/Eocene thermal maximum. *Nature* 441, 610–613.
- Smith, J.M.B., Bayliss-Smith, T.P., 1998. Kelp-plucking – coastal erosion facilitated by bull-kelp *Durvillaea antarctica* at subantarctic Macquarie Island. *Antarctic Science* 10, 431–438.
- Souchez, R., Bouzette, A., Clausen, H.B., Johnsen, S.J., Jouzel, J., 1998. A stacked mixing sequence at the base of the Dye 3 core, Greenland. *Geophysical Research Letters* 25, 1943–1946.
- Sowers, T., Bender, M., Raynaud, D., Korotkevich, Y.S., 1992. Delta-N-15 of N₂ in air trapped in polar ice – a tracer of gas-transport in the firn and a possible constraint on ice age/gas age-differences. *Journal of Geophysical Research* 97 (D14), 15,683–15,697.
- Sparrenbom, C.J., Bennike, O., Björck, S., Lambeck, K., 2006a. Holocene relative sea-level changes in the Qaqortoq area, southern Greenland. *Boreas* 35, 171–187.
- Sparrenbom, C.J., Bennike, O., Björck, S., Lambeck, K., 2006b. Relative sea-level changes since 15,000 cal. yr BP in the Nanortalik area, southern Greenland. *Journal of Quaternary Science* 21, 29–48.
- Spencer, M.K., Alley, R.B., Fitzpatrick, J.J., 2006. Developing a bubble number–density paleoclimatic indicator for glacier ice. *Journal of Glaciology* 52 (178), 358–364.
- Stastna, M., Peltier, W.R., 2007. On box models of the North Atlantic thermohaline circulation: intrinsic and extrinsic millennial timescale variability in response to deterministic and stochastic forcing. *Journal of Geophysical Research* 112, C10023.
- St. John, K., 2008. Cenozoic ice-rafting history of the central Arctic Ocean – terrigenous sands on the Lomonosov Ridge. *Paleoceanography* 23, PA1505. doi:10.1029/2007PA001483.
- St. John, K.E., Krissek, L.A., 2002. The late Miocene to Pleistocene ice-rafting history of southeast Greenland. *Boreas* 31, 28–35.
- Stein, A.B., 1996. Seismic stratigraphy and seafloor morphology of the Kangerlussuaq region, East Greenland – evidence for glaciations to the Continental shelf break during the late Weichselian and earlier. MSc thesis, University of Colorado, Boulder, 293 pp.
- Stein, R., Nam, S.-I., Grobe, H., Hubberten, H., 1996. Late Quaternary glacial history and short-term ice-rafted debris fluctuations along the East Greenland continental margin. In: Andrews, J.T., Austen, W.A., Bergseth, H., Jennings, A.E. (Eds.), *Late Quaternary Paleogeography of North Atlantic Margins*. Geological Society, London, pp. 135–151.
- Stirling, C.H., Esat, T.M., McCulloch, M.T., Lambeck, K., 1995. High-precision U-series dating of corals from Western Australia and implications for the timing and duration of the last interglacial. *Earth and Planetary Science Letters* 135, 115–130.
- Stirling, C.H., Esat, T.M., Lambeck, K., McCulloch, M.T., 1998. Timing and duration of the last interglacial – evidence for a restricted interval of widespread coral reef growth. *Earth and Planetary Science Letters* 160, 745–762.
- Stirling, C.H., Esat, T.M., Lambeck, K., McCulloch, M.T., Blake, S.G., Lee, D.-C., Halliday, A.N., 2001. Orbital forcing of the marine isotope stage 9 interglacial. *Science* 291, 290–293.
- Stocker, T.F., Johnsen, S.J., 2003. A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography* 18 Art. No. 1087.
- Stoner, J.S., Channell, J.E.T., Hillaire-Marcel, C., 1995. Magnetic properties of deep-sea sediments off southwest Greenland – evidence for major differences between the last two deglaciations. *Geology* 23, 241–244.
- Stuiver, M., Braziunas, T.F., Grootes, P.M., Zielinski, G.A., 1997. Is there evidence for solar forcing of climate in the GISP2 oxygen isotope record? *Quaternary Research* 48, 259–266.
- Stuiver, M., Reimer, P.J., Bard, E., Beck, J.W., Hughen, K.A., Kromer, B., McCormack, F.G., van der Plicht, J., Spurk, M., 1998. INTCAL98 Radiocarbon age calibration 24,000–0 cal BP. *Radiocarbon* 40, 1041–1083.
- Sugden, D., John, B., 1976. *Glaciers and Landscape*. Arnold, London, and Halsted, New York, 376 pp.
