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Antarctic sea ice and the control of Pleistocene climate instability

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Abstract. A hypothesis is presented for the origin of Pleistocene climate instability, based on expansion of Antarctic sea ice and associated changes in the oceans' salinity structure. The hypothesis assumes that thermohaline overturning is dominated by the reconfigured conveyor of *Toggweiler and Samuels* [1993b], in which deepwater upwelling is restricted to high southern latitudes. The reconfigured conveyor is shown to be potentially stabilized in an "on" mode by precipitation at high southern latitudes and potentially destabilized into "on" and "off" modes by the counteracting influence of Antarctic sea ice. The mechanism is clarified by the use of a hydraulic analogue. We hypothesize that this mechanism accounts for dominant patterns of thermohaline overturning and climate instability between Pleistocene warm and cold periods. The hypothesis is shown to be consistent with a range of paleoceanographic evidence and to potentially account for details of observed rapid climate changes during glacial and interglacial periods, including aspects of interhemispheric timing.

1. Introduction

Among the important discoveries from the Greenland ice cores is the unusual variability of Greenland climate during glacial periods of the late Pleistocene relative to the more recent warmer Holocene period [*Dansgaard et al.*,1993]. The abrupt character of the transitions between cold states (stadial events) and warm states (interstadial or Dansgaard/Oeschger (D/O) events) suggests that the climate state may have flipped repeatedly back and forth from one mode to another [*Oeschger et al.*, 1984]. Sediment records from the North Atlantic show that North Atlantic surface temperatures tended to vary in concert with Greenland temperatures, supporting the view that the Atlantic thermohaline circulation may have played a role in the transitions [*Lehman and Keigwin*, 1992; *Bond et al.*, 1993; *Broecker*, 1997].

The potential for the thermohaline circulation to exhibit multiple steady states was first demonstrated by *Stommel* [1961] using a hydraulic model consisting of two well-mixed basins connected by deep and shallow tubes, with imposed temperatures and freshwater transport, as shown in Figure 1. The model can be used to illustrate how the stability of the overturning depends on whether the freshwater flux between the basins promotes (Figure 1a) or opposes (Figure 1b) the density differences that drive overturning. In Figure 1a the system is subject to stabilizing feedbacks; if the flow slows for any reason, the salinity in the sinking basin increases, owing to longer residence time, and this feeds back to reinvigorate overturning. This stabilizing feedback allows only a single "on" steady state to exist. In Figure 1b the system is subject to destabilizing feedbacks; any slowing decreases the salinity in the sinking basin, which feeds back to cause further slowing. Depending on the strength of the destabilizing feedbacks, multiple "on" and "off" states are possible. The model thus shows how the potential for multiple steady states may be tied to the sign of the atmospheric freshwater transport [see also *Rahmstorf*, 1996].

Following *Stommel* [1961], a wide range of models of varying complexity have shown that thermohaline overturning circulation can exist in multiple steady states [*Rooth*,1982; *Bryan*, 1986; *Marotzke et al.*, 1988; *Manabe and Stouffer*, 1993; *Stocker and Schmittner*,1997]. On the basis of such models it is widely held [*Broecker*, 1997; *Stocker*, 1998] that the thermohaline circulation in the real ocean can exist in either of two states, one dominated by sinking in the North Atlantic, often known as the conveyor circulation, and a second dominated by sinking at high southern latitudes. This has further led to the hypothesis that the abrupt climate transitions seen in paleorecords might have been caused by transitions between these states. This "bipolar seesaw" hypothesis could potentially account [*Stocker*, 1998] for both the rapidity of Greenland climate changes during D/O events as well as the asynchrony of the temperature changes in the Northern and Southern Hemispheres as observed in sediment [*Charles et al.*, 1996] and ice core [*Blunier et al.*, 1998]

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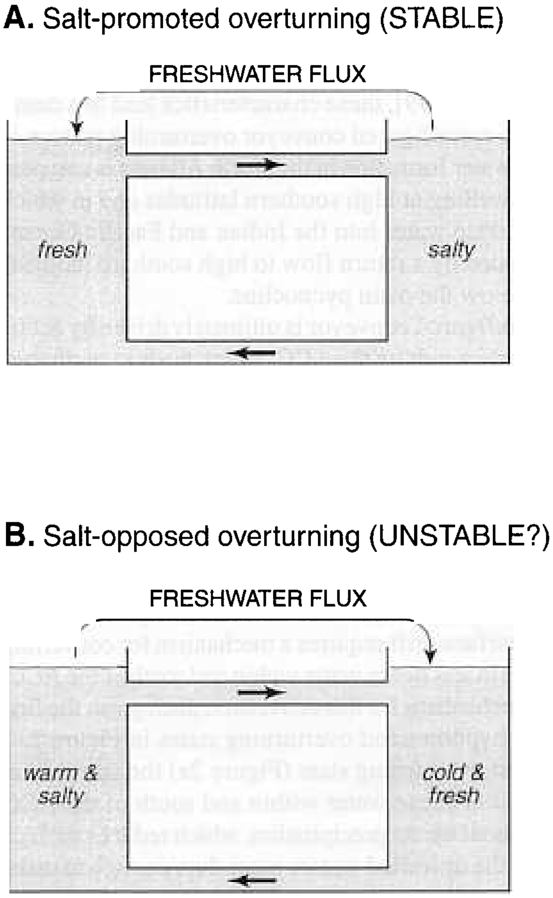

