MECHANISMS OF ABRUPT CLIMATE CHANGE OF THE LAST GLACIAL PERIOD

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[1] More than a decade ago, ice core records from Greenland revealed that the last glacial period was characterized by abrupt climate changes that recurred on millennial time scales. Since their discovery, there has been a large effort to determine whether these climate events were a global phenomenon or were just confined to the North Atlantic region and also to reveal the mechanisms that were responsible for them. In this paper, we review the available paleoclimate observations of abrupt change during the last glacial period in order to place constraints on possible mechanisms. Three different mechanisms are then reviewed: ocean thermohaline circulation, sea ice feedbacks, and tropical processes. Each mechanism is tested for its ability to explain the key features of the observations, particularly with regard to the abruptness, millennial recurrence, and geographical extent of the observed changes. It is found that each of these mechanisms has explanatory strengths and weaknesses, and key areas in which progress could be made in improving the understanding of their long-term behavior, both from observational and modeling approaches, are suggested. Finally, it is proposed that a complete understanding of the mechanisms of abrupt change requires inclusion of processes at both low and high latitudes, as well as the potential for feedbacks between them. Some suggestions for experimental approaches to test for such feedbacks with coupled climate models are given.

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1. INTRODUCTION

[2] Beyond the limited range of historical and instrumental data, the geologic record has provided abundant evidence of natural climate changes that are both abrupt and large in magnitude. The most dramatic insights into this aspect of climate system behavior have come from studies of ice and sediment cores from the northern North Atlantic. Ice core results from Greenland were the first to show that significant changes in regional climate occurred in the past over time scales of a few years to a few decades at most [e.g., *Alley et al.*, 1993; *Taylor et al.*, 1993]. The most recent and perhaps best studied of these events is the Younger Dryas (YD), an abrupt return to near-glacial temperatures in the high-latitude North Atlantic lasting roughly a millennium during the last deglaciation (Figure 1). Rapid temperature excursions were also a characteristic feature of Greenland climate during the last glacial period according to ice core records. These excursions, which have come to be known as Dansgaard-Oeschger or D-O events [*Dansgaard et al.*, 1984, 1993], recurred roughly every 1500 years and show up as abrupt warmings over Greenland of as much as 10°C followed by a somewhat more gradual return to cold glacial conditions (Figure 1). *Bond et al.* [1993] showed that D-O events in Greenland are closely matched by sea surface temperature (SST) changes recorded in high–accumulation rate sediments in the North Atlantic. Both the abrupt nature of the cold-to-warm shifts that terminate the cycles and the recurrence time scale (Figure 1) were found to be similar, suggesting a linkage between oceanic and cryospheric processes in this region. In addition, discrete layers of ice-rafted debris found in subpolar North Atlantic sediments [*Bond et al.*, 1992], Heinrich events, were deposited periodically during the coldest phases of the D-O cycles (Figure 1) and are thought to have originated from icebergs that came out of the Hudson Strait [e.g., *Hemming*, 2004].

[3] Though the amplitude and timing of abrupt climate change are best defined in the Greenland ice core records for the last glacial period between about 80,000 and 15,000 years ago, a large number of records from around the world show evidence of climate events with a similar temporal behavior [e.g., *Broecker and Hemming*, 2001; *Voelker and Workshop Participants*, 2002; *Hemming*, 2004]. The challenge is to determine whether the millennial variability that clearly shows up as abrupt change in Greenland is actually related to that at other locations around the globe or has the same cause. This issue can be addressed from both an

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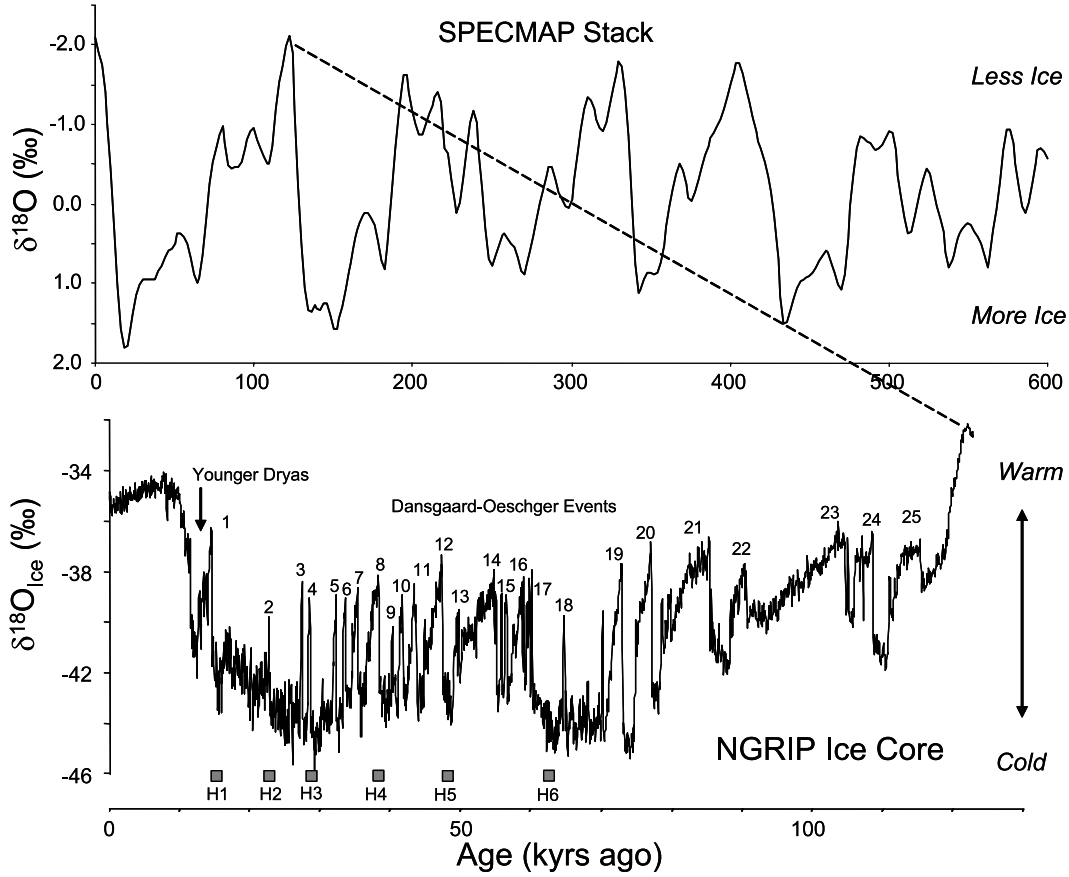

Figure 1. Comparison of (bottom) North Greenland Ice Core Project (NGRIP) ice core oxygen isotope record (a proxy for air temperature) to (top) a typical deep sea sediment oxygen isotope record (a signal dominated by global ice volume). The Mapping Spectral Variability in Global Climate Project (SPECMAP) stack [*Imbrie et al.*, 1984] is a composite record constructed from normalizing and averaging a number of low-latitude planktic foraminiferal isotope records and is representative of the resolution typically achieved in climate time series based on open ocean marine sediments with low to moderate sedimentation rates (i.e., less than about 5–8 cm ka^{-1}). The high-resolution NGRIP record [*North Greenland Ice Core Project Members*, 2004] spans the last 123,000 years and reveals pervasive century- to millennial-scale variability not resolvable in typical marine records. The abrupt cooling of the Younger Dryas and the rapid warmings that characterize the well-known Dansgaard-Oeschger interstadial events are numbered. Heinrich events (labeled H1 to H6) are discrete layers of ice-rafted debris found in marine sediment cores from a wide swath of the subpolar North Atlantic. They have been interpreted to reflect the episodic discharge of massive numbers of icebergs from the Hudson Strait region.

observational and a mechanistic approach. From the observations, we can ask whether the events in different records have the same timing, abruptness, and persistence. Many of the records from the North Atlantic region have been examined in this context [e.g., *Alley et al.*, 1999; *Hemming*, 2004], but this is less true for other regions, particularly the tropics. For the mechanistic approach, the question is whether there are plausible physical mechanisms that can link different regions, perhaps at a global scale, and what drives such mechanisms to produce abrupt changes.

[4] At present, the favored paradigm for explaining the abrupt climate events seen in ice cores and elsewhere centers on freshwater input to the high-latitude North Atlantic and its effect on heat transport into the region via disruption of the meridional overturning circulation (MOC) of the Atlantic [e.g., *Rooth*, 1982; *Broecker et al.*, 1985; *Bond et al.*, 1993; *Rahmstorf*, 1995; *Alley et al.*, 1999; *Ganopolski and Rahmstorf*, 2001; *Knutti et al.*, 2004]. It is argued that freshwater discharges into the North Atlantic Ocean from the surrounding landmasses abruptly weaken the ocean thermohaline circulation (THC) in the Atlantic and the delivery of heat from the tropics to the high latitudes causing cold events in Greenland. The most obvious source of this fresh water would be from periodic releases of meltwater derived from the surrounding ice sheets.

[5] A collection of paleochemical data, reviewed in section 2.1.2, support the premise that the THC has indeed undergone past changes in its intensity. However, it is not necessarily clear whether these changes were the cause or consequence of abrupt climate change. Furthermore, it is not obvious that THC changes alone can explain the climate changes recorded both within the North Atlantic and worldwide [*Wunsch*, 2006]. Are there other possible mechanisms that can explain the features of the observations? One such

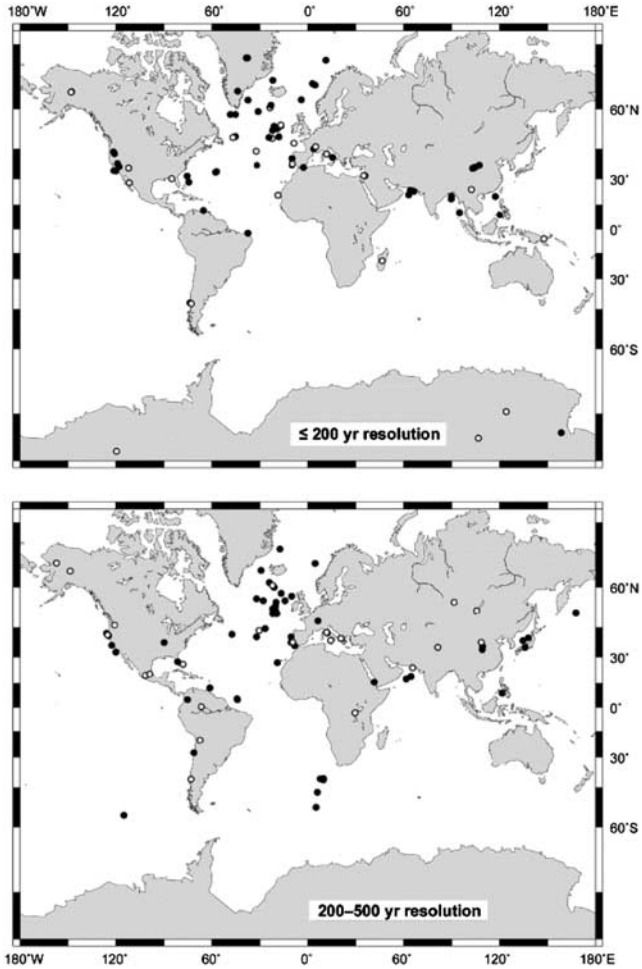

Figure 2. Spatial distribution of (top) sites with resolution of 1–200 years during MIS 3 and (bottom) sites with resolution of 200–500 years. Solid circles indicate sites with evidence of clear D-O-type climate oscillations, while open circles indicate locations where D-O cyclicity is unclear or absent. Reprinted from *Voelker and Workshop Participants* [2002], copyright 2002, with permission from Elsevier.

mechanism is a feedback involving sea ice in the North Atlantic. A number of papers have argued that the capacity of sea ice to affect climate both through albedo and air-sea heat exchange and also the ability of sea ice to rapidly change its distribution make this a good candidate mechanism for driving abrupt climate changes in the North Atlantic and perhaps worldwide [*Gildor and Tziperman*, 2003; *Timmermann et al.*, 2003; *Kaspi et al.*, 2004; *Li et al.*, 2005; *Wunsch*, 2006; *Denton et al.*, 2005]. Another possibility that has been proposed is that processes in the tropics play an important role in abrupt climate change [e.g., *Cane*, 1998; *Cane and Clement*, 1999; *Clement et al.*, 2001; *Pierrehumbert*, 2000; *Seager and Battisti*, 2007]. On the basis of the simple reasoning that the tropics are the largest source of interannual variability in the modern climate system and are also the main source of heat and water vapor for the global climate, the tropics would appear to be critical.

[6] In this paper, we first review and evaluate evidence for abrupt climate change during the glacial period from the paleoclimate record. We then review three different sets of physical mechanisms that have been proposed, thermohaline circulation, sea ice feedbacks, and tropical processes, to determine to what extent each can explain the main temporal and spatial characteristics of the observations. A final section suggests a somewhat different approach to understanding abrupt climate change by looking at feedbacks that could link these mechanisms on a global scale.

2. ABRUPT CLIMATE CHANGE OF THE LAST GLACIAL PERIOD: EVIDENCE

[7] The most clear and well-established evidence of abrupt climate change comes from Greenland ice cores (Figure 1). That the air temperature and rate of ice accumulation over the summit of the Greenland Ice Sheet can change abruptly is perhaps surprising in itself, but the more important issue is whether these climate events appear outside Greenland with the same dramatic and abrupt character.

[8] This issue was taken on by *Voelker and Workshop Participants* [2002], who made maps of the distribution of existing paleoclimate records with sufficient resolution in the time period of marine isotope stage (MIS) 3 to detect D-O events. Two maps were reconstructed (reproduced here as Figure 2), one showing the distribution of existing records with a resolution of <200 years, judged sufficient to identify D-O-scale events, and the other of records with a resolution of 200–500 years. The criteria used were that a record must have seven or more data points during the ~1500 year period of a D-O cycle to be included in the <200 year map and from five to seven data points per cycle to be included in the 200–500 year map. The quality of the age model for each record was also evaluated depending on the nature of the age control (e.g., ^{14}C, U/Th, and thermoluminescence) and the number of dated levels within the time period of MIS 3. One of the most striking aspects of the *Voelker and Workshop Participants* [2002] maps, but especially of the <200 year map, is that while the analysis indicates that D-O correlative events can be seen in regions all over the globe, including the tropics and Southern Hemisphere, there is an extremely disproportionate coverage of the North Atlantic, with enormous spatial gaps throughout the rest of the globe.

[9] Before going into our review of the records, we wish to point out a few issues for the reader to be aware of. First, we have not tried to apply a strict or concise definition of abrupt climate change in this paper. Examples of abrupt change from in and around Greenland (Figure 1) include the Younger Dryas cold event and the D-O events of the last glacial, as well as the Heinrich events recorded in North Atlantic sediments. These events all exhibit transitions that occur on decadal to centennial time scales and have recurrence time scales of roughly millennia. There is little reason to expect that they will look identical in different records because it is not clear to what extent these different types of events are actually analogous to each other and also because

different proxies are likely to record even the same event differently [e.g., *Hemming*, 2004]. Another issue that continually plagues the evaluation of abrupt change comes from limitations of the ability to date records precisely. In ice cores, for example, annual layer counting is used to provide absolute dating in calendar years down to a depth where compaction or basal deformation blurs or eliminates the visual layering [e.g., *Alley et al.*, 1997]. Below that, dating is usually based on a combination of techniques, including ice flow models, the identification of previously dated time markers (e.g., volcanic deposits), and synchronization to variations in the Earth's orbital parameters. In contrast, marine sediments typically are dated using radiocarbon methods back to approximately 45,000 years ago and then calibration of the measured ^{14}C ages to the calendar scale is relied upon [e.g., *Kitagawa and van der Plicht*, 1998; *Hughen et al.*, 2004a]. Foraminiferal oxygen isotope stratigraphies correlated to an orbitally tuned standard [e.g., *Imbrie et al.*, 1984] typically provide age control in older sequences. Speleothems and lake sediments each have their own dating methods and issues as well. Furthermore, differences in the temporal resolution between the different types of archives can dramatically affect the ability to correlate the respective climate proxies. The temporal resolution allowed by ice cores is often an order of magnitude or more better than what can be achieved in marine sediments, while archives like speleothems may provide high temporal resolution but can experience discontinuous growth. *Wunsch* [2006] has raised the above issues, suggesting that they present serious limitations to the ability to determine whether millennial variability is correlated in records around the globe.

[10] Thus, the definition of abrupt climate change will remain somewhat vague in this paper, and we will focus on the last glacial stage and temporal features in records that include decadal to centennial-time scale transitions with millennial-time scale recurrence. Where possible, the detailed timing, abruptness, and persistence of specific events at specific locations will be evaluated, and the records will be interpreted to determine to what extent the climate change or the proxy response is abrupt, but issues of proxy calibration and age control will still be a severe limitation.

[11] A final point we wish to make is that since our focus is on mechanisms, the goal of this section is not so much a comprehensive review of all the available climate records that show abrupt change but rather an identification of the records that place key constraints on the possible mechanisms. A more comprehensive review is given by the *National Research Council Committee on Abrupt Climate Change* [2002].

2.1. High Northern Latitudes

2.1.1. Records of Surface Change

[12] In many respects, the ice core records from Greenland have become the de facto "type-section" for late Pleistocene climate change, in that they serve as the measuring stick against which other climate records are inevitably compared. This is perhaps not surprising given their continuity, their intensive study at high resolution, and the simple fact that they were the among the first archives to yield clear evidence of climate changes that were startlingly rapid, including the aforementioned D-O events as well as the Younger Dryas (Figure 1). Furthermore, the long Greenland Ice Core Project (GRIP) [*Dansgaard et al.*, 1993] and Greenland Ice Sheet Project 2 (GISP2) [*Grootes et al.*, 1993] ice core records are virtually identical back to about 90,000 years B.P., giving confidence in their signals. The new North Greenland Ice Core Project ice core has extended the Greenland air temperature record back to about 123,000 years B.P. [*North Greenland Ice Core Project Members*, 2004] with the $\delta^{18}O_{ice}$ data over its length presented as 50 year mean values. In younger intervals where the annual layering is readily visible, large temperature changes, such as the warming experienced at the end of the Younger Dryas, have been shown to occur within decades or less [*Alley et al.*, 1993; *Taylor et al.*, 1993]. The rapid warmings in Greenland are also accompanied by equally abrupt increases (50–100%) in the rate of ice accumulation [*Alley et al.*, 1993; *Cuffey and Clow*, 1997]. The D-O events in Greenland recur on an approximately 1500 year time scale, though this is an average value [*Schulz*, 2002; *Rahmstorf*, 2003], and the recurrence interval can vary greatly from event to event (Figure 1).

[13] The first marine evidence for abrupt changes apparently synchronous with those in Greenland came from relatively high–accumulation rate sediment cores in the subpolar North Atlantic. Here, records of SST, based on abundance changes in a temperature-sensitive foraminifer, showed that the Younger Dryas and D-O climate events in Greenland were matched by corresponding SST changes of at least 5°C [*Bond et al.*, 1992, 1993]. Similar estimates have been derived by *Elliot et al.* [2002], who identified SST variations across D-O events of 3°C–5°C on Rockall Plateau. At the high-latitude locations of these investigations, SST variations can most easily be explained by north-south shifts in the position of the Polar Front. The initial studies by *Bond et al.* [1992, 1993] were also able to establish the temporal relationship between the temperature oscillations in Greenland and Heinrich events, discrete layers of ice-rafted debris (IRD) found in North Atlantic sediments [*Heinrich*, 1988] inferred by *Broecker et al.* [1992] and *Hemming* [2004] to reflect massive discharges of icebergs from the Hudson Strait region. Heinrich events, the timing of which is indicated in Figure 1, occurred during the coldest intervals of the last glacial at roughly 6000 year intervals. An extensive review of the origin, distribution, and timing of Heinrich event layers is given by *Hemming* [2004].

