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Global Climate and Sea Level

Enduring Variability and Rapid Fluctuations Over the Past 150,000 Years

Closely packed "steps" of massive coral reef structures near Bobongara Village, Papua New Guinea. Each terrace is over 10-m high and 10-m deep and extends for over a kilometer. They were constructed in direct response to 10 m to 30 m rapid sea level rises following large-scale iceberg discharges into the North Atlantic from the Laurentide Ice Sheet during the last glacial period. The large, 3.3 m ky^{-1}, uplift of the area is responsible for revealing their structure above present sea level. The monolithic block in the background skyline is the last interglacial terrace.

ABSTRACT. Although climate variations and sea level changes are often discussed interchangeably, climate change need not always result in sea level change. Perturbations in Earth's orbit cause major climate changes, and the resulting variations in the amount and distribution of solar radiation at ground level follow cycles lasting for thousands of years. Research done in the last decade shows that climate can change on centennial or shorter time scales. These more rapid changes appear to be related to modifications in ocean circulation initiated during the last glacial period either by injections of fresh meltwater or huge ice discharges into the North Atlantic. When first detected, these rapid climate changes were characterized as episodes decoupled from any significant change in sea level. New data clearly show a direct connection between climate and sea level, and even more surprising, this link may extend to times of glacial-interglacial transitions and possibly also to interglacials. The full extent of this sea level/climate coupling is unknown and is the subject of current research.

INTRODUCTION: Drivers of Rapid Global Change

The last 150,000 years span a fascinating time in climate history. The growth and decay of large Northern Hemisphere ice sheets acting in harmony with major variations in ocean circulation amplified climate variations and resulted in severe and rapid climate swings throughout this interval. The consequences extended beyond climate variability to include rapid, large-scale changes in sea level that are evident in tropical corals located thousands of kilometers from the North Atlantic and the adjacent continental ice sheets.

The fundamental cause of these major swings in climate is tied to changes in insolation, the effective solar heating of the planet. Insolation at any one location is tightly controlled by perturbations in Earth's orbit and the inclination of its rotational axis relative to the plane of the ecliptic (Milankovitch, 1930; Hays et al., 1976; Imbrie et al., 1984; Imbrie and Imbrie, 1979). There may be uncertainties in the exact details of the process, but there is widespread agreement that perturbations of Earth's orbit are a key driver of climate variability (Lisiecki, 2010; Muller and MacDonald, 2000). Earth's orbital parameters have periods of 23,000 (precession), 41,000 (tilt), and 100,000 (eccentricity) years, with the 23,000-year cycle dominating insolation. By contrast, the several most recent ice ages repeat with a period of ~ 100,000 years, which is the weakest cycle. By including a nonlinear response to insolation at times of critical transitions, it has been shown that the impact of 100,000-year cycles can strengthen to a level that agrees with observations (Imbrie and Imbrie, 1980; Muller and MacDonald, 2000).

These astronomically forced changes in insolation and, hence, ice history are clearly correlated to global sea level variations (Figure 1). But, due to the redistribution of mass accompanying ice sheet growth and decay, there are additional, localized effects that complicate efforts to recognize a uniform sea level response to orbital forcing/insolation changes on a global scale (Figure 2). Large ice sheets deform the crust and mantle as they grow and decay, and as a result, sea level measured at a particular location does not necessarily represent the global ocean level. This time-dependent process is a function of the proximity of the ocean to a large ice sheet and must be accounted for before global sea level can be derived. For example, the 3-km-thick Laurentide Ice Sheet over North America contained so much localized mass that it shifted Earth's axis of rotation and distorted sea level in its immediate vicinity. Crustal loading led to mantle flow away from the ice load, resulting in a peripheral bulge in the Atlantic Ocean some distance from the ice sheet. In addition to lowering globally averaged sea level, the increased ice on land reduced the seawater load over the ocean, permitting ocean basin uplift. Then, as the ice melted, any region previously affected by the peripheral bulge began to sink, and the water load returning to the ocean depressed the ocean basin once again. Clearly, the net effect of large buildups of continental ice is a complex mix of these processes, and the sea level outcome depends on the proximity of each location in question to the ice sheet.

While sea level variations of many

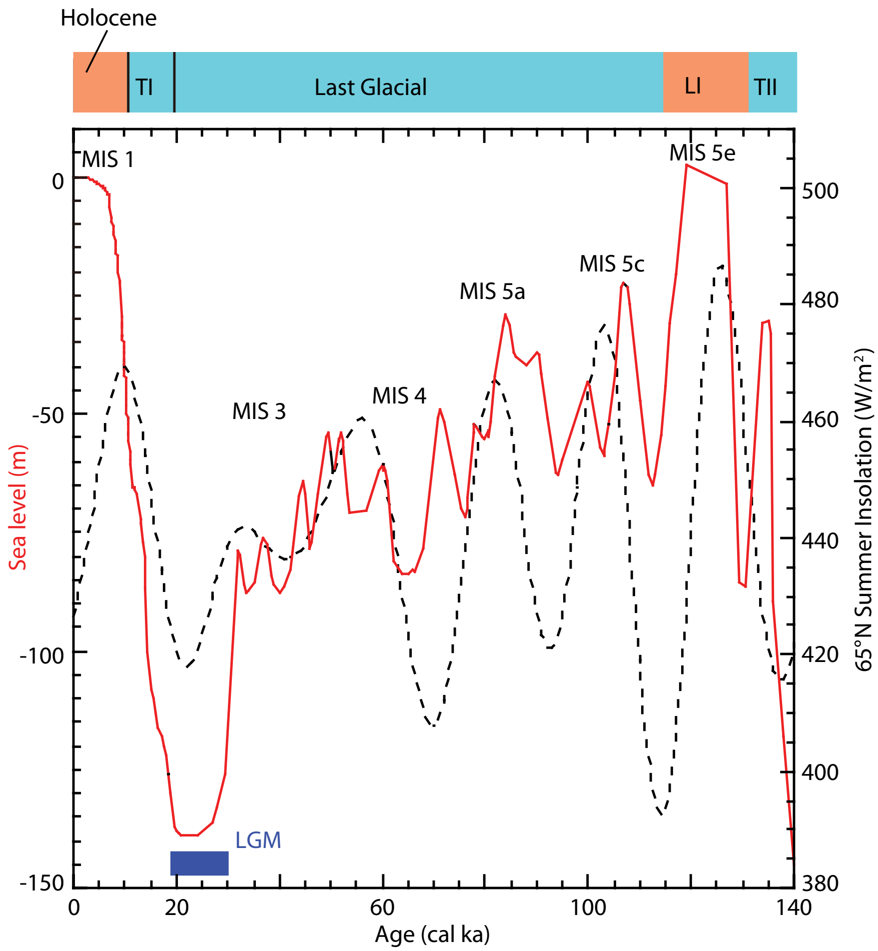

Figure 1. Prominent climate events of the past 140,000 years represented as variations in sea level (red curve). The 65°N insolation curve, based on the Milankovitch (1930) astronomical theory of orbital perturbations, closely follows major sea level highstands (dashed line) although there are leads and lags between the two in almost every case. Climate milestones cited in the text are identified as Holocene (0–10,400 years ago), Termination I (TI, 10,400–24,000 years ago), Last Deglaciation (LD, 10,400–19,000 years ago), Last Glacial Maximum (LGM, 19,000–26,000 years ago), Last Glacial (26,000–116,000 years ago), Last Interglacial (LI, 116,000–129,000 years ago), and Termination II (TII, 129,000–150,000 years ago). ka = thousands of years ago. MIS = Marine Isotope Stage.