Figure 1. A simplified version of the *Stommel* [1961] two-box model, consisting of two well-mixed basins, shown here with fixed temperatures and a fixed atmospheric freshwater flux between the basins. The overturning flow is taken as proportional to the density differences between the basins, which depends on both salinities and temperatures. The two cases are distinguished on the basis of whether the freshwater flux (a) promotes or (b) opposes the density difference that drives overturning. Multiple steady states are possible in Figure 1b but not Figure 1a [see also *Rahmstorf*, 1996].

records. Several modeling studies have also shown that a conveyor circulation, weakened by an enhanced hydrological cycle, can undergo large spontaneous fluctuations, resembling D/O events [*Sakai and Peltier*, 1997; *Tziperman*, 1997].

These studies do not, however, directly address the issue of why greater instability should necessarily be associated with colder climate. Under colder climate, the hydrological cycle should weaken, which eliminates rather than adds a source of instability, at least according to these climate models. Probably a better candidate for instability is the growth of the Laurentide or Fennoscandian ice sheets, which could destabilize the thermohaline circulation via meltwater discharges [*Birchfield and Broecker*, 1990; *Clark et al.*, 1999]. Indeed there is clear evidence for changes in thermohaline circulation associated with the catastrophic ice discharges, known as Heinrich events [*Bond et al.*, 1992]. However, the meltwater discharges cannot explain why interglacial periods are associated with an "on" mode of the conveyor, periods of intermediate cooling tend to be associated with an unstable, or fluctuating conveyor, and full glacial periods tend to be associated an "off" or at least weakened mode [e.g., *Alley and Clark*, 1999]. In particular, if meltwater exerts such a strong influence on the conveyor, it makes little sense that the major glacial terminations, when the meltwater inputs to the Atlantic are the greatest, should be associated with a transition from an "off" to an "on" mode. If anything, these periods should be among the most difficult of times for turning on the conveyor.

A possible connection between cooling and thermohaline instability has been suggested by Winton [1997], based on the nonlinearity of the equation state of seawater, particularly the weakened dependence of seawater density on temperature at low temperatures. A water column near the freezing point is thereby more easily stabilized by surface freshening than a warmer column. Winton showed how the onset of haline stratification can suppress or destabilize the process of deepwater formation at high latitudes in a cooler climate. While this mechanism is presumably pertinent to general circulation model (GCM) simulations of glacial climate [e.g. *Ganopolski et al.*, 1998], its effectiveness may depend critically on the dynamics of sea ice, which was not accounted for by Winton and is not reliably treated in climate models.

It is well recognized that the models used to study thermohaline circulation are highly uncertain, with the stability limits of the overturning depending on many poorly defined parameters [*Rahmstorf*, 1999; *Knutti et al.*, 2000]. A number of recent studies have highlighted the importance of the effective rate of vertical diffusion, which depends on the magnitude of the vertical diffusion parameter as well as on model resolution and the advection scheme [*Toggweiler and Samuels*, 1998; *Gnanadesikan*, 1999; *Manabe and Stouffer*, 1999]. The effective vertical diffusivity influences not only the strength of the overturning but also the location of deepwater upwelling. A high diffusivity promotes a strong overturning circulation with upwelling occurring predominately at low latitudes, while a low diffusivity allows only a weaker overturning circulation with upwelling restricted to high southern latitudes, [*Toggweiler and Samuels*, 1998; *Gnanadesikan*, 1999]. The low latitude upwelling route tends to be favored in most ocean models, while there is evidence [*Rintoul*, 1991; *Toggweiler and Samuels*, 1993b; *Gnanadesikan and Toggweiler*, 1999] that the southern upwelling route may be more important in the real ocean. Importantly, the stability characteristics of the overturning circulation dominated by southern upwelling have not been explored in models, largely because of the numerical difficulties of running models with sufficiently low vertical diffusion. The overturning pattern involving southern upwelling has been dubbed the "reconfigured conveyor" by *Toggweiler and Samuels* [1993b], to distinguish it from the more familiar conveyor emphasized by *Broecker's* [1991] logo, which involves wide spread upwelling in the Indian and Pacific Oceans.

Here we outline an alternative hypothesis for the origin of Pleistocene climate instability which is motivated by an examination of the stability characteristics of the reconfigured conveyor circulation. We suggest that the reconfigured conveyor can exist in either "on" or "off" modes depending on the freshwater budget of the ocean south of the Antarctic Circumpolar Current (ACC). Specifically, we argue (1) under warm climate conditions, when the budget is dominated by southward transport of atmospheric moisture, only an "on" mode exists; (2) under colder climate conditions, when the freshwater budget is dominated by northward transport of sea ice, both "on" and "off" modes are possible; and (3) under extreme cooling and strong ice transports, the "on" state becomes unstable and only the "off" mode is possible. We hypothesize that this progression accounts for the dominant patterns of thermohaline overturning and climate instability of the Pleistocene.