[14] In the subtropical North Atlantic, *Sachs and Lehman* [1999] presented a proxy SST record from the Bermuda Rise (33°N) for the time period between 60,000 and 30,000 years B.P. that clearly showed that abrupt temperature changes associated with the D-O events in Greenland and in the surrounding subpolar waters were experienced at this more southerly location (see Figure 3). Their record, based

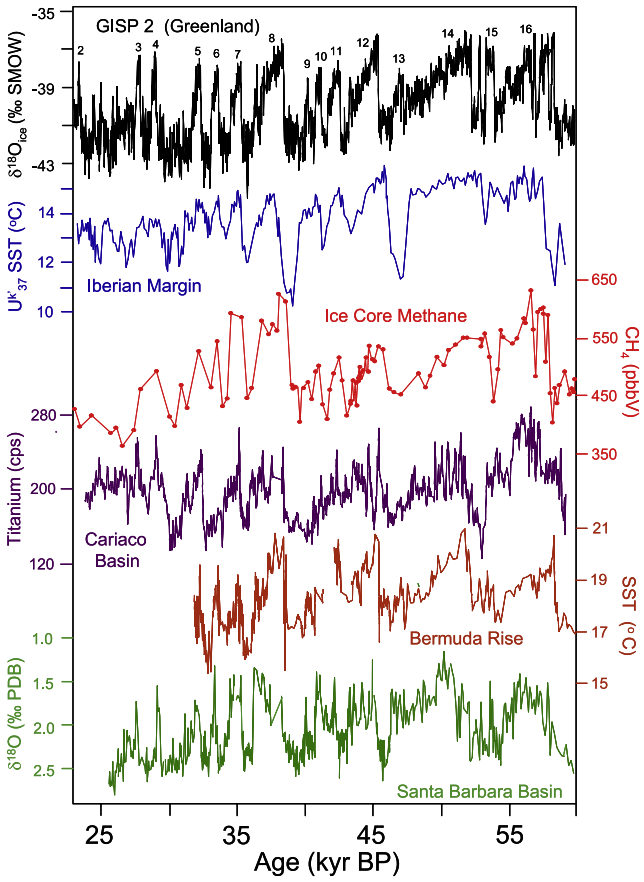

Figure 3. Paleoclimate time series showing examples of abrupt climate change during the last glacial period between 60,000 and 23,000 years ago. (top to bottom) Greenland Ice Sheet Project 2 (GISP2) δ^{18}O record [*Stuiver and Grootes*, 2000] showing D-O temperature variations over Greenland (warm interstadial events are numbered); alkenone ($U^{k\prime}_{37}$) sea surface temperature record from Iberian margin sediment core MD01-2443 [*Martrat et al.*, 2007]; GISP2 ice core record of atmospheric methane levels [*Brook et al.*, 1996], interpreted as a dominantly tropical wetland signal; titanium (Ti) content of Cariaco Basin (southern Caribbean) sediments, an index of regional hydrologic variability and riverine runoff [*Peterson et al.*, 2000]; reconstructed sea surface temperature (SST) from Bermuda Rise sediments [*Sachs and Lehman*, 1999]; and measured δ^{18}O of the thermocline-dwelling planktic foraminifer *N. pachyderma* from sediments of Santa Barbara Basin, California [*Hendy and Kennett*, 2000]. All time series are plotted on their original published age models.

on alkenone ($U^{k\prime}_{37}$) SST estimates, shows excursions of as much as 4°C–5°C across stadial-interstadial transitions. *Keigwin and Boyle* [1999] independently identified planktic foraminiferal δ^{18}O excursions in Bermuda Rise cores that could be matched to every D-O event within MIS 3; interpreting the δ^{18}O variations as solely the result of SST yielded an average of 4°C change across D-O events, a number in excellent agreement with the alkenone estimates. At approximately the same latitude to the east (37°N), *Shackleton et al.* [2000] similarly showed that every D-O event between 64,000 and 24,000 years ago could be identified in the planktic foraminiferal δ^{18}O record of a core, MD95-2042, recovered off the southern margin of Portugal. In a separate but nearby core, *Bard et al.* [2000] found evidence for sea surface cooling and lower salinities during the last three Heinrich (H) events (H1–H3) using a combination of alkenone proxies in a core from the Iberian margin. Depending on the alkenone calibration used, the amplitudes of cooling during the Heinrich events were estimated at 2°C–5°C. Unlike the longer Bermuda Rise and MD95-2042 cores, this record spans only the past 31,000 years, and so it misses nearly all of MIS 3 where the D-O events are most prominent. However, *Martrat et al.* [2007] recently presented long alkenone SST time series from the same region that clearly show temperature excursions associated with D-O events of the last glacial (Figure 3), as well as similar D-O-like variability over the past four glacial-interglacial climate cycles. In the western Mediterranean, in an impressive Alboran Sea record, *Martrat et al.* [2004] documented rapid SST changes of up to 6°C from alkenones that correlate to all of the D-O events of the last glacial. Similar millennial-scale variability, though of lower amplitude, also characterizes glacial MIS 6 sediments. Taken together, these and other regional records indicate that abrupt SST changes occurred synchronously over much of the midlatitude to high-latitude North Atlantic during the last glacial and that they were synchronous and correlative with D-O events within the limits of time resolution and available age control.

[15] One high-latitude quantity with clear climatic significance is the presence and extent of sea ice cover, but there is currently scant evidence for how sea ice has changed in the past. Sea ice is an important component of the climate system because of its high albedo and because it interferes with the exchange of gases and energy across the air-sea interface. In addition, the processes of freezing and thawing that accompany seasonal sea ice development play an important role in the surface salinity and thus on the density and stratification of water masses, vertical convection, and deepwater ventilation [*de Vernal and Hillaire-Marcel*, 2000]. Attempts to reconstruct sea ice from marine sediments have usually relied on micropaleontologic proxies since the presence of sea ice affects the distribution of plankton living in the photic zone. Studies based on dinoflagellate cysts [*de Vernal and Hillaire-Marcel*, 2000] and diatoms [*Koç et al.*, 1993] have generally indicated much greater sea ice extent in the high-latitude North Atlantic during the Last Glacial Maximum, but little if anything is known about sea ice fluctuations on millennial time scales. Given recent ideas about sea ice's potential role in the generation of abrupt climate change, discussed in more detail in section 3.2, the further development of sea ice proxies and their application in spatial and down-core reconstructions seems clearly warranted.

2.1.2. Records of Deep Ocean Change

[16] *Rooth* [1982] was the first to suggest that introduction of meltwater during deglaciation into the high-latitude North Atlantic could have triggered the Younger Dryas cooling by reducing the surface salinity and disrupting the

thermohaline circulation. *Broecker et al.* [1985] further developed this idea into the THC model that prevails today, with the rapid warmings (interstadials) and coolings (stadials) of the North Atlantic region interpreted to reflect a vigorous and weak THC, respectively. Considerable evidence exists to support the notion that the THC has varied with time. This evidence comes primarily in the form of sediment proxies that reflect either the paleonutrient content of the deepwater column or in some way record the strength or intensity of the deep flow. During the Younger Dryas cooling, for example, δ^{13}C and Cd/Ca data from benthic foraminifera suggest that an older and more nutrient-rich water mass of probable southern origins replaced North Atlantic Deep Water (NADW) [*Boyle and Keigwin*, 1987; *Keigwin et al.*, 1991; *Marchitto et al.*, 1998]. This inference is supported by measurements of grain size in the sortable silt fraction of several North Atlantic cores [e.g., *McCave et al.*, 1995; *Manighetti and McCave*, 1995], with reduced grain size interpreted to indicate a weakening of deep currents associated with lower North Atlantic Deep Water.

[17] On the basis of a collection of paleoceanographic proxies from the North Atlantic, *Clark et al.* [2002] argued that the THC has abruptly shifted between different "modes" of operation, which include a modern mode (strong overturning), a glacial mode (weak and shallower overturning), and a Heinrich mode (an essentially complete shutdown). In each of these modes, the sources of the deepwater masses in the North Atlantic vary geographically. Recent measurements of sedimentary ^{231}Pa/^{230}Th, interpreted as a kinematic proxy for the strength of the THC, would seem to support the interpretation of multiple modes. Using ^{231}Pa/^{230}Th in sediments from a core near Bermuda Rise, *McManus et al.* [2004] have argued that the meridional overturning of the Atlantic was nearly or completely eliminated during the last of the massive iceberg discharge events of the glacial, Heinrich 1 (H1) (~17,500 years ago), and declined sharply during the Younger Dryas event. In a separate study using ^{231}Pa/^{230}Th in a core from the Iberian margin, *Gherardi et al.* [2005] found that similar reductions in the MOC could be inferred at this location over the last 20,000 years, the same time interval spanned in the *McManus et al.* [2004] study. Since ^{231}Pa/^{230}Th can potentially be biased by variations in particle flux and composition, the observation of similar ^{231}Pa/^{230}Th patterns in different sedimentary environments is important because it tends to confirm that the ratio changes indeed record basinwide changes in deep circulation. However, timing differences between the two locations (i.e., western and eastern North Atlantic basins) indicate that the story of ^{231}Pa/^{230}Th may be more complicated than first thought. *Broecker and Barker* [2007], for example, note that the apparent reduction in deepwater production during the H1 event in the core studied by *Gherardi et al.* [2005] occurred more than a thousand years later than that found by *McManus et al.* [2004]. This suggests differences between renewal rates in the deep eastern and western basins, and perhaps location-dependent differences between shallow and deep overturning must be factored into understanding the history of the MOC.

[18] The paradigm of fresh water (meltwater) stratifying the surface ocean in the high-latitude North Atlantic and serving as the driver for THC changes and abrupt climate events is largely based on a reconstructed sequence of events linked in time to the Younger Dryas cooling, specifically the sudden eastward diversion of fresh water from glacial Lake Agassiz directly into the North Atlantic via the Great Lakes–St. Lawrence River system at the beginning of the Younger Dryas [e.g., *Clark et al.*, 2001; *Teller et al.*, 2002]. However, open issues remain, and a number of studies have raised questions about the existence and reported timing of freshwater diversion at the time of the Younger Dryas [e.g., *Lowell et al.*, 2005; *Teller et al.*, 2005; *Moore*, 2005; *Broecker*, 2006a]. Many of these concerns seem to have been addressed in a recent study that applied new proxies for freshwater sources (ΔMg/Ca, U/Ca, and ^{87}Sr/^{86}Sr) to planktonic foraminifera recovered from sediments at the mouth of the St. Lawrence estuary [*Carlson et al.*, 2007]. These new measurements seem to confirm that routing of fresh water from western Canada to the St. Lawrence River did indeed occur at the start of the Younger Dryas, and they allow for calculation of an initial base increase in freshwater flux of 0.06 ± 0.02 sverdrups (Sv) (where 1 Sv = 10^6 m^3 s^{-1}) and a maximum discharge of 0.12 ± 0.02 Sv. This estimate of initial discharge flux is in good agreement with an earlier flux estimate of 0.07 Sv made by *Licciardi et al.* [1999] for the Younger Dryas but is considerably smaller than estimates made by *Hemming* [2004] for freshwater inputs to the North Atlantic during Heinrich events (0.15–0.3 Sv over 500 years and 1.0 Sv if over 1 year, see Table 1).

[19] Though it is now reasonably well established that variations in the THC accompanied the Younger Dryas and at least the last of the Heinrich events, it is less clear whether the abrupt D-O events recorded in Greenland ice and elsewhere in surface records of the North Atlantic were all accompanied by similar THC changes. Longer sediment time series of some of the same paleonutrient proxies (e.g., benthic foraminiferal δ^{13}C) suggest that deepwater changes characterized earlier Heinrich events and at least some of the D-O events [e.g., *Oppo and Lehman*, 1995; *Charles et al.*, 1996; *Keigwin and Boyle*, 1999; *Curry et al.*, 1999; *Hagen and Keigwin*, 2002]. However, it is difficult to point to any single record that provides compelling and unambiguous evidence of a direct, one-to-one link between THC variations and all of the abrupt events of the last glacial. To some extent, this likely reflects the fact that the three-dimensional nature of the ocean's deep circulation makes it inherently more complex to reconstruct than the past properties of the surface ocean, which is essentially a two-dimensional problem. Evidence of past changes in the shallow meridional overturning may possibly be missed in deeper records and vice versa. *Elliot et al.* [2002], for example, in their study of two sediment cores from off southeastern Greenland and Rockall Plateau used benthic δ^{13}C variations to infer that Heinrich events were related to

TABLE 1. Different Values of Freshwater Input to the North Atlantic for Heinrich Events Based on Observations and Values Used in the Modeling Experiments Cited in This Review[a]

	Total Freshwater Volume (km^3)	Average Rate of Input (Sv)
Observation-based estimates		
Hemming [2004]	$\sim 3 \times 10^4$ (assuming 1 year duration)	1.0 (over 1 year)
	$\sim 2.5-5 \times 10^6$ (assuming 500 year duration)	0.15–0.3 (over 500 years)
MacAyeal [1993] and *Alley and MacAyeal* [1994]		~0.1
Licciardi et al. [1999]		~0.07
Carlson et al. [2007]		0.06–0.12
Model experiments		
Stouffer et al. [2006]	3×10^5	0.1 (over 100 years)
Zhang and Delworth [2005]	1×10^6	0.6 (over 60 years)
Vellinga and Wood [2002]	6×10^5	19 (over 1 year)

[a]Note that the estimate of *Licciardi et al.* [1999] is an average value for the Younger Dryas.

large-scale reductions in deepwater production but found that the cold stadial events in between were not associated with recognizable decreases in the MOC at least at the water depth of the two core sites (~2000 m). In contrast, *Keigwin and Boyle* [1999] reported δ^{13}C records from water depths close to 4600 m at Bermuda Rise that look more D-O-like in nature for at least a portion of MIS 3 but show no clear indication of an amplified Heinrich-type signal. *Vautravers et al.* [2004] present a benthic δ^{13}C record from a Blake Outer Ridge site at 3480 m water depth; at this location, changes in the deepwater δ^{13}C proxy appear to be associated with some, but not all, D-O events. Although not in itself a record of MOC changes, *Hill et al.* [2006] presented evidence from a Gulf of Mexico core that earlier glacial episodes of Laurentide Ice Sheet meltwater outflow from the Mississippi River fail to match the timing expected if meltwater diversion, in a situation analogous to the Younger Dryas excursion, is to explain the abrupt events during MIS 3.

[20] One of the more intriguing studies that addresses the relationship between the MOC and abrupt climate events is that of *Kissel et al.* [1999], who compiled bulk magnetic parameter data from a suite of seven North Atlantic cores ranging in latitude from 67°N to 33°N and covering different water depths. Short-term variations in the magnetic properties could be well correlated between cores and were interpreted by Kissel et al. to reflect changes in the transport efficiency of the magnetic particles by deep currents. Though the age control is at least partly based on wiggle matching, there is a fairly remarkable match between magnetic parameters and the Greenland temperature record, interpreted to indicate a weakening of current strength and particle transport during cold stadial and Heinrich events.

[21] In some cases, sampling and resolution issues complicate the interpretation of records. For example, *Piotrowski et al.* [2005] used neodymium isotopes, which can be applied as a conservative tracer of deepwater end-members, to reconstruct circulation changes from cores in the Cape Basin of the South Atlantic (~40°S). Their record nicely shows that changes in the relative proportions of northern-sourced North Atlantic Deep Water and southern-sourced Antarctic Bottom Water occurred at this location during some of the longer interstadial events of MIS 3 (e.g., interstadials 8, 12, and 14). However, the temporal resolution of the record is insufficient to show whether this linkage persists for the shorter D-O events as well. In other cases, the resolution is sufficient, but the time interval studied does not encompass the entire glacial. *Skinner and Elderfield* [2007], for example, recently presented the first benthic foraminiferal Mg/Ca record with the resolution necessary to see D-O variations in deepwater temperature. Their record, from 2637 m water depth on the Iberian margin, indicates a positive correlation between deepwater temperature change and Greenland temperatures but only spans the relatively short interval of Greenland interstadials 8 to 12 (~36,000 through 50,000 years ago) at present. Although abundant data exist to support the hypothesis that the Atlantic MOC was weaker during the Younger Dryas and perhaps weaker still during earlier Heinrich events, it can be argued that unequivocal evidence for a direct relationship between D-O events and the MOC has yet to emerge.

2.2. High Southern Latitudes

[22] Like their Greenland counterparts, Antarctic ice cores contain probably the best and most complete records of climate in the high-latitude Southern Hemisphere. However, slower ice accumulation rates tend to result in less highly resolved records that make it more difficult to assess the nature of millennial-scale climate events and especially their timing relative to high-latitude North Atlantic changes. Antarctic temperature variations (inferred from δ^{18}O and δD records) show a pattern of more gradual and less pronounced warming and cooling events over the same period that Greenland ice records the abrupt D-O events of the last glacial. *Blunier et al.* [1998], nevertheless, identified an out-of-phase behavior between temperature in the Byrd and Vostok (Antarctic) and Greenland ice cores over several of the longer-lasting interstadials (D-O events 8 and 12) and argued that this interhemispheric coupling was the result of THC variations. Underlying this interpretation is the premise that a vigorous THC draws heat from the Southern Ocean and that a weakening or collapse of the meridional overturning results in cooling of the North Atlantic and a concomitant warming in the Antarctic region, a mechanism that has come to be known as the "bipolar seesaw"

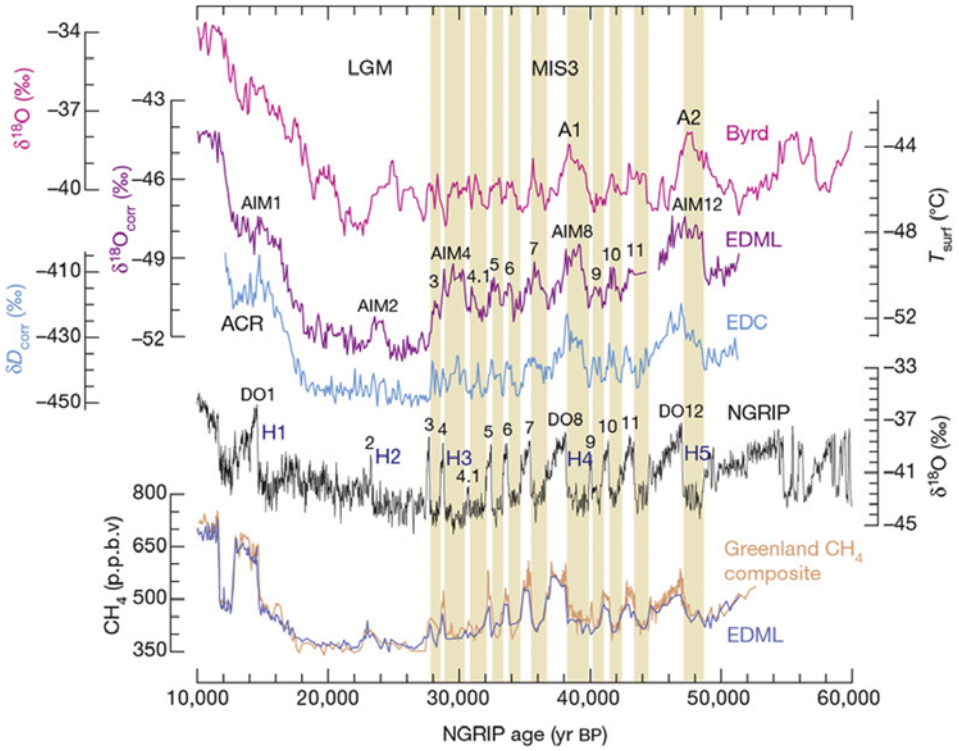

Figure 4. Comparison of Antarctic air temperature changes for the Byrd ice core, EDML ice core (from in the interior of Dronning Maud Land), and EDC ice core (from Dome C) (as inferred from δ^{18}O and δD records) to the northern Greenland δ^{18}O record from the NGRIP ice core for the time interval 10,000–60,000 years ago (reprinted by permission from Macmillan Publishers Ltd [*EPICA Community Members*, 2006], copyright 2006). All records are synchronized on the basis of their measured air bubble CH_4 (methane) contents. This comparison reveals a one-to-one match between Antarctic warming events and a corresponding cold stadial period in Greenland that has been attributed to the bipolar seesaw pattern in temperature predicted from models of variations in the ocean's meridional overturning circulation. Antarctic isotope maxima (AIM) are noted on the figure, as is the Antarctic cold reversal (ACR).

[*Crowley*, 1992; *Stocker et al.*, 1992; *Stocker*, 1998; *Broecker*, 1998]. On the basis of further analysis of the Byrd ice core, *Blunier and Brook* [2001] subsequently argued that at least seven Antarctic warming events over the glacial were matched by Greenland cooling, providing further support for the so-called bipolar temperature seesaw. Most recently, study of the new European Programme for Ice Coring in Antarctica (EPICA) Dronning Maud Land (EDML) ice core has yielded an Antarctic climate record with a resolution comparable to Greenland [*EPICA Community Members*, 2006]. With the enhanced resolution of the new oxygen isotope record from EDML, it is now clearly more evident that a one-to-one coupling exists between all Antarctic warm events and their cold stadial counterparts in the glacial D-O cycles of Greenland (Figure 4). Furthermore, though not abrupt and considerably dampened relative to Greenland, the amplitude of the Antarctic warm events was shown by the EPICA Community Members to be linearly related to the length of the concurrent cold stadial event in Greenland, suggesting that they all result from a similar reduction in the Atlantic meridional overturning. In the absence of definitive sediment records for THC changes during D-O events, the EDML observations may well provide the most compelling argument to date for a direct role of the deep ocean circulation in abrupt climate change.