tens to over 100 m are linked in some manner to Earth's orbital parameters, this cyclic forcing acts at periods of tens of thousands of years. For over a decade it has been clear that there are rapid swings in climate on centennial or shorter time scales, in particular, during long ice buildups culminating in glacial maxima (Alley, 1998), but not exclusively so. In 2001, we were the first to present evidence from U-series dating of coral terraces at Huon Peninsula, Papua New Guinea (PNG), that showed there were rapid and significant sea level variations during the ice buildup from 116,000 to 26,000 years ago (hereafter referred to as the "last glacial"; Figures 1 and 3; Yokoyama et al., 2001a). This finding was greeted with some skepticism (Hemming, 2004) until subsequent and novel analysis of Red Sea sediment cores (Siddall et al., 2003) confirmed the earlier results (Chappell, 2002). There is growing evidence that rapid climate changes are closely associated with rapid sea level variations (Deschamps et al., 2009; Weaver et al., 2003), although the full extent of this correlation is presently unclear.

BACKGROUND: The Challenge of Age Control

Part of the problem in linking rapid climate changes to sea level is the difficulty of determining the age of past sea levels with sufficient accuracy. Fractionation of oxygen isotopes during evaporation/precipitation, and subsequent accumulation of snow to form glaciers, provides a step toward meeting this challenge. Seawater at times of ice growth becomes increasingly enriched in ^{18}O, and this pattern is recorded in the $CaCO_3$ shells of marine

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microfossils and macrofossils. Although seawater temperature affects the $CaCO_3$ $\delta^{18}O$ value, it can be accounted for with enough reliability in Late Pleistocene records to be a precise measure of global ice volume, making it a powerful climate indicator and correlation tool. When tied to dated material, the marine $\delta^{18}O$ record provides much-needed detail between direct measurements of past elevations of sea level.

These direct measurements in the time interval considered here come from nearshore environments (see Englehart et al., 2011, this issue) and corals, although corrections are needed to return each to its elevation at the time of burial. Continuous records accumulate only at times of shoreline encroachment (transgression), so direct knowledge of sea level change is necessarily incomplete. This is where the proxy measurements of ice volume provided by $\delta^{18}O$ make a contribution by filling in the record between established tie points from marshes and corals. But this method still leaves the challenge of determining age control with enough accuracy to capture especially rapid changes in sea level.

Radiocarbon dating of carbon-bearing material (microfossils, molluscs, corals, organic matter) is effective back to about 50,000 years ago, though external calibrations are needed to account for variations caused by cycling of carbon through major reservoirs such as the atmosphere, ocean, and biota (Reimer et al., 2004, 2009; Fairbanks et al., 2005; Yokoyama et al., 2000a; Esat and Yokoyama, 2008). U-series dating of corals can extend knowledge of sea level events farther back in time, but samples older than 150,000 years are scarce and at present provide an incomplete record (Stirling and Anderson, 2009). High-precision U-series dating of corals depends on high-quality samples to ensure a closed system in which the parent and daughter nuclides in the decay chain remained undisturbed (Edwards et al., 1987; Gallup et al., 2002; Stirling et al., 1995). Unfortunately, this condition is difficult to satisfy; often there are systematic variations in age determinations apparently correlated with variations in the inferred composition of the ocean's U isotopes when the corals were growing (Gallup et al., 2002). These variations occur mostly

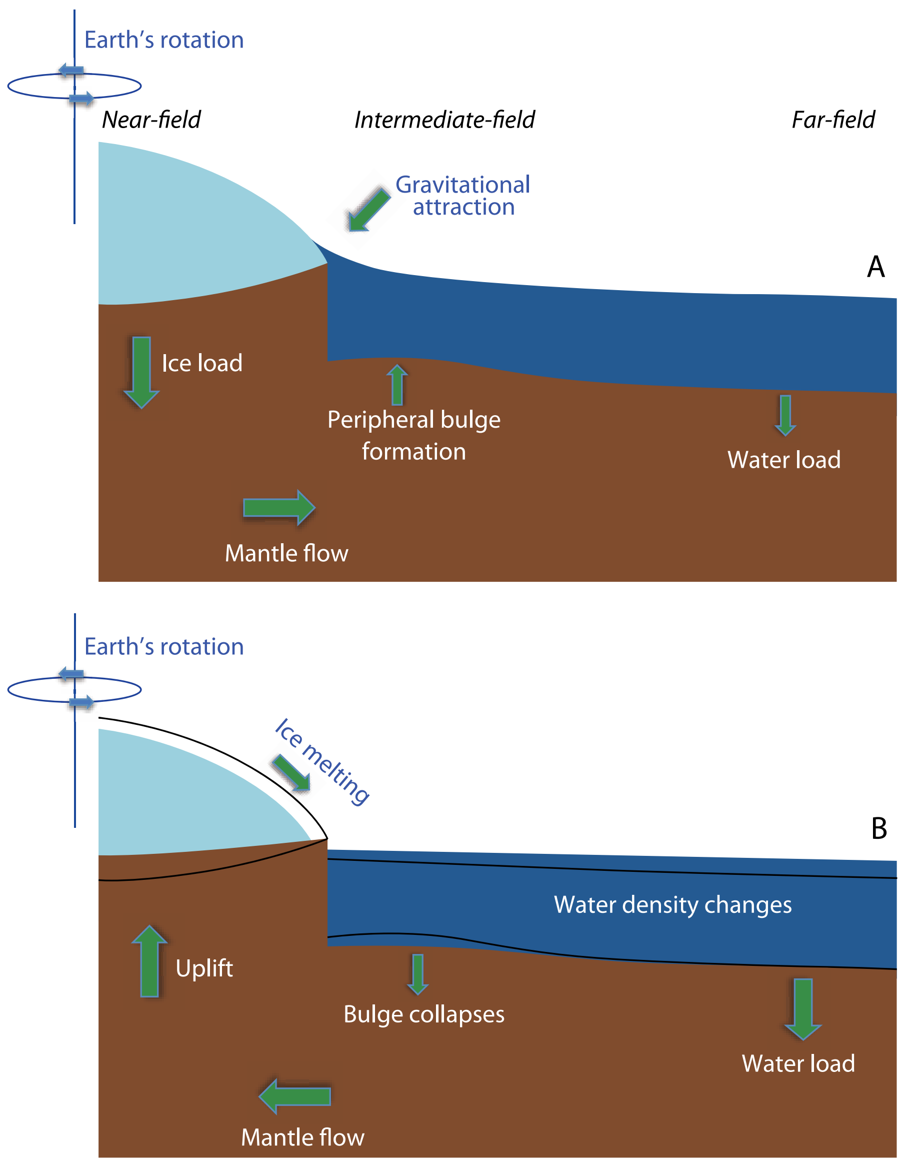


Figure 2. Several effects of changing the mass of large ice sheets. (A) During the growth stage, these are: local crustal loading and an intermediate crustal bulge due to mantle flow; gravitational attraction of seawater adjacent to and toward the ice margin; and perturbation of Earth's rotational axis. (B) During the melting stage, there is local uplift, bulge collapse, and a decrease in rotational disturbances and local gravitational effects on seawater.