The progression from a single "on" state, to multiple "on/off" states, and finally to a single "off" state as a function of freshwater forcing is familiar from *Stommel's* [1961] model. Our hypothesis is unique, however, in making a connection between this progression and the freshwater budget of high southern latitudes. Our hypothesis is also unique in emphasizing the importance of sea ice on the upwelling branch rather than the downwelling branch of the thermohaline circulation. Our hypothesis differs from the conventional "seesaw" hypothesis by not permitting a stable "off" mode except under cold climate conditions. Although previous studies have emphasized the sensitivity of the steady-state overturning to Southern Hemisphere moisture transport [*Rahmstorf*, 1996; *Wang et al.*, 1999], our study is the first to emphasize the stabilizing influence of freshwater inputs south of the ACC. This behavior was presumably not found in these previous studies because they were based on models in which the deep water upwells mostly at low latitudes.

In the remainder of this paper, we sketch our hypothesis in greater detail and examine the extent to which it is consistent with paleoceanographic evidence. We also explore its potential to account for details of the D/O events, including aspects of the interhemispheric timing. We emphasize that this is a highly speculative exercise based on qualitative reasoning and hydraulic analogues rather than on definitive models. Given the highly uncertain state of climate modeling, we suggest that much can still be learned through such an approach.

2. Hypothesis

Our basic hypothesis is sketched in Figure 2. The Atlantic overturning circulation is viewed as existing potentially in three qualitatively different steady states, with all three states sharing the characteristic that deep isopycnal surfaces rise to the sea surface (i.e., "outcrop") at high southern latitudes, as allowed by thermal wind balance within the ACC. All three states also share the characteristic that mixing throughout the main pycnocline is efficient along but not across isopycnal surfaces. As shown by *Toggweiler and Samuels* [1998] and *Gnanadesikan* [1999], these characteristics lead to a dominance of the reconfigured conveyor overturning pattern in which deepwater formation in the North Atlantic is compensated by upwelling at high southern latitudes and in which the flow of deep water into the Indian and Pacific Oceans is compensated by a return flow to high southern latitudes at depths below the main pycnocline.

The reconfigured conveyor is ultimately driven by action of the westerly winds on the ACC, which leads to northward surface Ekman drift and deep upwelling [*Toggweiler and Samuels*, 1993b, 1995, 1998). Geostrophic constraints dictate that the upwelling originates from below the depth of the topographic ridges that cross the latitude band of Drakes Passage. As shown by *Marshall* [1997], the magnitude of the northward surface drift depends on combined effects of winds, buoyancy forcing, and eddy-induced mixing and transport within the latitudes of the ACC. In particular, net northward surface drift requires a mechanism for converting dense waters to less dense water within and south of the ACC. Differing mechanisms for this conversion distinguish the first and second hypothesized overturning states in Figure 2.

In the first overturning state (Figure 2a) the conversion of dense to less dense water within and south of the ACC is brought about by net precipitation, which reduces surface salinities of the upwelled waters more than enough to make up for buoyancy losses due to cooling [see, e.g., *Warren et al.*, 1996]. The upwelled waters carried northward in the Ekman drift feed the formation of Antarctic Intermediate Water (AAIW), which is a major component of the return flow into the Atlantic. We assume this realization of the reconfigured conveyor mechanism dominates the Atlantic overturning during interglacial periods such as the Holocene, when the warm climate allows a strong net transport of atmospheric moisture to high southern latitudes.

We argue that this warm-period reconfigured conveyor circulation should not be vulnerable to permanent catastrophic collapse. The stabilization results because any slowing of the overturning leads to increasing density differences between the North and South Atlantic, which feeds back to reinvigorate the overturning. This stability can be understood by considering the freshwater budget of the deep ocean. The budget involves primarily two terms: (1) net input of freshwater due to precipitation at high southern latitudes, which penetrates into the oceans interior, mainly via mixing along isopycnal surfaces, and (2) net input of saltier North Atlantic Deep Water (NADW). If sinking in the North Atlantic were temporarily stalled, e.g., by a sudden discharge of glacial meltwater into the North Atlantic, the balance would be upset, and the deep oceans would then begin to freshen steadily with time. By salt conservation, the surface oceans to the north of the ACC, including, eventually, surface waters in the North Atlantic, would become saltier. These salinity changes, through their effect on density, would eventually

