Chapter Two

Challenges to our understanding of the general circulation: abrupt climate change

1

Richard Seager and David S. Battisti

2.1 INTRODUCTION

About 14,700 years ago (14.7kyr BP), towards the end of the last ice age, the climate warmed dramatically and abruptly around the North Atlantic- by as much as the difference between full glacial and interglacial conditions - in no more than a decade or two. This is all the more remarkable because it occurred in the presence of massive ice sheets and continuation of the albedo forcing that presumably had been helping maintain glacial conditions up to that point. But it was not to last. Sometime just after 13kyr BP this Bolling/Allerod wam period ended as climate first cooled, and then abruptly cooled, into the so-called Younger Dryas. As near glacial conditions returned, glaciers advanced around Europe and the forests that had established themselves in the preceding warm epoch died. The Younger Dryas ended with a second abrupt warming, that occurred over only a decade or so, that shifted temperatures back to those of the Holocene and of today.

The idea that the climate system goes through such abrupt shifts did not take the climate research community by storm but dribbled into acceptance in the 1980s and the early 1990s. Only when duplicate ice cores said the same thing and the evidence was found in multiple quantities within the ice - oxygen isotopes, dust concentrations, snow accumulation and so on - and it could be correlated with terrestrial and marine records did acceptance that abrupt climate change was a reality sink in.

This gradual acceptance is telling. When Hays et al. (1976) showed just how well climate records from deep sea cores could be matched to orbital cycles it was deeply satisfying: the gradual waxing and waning of the great ice sheets could be explained by equally gradual changes in the distribution of delivery of solar radiation to the Earth's surface. Insolation over high northern latitudes was deemed to be particularly important with reduction in summer leading to retention of winter snow and ice sheet growth. All that remained was to show exactly how the climate system accomplished the necccesary links.

Almost three decades later we are still far from understanding how orbital changes are converted into ice sheet growth and decay. While this is testimony enough to our limited understanding of the climate system and general circulations, abrupt climate change is now the star witness. In this case the climate changes occurred not only abruptly but, apparently, in the absence of any external forcing. The lack of any theory for how such changes could occur helps explain the gradual acceptance of what the data was saying.

In the two decades since the discovery of abrupt change two advances have been made. First the spatial pattern of abrupt climate change has been better delimited and it is now known that these events occurred essentially synchronously across much of the northern hemisphere (including the northern tropics) within the atmosphere, the surface ocean and the deep ocean. Abrupt changes are not found in the ice records from Antarctica and the southern hemisphere remains a question because of limited data. These spatial patterns place some severe constraints on proposed mechanisms of abrupt climate change. Second, mechanisms have been advanced that revolve around the thermohaline circulation (THC). Broecker et al. (1985) were perhaps the first to suggest that rapid warmings and cooling of climate around the North Atlantic were caused by rapid switchings on and off of North Atlantic Deep Water formation with 'on' states being associated with transport of warm waters into the subpolar North Atlantic. Despite difficulties explaining the paleoclimate record of abrupt changes with the THC theory, no competing idea has yet been offered.

The paleoclimate record poses many challenges to our understanding of the general circulation of the atmosphere and ocean, of which explaining abrupt change is just one. How orbital changes cause ice sheet growth and decay remains a major unsolved problem but will be probably be solved as it becomes computationally feasible to integrate coupled atmosphere-ocean general circulation models (GCMs) through orbital cycles. It is only in recent years that time snapshot simulations of the Last Glacial Maximum (LGM) with coupled GCMs have become commonplace (Shin et al. 2003, Hewitt et al. 2003). As suggested by Ruddiman and McIntyre (1981), it is probably a solar radiation distribution that allows increased winter export of tropical moisture into higher latitudes, there to fall as snow, and reduced summer insolation at high latitudes, which allows the snow to be retained until next winter, that will cause ice sheets to grow.

Further back in time, evidence of equable climates poses an enormous challenge to our understanding of the general circulation and the climate system. A particularly interesting example is the Eocene, when temperatures of high latitude northern continental interiors remained above freezing in winter, allowing crocodilians to survive in subpolar regions. More recently the greening of the Sahara in the mid-Holocene, when the worlds most impressive desert essentially became a moist savannah, remains a fascinating unexplained problem. Certainly it was triggered by orbital changes that increase summer insolation over the Northern Hemisphere but the apparently abrupt onset and demise of the African Humid Period (deMenocal et al. 2000), and the fact that other northern hemisphere monsoon regions show less impressive changes, suggest a non-local coupling between deserts and monsoons on paleoclimate timescales that is waiting to be elucidated.

Baffling though these problems are, the focus in this article will be on abrupt climate changes in glacial times. We will advance a case for an important role for the tropics in climate change. To begin we will review the evidence for abrupt climate change and summarize the current knowledge of the spatial footprint. We will also review evidence for the relationship between abrupt changes in surface climate and deep ocean circulation. This will form the basis for a critique of the THC theory of abrupt change before we advance a case for a mechanism that involves global atmosphere-ocean coupling and an active role for the tropics. This mechanism will be as sketchy as that of Broecker et al. (1985) but hopefully will inspire some future investigations.

2.2 ABRUPT CLIMATE CHANGE IN POLAR ICE CORES

The characteristics of abrupt climate change as recorded in polar ice cores from Greenland and Antarctica have been well described by Wunsch (2003). The problem can be seen in Figure 1, which is reproduced in that paper from the original of Blunier and Brook (2001). The $\delta^{18}O$ content of ice in the Greenland core shows frequent rapid drops and even more dramatic increases throughout the last glaciation. The Younger Dryas appears to have been the most recent of these. The $\delta^{18}O$ content of ice is supposed to be a proxy for the local temperature when the snow fell. Cuffey et al. (1995) used borehole temperature measurements from the Greenland ice core, and models of heat and ice flow, to infer that during the glacial period an 0.33 increase in $\delta^{18}O$ corresponded to a $1°C$ increase in temperature. The $\delta^{18}O$ content in the ice core is, however, also influenced by the isotopic composition of the water that was evaporated and the change in composition due to evaporation/condensation cycling during transport to the ice core. Either way the record shows dramatic climate changes as much as two thirds of the size of the difference between full glacial and interglacial conditions. Using the Cuffey et al. relation the inferred temperature changes are as large as $15 - 20°C$. These changes occurred in decades.

The Antarctic cores do not show any such rapid climate changes. Wunsch (2003) concluded that the two are uncorrelated on the millennial timescale (but correlated on the glacial-interglacial timescale), although others have suggested more complex relationships between the two (Roe and Steig 2004). The timescale of the Antarctic core was adjusted to bring the methane records, also shown in Figure 1, into agreement on the basis that methane is a well-mixed gas. The fact that this can be done indicates that the source regions for methane, especially in tropical wetlands, were affected by the rapid climate changes (Brook et al. 1999) and that the changes were not simply regional North Atlantic events.

Statistical analysis presented by Wunsch shows to be true what the eye perceives, that the Greenland climate is bimodal, having two preferred states which it switches inbetween. The histograms of $\delta^{18}O$ in the two Greenland cores (GRIP and GISP2) show obvious bimodality whereas the Antarctic core (Byrd) is unimodal, though with a long tail. This is striking evidence of nonlinearity and threshold behavior in the climate system, at least in Greenland. Wunsch suggests that this arises from switching between two different evaporative source regions for the water that falls

Figure 1

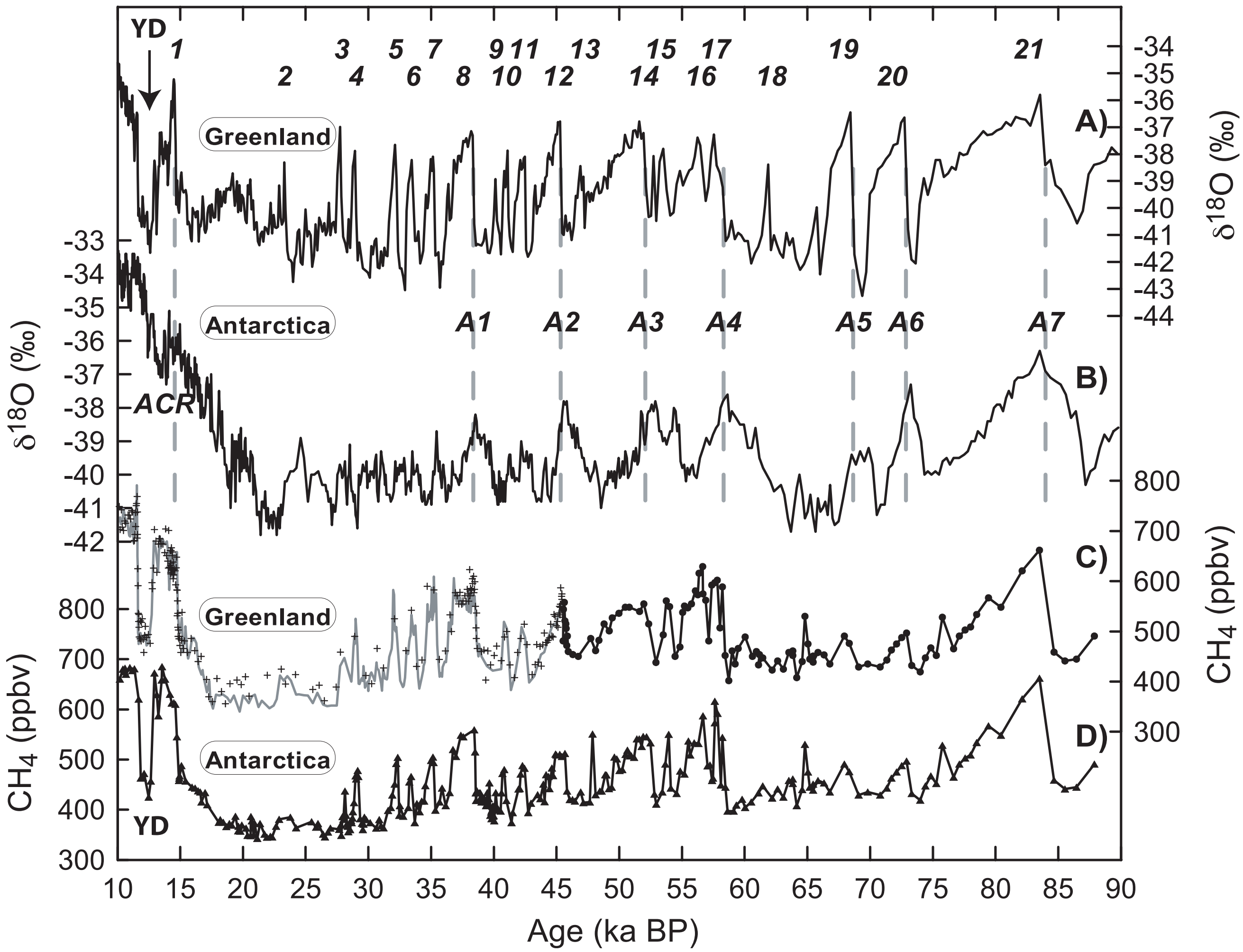

Figure 2.1 (A) $\delta^{18}O$ in the GISP2 Greenland ice core, (B) $\delta^{18}O$ in the Byrd Antacrtic ice core and, (C) and (D), the methane in the GRIP and GISP2 core and the Byrd core, respectively. The numbers at the top denote Dansgaard-Oeschger events in Greenland and the dashed vertical lines are Antarctic warm events. ACR refers to the Antarctic Cold reversal, a modest cooling during the deglaciation in Antarctica which preceded the Younger Dryas, labeled as YD. Taken from Blunier and Brook (2001).

as snow onto the Greenland core location and can reproduce similar behavior with a two source model, some noise and some simple rules for transitioning between sources.

Even were the two source explanation to be true it implies that the atmospheric circulation is capable of switching between circulation regimes with different trajectories of atmospheric water vapor and condensate between the oceans and Greenland. However it does appear that the original interpretion of the $\delta^{18}O$ in terms of temperature has some validity. Using an entirely independent methodology based on the temperature dependence of the diffusion of gases within the ice core, Severinghaus and Brook (1999) and Severinghaus et al. (1998) deduced a warming at the ice surface of as much as $9°C$ in a few decades at the transition into the Bolling warm period, an amount consistent with the Cuffey et al.'s oxygen isotope paleothermometry.

The kind of climate change seen in the Greenland ice cores almost perfectly fits Lorenz's description of an 'almost intransitive' system: 'a particular solution extending over an infinite time interval will possess successive very long periods with markedly different sets of statistics' (Lorenz 1968). Lorenz developed this concept of climate change, which seems more relevant today than ever, even before the ice core data was available!

In summary, abrupt climate changes, consisting of coolings followed by rapid coolings and then abrupt warmings, punctuated the entire glacial period at Greenland but no such thing happened in Antarctica. At Greenland there is enticing evidence of nonlinearity of the climate system with thresholds and switches. The Younger Dryas is the most recent of these abrupt changes. Much of the discussion to follow on the spatial and temporal structure of abrupt changes will concern the Younger Dryas (which we have labelled on Figure 1), because this is the best observed, but it is anticipated that the descriptions are generally valid for all the abrupt changes.

2.3 ABRUPT CLIMATE CHANGE IN THE SURFACE ATLANTIC OCEAN

Were abrupt climate change limited to Greenland it would not pose too much of a challenge to our understanding of the general circulation. It would be easy to imagine some alteration of the circulation, and movement of heat by stationary and transient eddies, that could accomplish the observed effects. However, as time has gone by, changes elsewhere in the global tropics and northern hemisphere have come to light and these changes appear to be synchronous with those in Greenland. The first evidence was from proxies for sea surface temperature (SST) preserved in ocean bottom sediment cores. Bond et al. (1993) claimed that when Greenland was cold the subpolar North Atlantic SSTs were also cold, although low sedimentation rates made the time resolution of the core too imprecise for easy cross comparison with the Greenland record.

Sachs and Lehman (1999) presented records from the Bermuda Rise, a region of high sedimentation that allows for good temporal resolution. They derived an SST from the alkenone unsaturation ratio measured in the sediments[2]. The agreement between the alkenone-derived SST and the Greenland record through numerous abrupt climate changes during the last glacial is startling (Figure 2): whenever Greenland is cold, the subtropical North Atlantic Ocean is cold too. While it is true that this conclusion partly depends on the fact that the age control on the Bermuda record was fitted in order to maximize the correspondence between the two records, the fact that it is possible to get such a high correspondence testifies to a real link between climate changes in these two locations. The typical warming here from stadial to interstadial was about $3°C$. It should be noted that the high sedimentation rate on the Bermuda rise occurs because of transport of material - and hence alkenones - to the site which can effect interpretation of the record (Ohkouchi et al. 2002).

The Cariaco basin just north of Venezuela is an invaluable treasure trove of climate records. The waters above are anoxic which prohibits mixing of the sediments by organisms living within. This, and very high sedimentation rates because of the proximity of the continent, has created a record with close to annual resolution extending back into the last ice age. Lea et al. (2003) used Mg/Ca ratios in Cariaco basin sediments to reconstruct the SSTs above during the last deglaciation. They found an abrupt warming at the Bolling/Allerod transition, an abrupt cooling of as much as $4°C$ at the beginning of the Younger Dryas and an abrupt warming at its termination. This result is consistent with Sachs and Lehman and Bond and, despite necessary concern about the proxy indicators, improves confidence that the entire North Atlantic surface ocean cooled dramatically, by a few to several $°C$ during Greenland stadials.