during sea level transitions (Esat and Yokoyama, 2006b, 2010). There are two diametrically opposed explanations for such behavior. In one, the variability is attributed to the movement of U and Th isotopes in and out of fossil corals in a well-defined pattern. The pattern of variations can then be used to derive a "corrected" age (Thompson et al., 2003; Thompson and Goldstein, 2005). In the other, the variability in U isotopes is attributed to actual changes in the U-isotope abundances in the ocean (Esat and Yokoyama, 2006b, 2010). The variations occur when rising sea levels access shoreline sediments containing significant amounts of U with a distinct U-isotopic composition. We believe the second explanation for the U-isotope variability to be more plausible. Briefly, the first model significantly affects the timing of many sea level transitions, such as the last interglacial (Thompson and Goldstein, 2005), and data indicate a highly variable period of particularly rapid sea level fluctuations rather than a regular pattern (Gallup et al., 1994; Stirling and Anderson, 2009; Esat and Yokoyama, 2006b, 2010).

Less-direct proxies than U-series dating of corals can be used to determine ages, but they often lack precise time control (Lisiecki and Raymo, 2005; Bintanja et al., 2005). One exception is speleothems (cave stalagmites and stalactites), which can provide precise dates but lack direct connection to sea level. Stable isotopes within speleothems, however, provide a connection with climate, which in turn may be related to the timing of major sea level transitions (Cheng et al., 2006, 2009). In rare cases, speleothems growing in periodically submerged caves can provide upper limits to the age of sea level highstands (Richards et al., 1994; Dutton et al., 2009). In verifying rapid changes in sea level, precision and accuracy of dating become important.

Considering the preceding discussion of rapid climate variability, its intriguing but incompletely documented link to sea level change, and the challenge of providing much needed age control, the following summaries focus on known instances of rapid sea level variation in the past 150,000 years. Our intent is to deduce overriding principles to help identify the triggers, nature, and course of such climate events. Measurements and records that can be clearly explained are not evenly spread across this time interval. Consequently, we summarize the record out of chronologic order, beginning with a discussion of the salient features of the long-term buildup of ice during the last glacial. We continue through the Last Glacial Maximum to Termination I before evaluating some of the less-well-documented aspects of the older record back as far as Termination II and the development of the last glacial period.

THE LAST GLACIAL: 116,000–26,000 Years Ago

Numerous instances of rapid and severe temperature changes taking place over a few years to decades were first revealed in analyses of Greenland ice core records of the last glacial period (Figure 3; Stuiver and Grootes, 2000). It was subsequently realized that North Atlantic and Pacific deep-sea cores contained evidence of similar events (e.g., layers of mineral grains—"ice-rafted debris"—first described by Heinrich [1988]). Broecker (1994, 2003) related the cold-warm climate fluctuations to the stop-start behavior of the North Atlantic circulation paced by periodic, massive iceberg discharges from major Northern Hemisphere ice sheets. Ice-rafted debris, scoured from continental bedrock and subsequently released to the seafloor as the icebergs melted in the open ocean, provided evidence in support of this hypothesis (Broecker, 1994, 2003). These so-called "Bond cycles" occurred at intervals of 6,000–7,000 years (Alley, 1998). Each cycle consisted of several relatively minor, successively cooler, climate oscillations lasting 1,000–1,500 years (previously known as "Dansgaard-Oeschger events) followed by a major, intense cold period called a "Heinrich event" (H1, H2, etc.). The physical presence of icebergs as well as the freshening of North Atlantic waters by melting of the icebergs prevented North Atlantic deepwater formation, initiating a cold snap and forming an extensive cover of winter sea ice that caused further cooling (Alley, 1998; Hemming, 2004; Clark et al., 2007; Broecker et al., 2010). The subsequent restart of circulation led to an abrupt, intense warm period followed by another Bond cycle (Hemming, 2004). Although the connections among iceberg discharges, changes in ocean circulation, and climate fluctuations are well understood, there is no obvious cause for periodic ice discharges. The emphasis during this period of discovery was on understanding rapid changes in climate; possible sea level rises due to ice sheet discharges were considered, but any sea level rise was expected to be minor, in the range of 1 to 2 m (MacAyeal et al., 1993; Hemming, 2004).

Then, progress came from an unexpected source: closely spaced ice age coral reef terraces that extend 80 km along the coast of the Huon Peninsula (Chappell et al., 1996; Yokoyama et al., 2001b; Esat and Yokoyama, 2006a). During the last glacial (116,000–26,000 years ago), global sea levels were 60–80 m below present (Yokoyama et al., 2007). Corals of that period are now again typically underwater and difficult to sample directly. However, in a few places, due to continuous tectonic uplift, they have been elevated above sea level. One of these reefs is near Bobongara village, Huon Peninsula, where the average uplift rate is ~ 3.3 m ky^{-1} (see photos on p. 54 of this article). Corals at this location constructed more than six terraces, each 10-m to 20-m high and 10-m to 20-m deep, currently ranging in elevation from 20 m to 130 m (Chappell, 2002; Yokoyama et al., 2001a). Early investigators speculated that these terraces resulted from episodic, co-seismic