[23] Ice core records elsewhere in the Antarctic do show some differences with respect to the EDML site, suggesting that regional variability exists in a continent of this size. The Taylor Dome ice core, recovered at a more coastal location in the western Ross Sea sector, shows a pattern of deglacial temperature change that was initially interpreted [*Steig et al.*, 1998] to be synchronous with that in Greenland (i.e., Bølling/Allerød warming and Younger Dryas cooling). However, more recent correlations to other ice cores have resulted in an altered age model [e.g., *Mulvaney et al.*, 2000] that brings its pattern of deglacial temperature change more in line with that at sites in the Antarctic interior. Nevertheless, the Taylor Dome record does show much greater variability than other Antarctic records, perhaps a consequence of its near-coastal location. *Morgan et al.* [2002] and *Jouzel et al.* [2001] have also raised questions about the bipolar seesaw concept based on deglacial temperature changes across the last deglaciation in the Law Dome and EPICA Dome C ice cores, respectively. At both sites, the pattern of cooling does not match the timing of the abrupt warming that occurred at the start of the northern Bølling transition about 14,500 years ago, suggesting that

considerable spatial heterogeneity in the record of temperature exists. Finally, while the amplitude of climate change in the Antarctic is generally dampened at best relative to events in Greenland, rapid shifts in climate are not undocumented for the region. At the site of the Siple Dome ice core, near the edge of the Ross Ice Shelf, *Taylor et al.* [2004] documented a dramatic rise in air temperature of ~6°C about 22,000 years ago. This sharp temperature increase apparently occurred within a few decades and has no obvious counterpart in Greenland nor within other Antarctic sites.

[24] Land-based records of climate change from the southern high latitudes are few and far between and suffer from age controversies and a resolution generally too low to preserve millennial-scale events, let alone abrupt changes. In addition, few available land-based records extend beyond the late glacial, and none that we can find are suitable for revealing evidence of D-O equivalent events. Even records that span the general time interval of the Younger Dryas event are equivocal. On the basis of radiocarbon dating of glacial deposits in the Southern Alps of New Zealand, *Denton and Hendy* [1994] and *Lowell et al.* [1995] argued for a glacial advance during the Younger Dryas in New Zealand. South Island pollen records suggest that increased moisture is more strongly indicated than cooling at this time [*Fitzsimons*, 1997; *Singer et al.*, 1998], while *Turney et al.* [2003] interpreted pollen records from both the North and South islands to indicate warming and an increase in westerly airflow during the interval correlative with the Younger Dryas chronozone. A more recent speleothem study of northern South Island sites by *Williams et al.* [2005] concluded that the Younger Dryas stadial was not well defined in New Zealand and that at the millennial scale a more convincing case can be made for an asymmetric climate response between the two hemispheres rather than a synchronous one. In southern South America, records are similarly contradictory. In the midlatitudes of southern Chile, glacial deposits have been used to argue for interhemispheric synchrony in temperature during deglaciation [e.g., *Lowell et al.*, 1995]. In contrast, pollen records from the same general area reveal no evidence for a cooling at the time of the Younger Dryas [*Bennett et al.*, 2000].

[25] Southern Hemisphere marine records are equally scarce. In a sediment core off New Zealand, *Pahnke and Zahn* [2005] found benthic foraminiferal δ^{13}C evidence for increases in intermediate water formation linked to Heinrich events and periods of inferred minimum NADW production in the North Atlantic. In the same core from Chatham Rise, *Sachs and Anderson* [2005] identified coincident episodes of increased surface productivity from concentration changes in sedimentary biomarkers diagnostic of coccolithophorids and diatoms. The sampling interval in both studies averaged about 200 years, which combined with bioturbation, makes it difficult to judge abruptness. There is no clear evidence of D-O-like excursions in either record. Off the Chilean continental margin, alkenone-based estimates of SST made by *Lamy et al.* [2004] show evidence of local millennial-scale SST changes of 2°C–3°C between 8000 and 50,000 years ago and a timing similar to Antarctic records, with warming of surface waters off Chile matching the Antarctic warmings that coincide with Heinrich events. Again, a sampling interval averaging 300 years makes it impractical to judge abruptness. More recently, *Kaiser et al.* [2005] extended the alkenone SST record of *Lamy et al.* [2004] back to 70,000 years and presented an improved age model for this site, which together reinforce the conclusion that millennial-scale SST variability at this midlatitude (41°S) location exhibits a distinctly Antarctic behavior.

[26] In summary, the relatively small number of existing high-latitude climate records from the Southern Hemisphere reveal potentially complicated regional patterns of climate change, and few, in truth, possess the length, temporal resolution, and age control needed to provide hard constraints on changes occurring around the time of the abrupt events in Greenland. The most obvious exception is the remarkable new EDML ice core record from the Antarctic (Figure 4), which appears to show a dampened but out-of-phase temperature pattern for all D-O events consistent with the bipolar seesaw hypothesis. Regardless of what one may conclude from the apparent climate patterns cited here, it is clear that the Southern Hemisphere remains woefully undersampled.

2.3. Tropics

[27] One of the first sites far removed from the North Atlantic to show a complete record of abrupt D-O equivalent events during the last glacial is the extraordinary Site 893, drilled by the Ocean Drilling Program in Santa Barbara Basin off southern California (Figure 3). Here, stable isotope and microfossil analyses reveal a remarkable correlation between sea surface conditions and basin ventilation in Santa Barbara Basin and the climate record from Greenland, with a clear record of the Younger Dryas and every D-O event preserved in multiple proxies [e.g., *Behl and Kennett*, 1996; *Hendy and Kennett*, 1999, 2000; *Hendy et al.*, 2002]. Though itself not a tropical location, the high-resolution climate records yielded from Santa Barbara Basin provided a major impetus in the mid-1990s to extend the search for evidence of abrupt climate change outside of the immediate circum–North Atlantic region. A number of important records have subsequently emerged from the tropics and near tropics that clearly indicate that the low latitudes were intimately involved in abrupt climate change.

[28] In central Florida, a pollen study of Lake Tulane sediments by *Grimm et al.* [1993] revealed a series of six peaks of *Pinus* (pine) pollen that within the limits of radiocarbon dating appeared to be coeval with the Heinrich events in the North Atlantic. These authors interpreted the *Pinus* peaks as representing cool-wet periods during the glacial. Recently, a more in-depth study of pollen and plant macrofossils from new cores recovered in Lake Tulane [*Grimm et al.*, 2006] confirmed that lake levels were higher and climate was wetter during the *Pinus*-rich intervals but confirmed also that Florida temperatures were actually warmer, not cooler, during these phases. *Grimm et al.* [2006] speculate that a reduction of the THC before and

during Heinrich events reduced northward heat transport and resulted in retention of warmth in the subtropical Atlantic and Gulf of Mexico.

[29] In the tropical Atlantic, the Cariaco Basin, an anoxic basin located off the northern coast of Venezuela, provided some of the first indications that the Younger Dryas and the complete set of D-O oscillations in Greenland could be traced event for event to the tropics [e.g., *Hughen et al.*, 1996, 2000, 2004b; *Peterson et al.*, 2000; *Haug et al.*, 2001; *Lea et al.*, 2003; *Peterson and Haug*, 2006]. Here, high sedimentation rates (40–50 cm ka^{-1}) and minimal to no bioturbation enable sampling at close to annual resolution [e.g., *Black et al.*, 1999]. Correlations between Greenland $\delta^{18}O$ and sediment color in Cariaco Basin (which is largely a function of organic matter content) imply that surface productivity along the Venezuelan coast varied in lockstep both in timing and abruptness with Greenland temperature variations, presumably in response to variations in trade wind strength and upwelling. Mg/Ca data from surface-dwelling foraminifera in a piston core record spanning the last 25,000 years [*Lea et al.*, 2003] indicate an abrupt warming of ~3°C at the onset of the Bølling and an equally abrupt decrease in SST of 3°C–4°C during the Younger Dryas, both synchronous within ±30–90 years of their correlative changes in Greenland air temperature. The measured amplitude of the Younger Dryas SST signal in Cariaco is as large as the local glacial-interglacial signal. In the same sediments, measured variations in the bulk Ti (Figure 3) and Fe contents have been interpreted to reflect dry conditions and reduced river runoff from northern South America during the Younger Dryas and cold stadial periods of the last glacial and wet conditions and increased runoff during warm interstadial times [*Peterson et al.*, 2000; *Haug et al.*, 2001]. Results of molecular and isotopic biomarker studies spanning the last deglaciation confirm rapid changes in drainage basin vegetation in response to changing hydrological conditions [*Hughen et al.*, 2004b]. These abrupt changes in the local precipitation regime are best explained by latitudinal shifts in the position of the Intertropical Convergence Zone (ITCZ) in the Atlantic, with the ITCZ farther south on average at times when the North Atlantic is cold [e.g., *Peterson and Haug*, 2006].

[30] Numerous land-based records from the circum-Caribbean region are consistent with the evidence from Cariaco Basin for a dry Younger Dryas and support the interpretation that the ITCZ shifted southward at this time (see *Peterson and Haug* [2006] for a review). Few records from this region are as highly resolved or long enough, however, to confirm the association of pronounced wet/dry cycles with D-O events inferred from Cariaco Ti and Fe records. *Schmidt et al.* [2004] used coupled Mg/Ca and foraminiferal $\delta^{18}O$ measurements in a low-resolution Caribbean core to show that Caribbean salinities increased during the last glacial period as a whole. In a later study using a higher sedimentation rate sequence from the Blake Outer Ridge in the western subtropical Atlantic, *Schmidt et al.* [2006] used similar techniques to show that surface salinity varied across a series of four D-O cycles in MIS 3, with reduced salinity associated with warm interstadials and higher salinity occurring during the intervening cold stadials in a pattern consistent with the Cariaco record.

[31] Farther south over the South American continent, longer records are available, though they tend to show the opposite hydrologic trends, implying abrupt, latitudinal shifts in the past footprint of rainfall over tropical South America. In the Salar de Uyuni record from the Bolivian Altiplano, wet conditions (e.g., high lake levels) were found to be generally coincident with dry conditions near Cariaco Basin and cold stadial events at high northern latitudes [*Baker et al.*, 2001]. Farther to the east, periods of speleothem growth in Brazilian caves from south of the present-day rain forest (~10°S) reflect episodic wet phases in a semiarid region that can be accurately dated using uranium/thorium methods. In these caves, speleothem growth occurred at times of Heinrich layer deposition in the North Atlantic, when conditions in and around Cariaco Basin to the north were dry [*Wang et al.*, 2004]. Pulses of terrigenous sediment input to marine cores off northeastern Brazil [*Arz et al.*, 1998] also appear to line up in time with the (cold) high-latitude Heinrich events; both the offshore and speleothem records from northeast Brazil are thus consistent with a southward ITCZ shift. Farther south yet, in subtropical southern Brazil (~27°S), a continuous speleothem oxygen isotope time series from Botuverá Cave spanning the last 116,000 years [*Cruz et al.*, 2005] shows dominant variability in the frequency band of orbital precession. Although this record does show evidence of millennial-scale variations in rainfall, the amplitude of the signal is greatly dampened, and individual excursions cannot be related to D-O or Heinrich events in any obvious way. In a higher-growth-rate stalagmite from the same cave system, however, *Wang et al.* [2006] have produced a high-resolution record spanning the last 36,000 years. This record reveals a much more striking correlation with Greenland, with dry conditions associated with warm interstadial events (3 through 7) in the latter portion of MIS 3. Since this site would appear to be too far south to be directly explained by ITCZ shifts, *Wang et al.* [2006] suggest that the antiphased pattern of rainfall between here and northern South America was the result of asymmetries in the Hadley circulation that accompanied changes in ITCZ mean position.

[32] As one surveys the growing literature on abrupt climate change in the rest of the tropics outside the Atlantic, it is apparent that changes in the hydrologic cycle emerge as a dominant theme in all ocean basins. In the Pacific Basin, *Stott et al.* [2002] used foraminiferal Mg/Ca and $\delta^{18}O$ from sediments from the western Pacific warm pool to delineate between SST and salinity change. They identified a first-order salinity signal that appeared to vary in accord with the Greenland D-O cycles with a pattern interpreted to be analogous to that produced by the modern El Niño–Southern Oscillation. More saline El Niño–like conditions (reduced east-west equatorial temperature gradient in the Pacific) were found to correlate with cold stadials at high latitudes, suggesting a shifting of tropical convection away from the western Pacific, while warm interstadials showed decreased

(La Niña–like) salinity patterns. The magnitude of the reconstructed salinity change across stadial-interstadial events is on the order of 1–2‰. Though the time scales are different, El Niño–like conditions during cold episodes are consistent with the observations of *Koutavas et al.* [2002] for the eastern Pacific, who inferred relaxed trade winds and a southward shift of the Pacific ITCZ during the Last Glacial Maximum. Another study from the eastern Pacific is that of *Ortiz et al.* [2004], who describe evidence for productivity variations that mirror D-O events in a high-deposition-rate core from near the coast of southern Baja California. This record, which spans the last 52,000 years, shows evidence of low productivity during cool stadial intervals and high productivity during warm interstadials of MIS 3. Ortiz et al. note that one possible explanation for these variations is a shift between the modern balance of El Niño–La Niña conditions, with cold intervals characterized by more El Niño–like conditions with a deep nutricline and lower surface production.

[33] In the far western Pacific, *Rosenthal et al.* [2003] and *Dannenmann et al.* [2003] used coupled δ^{18}O and Mg/Ca data from Sulu Sea sediments to show that millennial-scale δ^{18}O events during the deglaciation (e.g., Younger Dryas) and last glacial (D-O events) were primarily the result of changes in surface water salinity. They attribute this variability to the east Asian monsoon and its influence on the balance between surface water contributions from the South China Sea and western Pacific warm pool. Within the dating uncertainties that one faces when comparing records on independent age scales, the Sulu Sea record indicates that times of fresher surface conditions in the Sulu Sea coincide with similar conditions in the western Pacific warm pool [*Stott et al.*, 2002] and also with intensifications of the summer east Asian monsoon as recorded in the Hulu Cave record from China [*Wang et al.*, 2001] and thus by inference with warm interstadials in the Greenland ice core records (Figure 5). *Wang et al.* [1999] noted millennial-scale fluctuations in South China Sea sediments that they similarly posited to variations in the east Asian monsoon.

[34] Moving west over land to regions directly affected by the Indian/Asian summer monsoon, dry conditions have been found to correlate with cold periods in the high northern latitudes (e.g., the Younger Dryas and Heinrich events and cold stadials of the last glacial). This relationship has been documented in a number of speleothem δ^{18}O records covering a variety of different time windows and well dated by U/Th methods. In a Holocene-age speleothem from southern Oman, *Fleitmann et al.* [2003] found that early Holocene decadal to centennial variations in monsoon precipitation were in phase with temperature variations recorded in Greenland ice, indicating that monsoon intensity during this interval was tightly linked to high-latitude conditions. A decreasing monsoon intensity after about 8000 years ago, in contrast, appears to have been more sensitive to changing large-scale summer insolation patterns. From the western Himalayas of India, *Sinha et al.* [2005] produced a speleothem δ^{18}O record from Timta Cave that spanned the last deglaciation. Their record shows a clear increase in summer Indian monsoon precipitation during the Bølling/Allerød warm period and a sharp decrease during the Younger Dryas event that immediately followed. This deglacial pattern of change is similarly recorded in speleothems from Hulu Cave [*Wang et al.*, 2001] and Dongge Cave [*Yuan et al.*, 2004; *Kelly et al.*, 2006], located some 1200 km apart in eastern subtropical (~32°N) and south central (~25°N) China, respectively. The Hulu Cave and Dongge Cave records span all and/or portions of the last glacial as well, while the Dongge record additionally covers the last full interglacial (equivalent to MIS 5) and the penultimate deglaciation (e.g., Termination II). During the last glacial, both cave sites show clear evidence of hydrologic changes correlative to the Greenland D-O events (Figure 5), with increased precipitation from a stronger east Asian monsoon falling during warm interstadial times. Another key record of glacial monsoon variability comes from cave deposits of Socotra Island, located off South Yemen near the entrance to the Gulf of Aden in the western Arabian Sea. Though covering only the time period from 42,000 to 55,000 years ago, the speleothem δ^{18}O record of *Burns et al.* [2003] is the most highly resolved monsoon time series spanning this interval (Figure 5) and bears a striking relationship to that of Greenland climate, again with increased precipitation coinciding with high-latitude warming. The high-resolution sampling of this record also allows for an assessment of abruptness; for example, the largest shift in speleothem δ^{18}O, which occurred at the onset of interstadial 12, appears to have occurred in 25 years or less. Burns et al. interpret their record as evidence for changes in ITCZ position over the Arabian Sea, with southward shifts during cold stadial intervals reducing the northward penetration of monsoon rains during the East African–Indian summer monsoon. At any one location like Socotra Island, the apparent abruptness of the inferred hydrologic changes could simply be due to the fact that small latitudinal shifts in convective activity and precipitation place that location either under or out of the influence of summer monsoon rainfall. However, the observation that similar hydrologic changes occur at locations as geographically widespread as South Yemen, India, and southern and eastern China tends to suggest that large-scale changes in the strength of the Indian and east Asian monsoon systems are involved.

[35] In marine records from the Arabian Sea, correlated millennial-scale changes in proxies for upwelling and surface productivity, and for ventilation within the mid-depth oxygen minimum zone, have been similarly attributed to variations in the strength of the southwest (summer) Indian monsoon. The prevailing pattern is one of evidence for a stronger monsoon and increased upwelling and surface productivity during warm interstadials (Figure 5), coupled with an intensification of the oxygen minimum zone as measured by elevated total organic carbon contents of sediments [e.g., *Schulz et al.*, 1998] and by records of denitrification derived from sediment δ^{15}N [*Altabet et al.*, 2002]. This interpretation of a stronger summer monsoon at times of high-latitude warmth is consistent with the speleothem-based reconstructions of monsoon precipitation, and

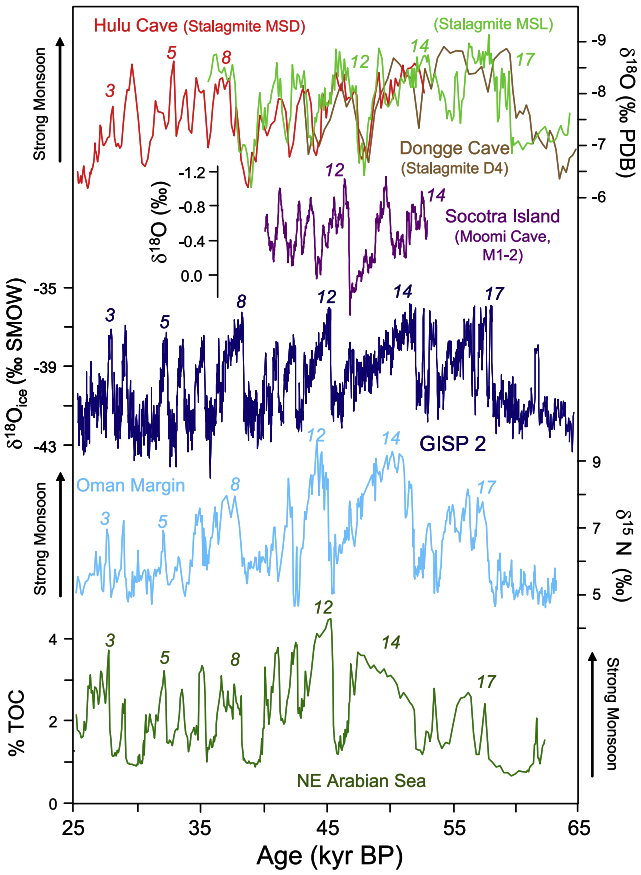

Figure 5. Paleoclimate time series showing examples of abrupt climate change in land and marine records influenced by the Indian/east Asian monsoon. Data are shown for the time period of MIS 3 spanning the interval from 65,000 to 25,000 years ago. (top to bottom) Speleothem δ^{18}O records of stalagmites MSD and MSL from Hulu Cave [*Wang et al.*, 2001] and Dongge Cave [*Yuan et al.*, 2004], located some 1200 km apart in China; speleothem δ^{18}O record from Moomi Cave located on Socotra Island near South Yemen and the Gulf of Aden [*Burns et al.*, 2003] (speleothem δ^{18}O at all sites is interpreted as a regional rainfall index (more negative δ^{18}O values equal wetter conditions and a stronger monsoon)); GISP2 δ^{18}O record [*Stuiver and Grootes*, 2000] showing D-O temperature variations over Greenland (selected warm interstadial events are numbered for comparison to other time series); sediment δ^{15}N record from Oman margin piston core RC27-23 in the northwestern Arabian Sea [*Altabet et al.*, 2002], interpreted to reflect denitrification changes resulting from variations in monsoonal upwelling/productivity and oxygen minimum zone intensity (more positive δ^{15}N values indicate stronger monsoon, increased monsoonal upwelling, and/or an intensified oxygen minimum zone); and percent total organic carbon content (TOC) of sediments from Arabian Sea piston core 111KL [*Schulz et al.*, 1998], with higher values interpreted to reflect stronger monsoonal upwelling and higher productivity/increased organic carbon preservation. All time series are plotted using their original published age models (data from *Burns et al.* [2003] are plotted on the corrected time scale of 10 September 2004).

the marine and land-based records together yield a coherent picture of monsoon behavior on millennial time scales. Within the Holocene, *Gupta et al.* [2003] have used variations in an upwelling-sensitive foraminiferal species to demonstrate that monsoon variations can also be linked to Greenland/high-latitude temperatures on centennial time scales as well.