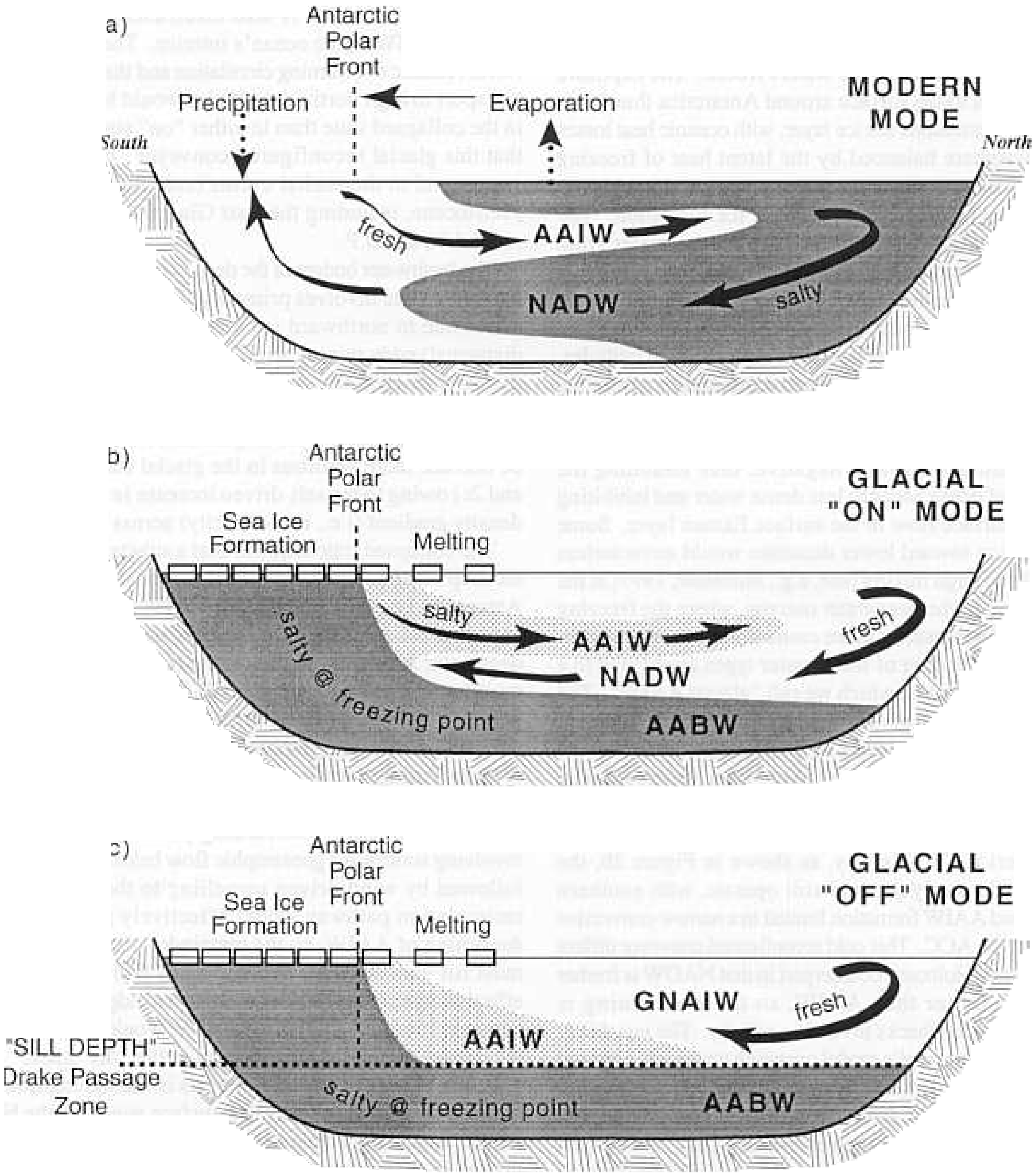

Figure 2. Hypothesized three states of the ocean's thermohaline overturning: (a) modern conditions prevailing in Holocene, (b) Glacial "on" state prevailing in interstadial (warm) events of the Pleistocene, and (c) glacial "off" state prevailing in stadial (cold) events of the Pleistocene.

lead to the onset of deep convection somewhere north of the ACC. The preferred site for convection would presumably be the North Atlantic, where low surface temperatures and high salinities precondition the water column for overturning, although deep convection elsewhere is also conceivable. Onset of convection would densify the water column in the North Atlantic, which in turn would initiate a net southward flow in the oceans interior and a shallower return flow at the surface. In this way, the reconfigured conveyor overturning pattern would start up again. Feedbacks involving southern freshwater input thereby stabilize this realization of the reconfigured conveyor against permanent collapse. The stabilization mechanism is essentially the same as that of *Stommel's* [1961] model for the salt-promoted case (Figure 1a).

The second state (Figure 2b, glacial "on" mode) corresponds to a slightly different realization of the reconfigured conveyor overturning pattern. Here the climate is assumed to have cooled sufficiently that the deepwater temperatures have virtually reached the freezing point, i.e., the potential temperature at which surface waters freeze. The exposure of these waters to the surface around Antarctica thus leads to a virtually continuous sea ice layer, with oceanic heat losses to the atmosphere balanced by the latent heat of freezing because no other significant heat source is available and with northward ice transport balancing net ice formation. The continuous sea ice layer shields the deeper water from atmospheric precipitation. The sea ice and overlying snow together melt once they reach warmer waters to the north. The net effect of this extreme cooling is thus to bring about a reversal of the freshwater budget of high southern latitudes, thereby concentrating salt in the deep ocean and concentrating freshwater in the surface oceans outside of the Antarctic.