2.4 ABRUPT CLIMATE CHANGE AWAY FROM THE NORTH ATLANTIC

2.4.1 The Caribbean, tropical Atlantic and Africa

Using other proxies, the Cariaco basin record off Venezuela has also been used to infer climate changes other than in surface ocean temperature. Hughen et al. (1996, 1998) showed that the sediments reveal a transition into and out of the Younger Dryas as abrupt as that observed in Greenland and interpreted as abrupt changes in trade wind strength. Peterson et al. (2000) show that almost all of the abrupt jumps seen in Greenland between 60 kyr BP and 25 kyr BP (and numbered in Figure 1) are also seen in the Cariaco record as shifts in sediment reflectance and major element chemistry. These sediment characteristics record changes in the rate of riverine influx from South America north of the Amazon basin, i.e. the balance of precipitation and evaporation in that region, suggesting that during Greenland cold stadials northern regions of South America received less rain than now. The most obvious way in which this could happen is through a southward shift of the Intertropical

[2]Alkenones are chemical compounds within the marine organisms that are resistant to decomposition and laboratory studies have shown the unsaturation ratio to vary linearly with the temperature of the water that the organism is living in.

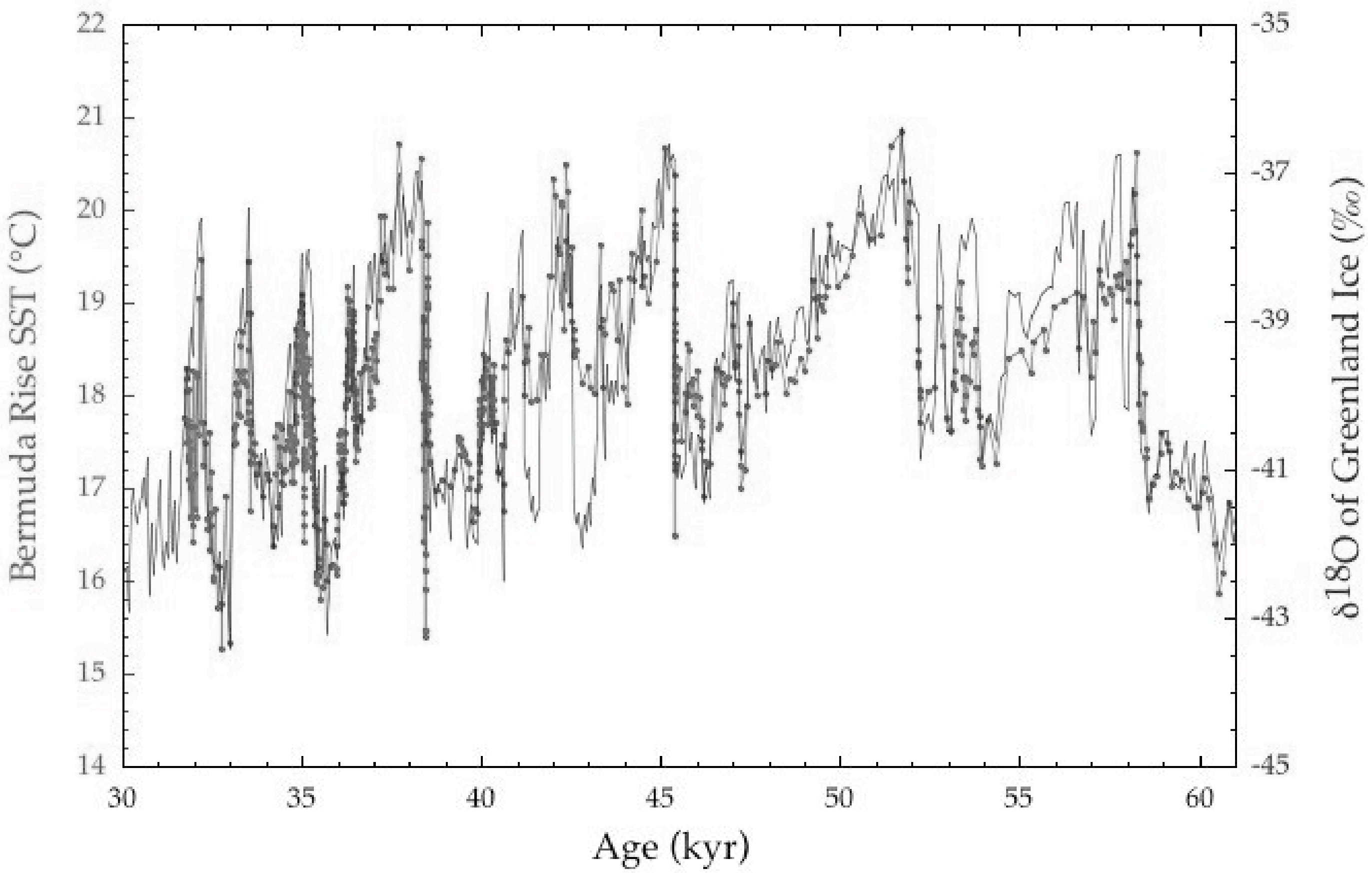

Figure 2.2 Reconstructed sea surface temperatures based on alkenone unsaturation ratios in sediments on the Bermuda Rise for the period from 30,000 to 60,000 years ago (line with dots, left axis), plotted together with Greenland $\delta^{18}O$ (plain line, right axis) on the GISP2 ice core time scale. Taken from Sachs and Lehman (1999).

Convergence Zone (ITCZ). Recent results from speleothems (calcium carbonate deposits in caves more commonly known as stalactites and stalagmites) have indeed indicated that, at times when northern South America was drier, northeast Brazil was wetter (Wang et al. 2004)[3]. These records make clear that abrupt climate changes impacted the tropical Atlantic region with a southward shifted ITCZ during cold stadials in Greenland.

There are numerous records that record the glacial and Holocene hydroclimate of tropical Africa. These come either from sediments in African lakes or from sediments offshore that record the rate of input of terrigenous material from the continent, assumed to be a proxy for precipitation over the land. Gasse (2000) reviews many of these records from the last glacial maximum and the deglaciation and found rapid increases in lake levels in tropical West Africa at the onset of the Bolling-Allerod warm interval and at the end of the Younger Dryas. This is consistent with results obtained from sediments offshore by the mouth of the Niger River that show a dry Younger Dryas across West Africa (Lezine et al. 2005). Adegbie et al. (2003) examined a sediment core off Cameroon in the Gulf of Guinea and deduced not just a weaker West African monsoon during the Younger Dryas but dry conditions correlating with Greenland stadials throughout the last 30,000 years of the glacial period.

2.4.2 Northern extratropics

The Younger Dryas abrupt climate change was first identified in pollen records in northern Europe and is named after a cold tolerant plant that reestablished itself during this time. Throughout Europe mountain glaciers advanced, as in Scandinavia and the Alps, or they reformed, as in the British Isles (see references in Denton et al. (2005)). Glacial advances require adequate snow in winter and cool enough summers for accumulation to exceed ablation. In North America the record is mixed. Shuman et al. (2002) show that across eastern North America vegetation changed dramatically at the beginning and end of the Younger Dryas but that these cannot simply be accounted for by a cooling, in part because summers may have been warmer as a consequence of increased insolation. In western North America alpine glaciers advanced during the Younger Dryas but with notable exceptions in the Sierra Nevada and around Mt. Rainier (Licciardi et al. 2004). Since precipitation patterns are highly spatially variable, the simplest explanation for glacial advances in Europe and North America would be a colder climate, including summers. Together with the Greenland record this suggests that the Younger Dryas abrupt climate change involved cooling across much of the northern extratropics.

Where longer records exist there is obvious correlation with all of the abrupt changes in Greenland during the last glacial. For example these events extended to

---

[3]The southward shift of the ITCZ during the Younger Dryas is not easy to reconcile with alkenone based SST reconstructions that show warming of the South American coast at $10^\circ N$ (Kim and Schneider 2003) and in the Caribbean Sea (Ruhlemann et al. 1999). Those SST reconstruction also disagree with the Mg/Ca-based ones of Lea et al. (2003) which show dramatic cooling north of Venezuela during the Younger Dryas. Curry and Oppo (1997) show cooling during Greenland stadials at $5^\circ N$ in the western tropical Atlantic based on the $\delta^{18}O$ in foramanifera which would also be consistent with a southward shifted ITCZ.

the northeast Pacific Ocean margin, appearing as changes in ocean oxygenation in the Santa Barbara basin and Gulf of California (Behl and Kennett 1996), although the interpretation of these records in terms of climate has proven difficult.

2.4.3 The tropics and the southern hemisphere

In the Atlantic sector, ice cores from Andean glaciers in Bolivia and Peru, both south of the Equator, show evidence of the Younger Dryas abrupt climate change (Thompson et al. 1998). On Sajama, in Bolivia, the abrupt shifts at the Bolling/Allerod transition and then into the Younger Dryas were as large as the shift between the depths of the LGM and the current climate. Younger Dryas era glacial advances have also been reported in the Andes but dating is uncertain. These results probably indicate that the cooling encompassed not just the northern hemisphere but also the tropics.

Further afield, abrupt climate changes throughout the last glacial cycle, and including the Younger Dryas, appear in speleothems from China (Yuan et al. 2004). Two speleotherms, from caves 1000km distant, record remarkably similar changes of $\delta^{18}O$ within them. According to the authors, the isotope content here does not reflect temperature changes but is indicating reduced precipitation during times when Greenland, and much of the northern hemisphere, was cold.

The China monsoon record is consistent with that derived from ocean sediments in the Arabian Sea which records changes in biological productivity, with higher productivity presumed to be caused by stronger monsoons and upwelling. Schulz et al. (1998) found that over the last 65,000 years there was an impressive correlation, independently dated in each record, of weaker monsoons going along with cold in Greenland (see also Altabet et al. 2002). Taken with the China record, this suggests that cold stadials in the northern hemisphere were associated with a weaker monsoon across all of Asia. The changes in monsoon strength are as large as the differences between glacial and interglacial climates. Thus abrupt climate changes were as strong here as around the North Atlantic.

An important record comes from the $\delta^{18}O$ content of surface dwelling foramanifera recorded in sediments on the eastern edge of the Indonesian archipelago and in the west Pacific warm pool with higher values interpreted to indicate increased salinity (Stott et al. 2002). The tropical Pacific salinity record does not clearly show the abrupt changes seen in Greenland but there is a visible correlation between increased salinity in the far west of the warm pool and stadial conditions in Greenland. This suggests reduced precipitation in this area during Greenland stadials.

The monsoon records and those from the Cariaco basin, the Bermuda rise, tropical Africa, the Santa Barbara basin and the west Pacific warm pool make clear that abrupt climate changes with impacts across the Northern hemisphere and throughout the global tropics occurred during the last deglaciation and throughout the last glacial. The Younger Dryas is just the best documented of these events.

Further south the trail begins to run dry. However a recent record developed by Farmer et al. (2005) on the basis of the Mg/Ca ratios within planktonic foraminifera in a well-dated core off the Namibian coast of south west Africa shows a potent Younger Dryas cooling of $2 - 3°C$. The SST evolution in this core during the Bolling-Allerod-Younger Dryas period correlates impressively with the Cariaco and Greenland records. This is the most important record to date documenting a Younger Dryas event in the southern extratropics. A report of a Younger Dryas glacial advance in New Zealand (Denton and Hendy 1994) has recently been claimed to be older by several hundred years (Broecker 2003). Both here and in the southern Andes much work remains to be done to identify and date late glacial advances.

2.4.4 The global mean climate

It is not yet clear whether the planet as a whole warmed and cooled during abrupt climate changes. If it did not then it makes sense to look for causes purely in changes in atmosphere and ocean circulations and heat transports. However, if the global mean temperature also changed then the circulation must have interacted with water vapor and/or clouds such that changes in greenhouse trapping and/or albedo allowed the planet to equilibrate with the incoming solar radiation at a different temperature. Changes in sea ice can also change the global mean temperature but have a lesser impact on planetary albedo than clouds because sea ice is most prevalent in locations and seasons with little solar radiation to reflect. Only improved temperature reconstructions from the tropics and southern mid-latitudes during times of abrupt changes will allow us to know if these were associated with planetary warming and cooling.

2.5 ABRUPT CLIMATE CHANGE AND THE DEEP OCEAN CIRCULATION

Even before there was a reliable means of determining past changes in ocean circulation, Broecker et al. (1985) presented the still-reigning paradigm of abrupt climate change: switches on and off of deep water formation in the North Atlantic Ocean and associated changes in the so-called thermohaline circulation (THC). When deep water formation does not occur the subpolar branch of the THC does not operate and surface currents do not bring warm, salty waters northward into the Nordic Seas, there to lose heat to the atmosphere. Consequently the climate around the North Atlantic cools.

The most compelling evidence to date of THC changes during abrupt climate changes is that of McManus et al. (2004) who showed that the THC was very weak between the LGM and the Bolling-Allerod warm transition when it abruptly 'turned on'. It then reduced again to half strength, quite sharply, during the Younger Dryas and then gradually returned to Holocene strength (Figure 3).

As the THC weakens and strengthens the transport of salty water from the subtropical Atlantic to the subpolar Atlantic would be expected to reduce and increase. Consequently during times of a weak THC, the salinity of the subtropical North Atlantic would be expected to increase. Recently Schmidt et al. (2004) reported there was indeed an inverse relationship between Caribbean Sea salinity and the THC that reinforces the evidence for THC changes. Together with earlier data (Hughen et al. 2000, Bond et al. 1997), these data indicate that abrupt climate changes *can* involve not just surface climate but also the deep ocean circulation or THC. It also indicates that the THC is capable of rapidly 'turning on' and that it can rapidly reduce in strength.

2.6 SEASONALITY OF ABRUPT CLIMATE CHANGE AROUND THE NORTH ATLANTIC OCEAN

Atkinson et al. (1987), using a method based on the current climate tolerances of hundreds of species of carnivorous beetles and the distribution of fossil remains of those species, concluded that over the British Isles the rapid warming of the Bolling-Allerod and the rapid changes of the Younger Dryas involved only modest summer temperature changes but enormous changes in winter temperature. The implied changes in seasonality were enough to convert the British Isles from a climate after the LGM (and after deglaciation of the Isles) similar to that of current day northeastern Russia, to one during the Bolling that was similar to that of today, then back again to one of great seasonality during the Younger Dryas, and then finally, back to a modern day climate. Each transition was accomplished in decades.

Denton et al. (2005) also show that around Greenland the Younger Dryas glacier advance is sufficiently limited that it can only be consistent with modest summer cooling. If the annual mean temperature reconstructions from ice cores are correct, then the winter cooling in Greenland must have been around $20°C$ - comparable to that in the British Isles. As they point out, all the paleoclimate data from around the North Atlantic Ocean, including Norway (Dahl and Nesje 1992), indicates that Younger Dryas cooling was largely contained within the winter while summers cooled more modestly.