[36] Is there evidence for cooling in the tropics coinciding with North Atlantic events? Tropical ice cores from high-altitude glaciers in the Andes of both Peru and Bolivia appear to show evidence of a temperature decrease during the Younger Dryas event [*Thompson et al.*, 2000], with temperature changes at Sajama in Bolivia inferred from the ice δ^{18}O to have been of similar magnitude to the glacial-Holocene signal. There are, however, age model concerns at some of these sites as well as debate as to what extent δ^{18}O in tropical ice cores is a temperature signal versus one reflecting a balance of local humidity/runoff [*Pierrehumbert*, 1999; *Baker et al.*, 2001]. To date, the only clear, quantitative evidence of Younger Dryas cooling within the tropics comes from the Cariaco Basin Ma/Ca record [*Lea et al.*, 2003]. Tropical temperature depressions have been an area of active debate for decades [*Crowley*, 2000]. Even for the Last Glacial Maximum (LGM), a time for which there is reason to expect a significant temperature signal because of lowered atmospheric CO_2, it is unclear how much the tropics cooled. Thus the question about how much colder the tropics may have been in the past, whether at the last glacial or during the stadials, is currently an open one.

[37] The observation that abrupt climate change is largely manifested in the tropics in the form of hydrologic variability has important implications for the tropic's potential role in driving climate and as a bridge between the hemispheres. In addition to the potential for altering atmospheric water vapor, itself a powerful greenhouse gas, variations in the concentration of atmospheric methane that accompany the D-O oscillations (high/low CH_4 during interstadials/stadials, Figure 3) have been well documented from ice core studies [*Stauffer et al.*, 1988; *Brook et al.*, 1996, 1999]. During the glacial, when the northern boreal regions that now constitute a significant source of methane were ice covered, low-latitude wetlands were likely the dominant source of variability in atmospheric methane levels. Indeed, *Ivanochko et al.* [2005] have argued that changes in the strength and structure of tropical convection systems throughout the tropics may have served to modulate terrestrial emissions of methane and provided a tropical mechanism for amplifying and perpetuating millennial-scale climate change.

3. PHYSICAL PROCESSES

[38] Our review of the paleoclimate data indicates that there is good evidence of abrupt climate changes timed with those that occurred in Greenland in regions as remote as the western Pacific Basin and Asian monsoon region, as well as coeval climate changes in parts of Antarctica (that are generally not abrupt). With these features in mind, we

now turn to physical mechanisms that can produce abrupt changes and possibly link these remote regions. We proceed by examining to what extent each mechanism can produce a change that (1) is abrupt, (2) can persist and possibly recur on millennial time scales, and (3) can link remote regions of the globe.

3.1. Thermohaline Circulation

3.1.1. Abruptness

[39] The first issue we address is whether the thermohaline circulation can change abruptly. The most common explanation for an abrupt change in the THC is that it is caused by a meltwater event. It is argued that if the fresh meltwater flowing into the Atlantic reduces the salinity sufficiently, then the waters will not be dense enough to sink. Without the formation of deep water in the North Atlantic, the balance of mass fluxes in the global ocean between upwelling and diffusion will have to adjust, thereby altering the thermohaline circulation globally [see *Kuhlbrodt et al.*, 2007]. The abruptness in this scenario can come about in two ways. First, the meltwater event is generally portrayed as an abrupt perturbation to the system. At least two different types of meltwater events have been suggested in the literature. The first pertains to the Younger Dryas. For that event, it has traditionally been argued that as the Laurentide Ice Sheet melted back at the end of the last glacial period, meltwaters flowing out of glacial Lake Agassiz were diverted from a southern route through the Mississippi drainage basin to the east and directly into the North Atlantic. The second type of meltwater event pertains to the Heinrich events. These events are defined by the presence of layers of ice-rafted detritus in North Atlantic sediments, indicating a significant source of fresh water from melting icebergs. *Hemming* [2004] has suggested that the provenance of these icebergs is the Hudson Strait, and the surges are likely to have been caused by episodic purging of the Laurentide Ice Sheet; however, other mechanisms are possible including jokulhlaup activity from Hudson Bay lake or ice shelf buildup/collapse. Several models of such behavior in glaciers have been constructed (see *Hemming* [2004] for references); however, fully interactive, dynamic glacial ice has not been built into most coupled climate models, so the connection between glacial dynamics and ocean/atmosphere circulation has not yet been tested.

[40] The second possible source of abruptness in the THC behavior is the possibility that there are multiple stable states of the THC. This was first suggested by *Broecker et al.* [1985] and then demonstrated with a climate model by *Manabe and Stouffer* [1988]. Since then, numerous idealized ocean-only and coupled models have been shown to contain multiple stable states [*Rahmstorf et al.*, 2005]. In such models, minimal stochastic perturbations in freshwater forcing can be sufficient to induce an abrupt change in the THC strength [*Ganopolski and Rahmstorf*, 2001; *Timmermann et al.*, 2003]. It remains to be seen, however, whether this behavior is an artifact of particular models [e.g., *Vellinga*, 1998], whether it appears in more complete models, or whether it is characteristic of the real climate system. Experiments with the latest state-of-the art coupled models do not appear to show such behavior. *Stouffer et al.* [2006] conducted a coupled model intercomparison project to test the response of a number of different models to a 0.1 Sv addition of fresh water to the North Atlantic over 100 years, after which the freshwater perturbation stops. The value of 0.1 Sv was chosen because it is comparable to the amount of additional fresh water that models simulate will be added to the North Atlantic because of large (i.e., 4 times the present) CO_2-induced climate changes. The results (Figure 6) show that there is a gradual weakening of the circulation over the 100 year period, but the models do not indicate a particularly abrupt or threshold-type behavior. Furthermore, the circulation recovers when the freshwater perturbation stops, with different degrees of rapidity in different models, though again, the behavior does not appear to be abrupt. Coupled model experiments with larger freshwater forcing do generate a more rapid change [e.g., *Vellinga et al.*, 2002; *Rind et al.*, 2001; *Zhang and Delworth*, 2005], though the models recover even from an essentially complete shutdown and return to the modern strength of the THC, suggesting that the "off" state is not a stable one, and overall, the responses are generally linear to the size of the freshwater perturbation [e.g., *Rind et al.*, 2001].

[41] The freshwater perturbation experiments with models raise the issue of how large or "catastrophic" the meltwater events actually were. That is, did enough meltwater flood into the North Atlantic to collapse the THC within decades to centuries in the models (as suggested by the observations for Heinrich events)? Several observation-based estimates have been made of the freshwater fluxes into the North Atlantic for the Younger Dryas and Heinrich events; these are summarized in Table 1. Earlier estimates were based on the flux of ice-rafted detritus [e.g., *MacAyeal*, 1993; *Alley and MacAyeal*, 1994] and on models of the history of the Laurentide Ice Sheet [*Licciardi et al.*, 1999]. These yielded values of about 0.07–0.1 Sv. A more recent estimate of *Carlson et al.* [2007] based on geochemical proxies from the mouth of the St. Laurence estuary gave similar values. *Hemming* [2004] made estimates based on oxygen isotope excursions in North Atlantic records. She argued that the magnitude of the freshwater input depends on the duration of the meltwater event because of ocean mixing; if it occurs over a longer time, one would expect more mixing and hence a larger value for the total input in order to explain the observed geochemical changes in the ocean. These estimates yield values higher than the previously accepted values of *MacAyeal* [1993], *Alley and MacAyeal* [1994], and *Licciardi et al.* [1999] and the more recent *Carlson et al.* [2007] estimate. In the model experiments of *Stouffer et al.* [2006], a 0.1 Sv input of freshwater per year reduces the THC strength by about 25% in 100 years (for the ensemble mean). This is clearly within the range of observed meltwater estimates (and is on the low side relative to Hemming's estimates) but does not shut down the circulation. Other modeling experiments have shown that larger values of meltwater are capable of

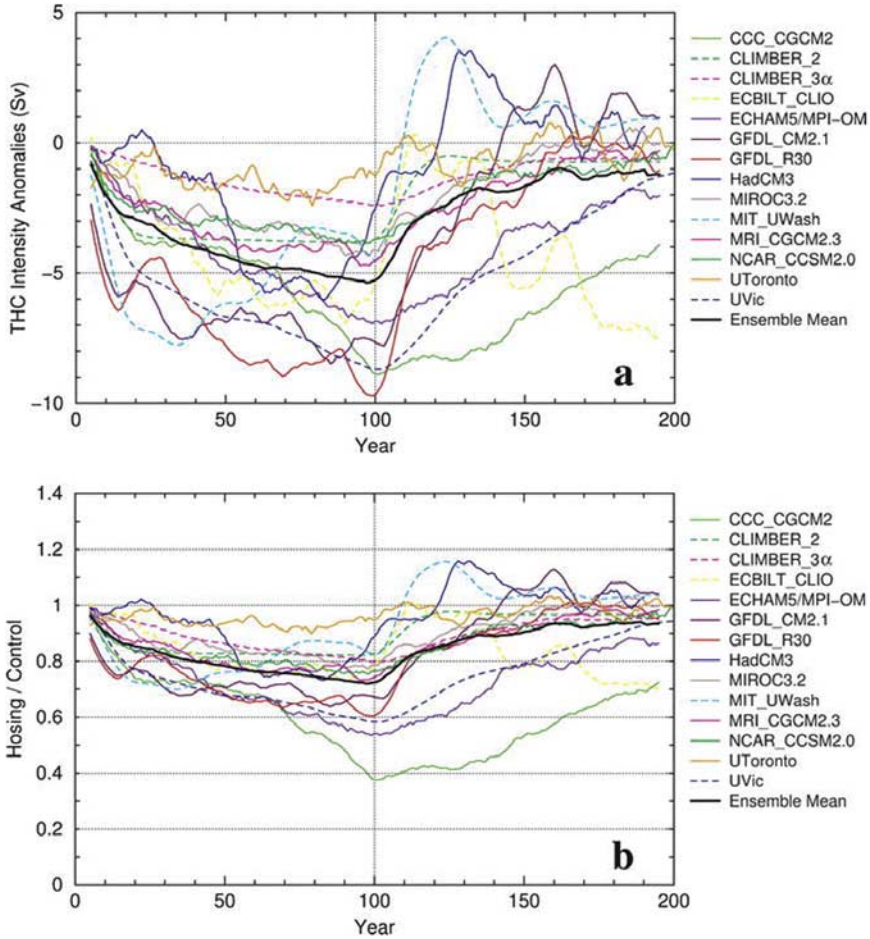

Figure 6. Time series of the THC intensity anomalies in the 0.1 Sv water hosing experiments: (a) absolute value of the anomalies and (b) relative anomalies compared with the long-term mean of the THC intensity in the control experiments. Solid curves denote coupled ocean-atmosphere general circulation models, and dashed curves denote Earth system models of intermediate complexity. From *Stouffer et al.* [2006]. Reprinted with permission courtesy of the American Meteorological Society.

shutting down the circulation completely: 0.6 Sv per year given by *Zhang and Delworth* [2005], 6×10^6 km^3 given by *Vellinga and Wood* [2002], and 0.15–0.2 Sv given by *Rind et al.* [2001]. While these values appear large relative to earlier estimates, they are within the range of Hemming's δ^{18}O-based estimates. Thus it appears that realistic values of freshwater input are capable of shutting down the THC if Hemming's estimates are included.

[42] One very important caveat to note about the model-simulated sensitivity of the THC to freshwater forcing is that all of these published model simulations use modern boundary conditions (i.e., land ice coverage, CO_2, surface albedo, and bathymetry). At present, freshwater perturbation experiments with a coupled general circulation model (GCM) for the glacial climate, which is more appropriate than the modern climate, have not yet been published. *Schmittner and Clement* [2002] showed in a simpler model that the THC response to freshwater forcing is sensitive to mean climate conditions, and it follows that the response in the rest of the world to THC changes could also be different for different mean climates. Thus, the question of whether the connections between the high and low latitudes are dependent on the mean climate state remains open.

[43] Another important point with respect to the abruptness of THC change that has been noted by *Seager and Battisti* [2007] is that the warmings in the ice core record tend to be more abrupt than the coolings (Figure 1). Experiments with ocean-only or idealized coupled models have suggested that a rapid recovery of the THC can occur after a gradual warming of the ocean by diffusion [*Weaver and Sarachik*, 1991; *Winton and Sarachik*, 1993; *Ganopolski and Rahmstorf*, 2001]. However, more sophisticated coupled GCMs (CGCMs) have, thus far, not shown abrupt THC strengthening leading to an abrupt warming of the North Atlantic.

[44] Finally, *Wunsch* [2006] has alternatively suggested that the THC can be altered by abrupt changes in the distribution of the surface winds. On the basis of a collection of idealized studies, he has argued that interaction between the atmospheric circulation and the large ice sheets of the glacial period could produce abrupt changes in the winds and hence impact both the temperature over Green-

land as well as the ocean circulation. These ideas certainly warrant further testing with coupled models.

3.1.2. Millennial Time Scale

[45] One of the most challenging features of the Greenland ice core record to explain is the time scale of the events. This includes the facts that the Greenland events appear to persist for several centuries and in the case of the YD to persist for as long as ~1000 years and also that there appears to be a recurrence of the D-O events on average about every 1500 years. On the basis of the CGCM results cited in section 3.1.1, it seems that if a meltwater event persists for centuries, then the THC response will persist for that time period. However, *Hemming* [2004] has pointed out that we have little idea of how long the Heinrich event meltwater pulses may have lasted. If the event is as short as 1 year (a minimum estimate), then the model results would suggest that the THC response would gradually recover within a century [*Vellinga and Wood*, 2002] or less [*Krebs and Timmermann*, 2007a, 2007b]. On the other hand, if the meltwater pulses do last for several centuries, what sets this time scale? Perhaps it is set by glaciological instability. The so-called "binge-purge" mechanism has been proposed, which suggests that ice sheets gradually accumulate until the weight of the ice sheet is sufficient to lubricate the base of the layer, causing the ice sheet to collapse [*MacAyeal*, 1993]. Then the cycle begins again, and the geographic scale of the glacier can produce a recurrence time scale of millennia. That study predicts a recurrence interval of about 7 ka, roughly consistent with that of Heinrich events. However, in reviewing the mechanisms that could explain Heinrich layers, *Hemming* [2004] notes that there are other possible mechanisms and suggests particular ways to test each of these. It is as yet not clear whether glaciological processes could produce meltwater events on the shorter D-O time scale.

[46] It has been suggested by *Alley et al.* [2001] and *Roe and Steig* [2004] that stochastic processes together with threshold crossings between stable climate states can explain the temporal behavior of the Greenland ice core records. These conceptual models require the specification of some characteristic time scales, which are typically chosen based on the best fit to observations. While these provide useful null hypotheses, they leave open questions about the mechanisms that can introduce these millennial time scales. For example, *Alley et al.* [2001] suggest that a weak periodic forcing of about ~1500 years when coupled with stochastic forcing can produce time series with similar statistical properties as the ice core records. However, they note the origin of such a periodic forcing, although it need only be a weak one, is entirely unknown.

[47] While there are no known climate forcings that have a time scale of 1500 years, there has been some suggestion that solar forcing could be responsible for millennial-time scale variability in the climate [*van Geel et al.*, 1999; *Bond et al.*, 2001]. For example, if the climate system is sufficiently nonlinear, the shorter-time scale solar cycles could produce some longer-term variability. This idea has been tested by *Braun et al.* [2005] in a model of intermediate complexity. By imposing a freshwater forcing to the North Atlantic at time scales of 87 and 210 years (two known solar cycles), they were able to produce 1500 year cycles in the THC. This study, however, begs the question of why a change in the Sun's irradiance should cause a meltwater pulse.

3.1.3. Global Linkages

[48] Simple reasoning would lead to the conclusion that a change in the THC could affect certain remote regions of the globe. Water that flows into the North Atlantic as warm water at the surface exits the basin as cold deep water resulting in a net import of heat into that region. This has been documented using a wide range of different oceanographic [*Talley*, 2003; *Ganachaud and Wunsch*, 2003] and atmospheric [*Trenberth and Caron*, 2001] data sets, all of which show that in the zonal mean, the ocean transports heat across the equator from south to north. If this water mass formation were to stop, it is reasonable to assume that the import of heat into the Northern Hemisphere from the south would also be reduced, which would lead to a cooling of the Northern Hemisphere and a warming of the Southern Hemisphere. This is the essence of the so-called bipolar seesaw originally proposed by *Crowley* [1992], *Stocker et al.* [1992], and *Stocker* [1998]. The idea is that a change in the strength of the THC that causes an abrupt change in temperature of the high-latitude Northern Hemisphere would produce a synchronous change of the opposite sign in the high-latitude Southern Hemisphere. *Knutti et al.* [2004] have shown in a model that in addition to this "thermal" seesaw, there is also a dynamic seesaw that arises because of the influence of meltwater on sea level. If freshwater flows into the western part of the basin, the zonal gradient in sea level gives rise to a reverse meridional overturning (with water flowing to the south in the surface and to the north in the subsurface), which contributes to the warming in the Southern Hemisphere and cooling in the north.

[49] It should be noted that *Wunsch* [2006] has used the THC heat transport argument to come to the opposite conclusion: that it is unlikely that THC changes can have an impact outside the Atlantic Basin as it contributes only a small fraction of the total poleward heat transport. One way to test the impact of the THC on the global climate is to use our best understanding of the global climate system as represented in coupled GCMs.

[50] The first experiments with coupled GCMs to test the effects of altered THC on the global climate were performed by *Manabe and Stouffer* [1988]. The model that they used was developed at the Geophysical Fluid Dynamics Laboratory (GFDL) to study a host of climate problems. The atmosphere model had a horizontal resolution of 4.5° latitude and 7.5° longitude and nine vertical levels. The ocean model had a horizontal resolution of 3.75° latitude and 4.5° longitude and 12 vertical levels. In order for the simulation of the climate to be realistic and stable, adjustments had to be made to the surface heat and freshwater fluxes to correct for biases in both the atmosphere and ocean models, a practice that has come to be known as adding "flux corrections."

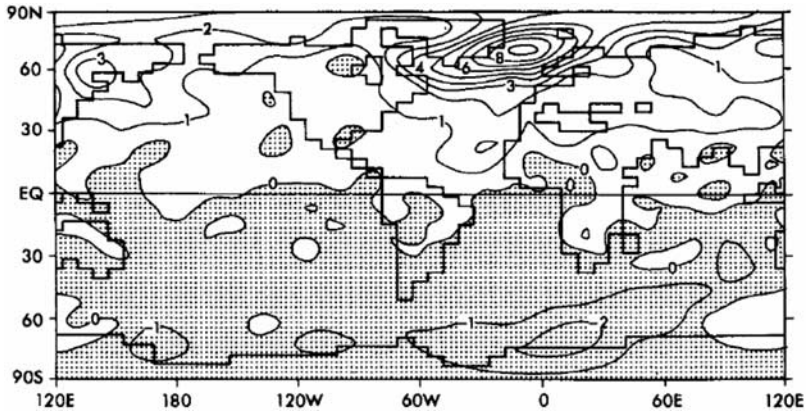

Figure 7. Surface air temperature difference between two stable states of the GFDL coupled model: strong THC (~12 Sv) minus weak THC (~0 Sv). Contour internal is 1 K. From *Manabe and Stouffer* [1988]. Reprinted with permission courtesy of the American Meteorological Society.

Experiments were then performed in which the freshwater forcing in the North Atlantic was altered to show that two stable states of the ocean circulation are possible for identical boundary conditions, one with a strong overturning and one with weak overturning. This had a large effect on the temperature in the North Atlantic region but less effect on temperature outside the region, notably in the Southern Hemisphere (Figure 7), a result consistent with other GCM simulations of a similar generation [e.g., *Rind et al.*, 1986]. On the basis of these results, one would have to conclude that the THC is not capable of explaining the observed abrupt climate changes outside the Atlantic Basin (nor over the Antarctic continent).