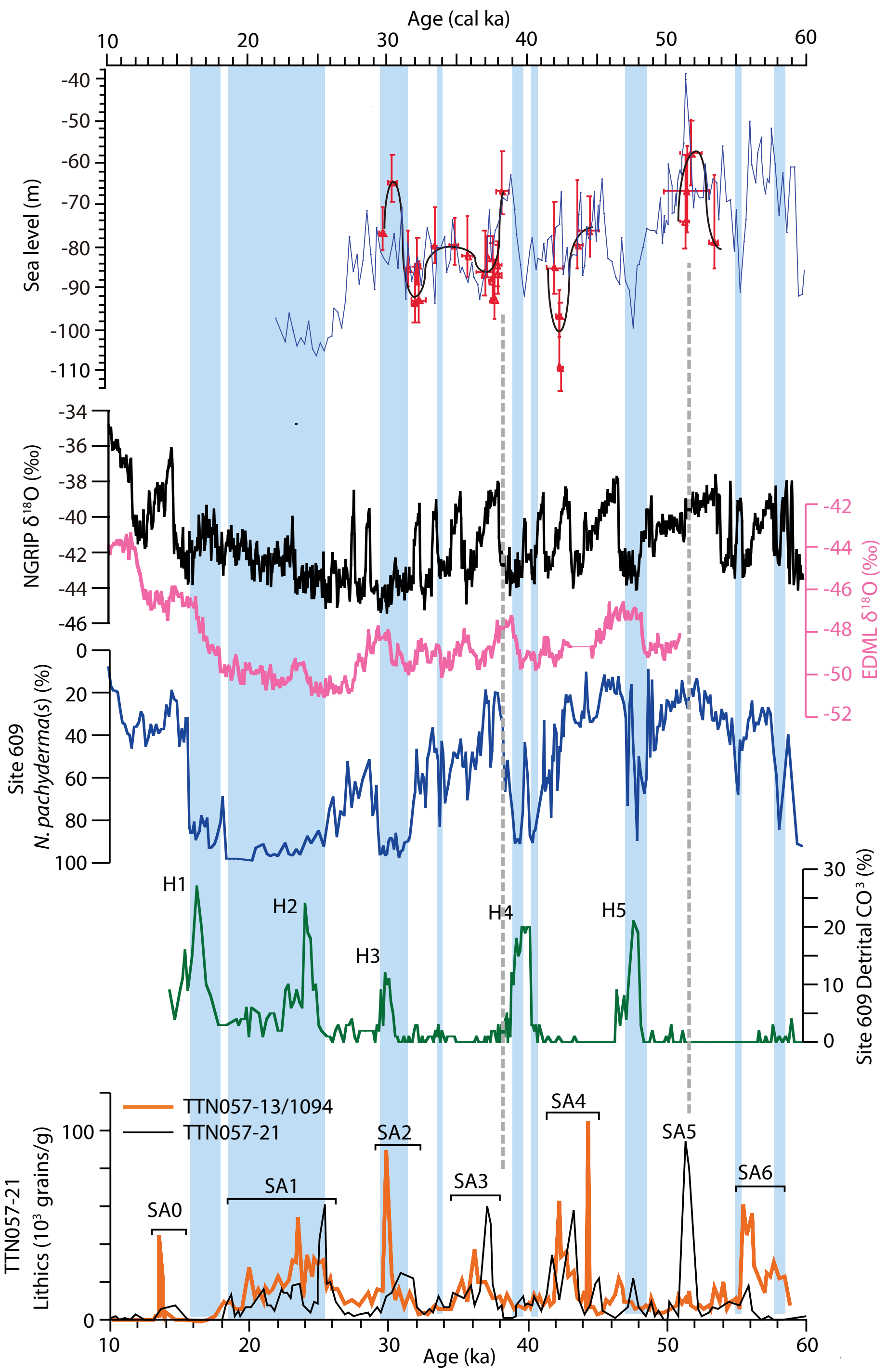


Figure 3. Rapid sea level variations in the interval 60,000–10,000 years ago compared with multi-parameter ice and deep-sea core records. The blue sea level curve in the upper panel is based on oxygen isotopes of foraminifera from the Red Sea corrected for salinity variations due to variable mixing with Indian Ocean seawater (from Siddall et al., 2003). There is close agreement with superimposed values of sea level shown in red (smoothed curve in black) that are based on isotopes in corals from the Huon Peninsula, Papua New Guinea (PNG), and dated with U-series techniques (Yokoyama et al., 2001a,b). The PNG coral ages have been recalculated due to their low $^{234}U/^{238}U$ ratios relative to current seawater values (Siddall et al., 2008; Thompson and Goldstein, 2005). These corrections increase the coral ages and cause stratigraphic distortion between the data sets. We believe the lower ^{234}U levels in the glacial ocean occur naturally, and therefore there is no need for correcting the coral ages (Esat and Yokoyama, 2006b, 2010). The North Greenland Ice Core Project (NGRIP) ice core record (black) is from Greenland (North Greenland Ice Core Project Members, 2004), and the EPICA Dronning Maud Land ice core (EDML, pink) is from Antarctica (EPICA Community Members, 2006). The abundance of *N. pachyderma* (sinistral) (blue) is data from the North Atlantic Deep Sea Drilling Project Site 609 (Obrochta, 2008); cold periods are indicated by sharp increases in *pachyderma* abundances that also correlate with ice-rafted carbonate debris from the same site and are labeled as Heinrich events H1 to H5 (green). The vertical blue bands identify cold events based on the *pachyderma* curve. Ice-rafted lithic grains from two South Atlantic deep-sea cores (bottom panel; Kanfoush et al., 2000) are in general agreement with Site 609 except for the timing of South Atlantic (SA) events SA3, SA4, and SA5 that point to possible problems with age calibration. ka = thousands of years ago.

uplift events without a discernible link to changes in sea level (Chappell et al., 1996; Ota and Chappell, 1996). Such large tectonic shifts, however, are inconsistent with the known uplift history of the region (Pandolfi et al., 1992). U-series dating of the coral terraces has since revealed a close match with the timing of North Atlantic Heinrich events (Figure 3), and, like other well-documented Huon terraces, these ice age terraces point to a sequence of increases in sea level. The work on dating the Huon terraces and their relationship to North Atlantic Heinrich events was published in 2001 (Yokoyama et al., 2001a), and subsequently confirmed and reinforced by additional work (Chappell, 2002; Siddall et al., 2003). Within the framework of a Bond cycle, the 10–15-m sea level increases documented near Bobongara record the discharge of ice into the ocean. This amount of ice is equivalent to the entire volume of the European ice sheet at the height of the last glacial, or to 20% of the Laurentide Ice Sheet. This amount of sea level rise is also likely to have destabilized the Antarctic ice sheet and perhaps led to additional increases in sea level (Kanfoush et al., 2000; Yokoyama et al., 2001a; Rohling et al., 2004). Although icebergs released into the ocean may not melt rapidly, they do cause an immediate rise in sea level. It is at this time that the intense Heinrich event cold snap occurred and North Atlantic circulation was interrupted (EPICA community members, 2006).

It is now well accepted that 15–20 m, and possibly up to 30 m, of sea level increases are associated with last glacial Heinrich events (Broecker, 1994). Conceptually, the rapid sea level increases had to occur concurrently with North Atlantic ice discharges, which also initiated a cold snap as armadas of icebergs interrupted thermohaline circulation. This is the reverse of the usual oxygen isotope-temperature relationship determined by analysis of ice cores and deep-sea sediment cores, which equates warmer temperatures and higher sea levels with $\delta^{18}O$ increases (Figure 3). Contrary explanations, where sea level rise occurs slowly during the warm period after termination of a Heinrich event, have also been proposed (Arz et al., 2007). However, these types of models do not fully account for all of the observations (Rohling et al., 2008). Although there is no agreement regarding the mechanism that triggers the Dansgaard-Oeschger-Heinrich cycles, the ensuing warm conditions at the end of each cycle are readily understandable and accepted as signaling the restart of thermohaline circulation and the release of ocean heat (Hulbe, 1997; Jackson, 2000; Flückiger et al., 2006; Clark et al., 2007; Alvarez-Solas et al., 2010).