Consequently the climate flip-flops measured in the ice cores - which do seem representative of the climate of the region - represent rapid transitions between two climate regimes. One is a maritime climate akin to the modern one and the other is a climate of marked seasonality in which winters are not tempered by the ameliorating effects of release of heat from the ocean or via atmospheric heat flux convergence. The most common explanation for how winters around the North Atlantic could become so severe is that sea ice expanded southward to the latitude of southern Britain. If winters around the North Atlantic got as cold as the reconstructions suggest they did during stadials and the Younger Dryas then sea ice would almost certainly encroach this far. How this could ever happen is a subject we will shortly turn to.

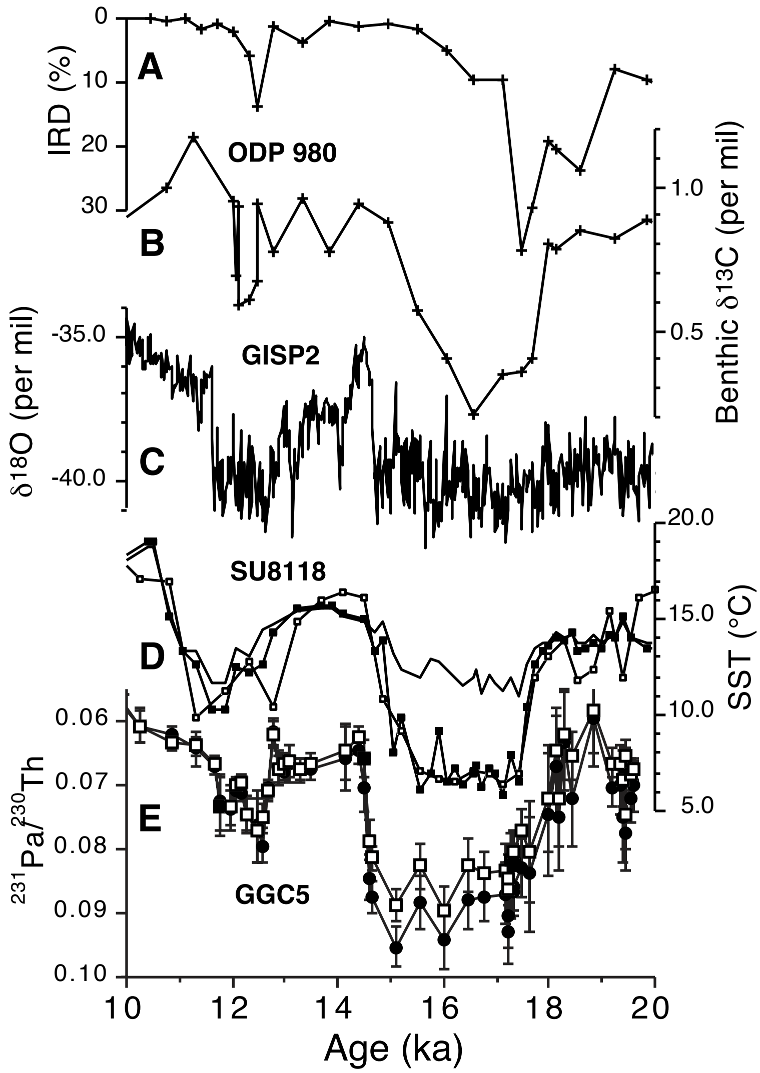

Figure 2.3 Records of ocean circulation and climate in the North Atlantic region during the last deglaciation from 20,000 to 10,000 years ago. (a) shows ice rafted debris from sediments in the subpolar North Atlantic, (b) shows the benthic $\delta^{13}C$, which is impacted by the strength of deep water formation, (c) shows the GISP2 $\delta^{18}O$ record, (d) shows reconstructed subpolar SSTs from a variety of sources and (e) shows the ratio $^{231}Pa/^{230}Th$ from the Bermuda Rise, a measure of the strength of meridonal overturning. The ratio $^{231}Pa/^{230}Th$ reaches 0.093 for a total cessation of the THC. Figure taken from McManus et al. (2004).

2.7 THE LACK OF MODERN ANALOGUES OF PAST ABRUPT CLIMATE CHANGES

The period of widespread instrumental observations of weather and climate, basically from the middle Nineteenth Century, when measurements from ships were begun, until the present, contains many climate changes but none that appear to be analogues, even in weakened form, of past abrupt climate changes. The North Atlantic region has experienced climate changes on decadal timescales. These have primarily been associated with changes in the atmosphere circulation. For example the trend from a low index to a high index state of the Northern Annular Mode (NAM) between the 1960s and late 1990s brought striking changes of climate to western Europe including drought in Spain and milder, snowier winters, and advancing glaciers in western Norway. The NAM trended in the opposite direction from the 1920s through to the 1960s. The more recent upward trend has been explained as a consequence of rising greenhouse gases (Shindell et al. 1999) or as a response to Indian Ocean warming (Hoerling et al. 2004). Both explanations make the earlier opposite trend hard to explain. Although basin-wide SST anomalies, which could be related to THC variations, do appear in the North Atlantic they do not seem to explain the NAM behavior. The NAM behavior has been gradual, free of abrupt shifts.

The most celebrated climate shift in the instrumental record is that of 1976/77. This winter ushered in an extended period in which the tropical Pacific Ocean was warmer than normal, as were the waters along the west coast of the Americas while the central North Pacific Ocean was cold (Zhang et al. 1997, Mantua et al. 1997). In the record of tropical tropospheric temperatures the transition does appear quite abrupt (Seager et al. 2004). More generally it appears as the dividing point between a period of strong ENSO events that began with the 1976/77 winter and a period of weaker ENSO activity in the decades before. This climate shift, and perhaps an opposite shift after the 1997/98 El Niño, does testify to the ability of the tropics to organize mid-latitude climate on multidecadal timescales (Hoerling et al. 2004, Huang et al. 2005), but falls short of being directly relevant to past climate changes.

The only other striking climate transition that has occurred in the instrumental record is the shift to a drier climate in the Sahel region of West Africa in the early and mid 1970s. This appears to be related to changes in tropical SSTs, including those in the Indian Ocean (Giannini et al. 2003), although the mechanisms remain obscure. Other monsoons have not gone through such clear transitions. The Sahel drying could be relevant to abrupt transitions into and out off the African Humid Period of the mid Holocene, a time when the region of the current Sahara Desert was grassland (deMenocal et al. 2000).

Neither indices of tropical climate variability (e.g. NINO3) nor extratropical circulation variability show evidence of bimodality such as that seen in the Greenland ice core data (Wunsch 2003) and efforts to locate regime-like behavior in the climate system have so far failed (Stephenson et al. 2004). Since regime-like behavior did occur in climates of the past, this emphasizes that past abrupt climate changes are different, not just quantitatively, but also qualitatively, from those in the instrumental record.

2.8 EXAMINING PROPOSED CAUSES OF ABRUPT CLIMATE CHANGE

Next we turn our attention to the possible causes of abrupt climate change. First we will summarize what needs to be explained.

2.8.1 The spatial character of abrupt climate changes

According to the studies described in the previous sections, the climate of the North Atlantic region, during glacial times, moved between two different states of operation. The transitions in between occurred rapidly, especially for the warming. These rapid climate changes involved striking temperature changes across western Europe and eastern North America, enormous in winter but modest in summer. Consequently there were abrupt changes in the degree of seasonality.

During North Atlantic cold stadials, surface ocean temperatures in the subtropical North Atlantic cooled, the ITCZ over South America shifted south, the tropical Americas and the South Atlantic Ocean off Africa cooled and the Asian monsoon weakened. In the tropical regions the transitions were as large as the difference between full LGM and modern states. Hence, the abrupt climate change signal does not appear to become more muted with distance from the North Atlantic. Any proposed mechanism must be able to explain these observed climate changes.

2.8.2 Strengths and weaknesses of the THC-driving theory

2.8.2.1 Role of the THC in today's climate

Because of the North Atlantic branch of the THC, with deep sinking at high latitudes compensated by northward flow at the surface and southward flow at depth, the North Atlantic Ocean moves heat northward at all latitudes. The following discussion is based on Seager et al. (2002). The North Atlantic Ocean moves about 0.8PW across $35°N$ (Trenberth and Caron 2001), enough to warm the area north of $35°N$ by $3-4°C$. This pales in comparison to a warming of $27°C$ due to the vastly greater atmosphere heat transport across $35°N$ and a warming of another $27°C$ in winter due to seasonal release of heat stored since the last winter. Not all of this ocean heat transport (OHT) is due to the THC as the subtropical and subpolar gyres and wind-driven overturning also contribute. Those circulations persist even when the THC is weakened or shuts down leaving a poleward OHT, though one that is greatly reduced.

The heat transported by the North Atlantic Ocean is released to the atmosphere primarily in two regions. The first is in the Gulf Stream region east of North America where, during winter, cold, dry air from the continent flows over the warm waters offshore extracting up to $400Wm^{-2}$, in the seasonal mean, from the ocean. A sizeable portion of this heat is picked up by transient eddies - atmosphere storms

- and converged over eastern North America ameliorating winters there. The transient eddy heat flux, as always, is acting diffusively to oppose the temperature contrasts created by stationary waves and the seasonal release of heat by the ocean (Lau 1979). The other region where the North Atlantic OHT is preferentially released is in the Norwegian Sea keeping it ice free in winter and warming the coast of Norway. Because of this geography of heat release the North Atlantic OHT warms the climate on *both* sides of the ocean, leaving the bulk of the temperature contrast of about $15°C$ to be explained by the more basic continental-maritime climate distinction and by stationary waves, especially those forced by flow over the Rocky Mountains.

2.8.2.2 The spatial pattern of THC-induced climate change

What is the global signature of a sudden THC shutdown or onset? Many modeling groups have performed experiments in which the THC in a coupled ocean-atmosphere GCM is forced to shut down, usually by addition of a massive amount of freshwater to the surface of the subpolar North Atlantic Ocean (e.g. Manabe and Stouffer (1997), Rind et al. (2001),Vellinga et al. (2002), Vellinga and Wood (2002), Zhang and Delworth (2005)). Freshwater dumping is usually justified in the paleoclimate context as an idealized representation of a meltwater discharge from glacial dammed lakes[4] . All these experiments agree that once the water columns of the subpolar North Atlantic have been stabilized by the addition of low density fresh water, deep water formation is reduced or ceases, poleward flow into the region is weakened, the heat transport reduces and the sea ice extent increases. They also all agree that the North Atlantic region cools by as much as several degrees Celsius in the region of Iceland and that the cooling extends into Europe and over Greenland although at reduced strength.

Figure 4 shows the change in annual mean air temperature caused by a THC shutdown in the Geophysical Fluid Dynamics Laboratory (GFDL) coupled model experiment performed by Zhang and Delworth (2005). The annual mean cooling over Greenland is only a few $°C$ (Manabe and Stouffer 1997, Zhang and Delworth 2005), much smaller than that observed for abrupt climate changes (Severinghaus and Brook 1999). The modeled cooling over western Europe is also at most a few degrees Celsius, much less than that reconstructed from beetles (Atkinson et al. 1987) or periglacial evidence (see Denton et al. (2005)). All models agree that the North Atlantic cooling extends down into the subtropics and perhaps as far as the Equator, but that south of $45°N$ the cooling is only of the order of $1°C$, less than that reproduced by Sachs and Lehman (1999). The models also do not produce a strong tropical cooling over South America, in conflict with ice core records there for the Younger Dryas (Thompson et al. 1998). Further, as shown in Figure 4, a THC shutdown causes warming in the South Atlantic Ocean which is contrast to

[4]The sudden discharge of ice from continental ice sheets directly into the ocean is another source of freshwater. These are called Heinrich events and can be traced in ocean sediments by the debris carried within the ice. It has been suggested that they occur at the end of a cooling cycle (Clarke et al. 1999) but in general it is not well understood how they fit into climate changes and they are not dealt with here. See review by Hemming et al. (2004).

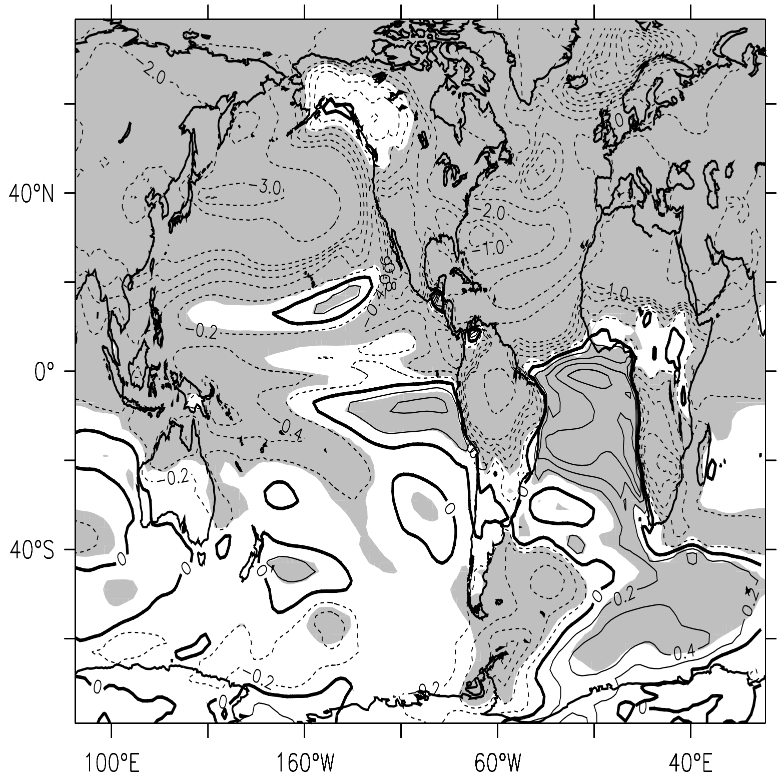

Figure 2.4 The change in annual mean surface air temperature between two states of a coupled GCM, one with a collapsed thermohaline circulation and one with an active thermohaline circulation. The model was the latest version of the GFDL coupled GCM. Shading indicates the change is significant at the 5% level. The figure is taken from Zhang and Delworth (2003).

the cooling there during the Younger Dryas reported by Farmer et al. (2005).

Rind et al. (2001), Vellinga and Wood (2002) and Zhang and Delworth (2005), using different coupled GCMs, all show that for a THC shutdown the ITCZ moves south in the tropical Atlantic, broadly consistent with cooling of the North Atlantic Ocean that occurred in the models, and interpretation of the Cariaco Basin record in terms of a reduction in precipitation over northern South America and the speleothem record of increased precipitation just south of the Equator. The precipitation changes reach up to a few mm per day in some regions. The models of Vellinga and Wood (2002) and Zhang and Delworth (2005, see figure 5) have reduced precipitation in the Asian monsoon region while the Rind et al. (2001) model has a very modest weakening of the Indian monsoon and an equally modest strengthening of the East Asian monsoon.