In this colder climate the net buoyancy forcing at high southern latitudes must be negative, thus inhibiting the conversion of dense water to less dense water and inhibiting northward surface flow in the surface Ekman layer. Some net conversion toward lower densities would nevertheless be possible through mixing [see, e.g., *Marshall*, 1997] at the northern limit of the deepwater outcrop, where the freezing waters would necessarily make contact with warmer waters to the north. Blending of these water types could lead to a transitional watermass, which we call "glacial AAIW." The circumstances would generally allow this glacial AAIW to be saltier than surface waters in the North Atlantic, but the water would not necessarily be colder. Deep waters forming in the North Atlantic could therefore potentially be dense enough to penetrate beneath the glacial AAIW stratum in the oceans interior. In this way, as shown in Figure 2b, the reconfigured conveyor could still operate, with southern upwelling and AAIW formation limited to a narrow convective band within the ACC. This cold reconfigured conveyor differs from its warmer Holocene counterpart in that NADW is fresher rather than saltier than AAIW, so the overturning is destabilized by feedbacks involving salinity. The instability is analogous to Stommel's model operating under salt-opposed conditions (Figure 1b), i.e., weakening overturing increasing the salinity excess of AAIW relative to NADW and this feeding back to further weaken overturning, etc. We suggest that this glacial reconfigured conveyor "on" mode may correspond to interstadial events (warm periods) of the late Pleistocene.

The third overturning state (Figure 2c, glacial "off" mode) corresponds to a collapsed reconfigured conveyor of state 2. Permanent collapse of the glacial reconfigured conveyor is possible because of the reversal of the salinity feedbacks relative to the modern conveyor. In the collapsed state, dense waters forming in the North Atlantic are insufficiently dense to sink beneath glacial AAIW, so that the connection between sinking in the north and wind-driven upwelling around Antarctica is broken. Following *Boyle and Keigwin* [1987], we designate these northern waters as Glacial North Atlantic Intermediate Water (GNAIW). The circulation of GNAIW can be likened to North Pacific Intermediate Water in the modern ocean, which is also insufficiently dense to sink beneath AAIW in the ocean's interior. The strength of the North Atlantic overturning circulation and the associated heat transport to high northern latitudes would be much weaker in the collapsed state than in either "on" state. We suggest that this glacial reconfigured conveyor "off" mode may correspond to the stadial events (cold periods) of the late Pleistocene, including the Last Glacial Maximum (LGM) around 20 kyr B.P.

The freshwater budget of the deep ocean in the hypothesized collapsed state involves primarily two terms: (1) freshwater losses due to northward ice transport and (2) lateral (i.e., diapycnal) eddy mixing within the ACC, which exchanges the salty deep water with less salty water to the north. Although eddy mixing in the ACC is also potentially important in the Holocene conveyor (Figure 2a), it can be expected to be become more vigorous in the glacial states (Figures 2b and 2c) owing to the salt-driven increase in the north-south density gradient (i.e., baroclinicity) across the ACC.

The collapsed state requires that a substantial fraction of the deep ocean be filled with bottom waters that formed around Antarctica, i.e., with Antarctic Bottom Water (AABW). The expansion of AABW is required because in the absence of deepwater formation in the Northern Hemisphere, AAIW must immediately overlie AABW. The maximum depth to which AAIW can descend en mass is bounded, however, by the depth of the topographic ridges that cross the latitudes of Drake Passage (designated "sill depth" in Figure 2c). Below that depth, AAIW will tend to recirculate back to the surface in a "Deacon cell" overturning pattern [*Bryan*, 1991], involving southward geostrophic flow below the ridge crests followed by wind-driven upwelling to the surface. This recirculation pathway would effectively prevent further deepening of AAIW, so the remainder of the deep oceans must fill with AABW. According to *Warren* [1990], the effective depth of the relevant topographic ridges is somewhere between 1500 and 2500 m, so AABW would necessarily fill the oceans to around these depths in the "off" state.

A key element of our hypothesis is the assumption that the salinity difference between surface waters in the North Atlantic and intermediate waters formed in the southern ocean may have changed sign during cold periods. In the modern ocean the difference amounts to >0.5 g kg^{-1}, so a major rearrangement of the salinity structure of the oceans is clearly required. The required rearrangement may have been possible, we suggest, by the cooling of the deep ocean to the freezing point and by associated changes in the circumstances of sea ice formation around Antarctica.

In the modern ocean the formation of sea ice in the open ocean requires stable haline stratification, otherwise the freezing water in contact with the ice cannot float over warmer waters below. As discussed by *Walin* [1993], however, ice formation over a stable halocline is allowed only for a limited amount of surface cooling. With too much cooling, the halocline is destroyed by the injection of salty brines from the ice, thus leading to enhanced mixing with deeper waters and an associated upward transport of heat that paradoxically, melts the ice. Once this point is reached, further cooling can lead to additional sea ice formation only once the entire water column is cooled to the potential temperature of surface freezing, thereby effectively exhausting the sensible heat reserve of the water column. In the modern ocean, cooling of the full water column to this degree is likely possible only where the bottom is shallower than a few hundred meters, thus restricting the formation of salty bottom waters from brine rejection to shelf regions. In our postulated glacial ocean, however, we suggest that the full water column in the open ocean around Antarctica (as well as the deep ocean to the north) may have been cooled to nearly the potential temperature of surface freezing, thus allowing salty brines to accumulate throughout the water column around Antarctica, including the surface Ekman drift feeding the formation of AAIW.

3. Evidence

3.1. Deepwater temperature and distribution

Our hypothesis calls for dramatic changes in the structure of the deep ocean during cold periods, including an expansion of the volume of the deep ocean filled with AABW, and the cooling of this deep layer to virtually the freezing point. Are these changes consistent with paleoceanographic evidence?