THE LAST GLACIAL MAXIMUM AND TERMINATION I: 26,000–10,400 Years Ago

The Last Glacial maximum (LGM) seems to follow a trend of climate transitions from cold to colder states before warming and deglaciation. There is no obvious explanation for this behavior and Heinrich event H2 about 24,000 years ago just after the start of LGM. From a compilation of all available data, Clark et al. (2009) determined that LGM extended from 26,500–19,000 years ago, with global sea level remaining throughout that period at 130 m below present. This timing agrees with the estimated duration of the maximum extent of most global ice sheets. Sea levels could have begun to fall as early as 32,000–30,000 years ago from their nominal lows of 50–70 m below present sea level during the last glacial period (Lambeck et al., 2002; Cutler et al., 2003; Siddall et al., 2003; Yokoyama et al., 2007). In contrast, LGM CO_2 levels appear to have decreased only marginally, by about 15 ppm (Clark et al., 2009).

Controversy surrounds the timing and nature of events at the end of LGM. Data from a Gulf of Bonaparte sediment core (North Australian continental shelf) indicate a rapid 10–15-m sea level rise lasting for about 500 years (Yokoyama et al., 2000b, 2001c; De Deckker and Yokoyama, 2009). Sediment cores from the Sunda shelf support the occurrence of a 10-m sea level rise, but place the start at the slightly earlier time of 19,400 years ago, with a duration of 800 years (Hanebuth et al., 2009). Similarly, observations from the western margin of the Irish Sea basin appear to show at least a 10-m sea level rise, but possibly at an earlier time, at about 19,800 years ago, due to uncertainties in reservoir ages associated with radiocarbon dating (Clark et al., 2004; McCabe et al., 2005, 2007). Coral data from Barbados further confuse the issue by indicating a sea level highstand at about 19,500 years ago, but with large depth uncertainties (Peltier and Fairbanks, 2006). Regardless of the exact timing of the nominal 19,000-year event, a 10-m rapid sea level rise likely affected North Atlantic deepwater formation and modulated the transition into the major deglaciation sequence leading to the Holocene (Severinghause, 2009). Indeed, a meltwater pulse at this time modulates ice sheet models that tend to otherwise underestimate global ice volumes (Yokoyama et al., 2000b, 2001c; Lambeck et al., 2000). Whether the 19,000-years-ago sea level fits in with a Heinrich-type climate event is uncertain. Clearly, additional data are required to better define the timing and details of these events (Figure 4).

The ~ 9,000-year duration (19,000–10,400 years ago) of the last deglaciation includes at least one clear climate reversal (Broecker et al., 2010; Thornalley et al., 2010). Detailed analysis indicates a close relationship between this reversal and Heinrich event H1 at around 16,500 years ago, both in terms of climate change and sea level (Figure 4). In turn, as the Heinrich events of the last glacial period were initiated by variations in North Atlantic deepwater formation, there is likely to be a link between the two (Broecker et al., 2010). The most prominent event of this period is the Younger Dryas intense cold period (12,900–11,700 years ago), whose onset was precipitated by the presence of fresh surface water in the North Atlantic, which prevented deepwater formation. Most accounts of the causes of the Younger Dryas assume a random catastrophic event, either freshwater release from the breach of glacial Lake Agassiz (Johnson and McClure, 1976; Rooth, 1982; Broecker, 2006) or a meteorite strike (Firestone et al., 2007). However, the key evidence against a random event is the occurrence of similar climate reversals at other glacial-interglacial transitions. In particular, the structure of termination TIII (~ 245,000 years ago) appears to be very similar to TI, and TII includes very large climate and sea level oscillations. Cortese et al. (2007) document sea surface temperature fluctuations during the last five terminations from analysis of a South Atlantic deep-sea core that are consistent with cold reversals being an integral part of glacial-interglacial transitions.

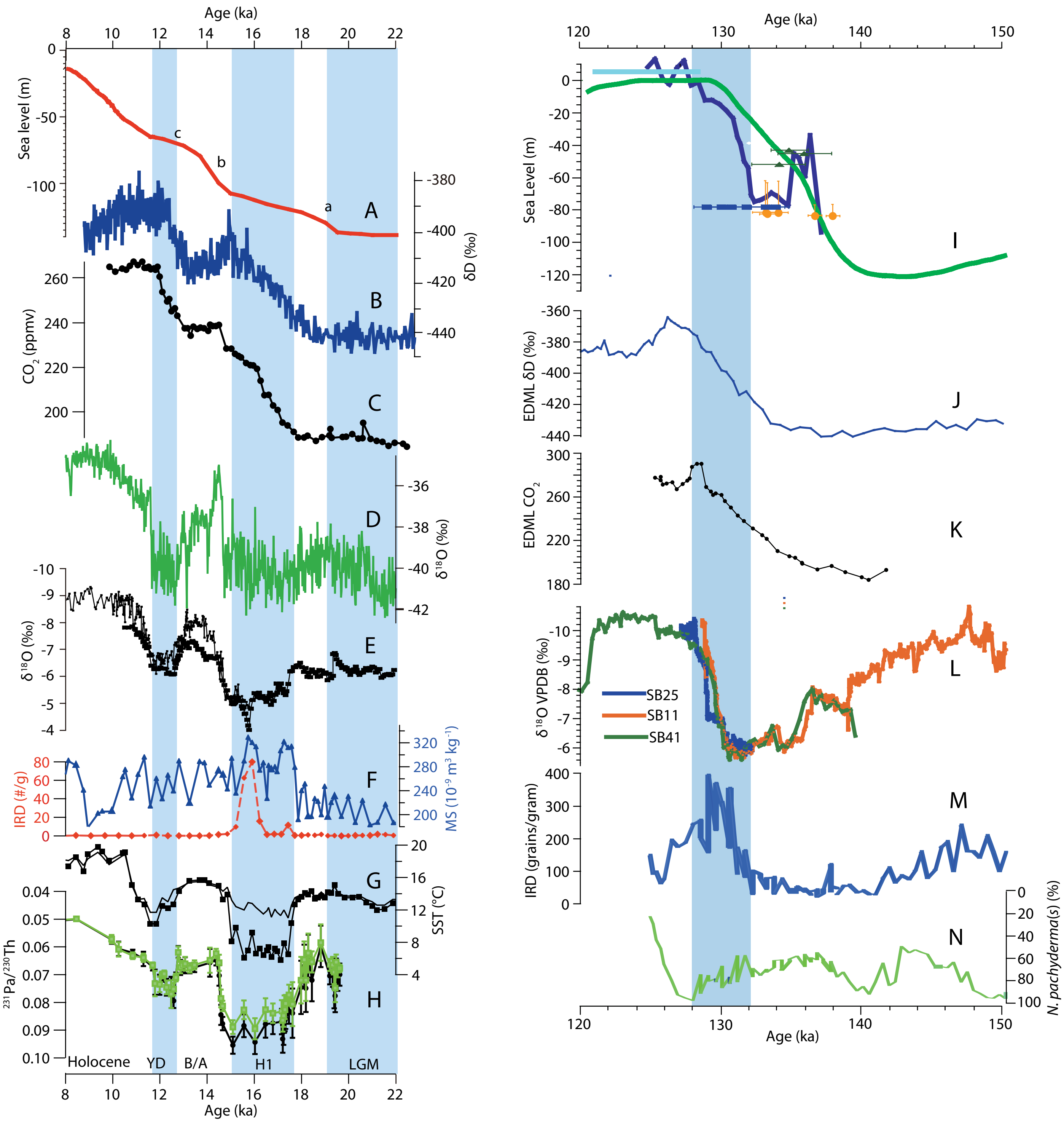
Ice cores from Antarctica and Greenland during a period labeled as the Antarctic Cold Reversal or the Bølling/Allerod (B/A; 14,500–12,900 years ago) show progressive cooling that terminates with the intense cold of the Younger Dryas (Figure 4). This structure is very similar to that of Dansgaard-Oeschger cycles followed by a Heinrich event. There is evidence for a small (< 6 m) sea level step at the onset of the Younger Dryas around 13,000 years ago in both Barbados and Tahiti coral records (Bard et al., 2010). The other major sea level fluctuation of this period is so-called Meltwater Pulse 1A around 14,600 years ago that coincides with the boundary between B/A (Deschamps et al., 2009) and the so-called "Mystery Interval" (MI; Denton et al., 2006), which includes Heinrich event H1 at 16,000–17,000 years ago, close to the beginning of MI. The exact timing and relationships among these sea level oscillations and changes in climate are not well established. However, Heinrich-type events, linked to variations in North Atlantic deepwater formation, appear to have continued from the last glacial into the last deglaciation. They may have occurred more frequently during TI than during the last glacial and may also have been present at other terminations.