[51] Much progress has been made in climate simulation with GCMs since the time of the *Manabe and Stouffer* [1988] study, resulting in a new generation of coupled GCMs with new parameterizations, higher spatial resolution, and an improved simulation of the modern climate. An indication of this progress is that the new version of the GFDL coupled model no longer needs flux corrections to

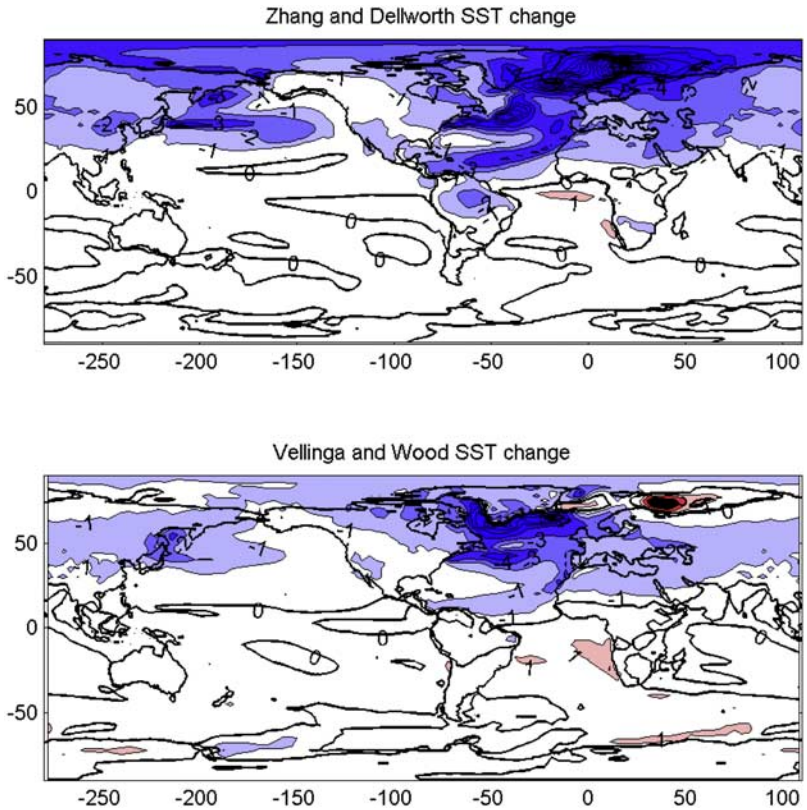

Figure 8. Surface temperature difference between weak THC and modern states from experiments with the (top) GFDL coupled model [*Zhang and Delworth*, 2005] and (bottom) HadCM3 [*Vellinga and Wood*, 2002]. Contour interval is 1 K.

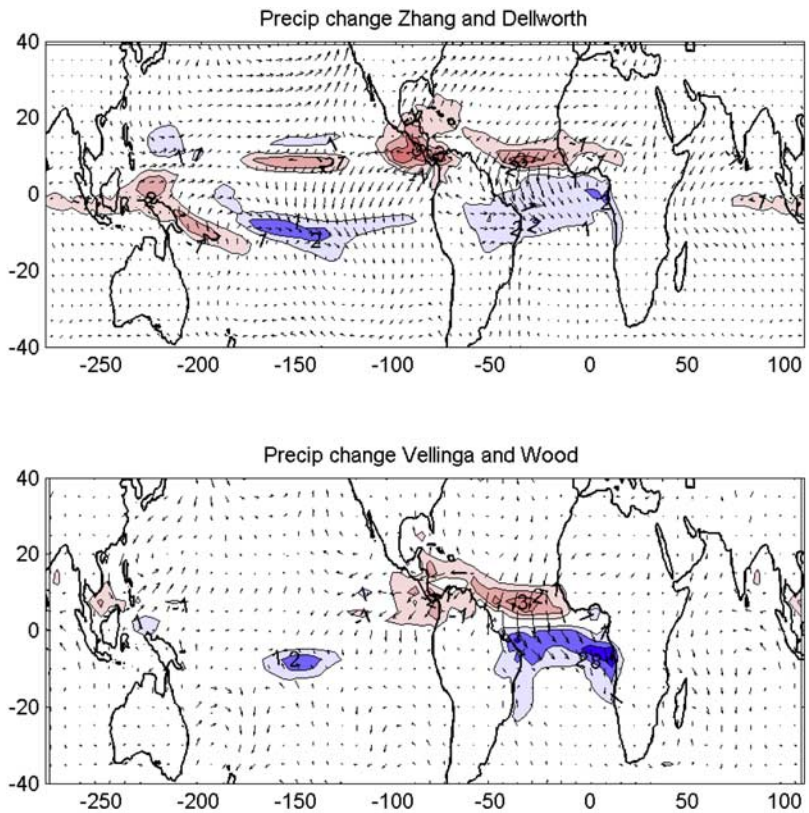

Figure 9. Precipitation (colors/contours) and surface wind (vectors) difference between weak THC and modern states from experiments with the (top) GFDL coupled model [*Zhang and Delworth*, 2005] and (bottom) HadCM3 [*Vellinga and Wood*, 2002]. Contour interval is 1 mm d^{-1}. Wind vectors are the same scale in both Figures 9a and 9b, and the maximum vector for the GFDL and HadCM3 models (on the equator in the tropical Atlantic in both cases) is 2 m s^{-1} and 2.5 m s^{-1}, respectively.

simulate a realistic climate [*Delworth et al.*, 2006]. The simulated climate still differs from the real climate in some significant ways, particularly in the tropics, but it is possible to produce a climate state that is essentially stable. The new version of the model has a horizontal resolution in the atmosphere of 2° latitude by 2.5° longitude and 24 vertical levels. The ocean has 50 vertical levels, 1° zonal resolution, and 1° latitude resolution outside the tropics, increasing to 1/3° at the equator. This higher resolution is designed for improved simulation of near-equatorial dynamics, which have a narrow meridional scale.

[52] *Zhang and Delworth* [2005] (referred to hereinafter as ZD05) used this model to study the climate response to a weakened THC. A flux of 0.6 Sv of fresh water is distributed over the northern North Atlantic over 60 years, which reduces the meridional overturning from 16 to 6 Sv. The difference in surface climate between the weak and strong THC conditions is shown for temperature in Figure 8a and for precipitation in the tropics in Figure 9a. The main difference between this result and the earlier results of *Manabe and Stouffer* [1988] is that there is now a stronger signal outside the North Atlantic, in particular in the tropical Atlantic and the northern Pacific Ocean. It is not possible to test exactly why the results with the two versions of the model are different because there are so many things about the model that have changed. However, an obvious difference is in the spatial resolution, which has more than doubled.

[53] Before entering into the mechanisms by which the THC affects the tropics, let us first compare the results with a similar experiment performed earlier using a different model in order to get some sense of the robustness of the results. *Vellinga and Wood* [2002] (hereinafter referred to as VW02) used a recent version of the U.K. Meteorology Office's Hadley Centre Model (HadCM3). It is also a coupled GCM that produces a stable climate simulation without flux adjustments. The resolution is 2.5° × 3.75° with 19 levels in the atmosphere and 1.25° × 1.25° with 20 vertical levels in the ocean. A freshwater pulse of 6 × 10^5 km^3 is put into the northern North Atlantic. In response, the THC collapses within 10 years. The model is then allowed to run with no additional forcing for 150 years during which the THC recovers to its control values. The change in surface climate is shown together with the results from ZD05 in Figures 8 and 9. We will also cite the published experiments of *Stouffer et al.* [2006], in which a more modest freshwater forcing was applied in the discussion of the robustness of some of the features of the global climate change in response to THC perturbations.

[54] There are a number of common features in the simulated climate response to a reduced THC in the GFDL CGCM and HadCM3. Both show significant cooling in the

North Atlantic, with some cooling throughout the Northern Hemisphere. The signal is larger in the ZD05 experiments, but that is perhaps because the THC perturbation is larger. The results shown here (and in the original papers) are for a change in THC of 16–4 Sv for ZD05 (a 75% reduction) and 20–8 Sv for VW02 (a 60% reduction). To the extent that the surface climate response depends on the absolute change in THC, these two experiments are essentially equivalent, although the fractional change is larger in the ZD05 work. Also, both models show a modest warming of the Southern Hemisphere, with the largest signal in the eastern South Atlantic, though at most that signal is 1°C, and there is practically no temperature signal in Antarctica.

[55] As for the tropics, the signal of cooling in the northern tropical Atlantic and warming in the southern tropical Atlantic is simulated by both models. The cooling in the north as given by ZD05 is somewhat larger than given by VW02, but both show a tongue of maximum cooling extending from the coast of Africa toward the southwest. Despite the somewhat larger northern tropical Atlantic cooling given by ZD05, there is a very similar response in tropical Atlantic precipitation in both models with decreased precipitation north of the equator and increased precipitation to the south (Figure 9). *Stouffer et al.* [2006] showed similar patterns of temperature and precipitation changes in the 13 models they used in their study, though magnitudes varied from model to model and were generally smaller in the models of intermediate complexity.

[56] Outside the Atlantic, another similar feature of the GFDL and HadCM3 models is the reduction in the magnitude of the temperature signal in the Pacific and Indian oceans compared with the Atlantic. Both models show less than 1°C temperature change anywhere in those basins south of 20°N. Despite little temperature change in the Pacific, both models simulate precipitation changes in that basin. ZD05 show the precipitation changes in the Pacific are of comparable magnitude and even in some places are larger than in the Atlantic. VW02 show that the Pacific precipitation signal is smaller in the Pacific than the Atlantic but not everywhere (e.g., the southern tropical Pacific).

[57] One area of particular interest is that which is affected by the summer Asian monsoon because several records of abrupt change come from that region. In Figure 10 the precipitation and surface winds for June, July, and August are compared for the two model simulations. Both models simulate comparable overall magnitudes for the monsoon precipitation for the modern climate (left plots), though the HadCM3 monsoon appears to be somewhat stronger. However, there is a great deal of detail in the spatial structure that is different between the two models. The impact of the change in THC is shown in the right plots. There is a reduction in the precipitation over some regions, but it is clear that not all of the simulated centers of monsoon precipitation are weakened by the reduced THC. Both models show a reduction in precipitation over the Indian subcontinent. However, the change is small (<2 mm d^{-1}) and does not coincide with the main regions of monsoon precipitation. It could be argued that the changes over India represent a slight contraction of the region most affected by monsoonal precipitation. ZD05 show a significant reduction in the strength of the southerly winds over the western Arabian Sea (~18% as reported in their work), but VW02 do not show such a response in the wind field. Both models show a reduction in the precipitation in the Southern Hemisphere between the equator and 10°S, which, unlike the Northern Hemisphere, does appear to be a weakening of the main center of convection. One final observation is that both models show some increased precipitation over the oceans north of the equator, which is opposite to the precipitation change at the same latitude in the eastern Pacific and Atlantic basins. The response of the Asian monsoon was not intercompared in the models used by *Stouffer et al.* [2006], though the average over all the models does show a decrease in precipitation over India with little signal in the rest of the monsoon region.

[58] These two different state-of-the-art models show that a connection between the THC and remote regions of the tropics is plausible, and despite the difference in resolution, numerical schemes, and model physics, some of the remotely forced signals are robust. It is difficult, however, to discern the mechanisms that give rise to the changes shown in Figures 8 and 9 because they are the result of changes in the ocean, atmosphere, and interactions between the two; a causal chain of events is therefore impossible to diagnose. We turn, instead, to other studies that have used idealized experiments with coupled and atmospheric GCMs to test how changes in the North Atlantic can reach the global tropics. Four different mechanisms have been proposed in the literature; these are reviewed in sections 3.1.3.1–3.1.3.4. We will focus on the connection between the THC and the tropics where the hydrological signals are significant and the data indicate that abrupt change occurred.

3.1.3.1. Zonal Teleconnections in the Tropics: Tropical Atlantic to Pacific

[59] Both CGCMs show a similar pattern of temperature and precipitation change in the Atlantic Basin. Other studies have shown how changes in the northern circum–North Atlantic region can affect the tropical Atlantic through both the ocean and atmosphere.

[60] *Dong and Sutton* [2002] used the same model as VW02 to study the adjustment process to a change in THC and found a fast connection between the North Atlantic and the tropical Atlantic through the ocean. Rather than looking at the equilibrated state, as was done by VW02, they looked at the sensitivity of the climate to an instantaneous change in the salinity of the North Atlantic Ocean in the same Hadley center coupled model. The results showed that the ocean adjusts quickly to the freshwater perturbation, and the THC goes from ~15 Sv to almost zero in about a month because of the capping of the North Atlantic with low-density water. The effect of the change in THC is felt in the tropical Atlantic within about a season, with the currents and transports being significantly affected. After ~5–6 years, the tropical SST response peaks with a cooling north of the

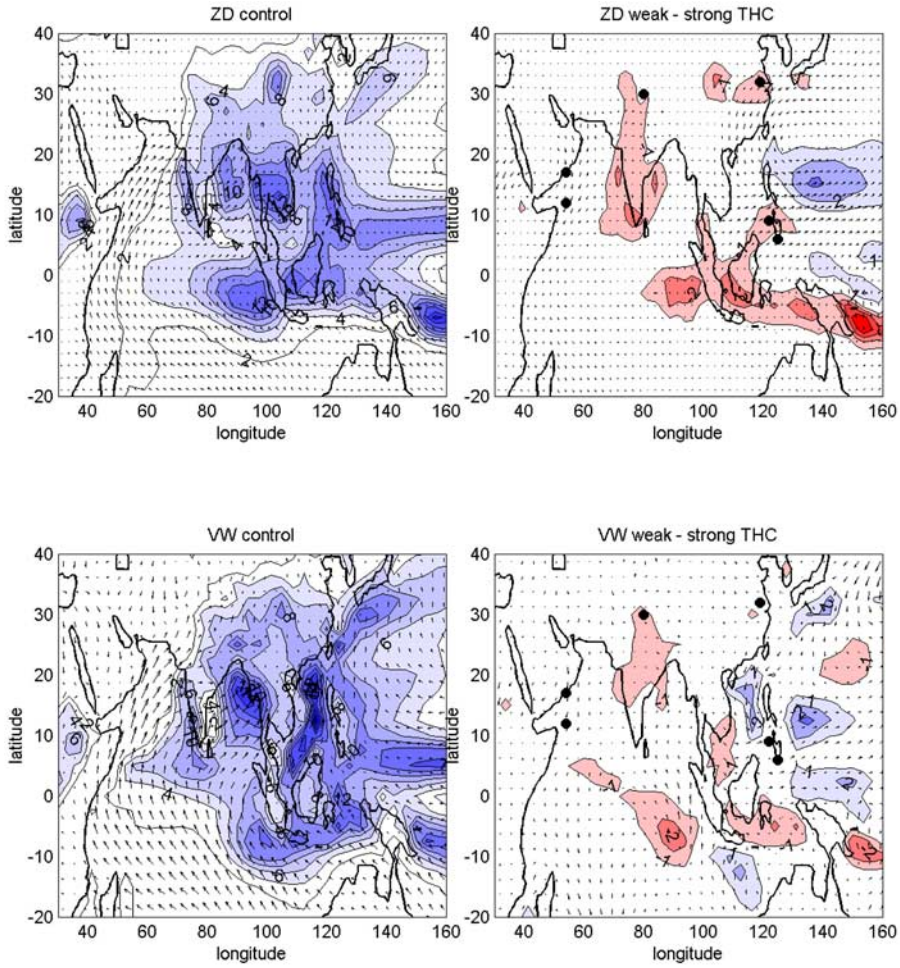

Figure 10. June, July, and August precipitation (colors/contours) and surface winds (vectors) from (top) *Zhang and Delworth* [2005] and (bottom) *Vellinga and Wood* [2002]. (left) Mean from the control run with each model. Contour interval is 2 mm d^{-1}, and maximum wind speed vector is 10 m s^{-1}. (right) Difference between the weak THC and the control. Contour interval is 1 mm d^{-1}. Wind vectors are the same scale in all plots, and the maximum wind speed vector is 2 m s^{-1} for both models. The circles in the right plots show the locations of the paleoclimate records discussed in section 3.1.3. They are from *Sinha et al.* [2005] for 30°N, 80°E; *Stott et al.* [2002] for 6°N, 125°E; *Fleitmann et al.* [2003] for 17°N, 54°E; *Burns et al.* [2003] for 12°N, 54°E; *Wang et al.* [2001] and *Yuan et al.* [2004] for 32°N, 119°E; and *Rosenthal et al.* [2003] and *Dannenmann et al.* [2003] for 9°N, 122°E.

equator and a warming south of the equator, similar to the pattern in Figure 8.

[61] *Chiang et al.* [2003] have argued that changes in the northern circum–North Atlantic region influence the tropical Atlantic through the atmosphere. They showed that glacial boundary conditions, in particular ice sheets over land, generate stronger trade winds and reduced SSTs in the northern tropical Atlantic. The altered SST pattern shifts the ITCZ to the south leading to significant precipitation anomalies in the Atlantic Basin. *Chiang et al.* [2003] note that such a pattern in meridional SST gradient and ITCZ position is analogous to modern interannual climate variability [*Moura and Shukla*, 1981; *Nobre and Shukla*, 1996; *Chiang and Vimont*, 2004]. Figure 9 shows that in both coupled models, there is indeed an increase in the trade winds in the northern tropical Atlantic and an anomalous southerly flow across the equator, which are both consistent with this "meridional mode" as well.

[62] In both models the tropical response to the THC is not confined to the Atlantic. How does the signal get out of the Atlantic and into the Pacific? *Dong and Sutton* [2002] find that after 7 years, their perturbed experiments showed an El Niño event developing in response to the freshening of the North Atlantic. Those authors hypothesized that the signal propagates from the tropical Atlantic into the Pacific via atmospheric Rossby waves and then alters the surface winds in the Pacific, initiating coupled feedbacks that lead to an El Niño event. ZD05 invoke a similar pathway to explain their results. They argue that the Pacific and Atlantic are linked through the Central American region. There is an increased sea level pressure (SLP) over the northeastern tropical Pacific as a result of the Atlantic changes; that is, the far northeastern ITCZ in the Pacific appears to be strongly influenced by conditions in the Atlantic, unlike the region just to the south that is strongly influenced by the El Niño–Southern Oscillation (ENSO).

ZD05 suggest that this SLP anomaly initiates coupled interactions that amplify the response. According to ZD05, reduced SLP over the far northeastern tropical Pacific induces a coupled response in the meridional plane: higher SLP to the north induces northerly cross-equatorial flow, which leads to cooling to the north of the equator and warming to the south. *Xie* [1994] previously argued for the presence of the coupled "meridional mode" in the eastern Pacific and has pointed out that this is fundamental to the annual cycle of temperature and precipitation in that region. We note, however, that there is little precipitation change to the south of the equator; that is, unlike the Atlantic, in this longitudinal sector the precipitation change is perhaps more a weakening of the ITCZ than a shift.

[63] ZD05 argue that changes in the tropical Pacific climatology can then influence the Indian monsoon. In the modern climate, warm El Niño events tend to coincide with weaker Indian monsoons [*Shukla*, 1987]. This connection, however, appears to vary on a decadal time scale for reasons that are not well known and may be the result of stochastic processes [*Kumar et al.*, 1999; *Gershunov et al.*, 2001; *Steerl et al.*, 2007]. Nevertheless, there is a dynamical connection between the western Pacific convective area and the Indian monsoon via westward propagating Rossby waves: when atmospheric heating is shifted eastward with El Niño, a Rossby wave signal tends to produce flow that would dry the Indian subcontinent [*Hoskins et al.*, 1999; *Lau and Nath*, 2000]. ZD05 invoke a similar connection to explain their model results: the shift of rainfall to the east in the southern tropical Pacific would weaken the Indian monsoon.

[64] The mechanisms described here by which the North Atlantic affects the global tropics, based on the studies of *Dong and Sutton* [2002] and ZD05, are illustrated schematically in Figure 11a. These will be contrasted with two other mechanisms that have been previously proposed in the literature.

3.1.3.2. Zonally Symmetric Adjustment to Interhemispheric Temperature Gradients

[65] Rather than propagating equatorward in the Atlantic Basin and then hopping over the isthmus into the Pacific as proposed by *Dong and Sutton* [2002] and ZD05, other studies have suggested that cooling of the high northern latitudes perturbs the planetary heat budget causing a hemisphere-wide and largely zonally symmetric adjustment. *Broccoli et al.* [2006] performed idealized experiments using an atmospheric GCM coupled to a mixed layer ocean model in which the Northern Hemisphere was cooled and the Southern Hemisphere was warmed (so that the global mean temperature was the same). They found that there was an increased transport of heat from the south (the warmer hemisphere) into the north (the cooler hemisphere), which was accomplished by a reorganization of the Hadley cell. The ascending branch of the annual mean Hadley cell shifted south into the warmer hemisphere resulting in an asymmetric cell that transports heat across the equator (Figure 12). Associated with this circulation change is an annual shift of the ITCZ and precipitation toward the south.