Studies of the Cd content and carbon isotopic composition of benthic foraminifera show that the deep Atlantic south of 45°N and below 2500-3000 m meters was largely filled during the LGM with nutrient rich bottom water originating in the Southern Ocean [*Curry and Lohmann*, 1982; *Boyle and Keigwin*, 1987; *Oppo and Fairbanks*, 1990; *Duplessy et al*, 1988]. Isotopic studies on benthic foraminifera in the Indian Ocean show the presense of a very sharp discontinuity at around 2000 m depth separating glacial intermediate water from bottom water [*Kallel et al.*, 1988]. A somewhat less pronounced front at around 2700 m is also evident in glacial reconstructions of the Pacific [*Duplessy et al*, 1988], although the interpretation in terms of nutrient distributions is complicated by conflicting Cd data [*Boyle*, 1992]. Although questions remain, these results are generally consistent with the changes in AABW distribution required for our "off" mode, i.e., the filling of the oceans with AABW to roughly the depth of the topographic ridges that cross the latitude band of Drake Passage.

Today, the deep oceans below 3 km depth are mostly filled with water with potential temperatures between 1° and 2°C, i.e. 3° to 4°C above the freezing point of surface waters (around -1.9°C). Until recently, it was believed that the deep oceans were ~2°C cooler during the LGM, in which case they would have contained a sensible heat reserve corresponding to a temperature excess of 1°- 2°C relative to the surface freezing point. These estimates were based on the change in oxygen isotopic composition of benthic foraminifera after correcting for the influence of changing continental ice volume [*Fairbanks*, 1989; *Guilderson et al*, 1994]. The recent study by *Schrag et al.* [1996] of sediment pore water at 3000 m depth in the tropical Atlantic suggests, however, that the ice volume contribution to the oxygen isotopic signal may have been overestimated, so that a greater glacial deepwater cooling is required to account for the foraminifera oxygen isotope record. The Schrag et al. estimate places glacial deep waters within 1°C of the surface freezing point. Although pore water studies from more sites are needed before firm conclusions can be drawn, these results suggest that a virtual complete exhaustion of the sensible heat reserve of the deep ocean probably did occur during LGM, as called for by our hypothesis.

3.2. Glacial Sea Ice Limits

Our proposed glacial "on" and "off" states require that the glacial sea ice limits extended at least as far north as the locations of AAIW formation. This raises two questions: (1) At what latitudes should we expect the postulated glacial AAIW to have formed? (2) Did sea ice limits extend this far north during cold periods?

In the traditional view [e.g., *Sverdrup et al.*, 1942, p. 619], AAIW forms from Antarctic surface waters that sink near the Antarctic Polar Front (APF) where they blend with surrounding waters. The APF, in turn, can be associated with the northern limit of deep Antarctic upwelling [*Wyrtki*, 1960; *Taylor et al.*,1978]. Assuming the southward subsurface flow that feeds the upwelling remains in geostrophic balance [*Warren*, 1990; *Toggweiler and Samuels*, 1993a], the location of the APF is constrained to an average latitude of around 55°S, the northern limit of Drake Passage. This constraint on the position of the APF is concordant with the average latitude of the APF estimated from satellite observations by *Moore et al.* [1999]. The position of the front in different sectors of the Southern Ocean is also closely controlled by bottom topography [*Gordon et al.*, 1978; *Moore et al.*, 1999]. These constraints must apply independent of the climate state, so the APF, defined on the basis of these dynamic constraints, cannot shift appreciably between glacial and interglacial periods. The identification of the APF in glacial periods is, of course, ultimately a matter of definition, and other definitions are possible.

The importance of the APF as the formation region for AAIW has been questioned by *McCartney* [1977, 1982], and the origins of AAIW remain controversial [*Molinelli*, 1978, 1981; *England et al.*, 1993]. No doubt, the issue is complicated by the possibility of positive surface buoyancy forcing in the frontal region, which may allow Antarctic surface waters to cross the APF without sinking. In our postulated glacial state, however, AAIW forms by mixing at the northern limit of deep upwelling under negative surface buoyancy forcing. Regardless of uncertainties in modern AAIW formation, the postulated glacial AAIW could only have formed near the position of the modern APF. A suitable test of our hypothesis therefore is that the sea ice limits during glacial periods extended at least as far north as the modern APF.

Seasonal sea ice extent during the LGM has been estimated on the basis of sediment type [*Hays et al.*, 1976, *Burckle et al.*, 1982], ice-rafted volcanic detritus [*Cooke and Hays*, 1982], and fossil plankton assemblages [*Crosta et al.*, 1998a, 1998b]. Although estimates differ in detail, there is a broad consensus that the winter sea ice limits were displaced at least as far north as the modern APF over most of the Atlantic sector and roughly as far north as the modern APF in the Indian and Pacific sectors. Whereas the winter limits seem reasonably robust, there is much less agreement over summer limits, with earlier estimates [*Hays et al.*, 1976; *Cooke and Hays*; 1982] suggesting a substantial northward displacement but with more recent reconstructions [*Crosta et al.*, 1998a,1998b] suggesting little change relative to modern limits. Although more uncertain, none of the reconstructions place the LGM summer ice limits as far north as the modern APF.