> "RESEARCH DONE IN THE LAST DECADE SHOWS THAT CLIMATE CAN CHANGE ON CENTENNIAL OR SHORTER TIME SCALES."

TERMINATION II: 150,000–129,000 Years Ago

The transition from the penultimate glacial to the last interglacial (TII) occurred from about 150,000–129,000 years ago (Gallup et al., 2002; Cheng et al., 2006, 2009; Thomas et al., 2009). In comparison to TI, details of this period are poorly understood. There has been a long-standing argument over the timing of the TII sea level rise compared to the timing of the warming predicted by the Milankovitch theory (Winograd et al., 1992; Karner and Muller 2000; Henderson and Slowey, 2000). The Devils Hole record from Nevada, based on isotopes within a calcite vein (Winograd et al., 1992; Ludwig et al., 1992), and dating of Bahamas carbonate sediments (Henderson and Slowey, 2000) suggest early warming and early deglaciation during TII prior to Northern Hemisphere insolation taking full effect. These data have often been cited as evidence of shortcomings in the Milankovitch theory (Karner and Muller, 2000; Henderson and Slowey, 2000). However, the early warming could have been a regional signal rather than a global phenomenon. There is some evidence supporting this hypothesis from offshore of the Devils Hole locality, along the California margin where sea surface temperatures may have risen early in the deglaciation in response to the resumption of the warm California Current at the end of the glacial maximum (Herbert et al., 2001).

Major climate and sea level oscillations



during TI and TIII (~ 245,000 years ago) suggest that similar variability should be expected during the penultimate deglaciation ending at TII. However, boreal summer insolation during TII was significantly stronger than at either other termination. This stronger insolation may explain why TII shows the fastest rate of deglaciation of all three intervals, and may explain the suppressed North Atlantic deepwater formation and lack of a Younger Dryas-type climate oscillation during TII (Carlson, 2008). Alternatively, the stronger boreal insolation during TII may have triggered an especially large volume of ice discharge compared to that of the Younger Dryas or the Heinrich events of the last glacial. This large insolation overprint may have altered the climate system so that the several climate and sea level oscillations seen in TI and TIII never developed. We pursue this point in the discussion that follows.

The discovery of an ancient, fossil-coral-lined sea cave at Huon Peninsula, PNG, provided unexpected insights. The 130,000-year age and the location of the cave about 80 m below the last interglacial reef indicated a very large (≥ 60 m) sea level fall at about 130,000 years ago from an earlier peak in sea level at ~ 135,000 years ago, ~ 20 m below the last interglacial sea level (Esat et al., 1999). In addition, coral temperature proxies indicated conditions at 130,000 years ago were about 7°C cooler than temperatures of the last interglacial (McCulloch et al., 1999). The cave was named Aladdin's Cave and the rapid climate and sea level fluctuation was termed the "AC transition." The early rise in the sea level, up to almost interglacial levels, could be attributed to an earlier-than-expected peak in insolation forcing at 135,000 years ago, predating Milankovitch predictions (Figure 4). However, the 135,000-years-ago sea level highstand could not be sustained, and sea level dropped precipitously more than 60 m by 130,000 years ago before rising back to interglacial levels at 129,000 years ago in step with expected insolation forcing.

The pattern of climate and sea level variations during TI and TIII have been related to changes in North Atlantic deepwater formation associated with Dansgaard-Oeschger and Heinrich events (Oppo et al., 2001; Cheng et al., 2006; Thomas et al., 2009). In this sense, the early sea level rise and the cold snap, defined by the AC transition at 130,000 years ago (Esat et al., 1999; Beets and Beets, 2003; Gallup et al., 2002; Siddall et al., 2008; Thomas et al., 2009; Fujita et al., 2010) could be due to changes in ocean circulation precipitated by iceberg discharges. The subsequent rapid warming and rapid sea level rise also fit in as part of a Heinrich-type sequence. In comparison with TI, the magnitude of the sea level rise and fall at TII is substantially larger, presumably due to the strength of TII insolation (Carlson, 2008). Therefore, the large sea level oscillation at TII is likely to be due to ice discharges and related ocean circulation changes rather than to any deficiency in the Milankovitch insolation theory.

Based on all these observations, we conclude that Dansgaard-Oeschger and Heinrich-type cycles form an uninterrupted backdrop to ice ages, but also extend through to glacial-interglacial transitions. All are likely to be associated with rapid sea level fluctuations.