Figures 4 and 5 show that in the GFDL model the impacts of a THC shutdown extend into the northern and tropical Pacific Ocean where the pattern of temperature change is oddly similar to that in the Atlantic i.e. significant cooling in mid-

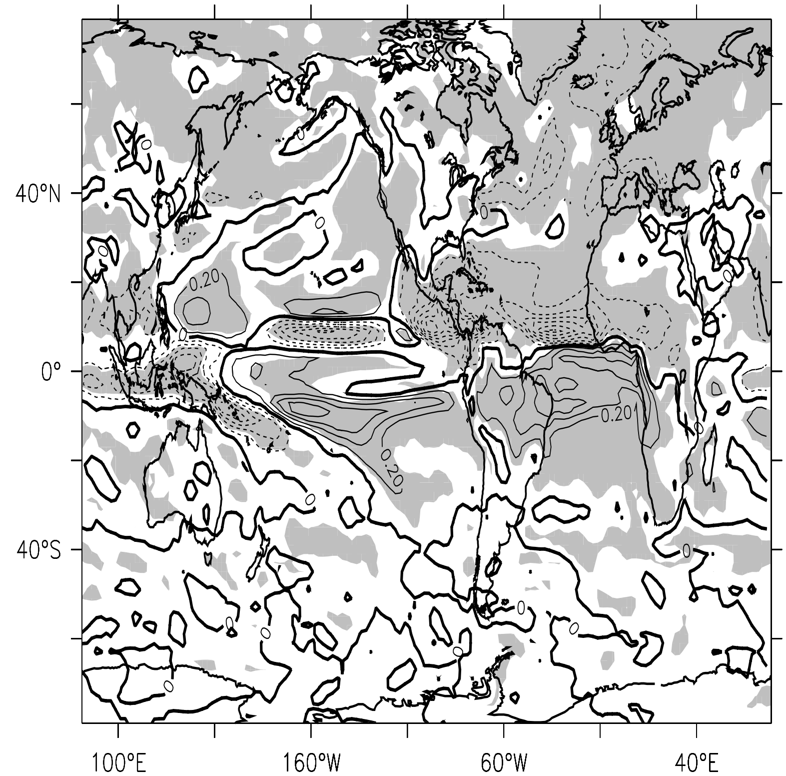

Figure 2.5 The change in annual mean precipitation (m/year) between two states of a coupled GCM, one with a collapsed thermohaline circulation and one with an active thermohaline circulation. The model was the latest version of the GFDL coupled GCM. Shading indicates the change is significant at the 95% level. The figure is taken from Zhang and Delworth (2003).

latitudes but warming immediately south of the Equator. It is yet to be determined why this occurs. The Hadley Centre model of Vellinga and Wood (2002) has a similar response. In the GFDL model tropical Pacific precipitation changes as expected given the change in surface temperature with a southward shift of the ITCZ.

In the North Atlantic region, therefore, there is sufficient agreement between models and paleoclimate data that changes in the THC were likely involved in abrupt climate changes. This said, changes in the THC, even shutdowns - at least as represented in GCMs - cannot explain the magnitude of the cooling around the North Atlantic, despite large increases in sea ice cover in the surrounding seas (Manabe and Stouffer 1997, Rind et al. 2001), or in the subtropics (Sachs and Lehman 1999). Therefore, even on its own home turf, the THC theory falls short of being able to offer a complete explanation of abrupt climate changes, unless all existing coupled GCMs are significantly in error (which they may be).

Outside of the North Atlantic region, the THC theory of abrupt climate change cannot entirely explain the paleoclimate data. According to GCMs, the impact of the THC on temperature and precipitation over equatorial and southern South America is too weak to explain the impressive documentation of the Younger Dryas in tropical ice cores. Changes in the THC also cannot fully explain the equally impressive reduction in monsoon strength, coherent across the Asian monsoon, that occurs during cold stadial events when the THC was weaker. They also cannot explain cooling in the southeast subtropical Atlantic Ocean. On the other hand the impacts of a THC shutdown do extend across the entire northern hemisphere and into the tropics.

2.8.2.3 The temporal behavior of the THC

McManus et al. (2004) showed, using radiochemical data, that the North Atlantic THC was operating during the LGM but in a 'drop dead' state during the beginning of the last deglaciation between the LGM and about 14.7 kyr ago. Then it abruptly increased to near modern strength, apparently in decades, coinciding with the dramatic Bolling-Allerod warming. At the beginning of the Younger Dryas the THC rapidly weakened, but remained active, and at the end it more gradually recovered to Holocene values. This remarkable result confirmed what had been suggested for a while, the THC is not a sluggish part of the climate system but can shift between modes of operation in years to decades. What can cause such changes?

Manabe and Stouffer (1988) showed that a coupled GCM (the GFDL one) had two stable modes of operation, one with an active North Atlantic THC and one with a 'drop dead' THC that could be induced by a massive addition of freshwater into the subpolar North Atlantic Ocean. This work inspired a generation of similar experiments and led to the development of the dominant paradigm of abrupt climate change: releases of glacial meltwater, whether it be from ice-dammed lakes or ice surges, caps the North Atlantic Ocean with fresh surface water shutting down deep water formation and turning off the flow of warm water from the subtropical North Atlantic into the subpolar Atlantic, thus cooling regional climate which is amplified by an increase in sea ice extent.

Rind et al. (2001), using a coupled GCM with a free-surface ocean model that allows for an increase in river flow at any chosen location to be directly specified, were able to induce a North Atlantic THC shutdown within a decade or two when the flow through the St. Lawrence was increased. However to do this required an inflow of 20Sv for five years. Licciardi et al. (1999) and Clark et al. (2001) have attempted to reconstruct the history of freshwater flux from North America into the Atlantic Ocean during the last deglaciation. They show jumps of a fraction of a Sverdrup in combined St. Lawrence and Hudson River runoff occurring as a result of reroutings of continental drainage and from sudden emptying of ice-dammed lakes (such as Lake Agassiz). For changes in freshwater flux to the North Atlantic of this magnitude, coupled GCMs agree that the THC would weaken by only a modest amount and only very gradually, say, over hundreds of years (Manabe and Stouffer 1997, Rind et al. 2001).

If the coupled GCMs are correct, then realistic freshwater influxes cannot explain the rapid decreases in the North Atlantic THC that, almost certainly, seem to have occurred in the past. Further, the paleoclimate record shows that in the

North Atlantic region, it was actually the *warmings*, and THC resumptions, rather than the *coolings*, and THC slowdowns or shutdowns, that were the most abrupt. Coupled GCMs can produce rapid cessations of deep water formation if given apparently unrealistic perturbations but none has yet produced an abrupt resumption of deep water formation and the THC. Instead, after a period of a weak or nonexistent THC, the THC gradually returns over a hundreds years or so (Vellinga et al. 2002). It is also not clear that all of the abrupt changes seen in Figure 1 were the result of freshwater perturbations. In fact that seems rather unlikely suggesting other mechanisms whereby the THC can be turned on and off.

Simple climate models (often called 'intermediate' models) that contain more than one stable mode of operation of the THC can reproduce many aspects of observed climate records including rapid THC resumptions and warmings (Ganopolski and Rahmstorf 2001). The finding of multiple equilibrium states in the GFDL coupled model lent credibility to such models. However, the key physics acting in the intermediate climate models that have produced regime like behavior in the THC include ocean convection, advection and diffusion. The realism with which these processes are represented is affected by the very coarse horizonal and vertical resolution within the intermediate models. Intermediate models have non-trivial problems in simulating the effects of topography on circulation which strongly influences where ocean convection takes place and its stability of location and strength. The ocean components of the simple models will also have difficulty representing the real processes by which the ocean moves heat which, as shown by Boccaletti et al. (2005), is largely contained within shallow, surface intensified, circulations and not within deep overturning circulations. Similarly, the atmospheric components of the intermediate models are, essentially, extended energy balance models. It is not clear that regime-like behavior would occur if the same ocean models were coupled to more realistic atmospheric models (such as those that allow for weather).

In general, it needs to be demonstrated that multiple regimes can be supported by models with higher resolution that adequately resolve the physics responsible for the multiple regimes in intermediate models. However to date, it is not even clear whether coupled GCMs other than the old Manabe and Stouffer GFDL model have multiple equilbrium states. Vellinga et al. (2002) say they have not found this behavior in the Hadley Centre coupled GCM. In general the GCM model results are at sufficient variance with the paleoclimate record to raise several questions that will be addressed shortly.

2.8.3 Summary

There is no doubt that abrupt climate changes, such as the stadial-interstadial transitions of the last ice age and the Bolling Allerod warming and Younger Dryas of the last deglaciation, involve changes in the North Atlantic THC and probably as an active component. However changes in the THC alone cannot explain the known global climate changes associated with these events, especially in the tropics, even accounting for its impacts on sea ice. Further, no state-of-the-art climate model has shown that regime-like behavior of the THC, with rapid transitions between states, is possible. On this basis we suggest that there is a lot more to the puzzle of abrupt climate change than changes of the THC.

2.9 ATMOSPHERIC CIRCULATION REGIMES AND GLOBAL ATMOSPHERE-OCEAN COUPLING AS POSSIBLE CAUSES OF ABRUPT CLIMATE CHANGE

The discussions of the spatial extent of abrupt climate changes in glacial times and during the last deglaciation should make it clear that the causes must be found in changes in the general circulations of the *global*, as opposed to *regional*, atmosphere and ocean circulation. The idea that the THC changes, directly impacting a small area of the globe, and that, somehow, most of the rest of the world piggybacks along in a rather systematic and reliable way seems decidedly dubious.

2.9.1 The problem of North Atlantic climate change

Consider the changes in seasonality documented around the North Atlantic Ocean going from the LGM through the Bolling-Allerod abrupt warm transition and then cooling into the Younger Dryas followed by the second abrupt warm transition that ended it. A wide collection of evidence indicates that winter climate in, for example, the British Isles changed up and down by about $20 - 30°C$ during these transitions and summer temperatures by $4 - 6°C$ (e.g. Atkinson et al. (1987)).

This could be accomplished if the sea ice edge extended south of the British Isles during winter and then retreated far north in summer. Imposing this seasonal cycle of sea ice cover under an atmosphere GCM, Renssen et al. (2001) reproduced, approximately, the observed different deglacial climates in a set of time snapshot experiments. When sea ice extends that far south, and if it is thick enough with few gaps, the surface temperature of the ice and the air above drops dramatically in winter as the atmosphere becomes insulated from the ocean below. It is likely that the sea ice extension, rather than the presence of ice sheets (which existed throughout the Bolling-Allerod and Younger Dryas over North America and Scandinavia), caused winter cooling in the North Atlantic sector.

But sea ice never extends as far south as the British Isles in coupled GCMs when the THC is forced to shutdown. This is consistent with the fact that in the North Pacific and Southern Ocean sea ice typically does not extend equatorward of $60°$ even though, unlike the current North Atlantic Ocean but similar to one with a THC shutdown, there is close to zero poleward ocean heat transport there. Sea ice is melted from below and so its equatorward extent depends in part on the disposition of the ocean heat transport *but* the placement of the ocean heat transport depends on the wind stress. In THC shutdown experiments the sea ice expansion is restricted by the atmospheric circulation.

Assuming that, by some means, sea ice did extend as far south as the British Isles during the winters of stadials, the relative summer warmth then requires it to retreat far to the north and for the ocean to warm up tremendously. As a point of comparison, in the current climate ocean areas that are ice covered at the end of winter do not achieve temperatures of more than about $2°C$ in summer.

Thus the problems posed by abrupt change in the North Atlantic region are:

1. How could sea ice extend so far south in winter during the stadials?

2. How, during the spring and summer of stadials can there be such an enormous influx of heat as to melt the ice and warm the water below by close to $10°C$? If 50m of water needs to be warmed up by this much in four months it would take an average net surface heat flux of $150Wm^{-2}$, more than twice the current average between early spring and mid summer and more than can be accounted for by any increase in summer solar irradiance (as during the Younger Dryas).

3. How can this stadial state of drastic seasonality abruptly shift into one similar to that of today with a highly maritime climate in western Europe? Remember that both states can exist in the presence of large ice sheets over North America and Scandinavia.

2.9.2 Required changes of atmospheric circulation regimes and heat transports

To solve the first two problems we must imagine a stadial climate in which the heat transport has almost the opposite seasonal cycle to that of today. Whereas now winter atmospheric and ocean heat transport holds back the sea ice in the North Atlantic, during stadials weak heat transport in the mid to high latitudes must have allowed the sea ice to advance south. In contrast, during the summers of stadials, there must have been strong transport to melt back the sea ice and establish mild summers.

In thinking of ways to reduce the winter convergence of heat into the mid and high latitude North Atlantic, we might begin with the storm tracks and mean atmosphere circulation. The Atlantic storm track and jet stream have a clear southwest to northeast trajectory whereas the Pacific ones are more zonal over most of their longitudinal reach (Hoskins and Valdes 1990). If the Atlantic storm track and jet could be induced to take a more zonal track, akin to its Pacific cousin, the North Atlantic would cool.

The cooling over western Europe and the North Atlantic Ocean would be driven by less southerly flow and advective warming in the winter stationary waves - the real reason for the mildness of west European winters (Seager et al. 2002). Transient eddy heat transports, which act diffusively on temperature, would oppose the cooling as the storm track - defined as the statistical mean of the routes taken by eddies or, for this purpose, as the location of the maximum eddy heat transport - relocated towards the Mediterranean, and Europe was placed on the eddy heat flux convergence side of the storm track.

2.9.3 Impacts of an hypothesized shift to zonal circulation during winter in the North Atlantic sector

An essentially zonal wind flow across the North Atlantic Ocean could set in motion a chain of events that establish a cold North Atlantic climate. First of all the removal of warm southwesterly advection into the North Atlantic will directly cool the ocean and coastal regions there. In the current climate the flow of water in the North Atlantic Drift from the Gulf Stream and into the Nordic Seas is controlled by the wind stress pattern and reflects the northward tilt of the Atlantic jet stream and storm track. This water is salty and, as it cools on its northward track, becomes dense enough to sink to the bottom. The northward flow of water was the ultimate explanation of Warren (1983) for why deep water is formed in the North Atlantic and not the North Pacific (see also Emile-Geay et al. (2002)). If the wind stress pattern becomes zonal the North Atlantic Drift will flow directly across the Atlantic towards France and Spain instead of towards the Nordic Seas, reducing the salt flux into the subpolar Atlantic and causing the sinking branch of the THC to shift southward and reduce ocean heat flux convergence in the Nordic Sea.

In the current climate the sea ice edge is very much controlled by the pattern of the winds through the influence they have on both the ice drift and the atmosphere heat transport. For example positive phases of the NAO go along with less ice in the Greenland and Barents Seas, where the anomalous winds are from the south, and more ice in the Labrador Sea where the anomalous winds are from the north (Deser et al. 2000). On longer timescales changes in wind will impact the ocean heat transport and this will also influence the sea ice cover. A shift to more zonal winds across the North Atlantic will, by all processes, allow sea ice to extend further south than it currently does, cooling the North Atlantic regional climate.