Would the lack of summer sea ice at the APF falsify our hypothesis? To address this question requires a seasonal perspective on AAIW formation, which raises complications we have not yet addressed. To avoid a lengthy digression, we defer this subject to Appendix A, where we have sketched a plausible scenario for the seasonal dependence of glacial AAIW formation. Within the context of this scenario, summer ice limits are not a sensitive test of our hypothesis, although the winter limits are clearly important. The scenario thus illustrates a way in which our hypothesis can be reconciled with an absence of summer sea ice at the APF. We merely conclude here that the winter sea ice reconstructions are consistent with the requirements of our hypothesis and accept that the summer data is equivocal.

3.3. Deepwater Temperature and Climate Instability

Our hypothesis predicts that the onset of thermohaline instability should be linked to the near exhaustion of the sensible heat reserve of the ocean below depths of around 2500 m. Can evidence of a link between deepwater cooling and climate instability be found in the paleorecords?

On the basis of a 0.5 million year sediment record from the North Atlantic, *McManus et al.* [1999] have found that climate variability tends to increase whenever the oxygen isotopic composition of benthic foraminifera exceeded a threshold of 3.5‰. The benthic isotopic record is influenced both by changing continental ice volume and by deepwater temperatures. McManus et al. interpreted the threshold as indicating that climate variability increases once the continental ice sheets exceed a particular critical mass. Essentially the same conclusion regarding climate variability and ice volume was arrived at by *Schultz et al.* [1999], comparing the Greenland proxy temperature record over the last 100 kyr with a record of global ice volume reconstructed from stacked planktonic oxygen isotopic records from the western Pacific and summer insolation at 65°N.

It is also possible, however, that the enhanced instability found by *McManus et al.* [1999] may reflect changing deepwater temperatures and not continental ice volume. On the basis of the study of *Chappel and Shackleton* [1986] the threshold of 3.5 ‰ found by McManus et al. was likely attained through a small growth in global ice volume relative to interglacial conditions plus the temperature effects on carbonate composition associated with virtually the full interglacial to glacial deepwater cooling. Given the *Schrag et al.* [1996] study showing that deepwater cooling proceeded to virtually the surface freezing point, the McManus et al. threshold in oxygen isotopic composition therefore appears consistent with our proposal that the onset of instability is linked to the near exhaustion of the sensible heat reserve of deep ocean. The planktonic threshold found by Schultz et al. may also represent water temperature (in this case surface water temperature), which could be reconciled with deepwater control of climate instability assuming surface and deepwater temperatures are sufficiently correlated.

The McManus et al. study also hints at the existence of a second threshold above 3.5 ‰, beyond which climate variability decreases. Although attaining higher values of benthic oxygen isotopic composition can only have been achieved through increased continental ice volume, this second threshold may, nevertheless, also be reconciled with our hypothesis. As clarified further by the model presented below, fluctuations between the glacial "on" and "off" states are only possible over a narrow range in Antarctic freshwater balance. If sea ice-driven losses of freshwater from the Antarctic ocean become too great, the thermohaline circulation becomes stabilized in the "off" state, thus eliminating the potential for sudden mode changes and thereby stabilizing climate. This suggests that the second threshold may correspond to a critical degree of Southern Hemisphere cooling, which is correlated on long timescales (e.g., 100 kyr) with global cooling and hence with global continental ice volume.

3.4. Glacial NADW Formation

An often-cited constraint on glacial NADW formation is the study of *Yu et al.* [1996], based on measurements of $^{213}Pa/^{230}Th$ ratios in glacial (LGM) and modern sediments. Their study indicates that the export of ^{213}Pa from the Atlantic into the Southern Ocean continued at roughly the modern rate during the LGM, and they infer from this that the export of deep (and intermediate) waters formed in the North Atlantic into the Southern Ocean also continued at comparable rates during LGM. If this were true, it would falsify our hypothesized collapse of NADW formation during glacial cold periods.

A weakness of the *Yu et al.* [1996] study, however, is that an export of ^{213}Pa from the deep Atlantic might alternatively have been sustained by a stronger input of southern deep water into the Atlantic which recirculated back to the Southern Ocean at depth. Since the input rate of southern deep waters during glacial periods is unconstrained and very plausibly may have increased [*Oppo and Fairbanks*, 1987], the

$^{213}Pa/^{230}Th$ data by themselves do not place tight constraints on the strength of deep and intermediate water formation in the North Atlantic. Furthermore, the low $^{213}Pa/^{230}Th$ ratios in Atlantic sediments, which Yu et al. use to infer the export rates to the Southern Ocean, might also have been influenced by variations in scavenging at the Atlantic continental margins (i.e., "boundary scavenging").

Probably a more pertinent test of our hypothesis is the glacial deep nutrient distributions. In the modern ocean the nutrient concentrations in the deep Atlantic ocean increase from north to south, following the direction of NADW flow [*Broecker and Peng*, 1982; pp. 28-40]. Reconstructions based on Cd/Ca and $^{13}C/^{12}C$ ratios of benthic foraminifera [*Duplessy et al., 1988; Oppo and Fairbanks, 1990; Boyle, 1992*] show that a nutrient gradient of the same sign persisted below 3000 m throughout glacial times, although the gradient was concentrated farther north [*Oppo and Fairbanks*, 1987]. The sign of the glacial nutrient gradient clearly demands that at least a small source of low nutrient water continued to flow into the deep Atlantic from high northern latitudes. The sign of the gradient is therefore seemingly at odds with our hypothesized glacial ocean in which surface waters in the North Atlantic are unable to sink to great depths because of the freshening of Atlantic surface waters caused by melting Antarctic sea ice.