Figure 4 (opposite page). (LEFT COLUMN) Sea level curves and associated measurements from the Last Glacial Maximum to the Holocene. Vertical light blue bars (labeled at bottom) distinguish intervals LGM (Last Glacial Maximum), H1 (Heinrich event 1), B/A (Bølling/Allerod), YD (Younger Dryas), and Holocene. (A) Generalized sea level with three times of note: (a) at 19,000 years ago marking the end of LGM, (b) at Meltwater Pulse 1A at 14,600 years ago from Barbados and Tahiti coral records, and (c) at the start of the Younger Dryas cold period (Lambeck et al., 2002; Yokoyama et al., 2000b, 2001c; Deschamps et al., 2009; Bard et al., 2010). (B and C) CO_2 from the Antarctic Dome C ice core showing pauses in deglaciation during the B/A and YD intervals (Monnin et al., 2001). (D) Greenland Ice Sheet Project Two (GISP2) ice core $\delta^{18}O$ measurements (Stuiver and Grootes, 2000). (E) The $\delta^{18}O$ speleothem record from Hulu and Donge caves (Wang et al, 2001; Yuan et al., 2004) showing similarities to the GISP2 data in (D). (F) Ice-rafted debris (IRD) and magnetic susceptibility from a sediment core off Portugal showing events near 16,000 years ago may be similar and include Heinrich event H1 (Bard et al., 2000). (G and H) Sea surface temperature (Bard et al., 2000) and $^{231}Pa/^{230}Th$ (McManus et al., 2004) records highlight times of intense cold when deepwater formation was interrupted. The first of these is associated with a meltwater pulse that occurred 19,000 years ago. (RIGHT COLUMN). Sea level curves and associated measurements from the penultimate glacial to last interglacial transition (150,000–120,000 years ago). (I) The dark blue line shows sea level based on $\delta^{18}O$ measurements from the Red Sea, shifted by 2,500 years to fit coral ages that show the highstand at about 135,000 years ago (Siddall et al., 2003). The green line is based on modeling and high-latitude air temperatures (Bintanja et al., 2005) and is made to fit the coral-based chronology. The light blue line shows sea level from the last interglacial for comparison. The green triangles and blue bars are from the Huon Peninsula (Stein et al., 1993; Esat et al., 1999) and the orange circles from Tahiti (Thomas et al., 2009). (J and K) EPICA Dronning Maud Land (EDML) Dome-C hydrogen isotope and CO_2 record (Lourantou et al., 2010; Masson-Delmotte et al., 2010). (L) $\delta^{18}O$ from speleothems (cave stalactites and stalagmites) (Cheng et al., 2009). (M and N) Ice-rafted debris and *pachyderma* records from North Atlantic Deep Sea Drilling Project Site 609 (Obrochta, 2008). The vertical blue rectangle highlights the sharp transition to the last interglacial following the 35,000-years-ago highstand. ka = thousands of years ago.

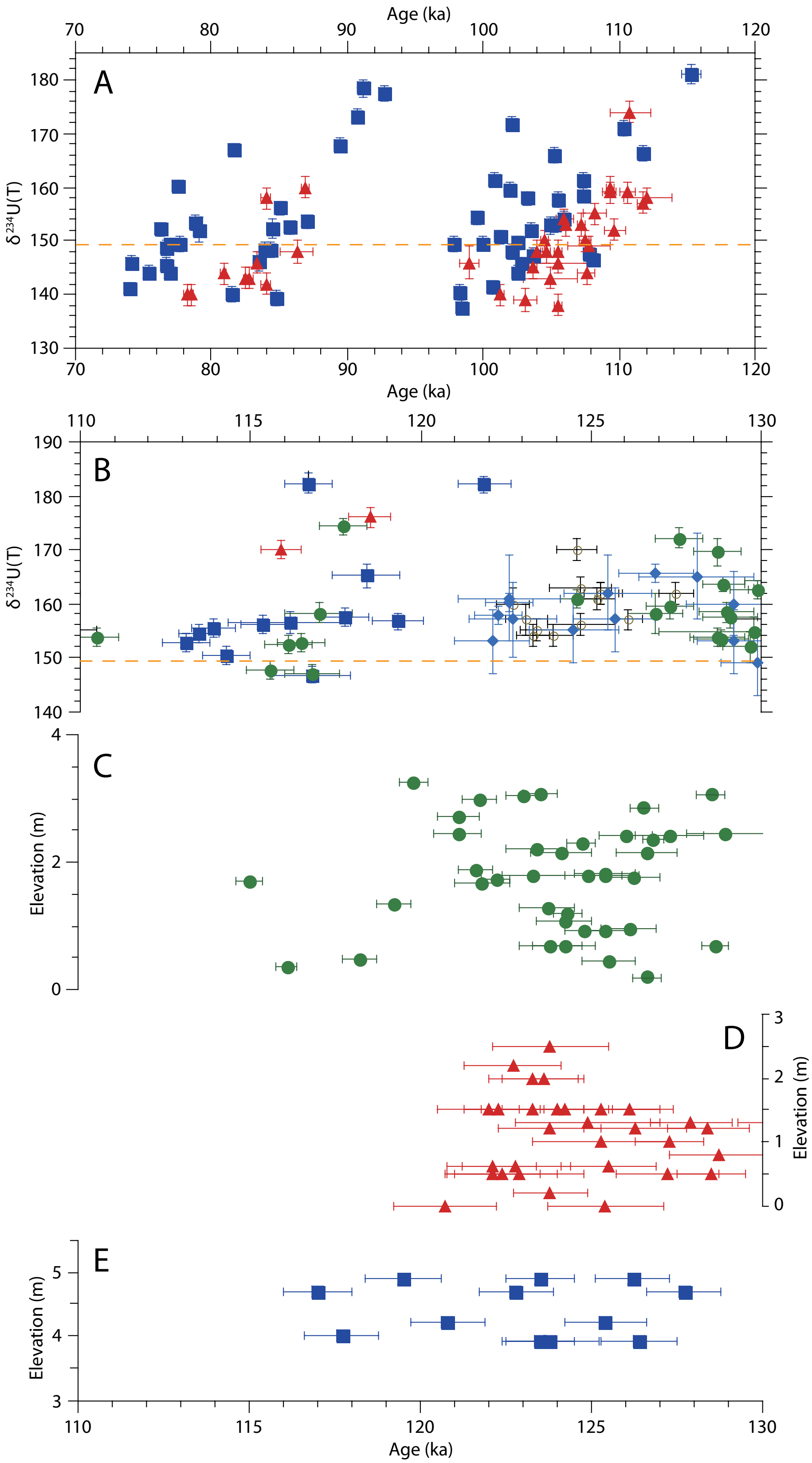


Figure 5. Coral-based measurements of seawater $^{234}U/^{238}U$ (A and B) and sea level (C–E) surrounding the time of the last interglacial. Data are from (A) Potter et al. (2004; blue boxes) and Thompson et al. (2005; red triangles), and (B) Cutler et al. (2003; blue boxes), Speed and Cheng (2004; red triangles), Thompson et al. (2003; open circles), and Edwards et al. (1987, 1997; blue diamonds). Interstadials 5a (80,000–74,000 years ago) and 5c (112,000–98,000 years ago) show internal structure in (A) that correspond to multiple sea level highstands with suborbital timing. Last interglacial coral ages from (C) Western Australia (Stirling et al., 1995, 1998) and (D) the Bahamas (Chen et al., 1991) show uninterrupted coral growth from 129,000–121,000 years ago and a gap in coral growth at 121,000–120,000 years ago. Coral growth in Western Australia resumes at 2–3-m lower elevations following the hiatus. Corals from Mexico (E) (Blanchon et al., 2009) show a highstand from 121,000–116,000 years ago that contradicts the West Australian results in (C). ka = thousands of years ago.