[66] Analogous results were found by *Chiang and Bitz* [2005], who examined the climate adjustment to imposed sea ice anomalies in the marginal seas of the North Atlantic and Pacific oceans and also in an atmospheric GCM coupled to a mixed layer ocean. They found that in response to sea ice anomalies, the ITCZ shifts southward at almost every longitude. Interestingly, they also found that there was a similar zonally symmetric ITCZ shift if the sea ice change was imposed only in the Atlantic or only in the Pacific. Moreover, they showed that this response is consistent with the ITCZ response to glacial land ice changes. The commonality in all of these climate perturbations is that they result in a cooling of the entire Northern Hemisphere high latitudes making it analogous to the *Broccoli et al.* [2006] results. *Chiang and Bitz* [2005] proposed a mechanism for the propagation of the cooling toward the equator. They argued that a signal originating in one or the other ocean basins, or over land, is quickly propagated zonally throughout the hemisphere via the westerly atmospheric flow and mixing. Then this high-latitude cooling signal propagates into the tropics via air-sea interactions in both the Atlantic and Pacific basins. The mechanism of equatorward propagation has been previously elaborated by *Xie* [1999] and applied to tropical Atlantic variability: cooler midlatitude oceans give rise to higher pressure, resulting in an anomalous pressure gradient that drives stronger trades winds. The stronger winds lead to increased evaporative heat flux and cooler SSTs, which subsequently move the anomalous pressure gradient equatorward, and the signal propagates. Once it reaches the region of the ITCZ, the anomalous north-south pressure gradient shifts the ITCZ to the south. Because the high-latitude cooling exists at all longitudes, this works in both basins. *Chiang and Bitz* [2005] showed further that the atmospheric heat transport was altered in their experiments such that there was a net transport into the Northern Hemisphere. Their results effectively show one mechanism by which high-latitude signals can reach the tropics, but what appears to be fundamental is the requirement for the atmosphere to transport heat into the cooler (or icier) hemisphere as described by *Broccoli et al.* [2006]. This mechanism is illustrated schematically in Figure 11b.

[67] The studies of *Broccoli et al.* [2006] and *Chiang and Bitz* [2005], however, both use atmospheric GCMs coupled to a mixed layer ocean and hence do not include feedbacks from ocean dynamics. To get an idea of whether these results would hold up under the presence of interactive ocean dynamics, we can return to the coupled GCM results shown in Figures 8 and 9. The THC transports heat from the Southern Hemisphere to the Northern Hemisphere in the Atlantic (and hence in the zonal mean). When this circulation is reduced, so is that heat transport. The atmosphere adjusts by increasing the atmospheric heat transport from the south to north (as shown by ZD05 in their Figure 1), which is consistent with a southward shifted Hadley cell and ITCZ. In a zonal mean sense, the results of ZD05 are generally consistent with this mechanism. However, in the coupled model used by ZD05, there is a strong double ITCZ, which is common to most CGCMs but absent in the

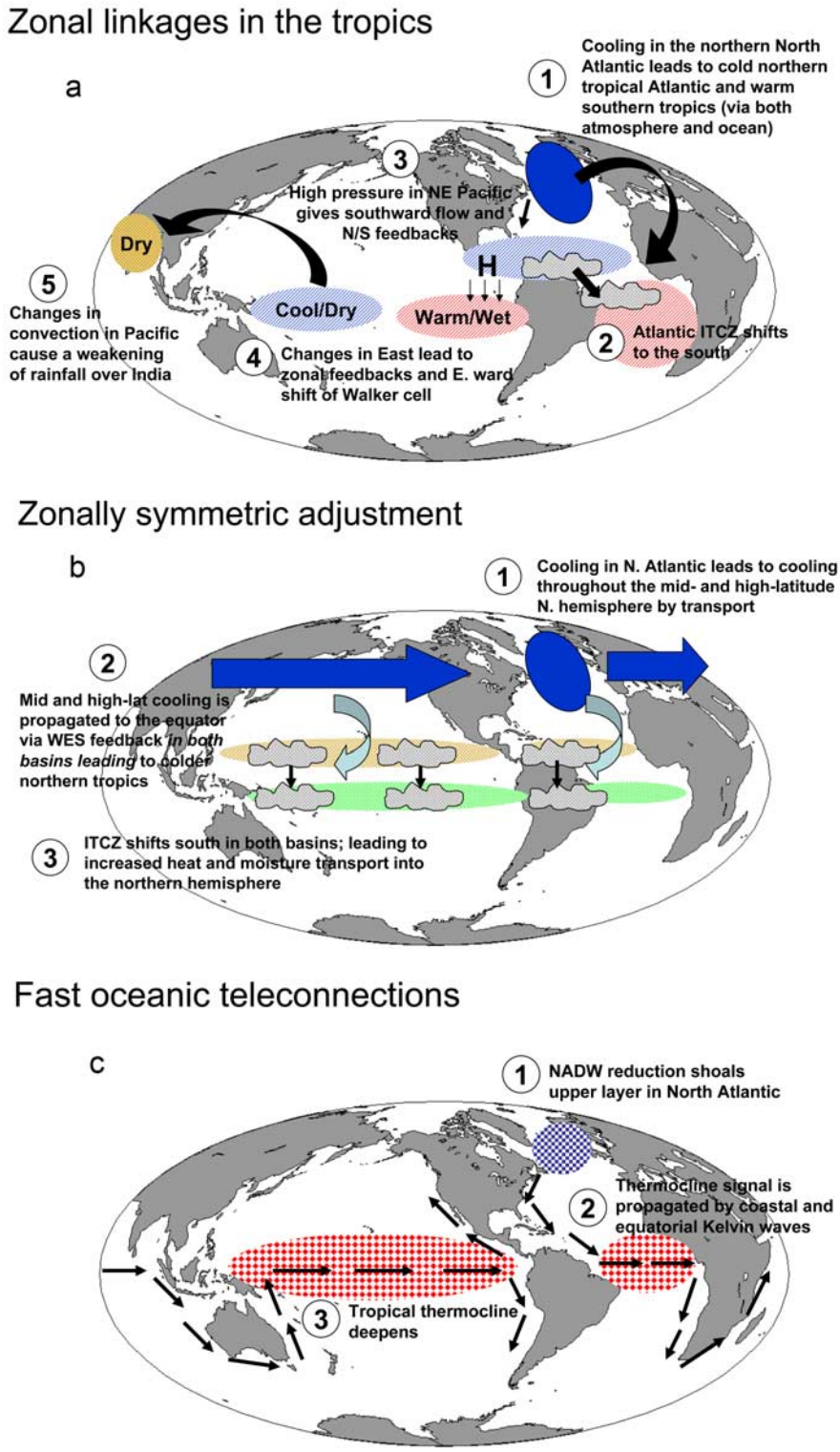

Figure 11. Schematic representations of the mechanisms discussed in the text linking THC changes to the tropics: (a) zonal linkages in the tropics, (b) zonally symmetric adjustment, and (c) fast oceanic teleconnections. The WES feedback labeled in Figure 11b is the wind-evaporation-surface temperature feedback described by *Xie* [1999].

real world. Thus, it is difficult to attribute changes in the ZD05 experiments to an ITCZ shift. It is clear from Figure 9, however, that there are significant zonal asymmetries to the precipitation response. ZD05 argue that these asymmetries arise from coupled interactions in the tropical Pacific, which are not present in the models used by *Broccoli et al.* [2006] and *Chiang and Bitz* [2005]. It is not clear at present to what extent the adjustment to a reduced THC in the tropical Pacific can be attributed to the zonally asymmetric teleconnections proposed by ZD05 versus a more zonally

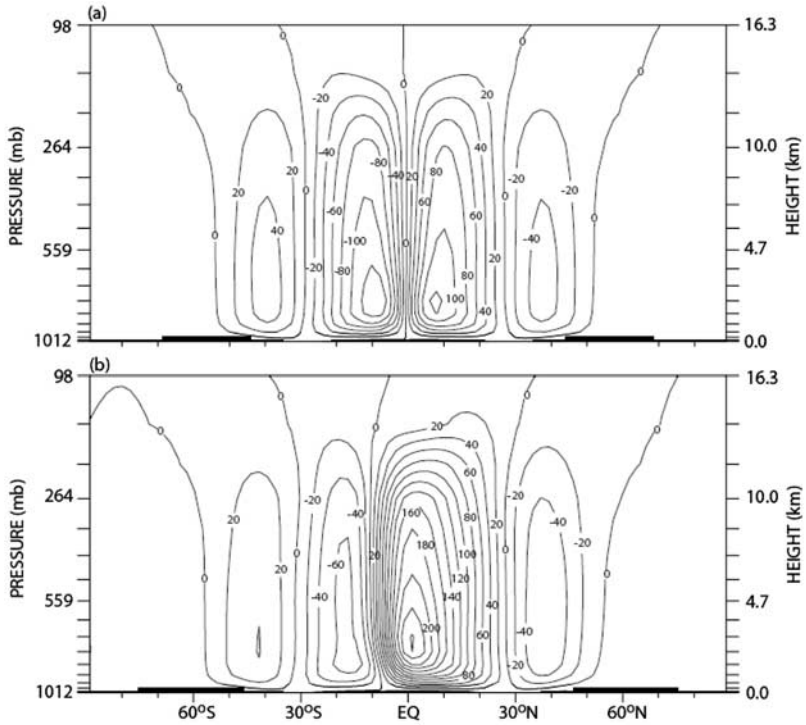

Figure 12. Latitude-height cross section of the annually averaged meridional stream function simulated with (a) axisymmetric surface temperatures and (b) asymmetric temperatures with a colder Northern Hemisphere and warmer Southern Hemisphere. From *Broccoli et al.* [2006].

symmetric adjustment, with the high-latitude changes influencing the tropics in both basins somewhat independently.

[68] The monsoons may be affected also by this zonally symmetric cooling of the Northern Hemisphere but in a slightly different way than for the ITCZ shift. Several studies have shown that more snow cover over the Tibetan plateau can weaken the monsoon [see *Zhao and Moore*, 2004] and that same result has been demonstrated with climate models [e.g., *Barnett et al.*, 1988]. However, analysis of observations of *Wu and Kirtman* [2007] suggests that the relationship between snow cover and the east Asian monsoon is not simple. It has not yet been examined in climate models whether a THC-induced cooling in the North Atlantic can produce an increase in snow cover over Tibet and whether that is the actual reason for the simulated monsoon weakening shown in Figure 10.

3.1.3.3. Global Ocean Adjustment

[69] We have seen clearly that signals from the North Atlantic can reach the tropics through the atmosphere in a short amount of time. Let us now consider whether the ocean can act as a mediator for transmission of signals from the North Atlantic to the tropics.

[70] It is well known that changes in the Atlantic THC impact the global ocean circulation. This result dates back to *Stommel*'s [1961] model, in which he showed that an increased salinity contrast between the tropics and high latitudes reduces the meridional overturning cell (MOC). This further enhances the salinity contrast and initiates a positive feedback that leads to a different state of ocean circulation with sinking in the tropics rather than in the high latitudes. Analogous behavior has been shown in global ocean GCMs coupled to simplified atmospheric models [*Saenko et al.*, 2004; *Knutti et al.*, 2004]. As an interesting example of the connectivity of the ocean circulation between basins, *Saenko et al.* [2004] showed that by removing fresh water from the North Pacific Basin and adding it to the Atlantic, it is possible to initiate a MOC in the Pacific and collapse the MOC in the Atlantic. It is clear from these model results that if the North Atlantic branch of the THC were to change, the signal would be communicated throughout the entire global ocean. However, it is not clear how these changes would affect the surface climate where many of the paleo-observations are from, particularly in the tropics.

[71] On the other hand, there is a component of ocean circulation that can respond quickly and have consequences for the surface climate: the global baroclinic adjustment. Numerous studies have shown that perturbations to the density of the North Atlantic Ocean affect the sea level and thermocline of the entire global ocean within a few years to decades [*Cessi et al.*, 2004; *Huang et al.*, 2000; *Johnson and Marshall*, 2002]. The reason the adjustment is so fast (unlike the advective time scales of the global ocean circulation that are on the order of 1000 years) is because the signal is carried by fast, large-scale oceanic waves. Coastal Kelvin waves carry a density perturbation signal equatorward from the North Atlantic; they reach the equator and travel eastward as equatorial Kelvin waves. At the eastern boundary, the Kelvin waves split into northern and southern branches. Rossby waves emanate from the bound-

ary and establish the interior ocean response. The southern branch of the coastal Kelvin wave wraps around the southern tip of Africa into the Indian Ocean and then by similar propagation travels through the Indian Ocean into the Pacific, with Rossby waves adjusting the interior basins. This connection between the North Atlantic and the global ocean is shown schematically in Figure 11c.

[72] While this may seem a convoluted path, it is clear that the ocean adjusts in this manner in simpler models. Using a shallow water (essentially two layers) model, *Huang et al.* [2000] predict that a 10 Sv reduction in the NADW formation would deepen the equatorial thermocline in the Atlantic by ~100 m and in the Pacific by ~50 m within decades. This is comparable to the thermocline changes associated with El Niño events. If an analogy with modern climate variability can be drawn here, one would expect an El Niño–like response (with a reduced east-west gradient) to be associated with reduced THC. *Timmermann et al.* [2005] have studied this mechanism in a coarse-resolution coupled GCM. They found a deeper Pacific thermocline in response to a reduced THC, but the response was small (~10 m) and had little effect on the SST. It should be noted that the model used by *Timmermann et al.* [2005] has a resolution of 3° on the equator in the ocean and as such does not simulate well equatorial upwelling, so it is not surprising that the thermocline changes have little influence on SST. On the other hand, the change in surface currents in response to this thermocline change, along with the change in trade winds associated with the atmospheric response to colder North Atlantic SSTs, altered the salinity of the warm pool, which they note is consistent with records from the warm pool region.

3.1.3.4. THC Impacts on ENSO Variability

[73] As a further illustration of the possible effects of THC changes in the tropics, *Timmermann et al.* [2007a] have shown that in meltwater hosing experiments, four different coupled GCMs (GFDL CM2.1, CCSM2, CCSM3, and EHCM5-OM1) simulate a strengthening of the ENSO variability when the THC is shut down. The mechanisms invoked to explain these results involve many of those shown schematically in Figure 11. Timmermann et al. argue that the THC shutdown shifts the annual mean position of the ITCZ to the south in both the Atlantic and Pacific basins (as in Figure 11b). In three of the GCMs (GFDL CM2.1, CCSM2, and CCSM3), this southward shift has an influence on the seasonal cycle in the eastern equatorial Pacific. In the modern climate, the ITCZ is to the north of the equator all year-round in the eastern Pacific [*Mitchell and Wallace*, 1992], a situation that is maintained through coupled interactions involving meridional winds, evaporation, mixing, and upwelling [*Philander et al.*, 1996; *Xie and Seki*, 1997]. There is a seasonal cycle in this asymmetry, which provides the "seeding" for the annual cycle of SST on the equator. The strong meridional asymmetry in Northern Hemisphere summer sets up the minimum in SST on the equator in boreal fall (October), and the weak asymmetry in Southern Hemisphere summer is followed by the maximum SSTs in the boreal spring (April). The SST signal in the eastern Pacific also sets up zonal coupled interactions that generate a westward propagating signal that amplifies the seasonal cycle [*Xie*, 1994; *Philander et al.*, 1996; *Li and Philander*, 1997]. *Timmermann et al.* [2007b] argue that when the meridional asymmetry is weakened, the seasonal cycle is also weakened. The strengthening of ENSO is then explained by Timmermann et al. via a frequency entrainment mechanism. When the seasonal cycle is weak, ENSO's own inherent variability, which occurs on an interannual time scale, dominates, whereas when the seasonal cycle is strong, it can drown out ENSO's inherent cycle. All of this happens nonlinearly, so the shift of variance between annual and interannual time scales is not necessarily smooth nor is it additive [e.g., *Chang et al.*, 1994; *Tziperman et al.*, 1994]. We note that coupled GCMs do have some difficulty in simulating ENSO properly, and so the connection between THC and ENSO should be viewed with some caution.

3.1.4. Are Simulated Global Changes in Response to the THC Consistent With Observations?

[74] Because of the prominence of the idea of the THC forcing abrupt climate change, let us address whether the model-simulated changes can explain the observations. The first question to ask is whether THC changes can explain the climate changes recorded in Greenland. While it is clear that when the THC is weakened, there is significant cooling throughout the North Atlantic region (Figure 8), most of that cooling occurs over the ocean in both models. The temperature change over Greenland is a maximum of about 4°C in ZD05 and 2°C in VW02, which is smaller than the observed changes that range from 5°C to 10°C. Also, *Stouffer et al.* [2006] have shown that the pattern and magnitude of cooling in the North Atlantic varies greatly among models. It is possible that this more modest simulated cooling is related to the fact that the topography of the Greenland Ice Sheet is not well resolved in these models, or perhaps local topography matters for the temperature at the Greenland summit. Also, there is the possibility that glacial boundary conditions (lower CO_2, more sea ice, and higher ice sheets) would make Greenland summit more sensitive to THC changes. For example, it has been argued by *Denton et al.* [2005] and *Broecker* [2006b] that more extensive winter sea ice during the coldest stadials would amplify the impacts of a THC change both in Greenland and around the globe. We will return to this idea in section 3.2.

[75] Do the observations support the idea of a bipolar seesaw? In addition to a cooling in the North Atlantic, *Stocker* [1998] predicted that a THC weakening should be accompanied by a synchronous warming of the high-latitude Southern Hemisphere. First, it should be noted that the coupled GCMs simulate little change in Antarctica in response to the THC change (Figure 8). A more idealized coupled model does simulate a large warming in Antarctica [e.g., *Knutti et al.*, 2004], but there is little reason to expect that model to have a more realistic simulation of the climate of Antarctica given the simplicity of the atmospheric component of the model. As for observational support for the bipolar seesaw, there does appear to be a warming (though it is not rapid) in Antarctic ice cores that coincides

with Heinrich events, which can be explained as an integrated response to the changes in the North Atlantic [*Roe and Steig*, 2004]. The EPICA core also shows the D-O events [*EPICA Community Members*, 2006], and the EPICA Community Members claim that the magnitude of the warming in Antarctica is proportional to the duration of the THC change in the North Atlantic. It should be noted, however, that the Antarctic record was compared with the GRIP record, not a proxy for THC variations, which are presently equivocal for D-O events.

[76] For the tropics, we will focus here on the distinction between the temperature and hydrologic signatures of the THC response in the tropics. As discussed in section 3.1.3, the models show a temperature change in the tropical Atlantic Basin, with little change outside. A cooling in the subtropical North Atlantic has been observed by *Sachs and Lehman* [1999] and *Bard et al.* [2000]; however the cooling predicted by those studies is larger than that simulated by the model, as pointed out by *Seager and Battisti* [2007]. The 3°C–4°C temperature change in Cariaco Basin is also larger than the model-simulated cooling, although local changes in upwelling that may not be well simulated by the model could potentially enhance the coastal signal. The absence of a large cooling associated with the Greenland events in other parts of the tropics may simply be due to a lack of reliable records. However, at this point the available observations appear to be consistent with the model in that the only clear signal of cooling is in the North Atlantic.

[77] The precipitation patterns in the tropical Atlantic Basin have been quite convincingly linked to THC-driven shifts in the ITCZ [*Peterson and Haug*, 2006; *Seager and Battisti*, 2007]. However, as one moves outside the tropical Atlantic, the consistency between models and observations becomes less compelling. The models simulate a robust increase in precipitation in the southern central Pacific, but there are no data available from that region. Both models simulate changes in precipitation in the western Pacific warm pool during boreal summer, but the pattern is not consistent. The record of *Stott et al.* [2002] and the Sulu Sea records both indicate saltier waters during stadials, which are consistent with the GFDL model decrease in precipitation, but the model signal is quite small (on the order of 10%). However, local salinity (which is how the Stott et al. record is interpreted) is not determined only by local precipitation, as illustrated by *Timmermann et al.* [2005]. Without a more complete analysis of the model results, it is difficult to link salinity changes at this site with large-scale THC-induced changes. Both models predict a weakening of Indian rainfall, which appears to be consistent with *Sinha et al.* [2005], though this record only covers the Bølling-Ållerød (15.2–11.7 ka). Neither model predicts much precipitation change over Southeast Asia where the speleothem records of *Wang et al.* [2001] and *Yuan et al.* [2004] are from, nor do they predict much change over Oman as do *Fleitmann et al.* [2003] or Socotra Island as do *Burns et al.* [2003]. The GFDL model does simulate a significant change in the southwesterly surface winds, which would be consistent with the change in productivity records from the Arabian Sea [*Schulz et al.*, 1998; *Altabet et al.*, 2002; *Gupta et al.*, 2003], but this does not appear in HadCM3. Thus it seems that the real-world events are more widespread in their monsoon response than the models simulate.

3.2. Sea Ice

[78] Several papers have suggested that a potential driver for abrupt climate change is sea ice. Sea ice has two different and strong influences on climate. The first is through its albedo, which can alter both the local and the global energy budget [*North*, 1984], and the second is through insulating the ocean from the atmosphere by cutting off the exchange of heat and moisture, thereby having a strong influence on the local climate. Here we review the studies suggesting sea ice as an important mechanism in abrupt change, again with attention to whether they can explain climate events that are abrupt, persistent/recurrent on millennial time scales, and have a global influence.

3.2.1. Abruptness

[79] The key to the possible abrupt behavior of sea ice is that (1) there is a temperature threshold for its formation and melt and (2) sea ice can melt and advance rapidly because of albedo feedback. *Gildor and Tziperman* [2001, 2003] further argued that there is another possible sea ice "switch," which is that as sea ice grows, the ability of the ocean to lose heat to the atmosphere is cut off, and at some point it cannot cool sufficiently, so it stops growing. Hence, Gildor and Tziperman argue that sea ice has both "on" and "off" switches. As such, it has been argued that sea ice can respond abruptly to changes in THC or atmospheric circulation and amplify the perturbation both locally and perhaps globally.