Currently much of the Gulf Stream water continues northward and becomes part of the deep overturning in the North Atlantic. As shown by Talley (2003), using data, and by Boccaletti et al. (2005), using a model, in the North Atlantic Ocean, unlike other basins, the deep overturning heat transport is equal in magnitude to that by the horizontal gyres and intermediate overturning. The northward flow of the Drift helps sustain winter heat release from the ocean, especially north of Scotland. The heat release and diabatic heating of the atmosphere above helps maintain the Icelandic Low. This was shown in the GCM experiments of Seager et al. (2002) and is implicit in the modeling of Hoskins and Valdes (1990). The Icelandic Low maintains the northward deflection of the storm track, jet and the North Atlantic Drift. Thus the interaction between the atmosphere and ocean circulations over the North Atlantic appears to be self-reinforcing. Perhaps forcing of this system from outside, maybe in the tropics, can set this reinforcing mechanism running in reverse and establish a stable zonal jet and storm track, a zonal North Atlantic Drift, a collapsed THC and a cold North Atlantic.

Although a shift to a zonal circulation across the North Atlantic would cause extensive cooling of the European region on its own it would be assisted by the induced THC shutdown. To date explorations of THC shutdowns have all focused on the impact of freshwater discharges from melting ice sheets or ice damned lakes. To our knowledge the impact on the THC of a sudden shift in the Atlantic to a more

Pacific-like wind stress regime has never been investigated. It would decrease the warm, salty water flow towards the Nordic Seas that, currently, as it cools, becomes dense enough to sink to the abyss. With that inflow shut off by the altered wind stress pattern it at least appears possible that the THC will shutdown too.

2.9.4 Summer climate and abrupt shifts in seasonality

During the Bolling warm period and the period after the Younger Dryas the climate of western Europe had a seasonality similar to that of today but during the cold spells while winters were extremely cold, summer temperatures remained close to $10°C$ (Atkinson et al. 1987, Denton et al. 2005). Not surprisingly such summer temperatures can be reproduced for these periods in an atmosphere GCM when the North Atlantic SSTs are specified to also be that warm (Renssen and Isarin 2001) During the cold periods the SSTs need to warm from freezing to about $10°C$. Currently the only regions of the world ocean that have a seasonal cycle that large are in the western boundary currents east of North America and Asia. Here warm moist advection around the summer subtropical anticyclones warms the SSTs, and cold dry advection of the continents, as well as transient eddies, cool the SSTs in winter (Seager et al. 2003a).

The summer warmth and increased seasonality require a much larger heat import into the North Atlantic region by the atmosphere and ocean during summer than currently occurs. One possible cause of this is the presence of extensive ice sheets over North America and Scandinavia. During winters ice sheets do not significantly perturb the surface or planetary radiation budget but in summer they are a vast radiative sink as they reflect solar radiation that otherwise would be absorbed at the surface. This radiative sink, and the associated cold temperatures, will induce an anomalous convergence of atmospheric heat transport over the ice. This could cause a much larger export of heat from the tropics into high latitudes during summer than is currently the case.

Few GCM studies have examined the the impact of ice sheets on seasonal energy transports. Hall et al. (1996) did find greatly increased summer atmosphere heat transport in an LGM simulation, although the fixed SSTs in the study make the results questionable. Summer ice sheets will also impact the stationary waves. Currently, during summers poleward warm advection around the Icelandic Low (which, unlike the Aleutian Low, is still present in summer, though weak) helps warm the coast of Scandinavia while further south cold advection around the North Atlantic subtropical high keeps the coasts of Portugal and North Africa cool (Seager et al. 2003a). If, during the summers of cold periods such as the Younger Dryas, the Icelandic Low was strong and south of its current location then it could cause advective warming of western Europe. It will be well worth examining how the Laurentide and Scandinavian ice sheets impacted the summer stationary wave climate, via their orography and albedo, and how this impacted summer temperatures and seasonality.

2.9.5 The tropical circulation and abrupt climate change

2.9.5.1 Tropical forcing of global climate variations

There are many reasons to think that the tropics play an active role, and maybe an organizing one, in abrupt climate change and are not a backwater responding passively to climate changes organized by the North Atlantic Ocean. For one thing the paleoclimate record shows that abrupt climate changes are, relatively, as large in the tropics (e.g. the Asian monsoon and Andean ice cores) as in the North Atlantic region (Denton et al. 2005). While models do show that weakening and strengthening of the Atlantic THC can inspire significant tropical climate change (Zhang and Delworth 2005, Vellinga and Wood 2002), the changes seem to fall short of those that actually did occur. It is equally plausible that past abrupt changes occurred as part of a global climate reorganization instigated within the tropics.

The other reason for looking to the tropics is that in the current climate many climate changes around the globe throughout the twentieth century have been forced from the tropics through varying SSTs and patterns of tropical convection. This includes the major droughts and pluvials over the Americas (Schubert et al. 2004, Seager et al. 2005, Huang et al. 2005), the drying of the Sahel (Giannini et al. 2004) and, quite possibly, the trends in the Northern Annular Mode (Hurrell et al 2004, Hoerling et al. 2004, Schneider et al. 2003). The tropical Pacific may also be able to exert a control on the Atlantic THC (Latif et al. 2000) through its influence on rainfall over the tropical Americas and the Atlantic Ocean and the vapor flux across Central America.

2.9.5.2 Limitations of the ENSO blueprint

Although all this is true, so far it has not been demonstrated that the tropics in any way control abrupt climate changes. The dominant mode of year to year and decade to decade climate variability in the tropics relates to the El Niño-Southern Oscillation (ENSO). Numerous attempts to explain past tropical SST changes have invoked changes with an ENSO-like spatial pattern (e.g. Stott et al. (2002), Koutavos et al. (2002)) and the global impacts of ENSO have been appealed to as a cause of climate change on glacial timescales (Cane 1998). Certainly ENSO responds to orbital and other external forcing and may even respond abruptly to gradual forcing (Clement et al. 1999, 2000, 2001, Mann et al. 2005).

Clement et al. (2001) suggested that the peculiar orbital forcing of the Younger Dryas interval stabilized the tropical Pacific atmosphere-ocean system resulting in long periods without interannual variability and a persistent change in the mean state with a La Niña-like pattern. While this result is compelling the global climate changes during the Younger Dryas do not fit the typical La Niña pattern. Of the available paleo records, only tropical cooling, as implied by Andean ice core records for the Younger Dryas, is consistent with a La Niña-like state. During stadials, reduced precipitation over northern south America, the weak Asian and west African monsoons and reduced salinity in the tropical west Pacific are all more typical of an El Niño like state. El Niño events also cool the mid-latitudes of each hemisphere (Seager et al. 2003b) which clearly happened during the Younger Dryas and glacial stadials. However an El Niño-like state would also reduce precipitation in the part of northeast Brazil where speleothems indicate wet conditions during stadials. This, in combination with the dry Caribbean coast of South America (as inferred from the Cariaco record), is more consistent with a weakened THC and a southward shift of the ITCZ (see Figure 5). Finally neither a THC shutdown nor an El Niño-like state can explain the cooling off Namibia during the Younger Dryas reported by Farmer et al. (2005).

Despite some evidence for an El Niño-like state during stadials currently the impacts of El Niño over the North Atlantic are far too weak to account for the dramatic climate changes that occured there during glacial stadials and interstadials. It is true that the climate response associated with persistent El Niño-like, or La Niña-like anomalies need not be the same as the one that goes along with interannual variability. However experiments with coupled models in which persistent El Niño or La Niña states were induced (Hazeleger et al. 2005 and unpublished results conducted by the first author with the GISS and CCM3 climate models) produced a response akin to the interannual one: during persistent El Niños the tropics warm, the mid-latitudes cool and the poles warm and there is a rather small global mean temperature change (Seager et al. 2003b). At least in these experiments persistent changes in ENSO did not excite positive feedbacks involving water vapor or clouds that significantly amplified climate change.

The inadequacy of the ENSO blueprint alone in explaining the spatial pattern of abrupt change indicates that in pursuing tropical involvement in abrupt climate change we need to think beyond ENSO. This need is highlighted by two problems of the model studies mentioned above. First, in interannual variability an El Niño warming of the eastern tropical Pacific is caused by a transient adjustment of the ocean circulation and upper ocean heat content. On longer timescales relevant to different climate regimes the changes in ocean circulation and heat transport must be in equilibrium with the atmospheric circulation. In this equilibrated case changes in SST must be sustained by different processes than for interannual changes in SST (Hazeleger et al. 2004). They will probably have different spatial patterns and different climate consequences.

Second, the models that were used to examine the global response to persistent El Niños and La Niñas (Hazeleger et al. 2005) did not allow for ocean circulation adjustment. It is an open question how persistent tropical climate changes impact the global ocean circulation, thermocline structure and heat transports. In both cases the challenge is to consider the full range of possible tropical and global atmosphere-ocean coupling and to think outside of ‘the ENSO box’.

2.9.5.3 Tropical heating and extratropical jets and storm tracks

Clearly, tropical climate changes have the *potential* to cause significant extratropical climate change. In an intriguing paper Lee and Kim (2003) have argued that the Northern Hemisphere contains two dynamically distinct jet streams. One is a subtropical jet (STJ) on the poleward flank of the Hadley Cell and the other is a polar front jet (PFJ) further poleward. As they point out, an STJ can be created in the absence of eddies by conservation of angular momentum in the poleward upper

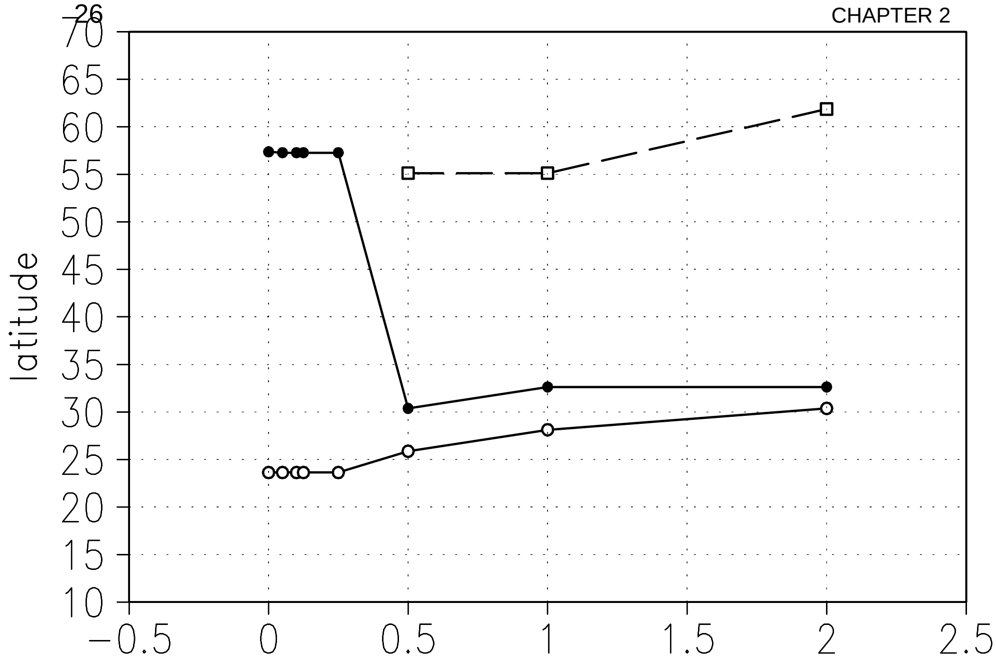

Figure 2.6 Latitudes of the subtropical jet (line with open dots), the primary (line with closed dots) and secondary (dashed line with open squares) eddy-driven jets as a function of the strength of an imposed tropical heating (horizontal axis). Results were generated from simulations with simplified models with the eddy-driven jet latitudes predicted with the unstable normal modes of an axisymmetric flow. Figure taken from Lee and Kim (2003).

level flow of the Hadley Cell (Schneider 1977, Held and Hou 1980). In the real world, while this process is relevant, the STJ is also shaped by eddies. In contrast, eddies can create a jet within a zone of uniform baroclinicity but without a prior existing jet (Panetta 1993, Lee 1997). Lee and Kim argue that the PFJ has the characteristics of a jet generated in this manner. Such jets are sometimes referred to as 'eddy-driven' although this term is potentially misleading as both the STJ and PFJ are in thermal wind balance.

Nonetheless the distinction between a subtropical jet with maximum westerlies on the poleward edge of the Hadley Cell and overlying a region of zero mean meridional surface wind and an eddy-driven jet further poleward, coincident with maximum eddy momentum flux convergence and lying between zero lines of the mean meridional surface flow, is valid (Son and Lee (2005), see also Robinson (2006), in this volume, for further discussion of the work of Son, Lee and Kim). The Asian sector has only a single jet with the character of an STJ. Over the Atlantic sector, during winter, both jets exist with an STJ that begins west of Africa and extends across Africa and Asia and a PFJ that begins over the southern United States and extends up to the British Isles and Scandinavia. This picture is broadly consistent with the observations of Palmen and Newton (1969).

Lee and Kim argue that the STJ and PFJ compete for the attention of transient eddies. When the STJ is strong enough, the meridional temperature gradient with which it must coexist, is potent enough to organize transient eddy activity resulting in a relatively southern storm track, as over the Pacific Ocean. Where the STJ is weaker the self-reinforcing interaction between eddies that feed off temperature gradients, the momentum fluxes off those eddies that drive jets - and associated temperature gradients - allows the establishment of a relatively northern jet, as over the Atlantic Ocean.

Lee and Kim show that the competition between the two jets is modulated by the strength of the tropical heating (Figure 6). When this is strong the STJ is strong enough to ’capture’ the transient eddies and merge the subtropical and eddy-driven jets. Their results are based on a collection of idealized and theoretical calculations. Son and Lee (2005) reproduced similar results with the statistically steady states of a primitive equation model subjected to different strengths of tropical heating and high latitude cooling. Both studies suggest that there are distributions of tropical heating and high latitude cooling that would cause the eddy-driven Atlantic jet to be captured by the STJ west of Africa. Were this to occur the Atlantic storm track would become reoriented to extend directly eastward from the Southern United States towards the Mediterranean region. In certain regimes, according to these studies, a modest change in the tropical heating could then allow reestablishment of the double jet structure and set the same chain reaction off in the reverse direction. These interactions between jets and thermal driving, when coupled to the ocean, may provide a means for creating climate transitions such as those seen in the paleoclimate records. As stated by Robinson (2006) the abrupt transitions of jet regimes here are governed by the dynamics of extratropical eddies and could also ’lurk in our future as climate warms’.

According to Yin (2003) southward movement of tropical convection during northern winter strengthens the northern hemisphere subtropical jet, which is consistent with the results of Lindzen and Hou (1988). A southward shift of the ITCZ in the longitudes of the Americas could lead to the Atlantic subtropical jet ‘capturing’ the transient eddies. This would cause a weakening of the eddy-driven jet over the North Atlantic, analogous to the mid-winter suppression of the North Pacific storm track (Yin 2003, but see Swanson (2006) in this volume for a purely mid-latitude explanation of this phenomena). This would reduce the atmospheric flux of heat into the Northeast Atlantic sector. Further the zonal elongation of the Atlantic jet would cause changes in the surface wind stress and curl and reduce, or eliminate, the flux of warm, salty water from the subtropical North Atlantic into the subpolar Atlantic, reducing the heat loss to the atmosphere there (and, hence, reinforcing zonal atmosphere flow), allowing sea ice to expand and also reducing, or shutting down, the THC.