THE LAST INTERGLACIAL: 129,000–116,000 years ago

Last interglacial (LI) U-series coral ages predominate in the literature because these fossil corals are located above present-day sea level and thus are easy to access. LI timing is well constrained to be from 129,000–116,000 years ago (Figure 5; Stirling et al., 1995; 1998; Gallup et al., 1994; Chen et al., 1991). Results from tectonically stable sites, well away from large ice sheets, indicate that sea level during LI was 3–5-m higher than present (Lambeck and Nakada, 1992; Stirling et al., 1998). A recent statistical analysis of a wide range of sea level proxies shows higher global LI sea levels peaking at +9 m relative to present, in a highstand around 125,000 years ago (Kopp et al., 2009). However, this type of statistical approach is not necessarily the best way to collate variable-quality age data.

Sea levels during LI were higher than

today for one or both of the following reasons: warmer global temperatures, or a greater flux of icebergs into the ocean (Overpeck et al., 2006; Otto-Bliesner et al., 2006). The latter is explained by any of several processes referred to as "ice sheet dynamics," which include (among others): buttressing of coastline and grounded glaciers, bedrock composition/bedrock topography, or rates of ice sheet flow as a function of thickness. A sea level rise approaching 130 m in as short a span as 10,000 years may require both processes. It should be noted that the observed difference of a few meters of sea level between TII and TI corresponds to less than 4% of the present ice volume, and could be explained not by temperature differences but instead by differences in ice sheet dynamics alone (e.g., Alley et al., 2010).

There have been persistent reports of a sea level highstand at the end of LI. In a study from Mexico, a 6-m highstand, about 3 m higher than the LI sea level, may have extended from 121,000–116,000 years ago (Figure 5; Blanchon et al., 2009). This study provides the only evidence for a warm-to-cold climate transition proceeding through a period of warmer climate and higher sea level. However, these data have large analytical uncertainties and are in conflict with western Australia results, which only allow a narrow window (117,000–118,000 years ago) for such an excursion (Stirling et al., 1998). In addition, Mexico is likely to have been subjected to isostatic movements due to its proximity to Northern Hemisphere ice sheets (Lambeck et al., 2002). Beyond 115,000 years ago, a further drop in sea level is expected at the start of the last glacial.

MARINE ISOTOPE STAGES 5a AND 5c: 105,000 and 77,000 Years Ago

Closely following the end of last interglacial, two periods of high Milankovitch insolation occurred at approximately 77,000 and 105,000 years ago. With lags of ~ 5,000 years, they correspond to Marine Isotope Stages 5a and 5c when global sea levels may have briefly increased to 20 m and 28 m below present, respectively (Figures 1 and 5). At Barbados, U-series dates of discrete coral terraces correspond to these two major interstadials, which are close to present-day sea level and show multiple sea level highstands (Potter et al., 2004). Similarly, uplifted coral terraces at Huon Peninsula show two to three discrete terrace structures for each interstadial, also pointing to multiple sea level highstands (Stein et al., 1993). Such suborbital-period events are clearly outside the scope of the Milankovitch orbital forcing theory (Milankovitch, 1930) and point to complex interactions among ice sheets, ocean circulation, and climate that are active during warm periods with reduced ice volumes and are not solely confined to glacial periods.

GENERAL OBSERVATIONS

We offer three general observations derived from this and related studies. The first involves how ice sheets grow. Records summarized in this article show that global changes in ice volume are not symmetric: ice sheets can decay rapidly while their growth is a longer-term process. This gradual accumulation of snow and ice at high latitudes requires significant amounts of moisture to be transported from warm equatorial latitudes to the poles by way of warm ocean currents. It also requires circumpolar fronts to be weak or limited in extent for this warm water to reach the polar ice sheets and provide the requisite snowfall. From this generalization, it appears that ice growth requires at least two conditions: relatively warm winters that allow tropical ocean currents to penetrate high latitudes and sustain ice sheet growth, and cool summers that prevent ice sheet melting.

A second generalization involves climate changeovers. Studies show that a transition from one climate or sea level state to another frequently includes a brief interval of extreme temperature. For example, in going from the last global ice buildup to the last

"NEW DATA CLEARLY SHOW A DIRECT CONNECTION BETWEEN CLIMATE AND SEA LEVEL, AND EVEN MORE SURPRISING, THIS LINK MAY EXTEND TO TIMES OF GLACIAL-INTERGLACIAL TRANSITIONS AND POSSIBLY ALSO TO INTERGLACIALS."

deglaciation, the climate passed through the colder Last Glacial Maximum (CLIMAP, 1976, 1981; Mix et al., 2001; MARGO Project members, 2009; Clark et al., 2009). Similarly, the warmth at the termination of a Heinrich event follows a much colder period at the end of a succession of cooler Dansgaard-Oeschger cycles (Alley, 1998; Hemming 2004; Clark et al., 2007). The same dynamic cycle appears to have been present during deglaciations as in the case of the Younger Dryas and for the Aladdin's Cave record of climate and sea level oscillation at the penultimate deglaciation. Most of these processes not only involved climate change but also sea level fluctuations.

As a third generalization, we note that it is possible that a transition from a warm interglacial to the next glacial may proceed through a period that is actually warmer than the ambient interglacial climate. Examples are rare and not well established (Neumann and Hearty, 1986; Blanchon et al., 2009). The best evidence can be seen at the end of the last interglacial when the climate may have gone through a warm period with higher sea levels at roughly 118,000–117,000 years ago (Stirling et al., 1998).

CONCLUSIONS

There is no question that major global climate and sea level variations are mainly in agreement with the theory of astronomical forcing (Milankovitch, 1930). However, climate also appears to be active over much shorter time scales than this theory predicts. As apparent in the records presented here, over the last 150,000 years there has been rapid climate variability acting on suborbital time scales, often associated with significant sea level fluctuations. A dominant cause appears to have been the interaction between thermohaline circulation in the North Atlantic and major Northern Hemisphere ice sheets. During the last glacial in particular (116,000–26,000 years ago), this variability was manifest as Bond cycles that encompassed sequences of Dansgaard-Oeschger oscillations and Heinrich events. This interaction appears to have continued through the transitions from glacial to interglacial periods as well. Suborbital variability is apparent within the last interglacial period and possibly the Holocene (Bond et al., 2001), although these periods include high sea level and stable, low ice volumes that presumably had little influence on climate. Much remains to be done to better understand the nature and cause of rapid climate and sea level variability. In this context, the need for high-quality, high-precision data from fossil and live coral reefs is paramount.

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