- Suwa, M., von Fischer, J.C., Bender, M.L., Landais, A., Brook, E.J., 2006. Chronology reconstruction for the disturbed bottom section of the GISP2 and the GRIP ice cores – implications for Termination II in Greenland. *Journal of Geophysical Research* 111 (D2) Art. No. D02101.
- Syvitski, J.P.M., Andrews, J.T., Dowdeswell, J.A., 1996. Sediment deposition in an iceberg-dominated glacial marine environment, East Greenland – basin fill implications. *Global and Planetary Change* 12, 251–270.
- Syvitski, J.P.M., Stein, A., Andrews, J.T., Milliman, J.D., 2001. Icebergs and seafloor of the East Greenland (Kangerlussuaq) continental margin. *Arctic, Antarctic and Alpine Research* 3, 52–61.
- Tang, C.C.L., Ross, C.K., Yao, T., 2004. The circulation, water masses and sea-ice of Baffin Bay. *Progress in Oceanography* 63, 183–228.
- Tarasov, L., Peltier, W. Richard, 2002. Greenland glacial history and local geodynamic consequences. *Geophysical Journal International* 150 (1), 198–229. doi:10.1046/j.1365-246X.2002.01702.x.
- Tarasov, L., Peltier, W.R., 2003. Greenland glacial history, borehole constraints, and Eemian extent. *Journal of Geophysical Research* 108 (B3), 2241–2244. Art. No. 2143.
- Tarduno, J.A., Brinkman, D.B., Renne, P.R., Cottrell, R.D., Scher, H., Castillo, P., 1998. Evidence for extreme climatic warmth from Late Cretaceous Arctic vertebrates. *Science* 282, 2241–2244.
- Thiede, J., Winkler, A., Wolf-Welling, T., Eldholm, O., Myhre, A.M., Baumann, K.H., Henrick, R., Stein, R., 1998. Late Cenozoic history of the Polar North Atlantic – results from ocean drilling. *Quaternary Science Reviews* 17, 185–208.
- Thomas, R.H., PARCA Investigators, 2001. Program for Arctic Regional Climate Assessment (PARCA) – goals, key findings, and future directions. *Journal of Geophysical Research* 106(D24) 33, 691–33705.
- Thomas, R., Abdalati, W., Frederick, E., Krabill, W., Manizade, S., Steffen, K., 2003. Investigation of surface melting and dynamic thinning on Jakobshavn Isbrae, Greenland. *Journal of Glaciology* 49, 231–239.
- Thomas, R., Frederick, E., Krabill, W., Manizade, S., Martin, C., 2006. Progressive increase in ice loss from Greenland. *Geophysical Research Letters* 33, L10503. doi:10.1029/2006GL026075.
- Thompson, W.G., Goldstein, S.L., 2005. Open-system coral ages reveal persistent suborbital sea-level cycles. *Science* 308, 401–404.
- Toniazzo, T., Gregory, J.M., Huybrechts, P., 2004. Climatic impact of a Greenland deglaciation and its possible irreversibility. *Journal of Climate* 17, 21–33.
- Vacco, D.A., Alley, R.B., Pollard, D., 2009. Modeling dependence of moraine deposition on climate history: the effect of seasonality. *Quaternary Science Reviews* 28 (7–8), 639–646.
- van den Broeke, M., Bamber, J., Ettema, J., Rignot, E., Schrama, E., van de Berg, W.J., van Meijgaard, E., Velicogna, I., Wouters, B., 2009. Partitioning recent Greenland mass loss. *Science* 326, 984–986.
- Vandermark, D., Tarduno, J.A., Brinkman, D.B., 2007. A fossil champsosaur population from the High Arctic: implications for Late Cretaceous paleotemperatures. *Paleogeography, Paleoclimatology, Paleogeology* 248, 49–59.
- van der Plicht, J., Van Geel, B., Bohncke, S.J.P., Bos, J.A.A., Blaauw, M., Speranza, A.O.M., Muscheler, R., Björck, S., 2004. The Preboreal climate reversal and a subsequent solar-forced climate shift. *Journal of Quaternary Science* 19, 263–269.
- van der Veen, C.J., 1999. *Fundamentals of Glacier Dynamics*. Balkema, Rotterdam, 462 pp.
- van Kreveld, S., Sarthein, M., Erlenkeuser, H., Grootes, P., Jung, S., Nadeau, M.J., Pflaumann, U., Voelker, A., 2000. Potential links between surging ice sheets, circulation changes, and the Dansgaard–Oeschger cycles in the Irminger Sea, 60–18 ka. *Paleoceanography* 15, 425–442.
- Velicogna, I., 2009. Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. *Geophysical Research Letters* 36 Art. No. L19503.