[80] *Kaspi et al.* [2004] used a box model to demonstrate how ice sheet dynamics, ocean, atmosphere, and sea ice can interact to produce Heinrich-like events. There are three phases of the oscillation that represent different states of North Atlantic sea ice cover; they are shown in Figure 13. Following a meltwater event (which in their formulation is caused by the "purge" phase of an ice sheet), the THC reduces and hence sea ice extends farther equatorward with a large reduction in high-latitude temperatures. Once the meltwater event ends, the THC begins to recover and melt back the sea ice, but the albedo and insulating effects of the reduction in sea ice amplify this, causing widespread melting and ultimately resulting in a minimum sea ice state. With sea ice extent at a minimum, the ocean temperature begins to cool because of heat loss to the atmosphere, which eventually leads to a readvance of the sea ice back to its initial levels. While highly simplified, this model nicely illustrates some of the key interactions between the ice sheets, ocean, atmosphere, and sea ice. *Gildor and Tziperman* [2001] used a similar model to argue that self-sustained oscillations involving these feedbacks can arise on 100 ka time scale and also reproduce the asymmetric, "sawtooth" pattern in glacial cycles. They did not find, however, that such oscillations would arise on millennial time scales. Thus, sea ice plays a key role in generating the abrupt swings as

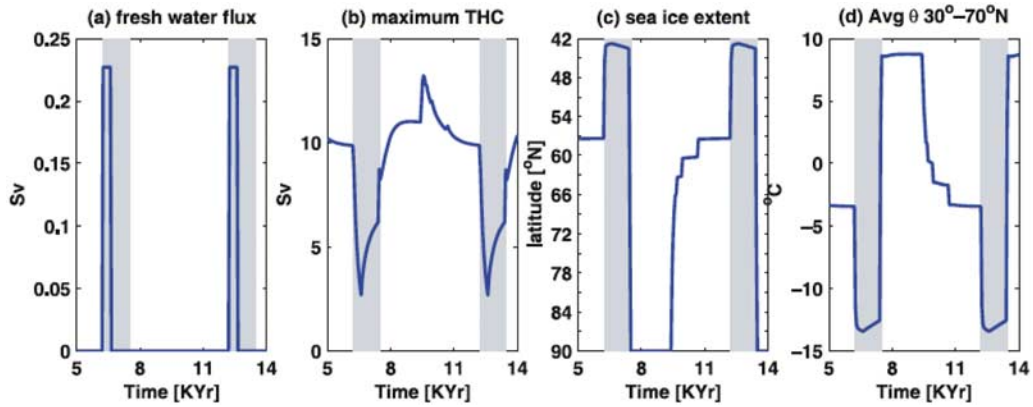

Figure 13. Three sea ice state mechanisms for the Heinrich cycle in a coupled model with a continuous meridional resolution. Periodic freshwater (glacier) discharge is specified in this model, and the response of the ocean, sea ice, and atmosphere is calculated and shown here. A detailed model description is provided by *Kaspi et al.* [2004, Appendix A]. This model experiment demonstrates that the triple sea ice state response to glacier discharges, which occurs in a simpler box model, also occurs in this fuller continuous model. Model time series for (a) the specified freshwater (glacial discharge) input into the northern region of the ocean model, (b) the maximum THC in the Northern Hemisphere, (c) sea ice extent in the Northern Hemisphere (latitude of the southernmost point covered with sea ice), and (d) averaged atmospheric temperature from 30°N to 70°N showing the atmospheric response to sea ice switches. The cold periods are highlighted by gray bands. From *Kaspi et al.* [2004].

the system adjusts itself back to an initial state, but the land-based ice sheets are required as a "trigger."

[81] Do these feedback mechanisms play out in more complete models? There is little discussion of sea ice feedback in the freshwater hosing studies cited earlier (VW02, ZD05, and *Stouffer et al.* [2006]). It seems likely that changes in sea ice are responsible for some of the large temperature changes (up to 10°C) over the North Atlantic Ocean (shown in Figure 8), though *Stouffer et al.* [2006] show that the pattern and magnitude of cooling (and perhaps sea ice) in the North Atlantic varies greatly among models. The issue of the quantitative relationship between THC and sea ice and its subsequent effects on regional climate appears to be an area ripe for exploration.

[82] The issue of rapid sea ice change has received considerable attention of late because of the possible effect of present and future increases in CO_2 on Arctic ice cover [*Vinnikov et al.*, 1999; *Stroeve et al.*, 2007]. *Winton* [2006] has explored the behavior of sea ice feedbacks in coupled models under increasing CO_2 and found that in two models (Max Planck Institute and National Center for Atmospheric Research (NCAR)), sea ice disappeared year round. He found that the ice-albedo feedback as demonstrated in simpler models [*North*, 1984] was responsible for rapid changes in sea ice. Thus, it does appear that this feedback is plausible, though further study is necessary to understand how sea ice behaves in these models, not to mention in the real climate system.

3.2.2. Millennial Time Scale

[83] Sea ice mechanisms do not in themselves introduce any millennial time scale behavior into the system, because the adjustment time scale is relatively short (as illustrated in Figure 13). It is possible that sea ice interacts with the ice sheets to introduce longer-term oscillations. *Gildor and Tziperman* [2001] have argued for this on a 100 ka time scale but not on millennial time scales. Aside from that, the only likely form of millennial recurrence comes from glacial dynamics with sea ice acting as a potential amplifier.

3.2.3. Global Linkages

[84] *Li et al.* [2005] argued on the basis of atmospheric modeling experiments that the large magnitude of warming during D-O events in Greenland can be explained by a removal of sea ice in the North Atlantic. They performed experiments with the NCAR CCM3 model with glacial boundary conditions (CO_2, ice sheets, SST, and sea ice) and then changed the sea ice to its modern extent, leaving all other glacial boundary conditions the same. Li et al. find that a reduction in winter sea ice, through both its albedo and insulation effects, can explain the magnitude of temperature and accumulation changes over Greenland during D-O warmings. *Denton et al.* [2005] argue further that wintertime sea ice is necessary to explain not only changes in ice core records from Greenland summit but also the pattern of moraines from the whole circum–North Atlantic region. A wintertime sea ice advance in the North Atlantic (e.g., as hypothesized for the YD) would make winters much colder throughout the region and hence would explain the significant seasonality shown in proxies from around the region.

[85] We have already discussed the study of *Chiang and Bitz* [2005] that showed that changes in sea ice anywhere in the Northern Hemisphere high latitudes could produce a climate change throughout the tropics. Their mechanism is illustrated in Figure 11b.

3.3. Tropical Ocean-Atmosphere Processes

3.3.1. Abruptness

[86] There is some evidence from the modern climate that the tropics are capable of undergoing abrupt change. A shift in surface temperatures in the tropical Pacific occurred in

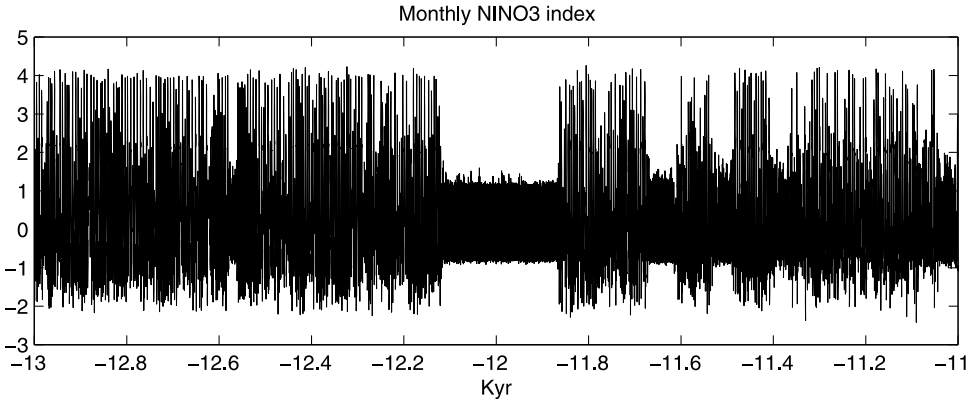

Figure 14. Monthly NINO3 index (the SST anomaly averaged over the region 5°S–5°N and 150–90°W) from an experiment in which the Zebiak-Cane coupled ocean-atmosphere model is forced with orbitally driven variations in solar insolation. How ENSO in the model abruptly locks to the seasonal cycle around the time of the Younger Dryas is shown. From *Clement et al.* [2001]. Reprinted with permission courtesy of the American Meteorological Society.

1976–1977, which coincided with atmospheric circulation changes [*Trenberth and Hurrell*, 1994; *Graham*, 1994; *Deser et al.*, 2004] as well as large ecosystem shifts [*Bakun and Broad*, 2003]. There is considerable controversy about what caused the change and the extent to which it represents a "regime shift." Some argue that it had its origins in the midlatitudes [*Barnett et al.*, 1999], but others have suggested that it originated in the tropics. For example, *Karspeck and Cane* [2002] have shown that the oceanic changes in the tropical Pacific can be reproduced by a model forced only by observed equatorial winds. Regardless of the origin of the wind changes, this result suggests that mechanisms within the tropical Pacific are responsible for the abrupt oceanic changes; that is, the trigger for change is in the tropical winds.

[87] Another mechanism for abrupt change originating in the tropics was postulated by *Clement et al.* [2001]. They examined the response of ENSO to orbital forcing over the last 500 ka using the Zebiak-Cane (ZC) coupled ocean-atmosphere model [*Zebiak and Cane*, 1987]. That model was designed to solve for anomalies in the upper ocean, SST, and surface winds relative to a specified mean state. When forced with the smoothly varying solar forcing, the model showed periods in which ENSO variability was abruptly "shut down" for several centuries (Figure 14). *Clement et al.* [2001] argued that the reason for this shutdown is that the solar forcing modulates the strength of the seasonal cycle, which pushes ENSO between two different modes of variability: one with a strong, regular 3 year period (strong seasonal cycle) and one with a weaker, irregular 4 year period (more like the observed seasonal cycle). In between those "regimes" of variability, ENSO locks into the annual frequency of the forcing, and the internally driven interannual variability is entirely removed. It should be noted that the abruptness of the response in the ZC model is sensitive to the specified drag coefficient.

[88] The ZC model is a highly idealized model, and so the results of *Clement et al.* [2001] could be called into question as arising from an unrealistically simple model. However, a recent study by *Timmermann et al.* [2007a] has shown some similar behavior in a coupled general circulation model. Timmermann et al. analyze an experiment with the ECHO-G model in which the orbital forcing is imposed but sped up by a factor of 100. In other words, 150,000 years of orbital forcing are simulated by running the model for 1500 years. Again, the smoothly varying forcing causes significant, abrupt changes in the behavior of ENSO (Figure 15). The mechanism that Timmermann et al. invoke to explain this behavior is similar to that discussed by Clement et al.: modulation of the strength of the seasonal cycle leads to regime changes in ENSO. Timmermann et al. argue that the specific mechanism is frequency entrainment: when the external forcing (i.e., the annual cycle) is strong, the internal variability (ENSO) becomes locked in phase and frequency to the forcing. The change is abrupt because the system undergoes a bifurcation.

[89] The possibility of abrupt changes is not limited to nonlinear ENSO dynamics. It has been argued that nonlinear behavior in the response of tropical convection to surface temperature can also lead to an abrupt response to a smoothly varying forcing. *Vavrus et al.* [2006] performed experiments with a coupled GCM in which atmospheric CO_2 is gradually increased at a rate of 1% year^{-1}. They find that the temperature increase of the deep tropics does not respond linearly to CO_2 concentrations: there is a modest warming for the first 80 years of the experiment after which the temperature increases much more rapidly. *Vavrus et al.* [2006] argue that the existence of a critical threshold for atmospheric convection in the model introduces this abrupt behavior. Once a large enough area of the deep tropics is convecting, water vapor and cloud feedbacks enhance the warming. They also argue that the impact of this expansion of atmospheric convection is felt outside the Pacific in the Aleutian Low and Pacific jet stream. It should be noted, however, that there is considerable debate surrounding the idea of a "threshold" for tropical convection, both about

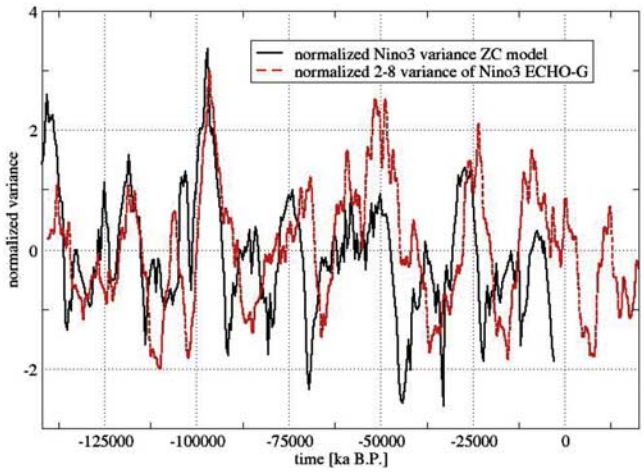

Figure 15. Normalized variance of the NINO3 index simulated by the Zebiak-Cane model forced with orbitally driven changes in the solar radiation [from *Clement et al.*, 1999] over the last 130 ka (black solid line) and normalized variance in the 2–8 year band of the NINO3 index simulated by the ECHO-G model forced with orbitally driven changes in solar radiation over the last 130 ka and the next 10 ka (red dashed line). See text for details of the model experiment. From *Timmermann et al.* [2007a]. Reprinted with permission courtesy of the American Meteorological Society.

whether one exists and how it would vary with climate change [*Dutton et al.*, 2000]. The way in which the model used by Vavrus et al. simulates the onset of tropical convection could be called into question on some of these grounds. Nevertheless, this type of experiment is useful in illustrating that there may be threshold behavior in the tropics that could introduce abruptness into the tropical climate system.

[90] The mechanisms for abrupt behavior in the tropical climate have thus far been illustrated with examples of an abrupt response to a smoothly varying forcing. However, it is not clear what that forcing would be for the abrupt events of the last glacial cycle. Thus far, it has not been shown how abrupt changes could arise from internal processes within the tropics.

[91] Are multiple stable states possible for the tropical climate? Thus far, this idea has been essentially untested. *Seager and Battisti* [2007] raise the possibility that there are two possible stable states for the subtropical jets, and a switch between those states would have large implications for regional and, possibly, global climate. Another interesting perspective on this issue comes from work with an ocean GCM by *Fedorov et al.* [2004]. They showed that perturbing the freshwater fluxes at the ocean surface can dramatically alter the structure of the tropical thermocline. For a strong freshening of the high-latitudes, the tropical thermocline deepens and approaches a state in which the tropics are permanently warm. This approach is also associated with a bifurcation between a state that is symmetric about the equator and one with north-south asymmetries. These results are provocative in that they suggest the possibility of multiple stable states in the tropical ocean. However, it is unclear whether such states would be maintained when coupled to the atmosphere. For example, *Hazeleger et al.* [2005] have argued that negative feedbacks between the ocean and atmosphere make the tropical climate resistant to change (and hence unlikely to be easily driven into a different stable state). Hazeleger et al. illustrated this idea with experiments showing that if, for example, the SSTs were to increase on the equator, the atmospheric circulation would speed up. The ocean response to this change in atmospheric circulation would be to export more heat from the equatorial region via an increased meridional overturning circulation, which would damp the original SST anomaly: hence a negative feedback. On the other hand, *Fedorov et al.* [2004] point out that salinity is less subject to such negative feedbacks. Perhaps the hydrological cycle coupled with ocean density structure can lead to some unexpected behavior in the tropical climate. Targeted experiments with coupled GCMs have not yet been performed to address this issue.

3.3.2. Millennial Time Scale

[92] The millennial time scale is perhaps the area in which ideas on the role of the tropics in abrupt climate change are least convincing. The main reason is simply that unlike high latitudes where glacial dynamics may introduce long time scales, there are no individual components of the tropical climate system that have such a long time scale. Nonetheless, *Clement and Cane* [1999] gave an example of how millennial variability could be generated internally in the tropics using the Zebiak-Cane model. The model was run for 150,000 years with no external forcing. The spectrum of the NINO3 index (the SST anomaly averaged over the region 5°S–5°N and 150–90°W) showed considerable power at century to millennial time scales. This behavior was tested for statistical significance against a null hypothesis of a linear damped oscillation. The coefficients of an AR(2) model that provide the best fit to the model NINO3 index were found. Then, the AR(2) model is run with noise to generate a 150,000 year time series. Comparing the power spectra of the ZC model and the linear, damped oscillator, it was found that the ZC model generates variability that is distinguishable from the linear damped oscillator at time scales of 100 years and longer. Because this model only simulates coupled interactions in the tropical Pacific, and these interactions are nonlinear, this result suggests that it is possible for such physical processes to generate millennial variability. While Clement and Cane did not determine the particular mechanisms in the model that give rise to the low-frequency variability, it has been shown by *Lorenz* [1991, p. 10] that

> . . . in chaotic dynamical systems in general, very-long-period fluctuations, much longer than any obvious time constants appearing in the governing laws, are capable of developing without the help of any variable external influences.

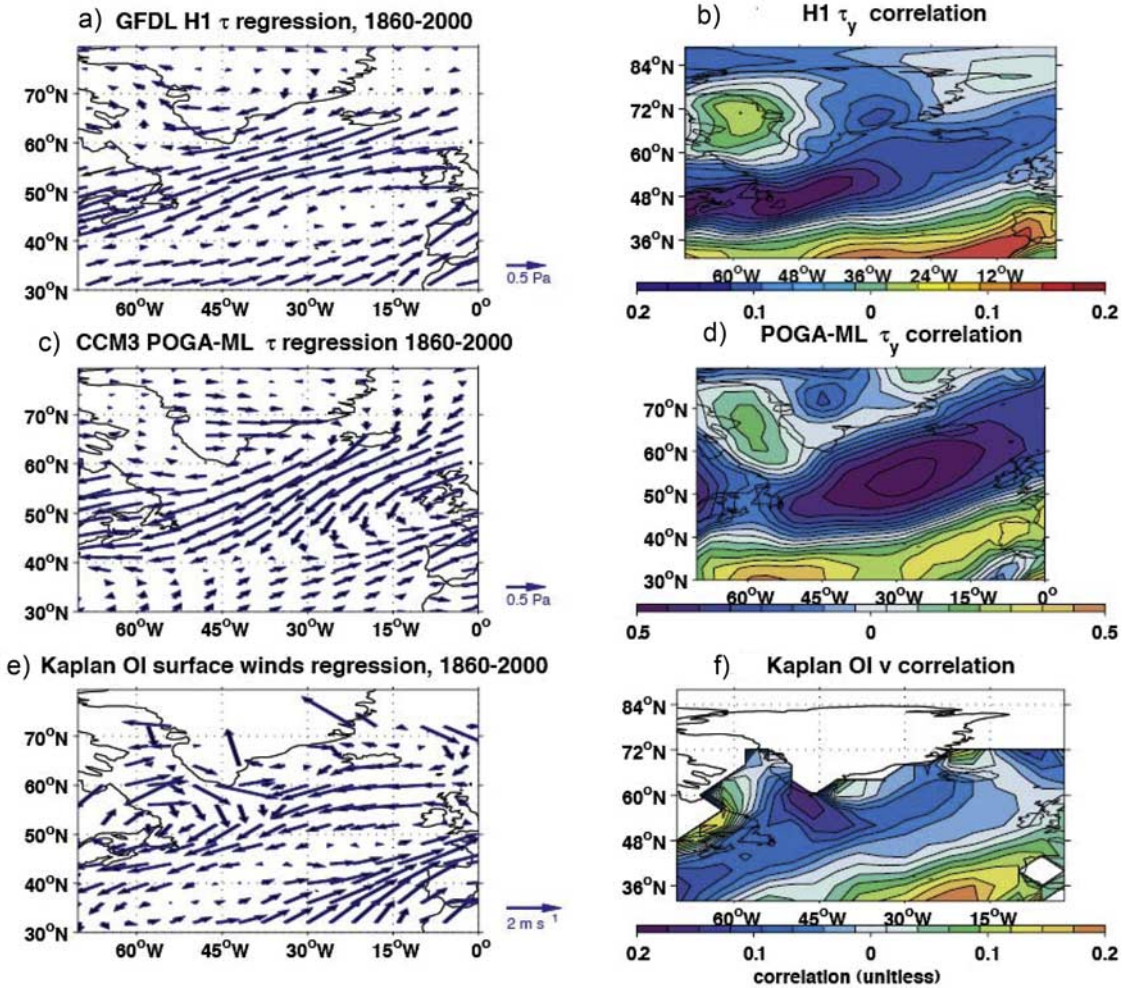

Figure 16. ENSO influence over the North Atlantic. (left) Regression patterns of wind vectors from the specified product, smoothed by a 3 month running average, on the NINO3 index, normalized to unit variance. Hence, units of regression coefficients are given per standard deviation of the index. (right) Corresponding correlation patterns, shown for the meridional component only. (a) GFDL H1 surface wind stress regression. (b) GFDL meridional wind stress correlation. (c) Pacific Ocean Global Atmosphere–Mixed Layer (POGA-ML) surface wind stress regression. (d) POGA-ML meridional wind stress correlation. (e) Analysis of International Comprehensive Ocean-Atmosphere Data Set (ICOADS) data surface wind regression. (f) Analysis of ICOADS data meridional wind correlation. From *Emile-Geay et al.* [2007].