These processes would also operate in the classical paradigm for abrupt climate change, which features abrupt shifts in the THC as the causal agent. For example, if there was a sudden resumption of the THC (for whatever reason) and the sea ice moved northward, the subtropical North Atlantic atmosphere and ocean would warm and the ITCZ would shift farther north (Chiang et al, 2003). A shift in the ITCZ northward across the equator would weaken the subtropical jet and cause the storm track in the North Atlantic to strengthen and shift northward into the Nordic Seas. This would reestablish the transport of warm, salty water into the subpolar North Atlantic, deflecting the jet and storm track northwards and further encouraging a strong THC.

It is also possible that the changes in the ITCZ could involve alterations in the latitudinal concentration of the heating such that, during stadials, as the ITCZ convection moves south it also becomes more longitudinally confined. According to Hou and Lindzen (1992) this is another process that can intensify the Hadley circulation and potentially set in motion the same set of atmosphere and ocean processes and feedbacks described above. In this case the southern hemisphere would be impacted to. A stronger STJ and weaker PFJ in the southern hemisphere could, if it lead to an equatorward movement of the surface westerlies, reduce the flux of salty water from the Indian Ocean into the south Atlantic. This salt flux helps sustain the North Atlantic branch of the THC (Gordon et al. 1992) so this is another means whereby tropical heating distributions could have the potential to alter the THC and create global climate changes.

Hence, interactions involving both the atmosphere and ocean in the North Atlantic-Americas sector could act as an amplifier of climate change linking the high latitudes and the tropics. We have already discussed evidence that the ITCZ over the Atlantic and Americas does move south during Greenland stadials (Peterson et al. 2000, Wang et al. 2004). To date this has been viewed as little more than a response to cooling of the North Atlantic Ocean but it would be immediately fruitful to examine the complete nature of two-way coupling between the ITCZ and the atmosphere and ocean circulation in the mid and high latitude North Atlantic region, both in modern and glacial climates.

Support for these ideas of reorganization of atmospheric circulation comes from simulations with coupled GCMs of the climate of the Last Glacial Maximum (LGM) such as the one conducted with NCAR's Climate System Model (CSM) (Shin et al. 2003) and further analyzed by Camille Li at the University of Washington who provided these figures. Figure 7 shows the upper troposheric zonal wind and the transient eddy heat transport in the lower troposphere for the winters of a simulation of the modern climate and for the winters of the LGM. Not surprisingly, in the presence of ice sheets, and with lower levels of carbon dioxide, the meridional temperature gradient was stronger at the LGM and, consistently, the jet stream was also. However, in a situation analogous to the modern day mid-winter suppression in the Pacific, the transient eddy heat flux was reduced throughout the winter. This could be a consequence of the orographic forcing of the Laurentide ice sheet generating a barotropic stationary wave that influences the structure of the Atlantic jet (see Swanson (2006) for a simple model demonstration of this). This atmospheric state, with reduced eddy heat transport, is suggestive of the stadial state during glacials. It will be interesting to see if some change can induce this state to flip over to one akin to the modern - or interstadial - state because, if it can, then this is a viable means for explaining the observed abrupt changes of the glacial period. Figure 8 provides a highly idealized schematic of the circulations proposed for the stadial and interstadial states of the glacial period.

An intriguing aspect of this idea is that during summer the stationary wave forced

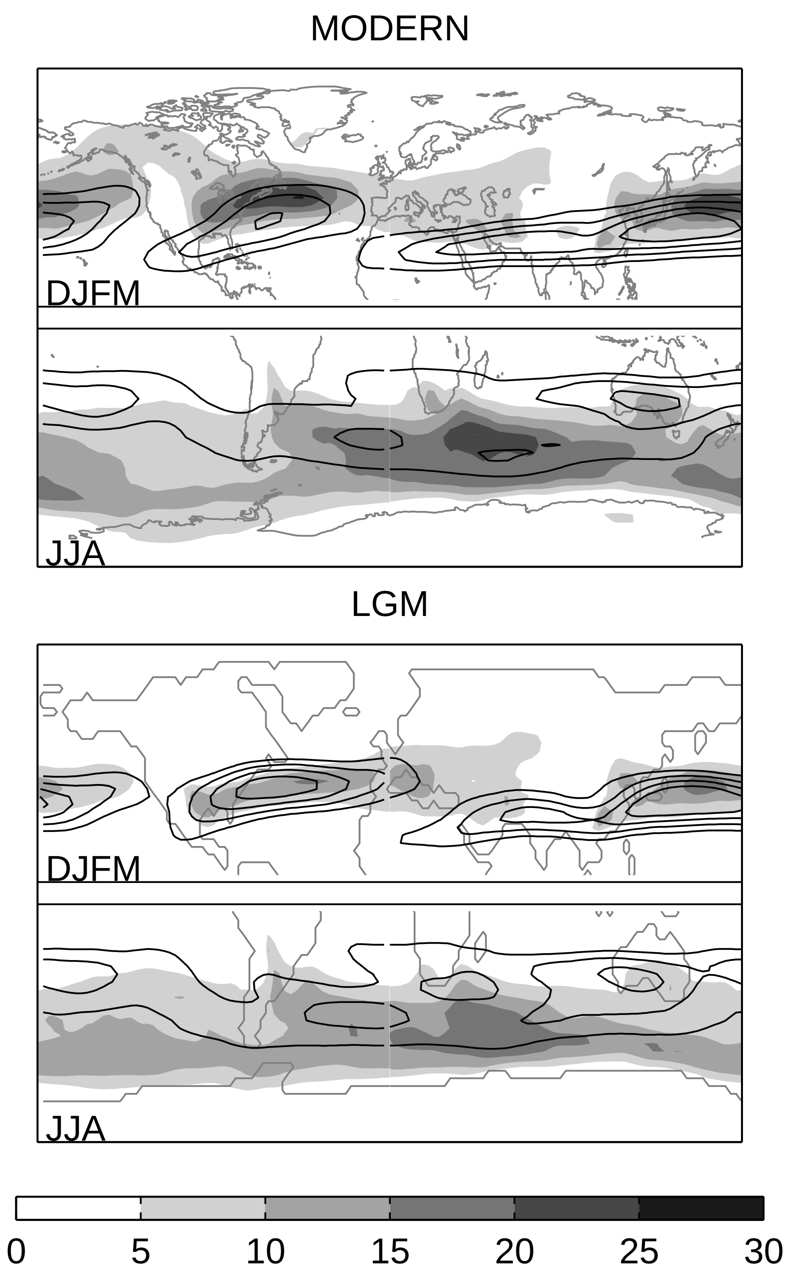

Figure 2.7 The northern hemisphere wintertime zonal wind in the upper troposphere (contour interval of 10 ms^{-1}, contouring beginning at 30ms^{-1}) and the transient eddy heat transport in the lower troposphere (shading, mKs^{-1}) from coupled GCM simulations of the modern climate (above) and the Last Glacial Maximum (below). Analysis conducted, and figure provided, by Camille Li of the University of Washington.

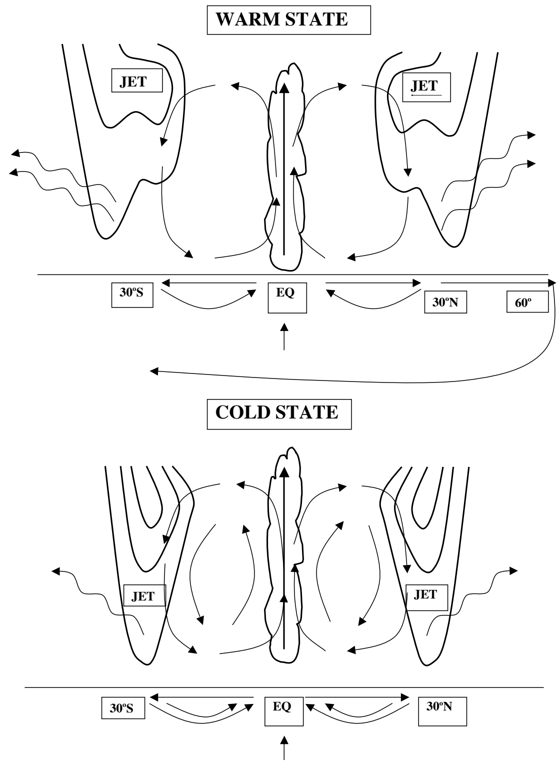

Figure 2.8 Schematic of the meridional cross section of the proposed atmosphere and ocean circulation for interstadial (above) and stadial (below) states of the glacial period. The stadial states have stronger overturning in the tropics, a stronger STJ and weaker PFJ, weaker eddy heat transports (indicated by wavy arrows) and, as a consequence of the changes in North Atlantic wind stress, a reduced THC.

by the Laurentide ice sheet will be weaker but, because of the albedo effect, the overall baroclinicity in the North Atlantic region will remain. This could allow for a strong summer storm track stretching into the subpolar North Atlantic, much as we see in winter today, that would effectively warm European summers in agreement with the proxy evidence. If it is true that the ice sheet orography is important for allowing shifts in jet regimes then it might be expected that at its most southward extent and largest volume the orographic forcing of the flow may become sufficiently dominant to allow only one jet and eddy regime, that of the stadial state. Such stability would lead to a cessation of abrupt changes as actually did happen, broadly speaking, between Dansgaard-Oeschger event 3 and the Bolling/Allerod warm jump (see Figure 1).

If abrupt climate change during the glacial between these two coupled atmosphere-ocean states in the tropics are responsible then changes in the global ocean are likely to be important in explaining the long term nature (centuries to a millennium or so) of each regime, a subject to which we now turn.

2.9.5.4 Near-global atmosphere ocean coupling and tropical climate reorganization

ENSO and decadal ENSO are oscillations of the tropical Pacific atmosphere and ocean. The climatological winds over the tropical Pacific Ocean can quite closely explain the spatial variation of tropical Pacific thermocline depths according to Sverdrup dynamics (Veronis 1973) and, with the inclusions of Ekman dynamics, the creation of a warm pool and a cold tongue (Clement et al. 2005). The same dynamics can explain the transient adjustment of the thermocline to varying winds (Cane 1984). A central element is an adjustment to balance at the Equator between the thermocline (or sea level height) tilt and the zonal wind stress. What this dynamics cannot explain is the mean thermocline depth.

Boccaletti et al. (2004) point out that the mean depth has to be related to the global surface heat budget and ocean heat transport. This is because a shallow equatorial thermocline allows upwelling to expose cold water which causes weak latent heat loss and, consequently, a net downward surface heat flux. Wind-driven overturning circulation exports this heat poleward. A deeper thermocline would allow less poleward ocean heat transport. While this is true the details of the adjustment between the thermocline, winds, currents and the ocean and atmosphere heat transports are opaque.

Huang et al. (2000), Johnson and Marshall (2004), Cessi et al. (2004) and Timmermann et al. (2005) have pointed out that, by reducing the transfer of mass from the upper layer of the ocean to deeper layers, cessation of North Atlantic Deep Water formation would cause the tropical Pacific thermocline to migrate downward. The adjustment occurs by coastal and equatorial Kelvin waves and begins within a decade or two although the splitting of the equatorial Kelvin wave in the Atlantic into coastal Kelvin waves that propagate both north and south limits the strength of the signal outside of the North Atlantic. Nonetheless a deeper tropical Pacific thermocline should cause a reduction of poleward tropical ocean heat transport. The tropical thermocline depth and structure must also be impacted by surface buoyancy fluxes, water mass transformation and subduction within the extratropics where water that eventually upwells at the Equator leaves the surface. Relatively little work has been done on this beyond theoretical explorations but Hazeleger et al. (2001, 2004) have shown that variations in the mid-latitude storm tracks and circulation can have a significant impact on the tropical Paciific thermocline. It remains to be seen how, for example, plausible shifts in the the storm tracks and jet streams in past climates impacted the tropical oceans.

2.9.5.5 Ocean and atmosphere heat transports and the global climate

If it is really true that global atmosphere-ocean coupling involving changes in storm tracks, the mid-latitude westerlies and the THC can impact the depth and structure of the tropical Pacific thermocline, then the potential exists to shift the partitioning between the tropical atmosphere and ocean heat transports. Held (2001) argued that this partitioning within the tropics cannot change because of the dynamical coupling, by Ekman transports, between the Hadley Cell and the meridional ocean overturning. However, the constraint is not nearly so tight once the tropical oceanic gyre transport, which moves heat equatorward, is considered (Hazeleger et al. 2004, 2005).

It is the total transport by atmosphere and ocean that is most tightly constrained, apparently by the radiation budget at the top of the atmosphere (Clement and Seager 1999). Consequently a reduction of wind-driven tropical ocean heat transport causes a compensating increase in atmosphere heat transport, with the total heat transport varying by little, a universal result with all manner of models in all types of experiments including removal of the ocean compoent, removal of mountains and removal of continents (Manabe et al. 1975, Cohen-Solal and LeTreut 1997, Clement and Seager 1999, Czaja and Marshall 2006, Winton 2003). It is plausible that a deeper tropical thermocline, by reducing the poleward ocean heat transport, could cause the atmosphere to carry a larger share of the total transport.

It is also a common model result that, as the tropical ocean heat transport is reduced, the subtropical SSTs cool and the marine low level cloud cover and planetary albedo increase. In addition, tropical deep convection becomes more confined towards the Equator resulting in reduced atmospheric water vapor and reduced greenhouse trapping in the subtropics[5]. Both effects cool the planet (Clement and Seager 1999, Winton 2003, Herweijer et al. 2005). Thus an ocean adjustment to a deeper equatorial thermocline, as could be induced by a THC shutdown or some unknown change of the extratropical atmosphere circulation, would be expected to cause a cooling of the global climate.

2.9.6 Discussion

The global atmosphere-ocean coupling idea of abrupt climate change, with an active or organizing role for the tropics, is pure speculation. To date no one has shown that the tropical-extratopical atmosphere-ocean circulation has different modes of operation with radically different climates but, then again, no one has tried. The tropical equivalent of a North Atlantic ‘hosing’ experiment capable of causing a model climate to flip into an alternate state has yet to be devised.

There also remains the tricky problem of what causes the climate to remain stuck in one regime or the other for centuries before rapidly switching back to the alternate state. Here the THC idea has little advantage over the global coupling idea. Although simplified climate models can simulate regime-like climate changes with abrupt transitions in between (Ganopolski and Rahmstorf 2001), this kind of behavior has not been found in coupled GCMs. Instead, after being forced to shutdown by a very large forcing, the THC in current coupled GCMs tends to dribble back to full strength over the following few centuries. In contrast, in the paleoclimate record, the resumptions of deep water formation appear more abrupt than the shut downs. For the global coupling idea the long residence in one climate state or the other and then a switch would, presumably, have to involve the deep ocean circulation. At some point deep ocean climate change reaches a point whereby its’ influence on

[5]It is also possible that changes in the fluxes of moisture by transient eddies and the eddy-driven mean meridional circulation cause drying in the subtropics but this possibility was not addressed in these studies.

the coupled climate of the atmosphere and upper tropical oceans causes a switch between the different tropical climate states.