- Velicogna, I., Wahr, J., 2005. Greenland mass balance from GRACE. *Geophysical Research Letters* 32, L18505. doi:10.1029/2005GL023955.
- Velicogna, I., Wahr, J., 2006. Acceleration of Greenland ice mass loss in spring 2004. *Nature* 443, 329–331.
- Verplanck, E.P., Farmer, G.L., Andrews, J., Dunhill, G., Millo, C., 2009. Provenance of Quaternary glacial and glacial marine sediments along the southeast Greenland margin. *Earth and Planetary Science Letters* 286, 52–62.
- Vinther, B.M., Andersen, K.K., Jones, P.D., Briffa, K.R., Cappelen, J., 2006. Extending Greenland Temperature Records into the late 18th Century. *Journal of Geophysical Research* 111, D11105.
- Vinther, B.M., Buchardt, S.L., Clausen, H.B., Dahl-Jensen, D., Johnsen, S.J., Fisher, D.A., Koerner, R.M., Raynaud, D., Lipenkov, V., Andersen, K.K., Blunier, T., Rasmussen, S.O., Steffensen, J.P., Svensson, A.M., 2009. Holocene thinning of the Greenland ice sheet. *Nature* 461, 385–388.
- Vorren, T.O., Laberg, J.S., 1997. Trough mouth fans – palaeoclimate and ice-sheet monitors. *Quaternary Science Reviews* 16, 865–886.

- Weidick, A., 1993. Neoglacial change of ice cover and the related response of the Earth's crust in West Greenland. *Grønlands Grønlands Geologiske Undersøgelse* 159, 121–126.
- Weidick, A., 1996. Neoglacial changes of ice cover and sea level in Greenland – a classical enigma. In: Gronnow, B., Pind, J. (Eds.), *The Paleo-Eskimo Cultures of Greenland*. Danish Polar Center, Copenhagen, pp. 257–270.
- Weidick, A., Oerter, H., Reeh, N., Thomsen, H.H., Thorning, L., 1990. The recession of the Inland Ice margin during the Holocene Climatic Optimum in the Jakobshavn-Isfjord area of west Greenland. *Global and Planetary Change* 82, 389–399.
- Weidick, A., Kelly, M., Bennike, O., 2004. Later Quaternary development of the southern sector of the Greenland Ice Sheet, with particular reference to the Qassimiut lobe. *Boreas* 33, 284–299.
- Whillans, I.M., 1976. Radio-echo layers and recent stability of West Antarctic ice sheet. *Nature* 264, 152–155.
- Wilken, M., Mienert, J., 2006. Submarine glacial debris flows, deep-sea channels and past ice-stream behaviour of the East Greenland continental margin. *Quaternary Science Reviews* 25, 784–810.
- Willerslev, E., Cappellini, E., Boomsma, W., Nielsen, R., Hebsgaard, M.B., Brand, T.B., Hofreiter, M., Bunce, M., Poinar, H.N., Dahl-Jensen, D., Johnsen, S., Steffensen, J.P., Bennike, O., Schwenninger, J.L., Nathan, R., Armitage, S., de Hoog, C.J., Alimov, V., Christi, M., Beer, J., Muscheler, R., Barker, J., Sharp, M., Penkman, K.E. H., Haile, J., Taberlet, P., Bilbert, M.T.P., Casoli, A., Campani, E., Collins, M.J., 2007. Ancient biomolecules from deep ice cores reveal a forested Southern Greenland. *Science* 317, 111–114.
- Wouters, B., Chamfers, D., Schrama, E.J.O., 2008. GRACE observes small-scale mass loss in Greenland. *Geophysical Research Letters* 35, L20501. doi:10.1029/2008GL034816.
- Zielinski, G.A., Mayewski, P.A., Meeker, L.D., Whitlow, S., Twickler, M.S., Morrison, M., Meese, D.A., Gow, A.J., Alley, R.B., 1994. Record of volcanism since 7000 B.C. from the GISP2 Greenland ice core and implications for the volcano-climate system. *Science* 264, 948–950.
- Zwally, H.J., Abdalati, W., Herring, T., Larson, K., Saba, J., Steffen, K., 2002. Surface melt-induced acceleration of Greenland ice-sheet flow. *Science* 297, 218–222.
- Zwally, H.J., Giovinetto, M.B., Li, J., Cornejo, H.G., Beckley, M.A., Brenner, A.C., Saba, J.L., Yi, D.H., 2005. Mass changes of the Greenland and Antarctic ice sheets and shelves and contributions to sea-level rise: 1992–2002. *Journal of Glaciology* 51 (175), 509–527.