These results are, of course, subject to the criticism that the real world may not behave as a "chaotic dynamical system" as do the ZC and Lorenz models, and there is no good empirical test of what regime the real-world tropical climate system is in. A good example of this is the fact that there is considerable debate as to whether the ENSO system (perhaps the most well-observed and extensively modeled phenomenon in the climate) is a nonlinear self-sustaining oscillation or whether it is a linear, damped system that is stochastically forced [*Thompson and Battisti*, 2001]. It should be noted that the spectral properties of the tropical climate at low frequencies have simply not yet been studied in more complete models. It would be useful to have multimillennia control simulations with coupled GCMs to study such an issue, but this may not yet be possible because of the computational cost of such experiments.

3.3.3. Global Linkages

[93] There is, however, clear evidence that changes in the tropics can exert a global influence on the climate. The most well-known and well-observed example of this is at interannual time scales where SST changes in the tropical Pacific associated with ENSO impact regional climates around the world [*Ropelewski and Halpert*, 1987]. There is also evidence for an influence of tropical SSTs on the high latitudes on longer time scales. *Hoerling et al.* [2001, 2004] used experiments with an atmospheric GCM to show that climate changes in the North Atlantic since 1950 can be linked to a warming of the tropical oceans over that time.

[94] One strong control from the tropics on the global climate is through the energy budget. To balance the large radiative energy gain in the tropics, some energy is radiated

locally to space, and the residual is exported to and warms the high latitudes. *Pierrehumbert* [1995, 2000] suggested that one of the main controls on the tropical radiative balance (and hence what is available for export) is the water vapor content of the subtropical free troposphere (i.e., above the convective boundary layer). First, he argues that the existence of this dry region is essential to maintain energy balance in the tropics. Without these regions in which the clear-sky outgoing longwave radiation (OLR) is high, the tropics would be in a state of runaway greenhouse. The clear-sky OLR is related to both the temperature and water vapor content of the atmosphere. If, for example, a surface warming were to moisten the subtropical atmosphere, this would lead to a reduction in OLR and provide a strong amplifier for the warming. There are still large uncertainties in what determines the water vapor content of the subtropical troposphere [*Held and Soden*, 2000], but it is now quite clear from satellite observations that water vapor is acting as a positive feedback in the global mean warming of the last several decades [*Soden et al.*, 2005]. It has also been shown that a transition from a La Niña– to an El Niño–like mean tropical state does significantly increase the net amount of water vapor in the troposphere contributing to net warming of the Earth's surface [*Barreiro et al.*, 2006]. These ideas suggest that a perturbation to the tropical climate that influences the water vapor content could be strongly amplified and lead to a globally synchronous warming or cooling.

[95] Another way to perturb the tropical energy balance is via changes in subtropical stratocumulus clouds. These clouds have a significant cooling effect on the climate because they have a high albedo, but they do not affect the OLR much because they are low and have cloud top temperatures that are close to the surface temperature [*Hartmann et al.*, 1992]. A perturbation to the tropical climate that alters the amount or spatial extent of these clouds would also lead to a global cooling or warming [e.g., *Miller*, 1997; *Clement and Seager*, 1999; *Clement and Soden*, 2005; *Barreiro et al.*, 2006]. However, it is not well known what controls the low cloud cover even for the modern climate, and it remains difficult to simulate these clouds in climate models [*Bretherton*, 2007]. Another limitation to the arguments suggesting a potential role for water vapor and low clouds is that those processes are likely to affect the tropical or global mean temperature, yet it remains to be seen to what extent the abrupt events of the last glacial are associated with a mean temperature change.

[96] A second idea derives from the example of modern teleconnections from the tropics [e.g., *Ropelewski and Halpert*, 1987], but this phenomenon is explored under glacial climate conditions during which the largest known abrupt climate changes occurred. *Yin and Battisti* [2001] performed a number of experiments with an atmospheric GCM in which the glacial SSTs were specified at values indicated by *CLIMAP Project Members* [1981]. The LGM boundary conditions for ice sheet elevation and extent (from the *Peltier* [1994] reconstruction), the seasonally varying insolation (calculated for 21 ka B.P.), and the concentrations of greenhouse gases (including a LGM CO_2 concentration of 200 ppm) were also prescribed. Yin and Battisti found that slight changes in the pattern of SST change in the tropical Pacific had a large impact on the temperature at the southern boundary of the Laurentide Ice Sheet and on its annual mean mass balance. Because meltwater from the Laurentide Ice Sheet is often suggested as playing a role in NADW formation, it could be argued that this is a mechanism for the tropical Pacific to influence the North Atlantic Ocean. However, Yin and Battisti point out that this large sensitivity to SST pattern, in particular SST gradients, implies that the spatial pattern of SST change must be known in detail in order to evaluate its impacts outside the tropics. As noted in section 2, the current paleodata coverage in the tropics is far from being able to address this issue. Hence, all that can be said at this point is that it is physically plausible that changes in the tropical SST influence the Laurentide Ice Sheet mass balance.

[97] Another teleconnection has been postulated linking the tropical Pacific and North Atlantic wind patterns. *Emile-Geay et al.* [2007] have shown that there is a statistically significant correlation between ENSO and winds flowing out of Fram Straits in the wintertime. The observed behavior is also captured in atmospheric GCMs forced by observed SST, though the simulated connection appears to be somewhat stronger than in observations (Figure 16). *Bond et al.* [2001] argued that it is the winds over Fram Straits that drive sea ice out of the Arctic, causing large-grained sediments (IRD) to be deposited in the open North Atlantic and coinciding during the last glacial with abrupt dips in Greenland temperatures. This atmospheric bridge represents another pathway by which the tropical Pacific can alter the North Atlantic climate, in particular the ocean, which should be further explored.

[98] The tropical Pacific can also influence the North Atlantic Ocean through freshwater transport in the atmosphere. *Schmittner et al.* [2000] show that there are sizable interannual variations in the freshwater export from the tropical Atlantic. In particular, during El Niño events, more fresh water is removed from the tropical Atlantic Basin. The exact pathway of this anomalous transport was not identified and is difficult to show because of the lack of complete observations of the large-scale circulation and distribution of water vapor. However, it is well known that there is a large flux of water vapor carried by easterly airflow over the Isthmus of Panama [*Zaucker et al.*, 1994], and this is a likely exit route for Atlantic water vapor. It has been argued that the changes in freshwater export have implications downstream for NADW formation. That is, if the mean state of the tropical Pacific were to shift to a more El Niño–like state, the atmosphere would export more water vapor from the tropical Atlantic, which would become saltier. This would then lead to an increased vigor of deepwater formation downstream in the North Atlantic. *Latif* [2003] suggested that such a mechanism was operating in increasing CO_2 experiments with the ECHO-G coupled ocean-atmosphere GCM. He pointed out that under increased CO_2, the mean state of the tropical Pacific in that model becomes

more El Niño–like and that a subsequent increase in salinity of the tropical Atlantic offsets the tendency of models to reduce NADW formation under global warming, leading to no change in that model's THC. *Schmittner and Clement* [2002] use a simplified ocean model to show that a persistent La Nina–like change in the tropical Pacific (which leads to a fresher Atlantic) could lead to a shutdown of NADW formation within 70 years.

3.4. Interactions Between the Tropics and High Latitudes

[99] In this section, we take a different approach to the problem of abrupt climate change. Thus far, each of the mechanisms has been tested essentially as a "driver" of abrupt change; that is, if that mechanism were to produce an abrupt change at the appropriate time scale, how would that particular change influence the rest of the globe. However, because evidence of abrupt, millennial-time scale changes appears in regions throughout the globe and in both oceanic and atmospheric quantities, it is perhaps more appropriate to think about these processes as being components of a global, coupled feedback. The focus thus far has been on either high-latitude processes (deepwater formation and sea ice) or low-latitude processes. Here we lay the groundwork for examining interactions, or feedbacks, between the low and high latitudes.

[100] One aspect of low-to-high-latitude coupling that has been explored is the linkage between the salinity of the tropical Atlantic and the THC. It is well known that the high salinity of the tropical Atlantic preconditions the surface waters for sinking downstream in the northern Atlantic [*Warren*, 1983]. The studies of THC shutdowns, however, show that the state of the THC can influence tropical rainfall patterns (Figure 9). *Vellinga and Wu* [2004] and *Krebs and Timmermann* [2007a, 2007b] have shown that the southward shift of the ITCZ in response to a THC shutdown causes a large increase in the net evaporation minus precipitation, which leads to a saltier northern subtropics. This signal is advected into the North Atlantic and leads to a rapid recovery of the THC. It is important to note that this is a negative feedback loop, which could lead to centennial variability [*Vellinga and Wu*, 2004; *Knight et al.*, 2005] but actually limits the time scale of a THC shutdown to centuries. This mechanism would have to be overcome in order for a climate event to persist for centuries to millennia as suggested by the observations.

[101] Exploration of the mechanisms must therefore reveal positive feedbacks that will drive the climate into a different state (not restore it to its initial state) in order to explain these abrupt events, and we suggest that such feedbacks may exist between the low and high latitudes. *Seager and Battisti* [2007] have hypothesized that there are possible positive feedbacks in the atmosphere between the low and high latitudes within the Atlantic sector. They suggest that a southward shift in the ITCZ (caused, for example, by a reduction of the THC) could affect the trajectory of the midlatitude storm tracks in the North Atlantic. If the storm tracks were to become more zonal, this would lead to a reduction of the warm, salty water flowing into the northern North Atlantic, preconditioning it for deepwater formation there. A more zonal jet would also cool Europe and allow ice to advance equatorward, which could explain some of the other records of abrupt climate change in the region. How this interaction could produce a global event including a widespread reduction of net precipitation in the Asian monsoon region, however, remains to be seen.

[102] *Fedorov et al.* [2006] put some of the global ocean-atmosphere linkages together in a review of the Pliocene paradox. They start by noting that while the atmospheric carbon dioxide was only slightly higher during the Pliocene than today and other boundary conditions such as the continental configuration and the incoming solar radiation were basically the same as present, the Earth was as much as 3°C warmer on average, and there was very little ice at high latitudes. Fedorov et al. then gather a collection of experiments with atmosphere and ocean (though not coupled) models to illustrate how processes connecting the low and high latitudes can give rise to an altered mean climate state under similar external forcing. That state in the Pliocene, they hypothesize, was a permanent El Niño state.

[103] Their argument centers on the mean depth of the equatorial thermocline: the shallow thermocline of the modern climate allows the ocean to gain heat in the equatorial cold tongue because evaporation does not balance the incoming solar radiation. *Boccaletti et al.* [2004] showed using an ocean model that the tropical thermocline adjusts in order to satisfy a balanced heat budget (energy going into the ocean must equal that going out in equilibrium). That is, when heat loss at high latitudes is reduced, the equatorial thermocline deepens so that the ocean does not gain as much heat in the tropics. *Fedorov et al.* [2006] argue that in the Pliocene, reduced heat loss at high latitudes (linked to the warmer temperatures) would lead to a deeper thermocline and a permanent El Niño state. They argue further that a permanent El Niño state would be associated with reduced stratus decks and increased atmospheric water vapor, both of which would warm the climate, thereby providing the feedback to conditions at high latitudes. Another way for the tropical thermocline to be altered has been suggested by *Fedorov et al.* [2007]. They use ocean GCM experiments to show that a freshening of surface waters in high latitudes can reduce the meridional density gradient between the equator and the subtropics and modify the wind-driven circulation with which the ventilated thermocline is associated. The reduction in the meridional density gradient leads to the deepening of the tropical thermocline and consequently to reduced zonal (east-west) SST gradient along the equator.

[104] While this is a truly global approach to understanding long-term climate changes, there are some places where these arguments need to be fleshed out. First, and most important, these ideas have not been tested in a coupled GCM. Thus, while the ocean may be happy with a deeper mean thermocline, coupling with the atmosphere may not allow for this state. Also, other studies have shown that in

the extreme case where the ocean transports no heat poleward, the climate is actually colder than present not warmer [*Clement and Seager*, 1999; *Winton*, 2003; *Herweijer et al.*, 2005]. The arguments advanced in those papers were that stratus decks are increased and atmospheric water vapor reduced when the ocean transports no heat, leading to a cooler climate, opposite to what is suggested by *Fedorov et al.* [2007]. These issues will only be sorted out when they can be tested in carefully designed experiments with coupled GCMs.

[105] Without a specific forcing, however, it is difficult to design targeted experiments with CGCMs. One possibility is to simply run the state-of-the-art models for multiple millennia with constant external forcing and see whether the models produce millennial variability similar to the observed. For the present version of the fully coupled models, this is simply not feasible because of the computational cost. Another approach is to perturb the model in some way and study the transient response. The freshwater forcing experiments that have been extensively performed are only one way of perturbing the system, and given the uncertainty about the evidence for and causes of meltwater events, this may not even be the most relevant perturbation. *van der Schrier et al.* [2007] provide an example of another way to perturb the system in order to test a specific hypothesis. They perform experiments with a simplified coupled model in which the position of the subtropical jet in the North Atlantic is strengthened artificially. The model response to this includes an advance of the sea ice margin and cooling in the North Atlantic. Other hypotheses could be tested within these types of experimental frameworks, such as the coupled climate response to a deeper equatorial thermocline, a weaker Asian monsoon, and more extensive North Atlantic sea ice to name a few. While computationally expensive, we suggest that the use of CGCMs in this way is likely to provide important information about global coupled feedbacks within the climate system, much as the pioneering work of *Manabe and Stouffer* [1988] did for the THC.

4. SUMMARY AND CONCLUSIONS

[106] Our review of the paleoclimatic evidence suggests that despite large spatial gaps in the data and limitations related to the age control for different proxies, there is considerable evidence that abrupt, millennial-time scale climate events in Greenland coincide with changes in many other parts of the world. Climate signals in the tropical Atlantic appear to be consistent with changes in the hydrological cycle and, in particular, a southward shift in the latitude of the ITCZ coinciding with cooling in Greenland. In records with sufficient resolution, these changes do appear to show the same abrupt character as those in Greenland. While data are sparse across most of the Pacific, there are several records from the Asian monsoon region that indicate that changes there coincide both in timing and abrupt character with events in Greenland as well. Because many of these proxies are likely related to local hydrological quantities that have strong small-scale spatial variability, it is difficult to distinguish between a change that is abrupt in time and a spatial shift. Nevertheless, the fact that abrupt events can be identified in different proxies and in different areas affected by the monsoon is suggestive that there is a large-scale change coinciding roughly in time with D-O events in Greenland. The Southern Hemisphere evidence for abrupt change remains quite unclear. There do appear to be millennial-time scale events in Antarctic ice cores that coincide roughly in time with the Greenland events, but they are not abrupt. Land and ocean records from the rest of the Southern Hemisphere are few and far between and generally do not have the appropriate resolution to characterize whether changes are abrupt.

[107] What are the mechanisms that can give rise to these features of the paleoclimate record? There is certainly support for a major role for the THC in abrupt climate change. There is paleoclimatic evidence for meltwater events flooding into the North Atlantic, and models suggest that realistic values of freshwater input could change the THC, even shutting it down completely. The spatial response to a THC shutdown simulated by models appears to be consistent with paleoclimate evidence for temperature and hydrologic changes in the tropical and North Atlantic, and models predict a weakening of the Asian monsoon, which is also generally consistent with observations. However, there are still major unresolved issues with the details of these statements, which we enumerate here:

[108] 1. It is still unknown what causes the meltwater pulses. *Hemming* [2004] suggested three different possibilities that apply to Heinrich events based on a review of the literature. All are basically internal to the glacial system and do not require any external forcing. She notes that there is little known about how long the meltwater pulses may last, and given the linearity of the model-simulated THC response to the freshwater forcing, this seems to be an important area to develop for constraining the time evolution of the THC change. In the case of the Younger Dryas, although there is a clear mechanism for producing a meltwater pulse into the North Atlantic, there is only limited geologic evidence for it. Additional records from the ocean [e.g., *Carlson et al.*, 2007] will no doubt shed light on this issue. For D-O events, the story is much less developed, mainly because there is weak evidence for meltwater pulses on this time scale. Thus, while plausible, it seems that there is still considerable work to be done on this first link in the chain. One way forward would be to incorporate glacial dynamics into coupled climate models. This would be of benefit not only for the study of past abrupt climate changes but also for understanding the future climate response to greenhouse gas forcing. Some modeling centers are already putting efforts toward this. Such modeling efforts could lead to progress in understanding the link between glacial dynamics and ocean circulation, but they would also be of use in testing ideas related to the connection between modes of atmospheric circulation and glacial dynamics, such as suggested by *Wunsch* [2006].

[109] 2. There is as yet incomplete evidence linking THC changes with D-O events in Greenland. As noted in section 2.1.2, it may be related to the fact that the three-dimensional nature of the ocean's deep circulation makes it inherently more complex to reconstruct than the past properties of the surface ocean, which is essentially a two-dimensional problem. However, the more compelling evidence for YD and H changes in THC does beg the question of whether what happens during those events is fundamentally different than for D-O events.

[110] 3. Regarding the spatial response to THC change, there are a few key issues that need further attention. First, the simulated magnitude of cooling over Greenland does not appear to be consistent with the model response. This may be rectified by inclusion of more extensive sea ice (as suggested by *Denton et al.* [2005]), but this has not been shown in a quantitative way in a coupled model. Second, the story in the Southern Hemisphere is as yet unclear. While there is evidence for warming (though not abrupt) in parts of Antarctica that coincide with H and D-O events, there are large parts of the Southern Hemisphere that are not sampled. The model-simulated response to a THC shutdown is warming primarily in the South Atlantic Ocean, with little warming elsewhere in the Southern Hemisphere and none over Antarctica. Last, a THC shutdown does appear to weaken the Asian monsoon, but the simulated change does not appear to be as widespread as the data suggest. Also, the mechanisms linking the THC and the monsoons are as yet unclear, and further work is necessary to understand this remote linkage. Finally, all of these spatial patterns of the response to a THC shutdown may well change for a colder mean state, and meltwater experiments should be performed in coupled models with an equilibrated glacial state.

[111] 4. The final issue concerns the abrupt warmings. Thus far, THC changes have been invoked to explain abrupt cooling in the North Atlantic. In models, when the freshwater input stops, the THC gradually recovers, showing no abrupt behavior. This could be different if there were multiple stable states for the THC, but that has not yet been shown with the latest generation of coupled climate models.

[112] Feedbacks involving sea ice are another mechanism that has appeal for producing abrupt change because of a positive albedo feedback, its strong influence on air-sea heat and moisture exchange, and possible "on" and "off" switches. However, sea ice also requires some forcing to cause it to change, and those that have been suggested in the literature are THC and atmospheric circulation changes. Thus far, the interaction between sea ice and ocean and atmospheric circulation has been studied only in simplified models, yet these ideas clearly need to be tested in more complete models. Another thorny issue regarding sea ice is the lack of any direct proxy for it; this is clearly an area where proxy development would be of great value. Again, better quantitative understanding of the controls on sea ice for past climate changes (both observational and modeling perspectives) will certainly be of benefit to understanding the ongoing, and possible future, changes in Arctic sea ice.

[113] The strength of the arguments for a role for tropical processes in abrupt climate change is their ability to influence the global climate. Small changes in the tropics can potentially alter the radiation budget of the planet, the hydrological cycle, the atmospheric circulation, and even possibly the THC. While there are some mechanisms that could produce abrupt changes in the tropical climate, they have only been demonstrated in response to some forcing (e.g., orbital changes or CO_2). One example for generation of millennial-time scale variability by tropical processes was given [*Clement and Cane*, 1999], but it was demonstrated with a highly idealized model. It is as yet unclear whether this would occur in a more complete model. Nevertheless, given the paleoclimate evidence from the Indo-Pacific region, any mechanism to explain abrupt change must account not only for what processes may influence the monsoon but also what the impact of a change in the monsoon would have on the rest of the globe. Furthermore, the absence of high-resolution records from the tropical Pacific Ocean leaves open questions about the role of coupled interactions (akin to ENSO) in abrupt change. Given the importance of this region in generating modern variability, it seems clear that this region should not be overlooked.

[114] Finally, given the potential role for processes occurring in both low and high latitudes, we suggest that a global approach is necessary for understanding the problem of abrupt change. Coupled GCMs certainly offer this kind of perspective, but they have been used only in limited applications to this problem, primarily in studies of the climate response to freshwater forcing in the Atlantic. While this has been useful, there are other ways to perturb the climate (e.g., different initial conditions or forcing persistent changes in particular phenomena) that may help to reveal the global-scale coupled feedbacks that can cause the climate to change abruptly around the globe.

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