At this point the global coupling idea is largely based on intuition. However there are interesting recent studies on 1) how the tropics both respond to, and organize, global circulation and climate change, 2) new theories on the role of tropical heat transports in global climate and 3) new ideas on the global controls on the tropical thermocline, all ensure that this idea will be actively pursued in the future.

2.10 CONCLUSIONS

The abrupt climate changes that occurred during the last glaciation and deglaciation are mind-boggling both in terms of rapidity and magnitude. That winters in the British Isles could switch between mild, wet ones very similar to today and ones in which winter temperatures dropped to as much as $20°$ below freezing, and do so in years to decades, is simply astounding. No state-of-the-art climate model, of the kind used to project future climate change within the Intergovernmental Panel on Climate Change process, has ever produced a climate change like this. The normal explanation of how such changes occurred is that deep water formation in the Nordic Seas abruptly ceased or resumed forcing a change in ocean heat flux convergence and changes in sea ice. However coupled GCMs only produce such rapid cessations in response to unrealistically large freshwater forcing and have not so far produced a rapid resumption. Even when they do produce cessations of deep water formation the climate change around the North Atlantic region is much less than the proxy reconstructions indicate, even though the sea ice cover in the models increases.

According to coupled GCMs, cessations and resumptions of deep water formation do cause climate changes around the world qualitatively akin to those reconstructed within the paleoclimate record. When the deep sinking branch of the THC in the North Atlantic is forced to shutdown, the Atlantic ITCZ moves south and the Asian monsoon weakens, both of which agree with reconstructions. However the modeled change in the Asian monsoon is weaker than that reconstructed while the North Atlantic Ocean circulation changes do not seem capable of causing - for the most recent such abrupt change (the Younger Dryas) - the South American cooling seen in Andean ice cores and the cooling in the southeast Atlantic.

Here we have argued that the abrupt changes must involve more than changes in the North Atlantic Ocean circulation. In particular it is argued that the degree of winter cooling around the North Atlantic must be caused by a substantial change in the atmospheric circulation involving a great reduction of atmospheric heat transport into the region. Such a change could, possibly, be due to a switch to a regime of nearly zonal wind flow across the Atlantic, denying western Europe the warm advection within stationary waves that is the fundamental reason for why Europe's winters are currently so mild. Such a change in wind regime would, presumably, also cause a change in the North Atlantic Ocean circulation as the poleward flow of warm, salty waters from the tropics into the Nordic Seas is diverted south by the change in wind stress curl. This would impact the location and strength of deep water formation and allow sea ice to expand south.

Changes in the distribution and strength of tropical convection are capable of causing such changes in the mid-latitude wind regime, according to idealized GCM experiments. Tropical forcing route is appealing because it could help explain the large abrupt changes in the monsoons and tropical climate that are known to have occurred as well as force changes in mid-latitude atmosphere and ocean circulation and climate.

That the period of instrumental records has been free of dramatic abrupt changes, and that even the Holocene was quiet compared to the glacial period, argues that climate instability arises when there are continental ice sheets in North America and Eurasia and/or when the climate is colder. During glacial periods the climate can be described as fitting with Lorenz's (1968, 1970) concept of 'almost-intransitivity' in which the climate possesses successive very long periods with remarkably different states and abrupt transitions inbetween. It must be that either the presence of ice sheets, with their albedo and orographic forcing, or the colder mean state, allows the atmosphere and ocean circulations to adopt almost-intransitivity. It is possible that long integrations of coupled GCMs with glacial boundary conditions will reveal these states. Climate modelers should hesitate before discarding a model simulation that produces a climate that is distinctly warm in many parts of the world even in the presence of glacial boundary conditions!

Currently our knowledge of the general circulations of the atmosphere and ocean does not provide a means whereby we can imagine alternative states of the tropical and global climate, and the ability to move rapidly between them. However there has been enough recent work on the relationships between tropical atmosphere and ocean heat transports and the global controls on heat transports and the tropical thermocline to provide some hints that rearrangment may be possible. The situation is poised for a 'Manabe and Stouffer' moment when, with some clever, or fortuitous, experiment, alternative states are demonstrated in a coupled model.

When, if ever, this will occur is unclear. The problem for dynamicists working in this area is that the period of instrumental observations, and model simulations of that period, do not provide even a hint that drastic climate reorganizations can occur. Our understanding of the general circulations is based fundamentally on this period, or, more correctly, on the last 50 years of it, a time of gradual climate change or, at best, more rapid changes of modest amplitude. So it is not surprising that our encyclopedia of knowledge of the general circulations contains many ideas of negative feedbacks between the circulations that may help explain climate variability but also stabilize the climate (Bjerknes 1964, Hazeleger et al. 2005, Shaffrey and Sutton 2005). The modern period has not been propitious for studying how the climate can run away to a new state. Because of this our understanding has to be limited. Possibly the extent of our understanding will be brought into question by climate change itself as the Earth's climate changes more rapidly than we can extend our understanding of it. But we do not have to wait for that unfortunate event as the past is already full of events that simply cannot be placed within our current understanding of the general circulations but are there, waiting to be explained.

Acknowledgement We wish to thank Amy Clement, Mark Cane, Peter deMenocal, Gavin Schmidt and George Denton for many useful conversations and T. Blunier, J. Sachs, J. McManus, R. Zhang, M. Vellinga, S. Lee and C. Li for kindly providing figures. We also thanks Tapio Schneider and Walter Robinson for excellent critical reviews of the manuscript. This work was supported by NOAA grant NA030AR4320179 PO7 (RS) and and NSF grant ATM-0502204 (DSB).

Bibliography

Adegbie, A. T., Schneider, R. R., Rohl, U. and Wefer, G. (2003). Glacial millennial-scale fluctuations in central African precipitation recorded in terrigenous sediment supply and freshwater signals offshore Cameroon, *Palaeogeog, Palaeoclim. Palaeoecol.*, **197**, 323-333.

Altabet, M. A., Higginson, M. J. and Murray, D. W. (2002). The effect of millennial-scale changes in Aabian Sea denitrification on atmospheric CO_2, *Nature*, **415**, 159-162.

Atkinson, T. C., Briffa, K. R. and Coope, G. R. (1987). Seasonal temperatures in Britain during the past 22,000 years, reconstructed using beetle remains, *Nature*, **365**, 587-592.

Behl, R. J. and Kennett, J. P. (1996). Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr, *Nature*, **379**, 243-246.

Bjerknes, J. (1964). Atlantic air-sea interaction, *Adv. Geophys.*, **10**, Academic Press, 1-82.

Blunier, T. and Brook, E. J. (2001). Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period, *Science* , **291**, 109-112.

Boccaletti, G., Pacanowski, R. C., Philander, S. G. H. and Fedorov, A. V. (2004). The thermal structure of the upper ocean, *J. Phys. Ocean*, **34**, 888-902.

Boccaletti, G., Ferreira, R., Adcroft, A., Ferreira, D. and Marshall, J. (2005). The vertical structure of ocean heat transport, *Geophys. Res. Lett.*, **32**, doi:10.1029/2005GL022474.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I. and Bonani, G. (1997). A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates, *Science*,**278**,1257-1266.

Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J., and Bonani, G. (1993). Correlations between climate records from North Atlantic sediments and Greenland ice, *Nature*, **365**, 143-147.

Broecker, W. S., Peteet, D. M. and Rind, D. (1985). Does the ocean-atmosphere system have more than one stable mode of operation? *Nature*, **315**, 21-26.

Broecker, W. S. (2003). Does the trigger for abrupt climate change reside in the ocean or in the atmosphere? *Science*, **300**, 1519-1522.

Brook, E. J., Harder, S., Severinghaus, J. and Bender, M. (1999). Atmospheric methane and millennial-scale climate change, in *Mechanisms of Global Climate Change at Millennial Time Scales*, Clark, P.U., Webb, R.S. and Keigwin, L.D., Eds., American Geophysical Union, Washington D.C., pp165-175.

Cane, M. A. (1984). Modeling sea level during El Niño, *J. Phys. Oceanogr.*, **14**, 1864-1874.

Cane, M. A. (1998). A role for the tropical Pacific, *Science*, **282**, 59-61.

Cessi, P., Bryan, K. and Zhang, R. (2004). Global seiching of thermocline waters between the Atlantic and the Indian-Pacific Ocean basins, *Geophys. Res. Lett.*, **31**, L06201, doi:10.1029/2003GL019091.

Chiang, J. C. H., Biasutti, M. and Battisti, D. S. (2003). Sensitivity of the Atlantic Intertropical Convergence Zone to Last Glacial Maximum boundary conditions, *Paleoceanogr.*, **18**, doi:10.1029/2003PA000916.

Clark, P. U., Marshall, S. J., Clarke, G. K. C., Hostetler, S. W. , Licciardi, J. M. and Teller, J. T. (2001). Freshwater forcing of abrupt climate change during the last glaciation, *Science*, **293**, 283-287.

Clarke, G. K. C., Marshall, S. J., Hillaire-Marcel, C., Bilodeau, G. and Veiga-Pires, C. (1999). A glaciological perspective on Heinrich events, in *Mechanisms of Global Climate Change at Millennial Time Scales*, Clark, P.U., Webb, R.S. and Keigwin, L.D., Eds., American Geophysical Union, Washington D.C., pp 243-262.

Clement A. C., Cane M. A. and Seager R. (2001). An orbitally driven tropical source for abrupt climate change, *J. Climate*, **14**, 2369-2375.

Clement A. C., Seager R. and Cane M. A. (1999). Orbital controls on the El Niño/Southern Oscillation and the tropical climate, *Paleoceanogr.*, **15**, 731-737.

Clement A. C.and Seager R. (1999). Climate and the tropical oceans, *J. Climate*, **12**, 3383-3401.

Clement, A. C., Seager, R. and Cane, M. A. (2000). Suppression of El Niño during the mid-Holocene by changes in the Earth's orbit, *Paleoceanogr.*, **14**, 441-456.

Clement, A. C., Seager, R. and Murtugudde, R. (2005). Why are there tropical warm pools?, *J. Climate*, **18**, 5294-5311.

Cohen-Solal, E. and Le Treut, H. (1997). Role of the oceanic heat transport in climate dynamics: A sensitivity study with an atmospheric general circulation model, *Tellus*, **49A**, 371-387.

Cuffey, K. M., Clow, G. D., Alley, R. B., Stuiver, M., Waddington, E. D. and Saltus, R. W. (1995). Large arctic temperature change at the Wisconsin-Holocene glacial transition, *Nature*, **270**, 455-458.

Curry, W. B. and Oppo, D. W. (1997). Synchronous, high frequency oscillations in tropical sea surface temperatures and North Atlantic Deep Water production during the last glacial cycle, *Paleoceanogr.*, **12**, 1-14.

Czaja, A. and Marshall, J. C. (2006). The partitioning of poleward heat transport between the atmosphere and ocean, *J. Atmos. Sci.*, in press.

Dahl, S. O., and Nesje, A. (1992). Paleoclimatic implications based on equilibrium-line altitude depressions of reconstructed Younger Dryas and Holocene cirque glaciers in inner Nordfjord, western Norway, *Palaeogeorg. Palaeoclimatol. Palaeocecol.*, **94** 87-97.

deMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Sarnthein, M., Baker, L. and Yarusinsky, M. (2000). Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing, *Quaternary Science Reviews*, **19, 1-5**, 347-361.

Denton, G. H., Alley, R. B., Comer, G. C. and Broecker, W. S. (2005). The role of seasonality in abrupt climate change, *Quaternary Science Reviews*, **24**, 1159-1182.

Denton, G. H., and Hendy, C. H. (1994). Younger Dryas Age Advance of Franz Joseph Glacier in the Southern Alps of New Zealand, *Science*, **264**, 1434-1437.

Deser, C., Walsh, J. E., and Timlin, M. S. (2000). Arctic sea ice variability in the context of recent atmospheric circulation trends, *J. Climate*, **13**, 617-633.

Emile-Geay, J., Cane, M. A., Naik, N., Seager, R., Clement, A. and van Geen, A. (2003). Warren revisited: Atmospheric freshwater fluxes and Why is no deep water formed in the North Pacific?" *J. Geophys. Res.*, *Oceans*, **108**, No. C6, Art. No. 3178, 10.1029/2001JC001058.

Farmer, E. C., deMenocal, P. B. and Marchitto, T. M. (2005). Holocene and deglacial ocean temperature variability in the Benguela upwelling region: Implications for low latitude atmospheric circulation, *Paleoceanogr.*, **20**, PA2018, doi:10.1029/2004PA001049.

Ganapolski, A. and Rahmstorff, S. (2001). Rapid changes of glacial climate simulated in a coupled climate model, *Nature*, **409**, 153-158.

Gasse, F. (2000). Hydrological changes in the African tropics since the Last Glacial Maximum, *Quat. Sci. Rev.*, **19**, 189-211.

Giannini, A., Saravanan, R. and Chang P. (2003). Oceanic forcing of Sahel rainfall on interannual and interdecadal time scales, *Science*, **302**, 1027-1030.

Gordon, A. L., Weiss, R. F., Smethie, W. M. and Warner, M. J. (1992). Thermocline and intermediate water communication between the South Atlantic and Indian Oceans, *J. Geophys. Res.*, **97**, 7223-7240.

Hall, N. M. J., Dong, B. and Valdes, P. J. (1996). Atmospheric equilibrium, instability and energy transport at the last glacial maximum, *Climate Dynamics*, **12**, 497-511.

Hays J. D., Imbrie J. and Shackleton N. J. (1976). Variations in the Earth's orbit, pacemaker of the ice ages, *Science*, **194**, 1121-1132.

Hazeleger, W., Seager, R., Visbeck, M., Naik, N. H. and Rodgers, K. (2001). On the impact of the mid-latitude stormtrack on the upper Pacific, *J. Phys. Oceanogr.*, **31**, 616-636.

Hazeleger, W., Seager, R., Cane, M. A., and Naik, N. H. (2004). How can tropical Pacific ocean heat transport vary, *J. Phys. Oceanogr.*, **24**, 320-333.

Hazeleger, W., Severijns, C., Seager, R. and Molteni, F. (2005). Tropical Pacific-driven decadal energy transport variability,*J. Climate*, **18**, 2037-2051.

Held, I. M. (2001). The partitioning of the poleward energy transport between the tropical ocean and atmosphere, *J. Atmos. Sci.*, **58**, 943-948.

Held, I. M. and A. Y. Hou (1980). Nonlinear axially symmetric circulations in a nearly inviscid atmosphere, *J. Atmos. Sci.*, **37**, 515-533.

Hemming, S. R. (2004). Heinrich events: massive late Pleistocene detritus layers of the North Atlantic and their global imprint, *Rev. Geophys.*, **42**, RG1005,doi:10.1029/2003RG000128.

Herweijer, C., Seager, R., Winton, M. and Clement, A. C. (2005). Why ocean heat transport warms the global mean climate, *Tellus*, **57**, 662-675.

Hewitt, C. D., Stouffer, R. J., Brocolli, A. J., Mitchell, J. B. and Valdes, P. J. (2003). The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum, *Clim. Dyn.*, **20**, 203-218.

Hoerling, M. P., Whitaker, J. S., Kumar, A. Wang, W. (2004). Twentieth Century climate change. Part II: Understanding the effect of Indian Ocean warming, *Clim. Dyn.*, **23**, 391-405.

Hoskins, B. J. and Valdes, P. J. (1990). On the existence of storm tracks, *J. Atmos. Sci.*, **47**, 1854-1864.

Hou, A. Y. and Lindzen, R. S. (1992). The influence of concentrated heating on the Hadley circulation, *J. Atmos. Sci.*, **49**, 1233-1241.

Huang, R.- X., Cane, M. A., Naik, N., and Goodman, P. (2000). Global adjustment of the thermocline in response to deepwater formation, *Geophys. Res. Let.*, **27**, 759-762.

Huang, H.- P., Seager, R. and Kushnir, Y. (2005). The 1976/77 transition in precipitation over the Americas and the influence of tropical SST, *Climate Dynamics*, **24**, 721-740.

Hughen, K. A., Overpeck, J. T. , Peterson, L. C. and Trumbore, S. (1996). Rapid climate changes in the tropical Atlantic region during the last deglaciation, *Nature*, **380**, 51-54.

Hughen, K. A., Overpeck, J. T., Lehman, S. J., Kashgarian, M., Southon, J., Peterson, L. C., Alley, R. and Sigman, D.M. (1998). Deglacial changes in ocean circulation from an extended radiocarbon calibration, *Nature*, **391**, 65-68.

Hughen, K. A., Southon, J., Lehman, S. J. and Overpeck, J. T. (2000). Synchronous radiocarbon and climate shifts during the last deglaciation, *Science*, **290**, 1951-1954.

Hurrell, J. W. Hoerling, M. P. Phillips, A. S. and Xu, T. (2004). Twentieth Century North Atlantic climate change. Part I: assessing determinism , *Clim. Dyn.*, **23**, 371-389.

Johnson, H. L. and Marshall, D. P. (2004). Global teleconnections of meridional overturning circulation anomalies, *J. Phys. Oceanogr.*, **34**, 1702-1722.

Kim, J.- H. and Schneider, R. R. (2003). Low-latitude control of interhemispheric sea-surface temperature contract in the tropical Atlantic over the past 21 k years: the possible role of SE trade winds, *Climate Dynamics*, **21**, 337-347.

Koutavas, A., Lynch-Stieglitz, J., Marchitto Jr., T. M. and Sachs, J. P. (2002). El Niño-like pattern in ice age tropical Pacific sea surface temperature, *Science*, **297**, 226-230.

Latif, M., Roeckner, E., Mikolajewicz, U. and Voss, R. (2000). Tropical stabilization of the thermohaline circulation in a greenhouse warming simulation, *J. Climate*, **13**, 1809-1813.

Lau, N.-C. (1979). The observed structure of tropospheric stationary waves and the local balances of vorticity and heat, *J. Atmos. Sci.* ,**36**, 996-1016.

Lea, D. W., Pak, D. P., Peterson, L. C. and Hughen, K. A. (2003). Synchroneity of tropical and high-latitude Atlantic temperatures over the last glacial termination, *Science*, **301**, 1361-1364.

Lee, S. (1997). Maintenance of multiple jets in a baroclinic flow, *J. Atmos. Sci.*, **54**, 1726-1738.

Lee, S. and Kim, H.-K. (2003). The dynamical relationship between subtropical and eddy-driven jets, *J. Atmos. Sci.*, **60**, 1490-1503.

Lezine, A. -M., Duplessey, J. C. and Cazet, J. P. (2005). West African monsoon variability during the last deglaciation and the Holocene: Evidence from fresh water algae, pollen and isotope data from core KW31, Gulf of Guinea, *Palaeogeog. Plaeoclim. PalaeoEcol.*, **219**, 225-237.

Licciardi, J. M., Clark, P.U., Brook, E. J., Elmore, D. and Sharma, P. (2004). Variable responses of western U.S. glaciers during the last deglaciation, *Geological Society of America*, **32-1**, 81-84.

Licciardi, J. M., Teller, J.T. and Clark, P. U. (1999). Freshwater routing by the Laurentide ice sheet during the last deglaciation, *Mechanisms of Global Climate Change at Millennial Time Scales*, **AGU**, 177-201.

Lindzen, R. S. and Hou, A. Y. (1988). Hadley circulation for zonally averaged heating centered off the equator, *J. Atmos. Sci.*, **45**, 2416-2427.

Lorenz, E. N. (1968). Climatic determinism, *Meteor. Monogr.*, **8**, 1-3.

Lorenz, E. N. (1970). Climatic change as a mathematical problem, *J. Appl. Meteor.*, **9**, 325-329.

Manabe, S. and Stouffer, R. J. (1988). Two stable equilibria of a coupled ocean-atmosphere model, *J. Climate*, **1**, 841-866.

Manabe, S. and Stouffer, R. J. (1997). Coupled ocean-atmosphere model response to freshwater input: Comparison to Younger Dryas event, *Paleoceanography*, **12**, 321-336.

Manabe, S., Bryan, K., and Spelman, M. J. (1975). Global Ocean-Atmosphere Climate Model. 1. Atmospheric Circulation, *P. Phys. Oceanography*, **5**, 3-29.

Mann, M. E., Cane, M. A., Zebiak, S. E. and Clement, A. (2005). Volcanic and Solar Forcing of El Nio over the past 1000 years, *J. Climate*, **18**, 447-456.

Mantua, N. J., Hare, S. R., Zhang, Y., Wallace, J. M., and Francis, R. C. (1998). A Pacific interdecadal climate oscillation with impacts on salmon production, *Bull. Amer. Meteor. Soc.*, **78**, 1069-1079.

McManus, J. F., Francois, R., Gherardi, J.- M., Kelgwin, L. D. and Brown-Leger, S. (2004). Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, **428**, 834-837.

Ohkouchi, N., Eglington, T. L., Keigwin, L. D. and Hayes, J. M. (2002). Spatial and temporal offsets between proxy records in a sediment drift, *Science*, **298**, 1224-1227.

Palmen, E. and Newton, C. W. (1969). Atmospheric Circulation Systems: Their structure and physical interpretation, Academic Press, New York and London, 603pp.

Panetta, R. L. (1993). Zonal jets in wide baroclinically unstable regions: Persistence and scale selection, *J. Atmos. Sci.* **50**, 2073-2106.

Peterson, L. C., Haug, G. H., hughen, K. A. and Rohl, U. (2000). Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial, *Science*, **290**, 1947-1951.

Renssen, H. and Isarin, R. F. B. (2001). The two major warming phases of the last deglaciation at 14.7 and 11.5 kyr cal BP in Europe: climate reconstructions and AGCM experiments., *Global and Planetary Change*, **30**, 117-154.

Renssen, H., Isarin, R. F. B., Jacob, D., Podzun, R. and Vandenberghe, J. (2001). Simulation of the Younger Dryas climate in Europe using a regional climate model nested in an AGCM: preliminary results, *Global and Planetary Change*, **30**, 41-57.

Rind, D., Russell, G., Schmidt, G., Sheth, S., Collins, D., deMenocal, P. and Teller, J. (2001). Effects of glacial meltwater in the GISS coupled atmosphere-ocean model 1, North Atlantic Deep Water response, *J. Geophys. Res.*, **106**, 27335-27353.

Robinson, W. A. (2006). Eddy-mediated interactions between low latitudes and the extratropics, in *The Global Circulation of the Atmosphere*, T. Schneider and A.S. Sobel, Eds., Princeton University Press, Princeton, NJ.

Roe, G. H. and Steig, E. J. (2004). Characterization of millennial-scale climate variability, *J. Climate*, **17**, 1929-1944.

Ruddiman, W. F. and McIntyre A. (1981). Oceanic mechanisms for amplifying the 23,000 year ice-volume cycle, *Science*, **212**, 617-627.

Ruhlemann, C., Mulitza, S., Mller, P. J., Wefer, G. and Zhan, R. (1999). Warming of the tropical Atlantic Ocean and slowdown of thermohaline circulation during the last deglaciation, *Nature*, **402**, 511-514.

Sachs, J. P. and Lehman, S. J. (1999). Subtropical North Atlantic Temperatures 60,000 to 30,000 Years Ago, *Science*, **286**, 756-759.

Schmidt, M. W., Spero, H. J. and Lea, D. W. (2004). Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation, *Nature*, **428**, 160-163.

Schneider, E. K. (1977). Axially symmetric steady-state models of the basic state for instability and climate studies. Part II: Nonlinear calculations, *J. Atmos. Sci.*. **34**, 280-297.

Schneider, E. K., Bengtsson, L., and Hu, Z. Z. (2003). Forcing of Northern Hemisphere climate trends, *J. Atmos. Sci.*, **60**, 1504-1521.

Schubert, S. D., Suarez, M. J., Region, P. J., Koster, R. D. and Bacmeister, J. T. (2004). Causes of long-term drought in the United States Great Plains, *J. Climate*, **17**, 485-503.

Schultz, H., von Rad, U. and Erlenkeuser, H. (1998). Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years, *Nature*, **393**, 54-57.

Seager, R., Kushnir, Y., Herweijer, C., Naik, N. and Velez, J. (2005). Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856-2000, *J. Climate*, **18**, 4068-4091.

Seager, R., Karspeck, A. R., Cane, M. A., Kushnir, Y., Giannini, A., Kaplan, A., Kerman, B. and Velez, J. (2004). Predicting Pacific decadal variability, In *Earth's climate: The Ocean-Atmosphere Interaction*, Wang, C., Xie, S.P. and Carton, J.A., Eds., American Geophysical Union, Washington D.C., pp105-120.

Seager, R., Murtugudde, R., Naik, N., Clement, A., Gordon, N. and Miller, J. (2003a). Air-sea interaction and the seasonal cycle of the subtropical anticyclones, *J. Climate*, **16**, 1948-1966.

Seager, R., Harnik, N., Kushnir, Y., Robinson, W. and Miller, J. (2003b). Mechanisms of hemispherically symmetric climate variability, *J. Climate*, **16**, 2960-2978.

Seager, R., Battisti, D. S., Yin, J., Gordon, N., Naik, N. H., Clement, A. C. and Cane, M. A. (2002). Is the Gulf Stream responsible for Europe's mild winters, *Quarterly Journal of the Royal Meteorological Society*, **128**, 2563-2586.

Severinghaus, J. P. and Brook E. J. (1999). Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice, *Science*, **286**, 930-934.

Severinghaus, J. P., Sowers, T., Brook, E. J., Alley, R. A. and Bender, M. L. (1998). Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice, *Nature*, **391**, 141-146.

Shaffrey, L. and Sutton, R. (2004). The interannual variability of energy transports within and over the Atlantic Ocean in a coupled climate model, *J. Climate*, **17**, 1433-1448.

Shin, S. I., Liu, Z., Otto-Bliesner, B., Brady, E. C., Kutzbach, J. E. and Harrison, S. P. (2003). A simulation of the Last Glacial Maximum climate using the NCAR-CCSM, *Clim. Dyn.*, **20**, 127-151.

Shindell, D. T., R. L. Miller, G. A. Schmidt and Pandolfo, L. (1999). Simulation of recent northern winter climate trends by greenhouse-gas forcing, *Nature*, **399**, 452-455.

Shuman, B., Webb III, T., Bartlein, P. and Williams, J. W. (2002). The anatomy of a climatic oscillation: vegetation change in eastern North America during the Younger Dryas chronozone, *Quaternary Science Reviews*, **20**, 1777-1791.

Son, S.-W. and S. Lee (2005). The response of westerly jets to thermal driving in a primitive equation model, *J. Atmos. Sci.*, **62**, 3741-3757.

Stephenson, D. B., Hannachi, A. and O'Neill, A. (2004). On the existence of multiple climate regimes, *Quart. J. Royal. Meteor. Soc.*, **130**, 583-605

Stott, L., Poulsen, C., Lund, S. and Thunell, R. (2002). Super ENSO and global climate oscillations at millennial time scales, *Science*, **297**, 222-226.

Swanson, K. (2006). Storm track dynamics, in *The Global Circulation of the Atmosphere*, T. Schneider and A.S. Sobel, Eds., Princeton University Press, Princeton, NJ.

Talley, L. D. (2003). Shallow, intermediate and deep overturning components of the global heat budget, *J. Phys. Oceanogr.*, **33**, 530-560.

Thompson, J. G., Davis, M. E., Mosley-Thompson, E., Sowers, T. A., Henderson, K. A., Zagorodnov, V. S., Lin, P.- N., Mikhalenko, Campen, R. K., Bolzan, J. F., Cole-Dai, J. and Francou, B. (1998). A 25,000- Year Tropical Climate History from Bolivian Ice Cores, *Science*, **282**, 1858-1864.

Timmermann, A., An, S. -I., Krebs, U. and Goosse, H. (2005). ENSO suppression due to weakening of the North Atlantic thermohaline circulation, *J. Climate*, **18**, 3122-3139.

Trenberth, K. and Caron, J. M. (2001). Estimates of meridional atmosphere and ocean heat transports, *J. Climate*, **14**, 3433-3443.

Vellinga, M., Wood, R. A. and Gregory, J. M. (2002). Processes governing the recovery of a perturbed thermohaline circulation in HadCM3, *J. Climate*, **15**, 764-780.

Vellinga, M. and Wood, R. A. (2002). Global climatic impacts of a collapse of the Atlantic thermohaline circulation, *Climatic Change*, **54**, 251-267.

Veronis, G. (1973). Model of world ocean circulation: 1. Wind-driven, two layer, *J. Mar. Res.*, **31**, 228-288.

Wang, X., Auler, A. S., Edwards, R. L., Cheng, H., Cristalli, P. S., Smart, P. L., Richards, D. A. and Shen, C.- C. (2004). Wet periods in northeastern Brazil over the past 210 kyr linked to distant climate anomalies, *Nature*, **432**, 740-743.

Warren, B. A. (1983). Why is no deep water formed in the North Pacific? *J. Mar. Res.*, **41**, 21-26.

Winton, M. (2003). On the climatic impact of ocean circulation, *J. Climate*, **16**, 2875-2889.

Wunsch, C. (2003). Greenland-Antarctic phase relations and millennial time-scale climate fluctuations in the Greenland ice-cores, *Quat. Sci. Rev.*, **22**, 1631-1646.

Yin, J. (2002. The peculiar behavior of baroclinic waves during the midwinter suppression of the Pacific storm track. Ph.D. thesis, University of Washington, 121pp.

Yuan, D., Cheng, H., Edwards, R. L., Dykoski, C. A., Kelly, M. J., Zhang, M., Qing, J., Lin, Y., Wang, Y., Wu, J., Dorale, J. A., An, Z. and Cai, Y. (2004). Timing, duration, and transitions of the last interglacial Asian monsoon *Science*, **304**, 575-578.

Zhang, Y., Wallace, J. M. and Battisti, D. S. (1997). ENSO-like decade-to-century scale variability: 1900-93, *J. Climate*, **10**, 1004-1020.

Zhang, R. and Delworth, T. L. (2005). Simulated tropical response to a substantial weakening of the Atlantic thermohaline circulation, *J. Climate*, **18**, 1853-1860.