We may be able to reconcile this observation with our hypothesized glacial ocean if we recognize that while the freshening of the Atlantic surface would shut down open ocean convection, it would not necessarily prevent convection driven by brine rejection under sea ice. In ice-covered shelf areas, or under perennial sea ice, which may have existed in the Norwegian Sea during glacial times [*Kellogg*, 1980], freezing waters may have formed locally with salinities as great as those in the Southern Ocean and thus have been sufficiently dense to sink into the abyssal North Atlantic. Small amounts of deep water formed this way may have accounted for the low-nutrient water present in the deep glacial North Atlantic. This perspective on NADW shutdown is consistent with the study of *Dokken and Jansen* [1999], who use benthic isotopic measurements from high-resolution sediment cores in the Nordic Seas to argue that deepwater formation in the North Atlantic occurred exclusively by brine rejection from sea ice during the stadial events.

Sarnthein et al [1994] have noted the existence of three dominant patterns (modern, glacial, Heinrich) in the distribution of low-nutrient waters in the deep North Atlantic [see also *Alley and Clark*, 1999]. Owing to the complicating effects of northern sea ice, we should not expect a simple mapping to exist between these nutrient patterns and the three hypothesized overturning modes depicted in Figure 2.

3.5. Modern Ocean Controls

One clearly contestable aspect of our hypothesis is the depiction of the modern ocean. We argue that under warm climate conditions such as the Holocene, the Atlantic overturning circulation is self-stabilizing in an "on" mode; the overturning has no stable "off" mode and therefore no latent potential for permanent collapse. This self-stabilization is linked to feedbacks involving freshwater input at high southern latitudes.

We are well aware that this depiction is in clear contradiction to the results from numerous ocean general circulation models (OGCM) which suggest that both "on" and "off" states are possible under warm climate conditions (e.g., *Manabe and Souffer*, 1993; *Rahmstorf*, 1996; *Stocker and Schmittner*, 1997]. Using a representative model exhibiting multiple steady states, *Rahmstorf* [1996] has shown that the origin of multiple states in these models, as by *Stommel's* [1961] simple model, is tied to the overturning being salt opposed. Rahmstorf further argues that the available hydrographic data suggest that the same is true for the Atlantic overturning in the real ocean, thus contradicting our hypothesized ocean as drawn in Figure 2a.

We suggest, however, that the case is not closed. Throughout the North and South Pacific Oceans and most of the south Indian Ocean, the NADW stratum in the real ocean is overlaid by a continuous salinity minimum layer which results from the penetration of freshwater from high latitudes via isopycnal mixing into the oceans interior [see, e.g., *Ried and Lynn*, 1971]. Given this salinity structure, the first and possibly most crucial step in the conversion of NADW to less dense water is indisputably reinforced by salinity differences in the real ocean. Although the salinity minimum layer is absent in the north Indian Ocean, transect studies constrained by silica indicate that this is not a major region for NADW upwelling [*Robbins and Toole*, 1997].

The inability to simulate deep and intermediate salinity distributions in the oceans is a common failing of three-dimensional OGCMs [see *Duffy and Caldeira*, 1997, and references therein]. The deepest waters are typically too fresh and the overlying intermediate waters too salty. *Duffy and Caldeira* [1997] suggest that this deficiency may be related to sea ice effects. An improper representation of the relative rates of isopycnal versus diapycnal mixing in the main pycnocline may also be relevant, as suggested by the tendency for models to have too much upwelling of deep water at low latitudes [*Toggweiler and Samuels*, 1993b; *Gnanadesikan*, 1999; *Gnanadesikan and Toggweiler*, 1999]. Excessive vertical mixing will exaggerate the tendency of salty low-latitude surface to penetrate downward into depth range of the intermediate waters. These deficiencies indicate that OGCMs have difficulty in properly simulating the influence of freshwater forcing on deep overturning and therefore very possibly also misrepresent the stability characteristics of the overturning. Although our depiction of the controls on modern overturning in Figure 2 is admittedly oversimplified, it may nevertheless illustrate a valid stabilizing feedback involving the freshening of the deep ocean around Antarctica by isopycnal mixing that is underrepresented in many models.

4. Model Illustrations

4.1. Hydraulic Analogue

To help clarify our basic hypothesis and to illustrate some of its explanatory power, we present in Figure 3 a simple model intended as a possible hydraulic analogue to the reconfigured conveyor. The model consists of a container that is partitioned into three basins by barriers which penetrate to intermediate depths. From south to north the basins are intended to roughly represent (1) the Antarctic Ocean water column south of the APF, (2) subantarctic waters, specifically waters throughout the ocean's interior that outcrop near the APF, and (3) the water column in the North Atlantic. The Antarctic and subantarctic waters are separated by a weir, which blocks all flow to an intermediate depth. This weir represents the barrier to north-south geostrophic transport in the latitudes of Drake Passage, so the weir extends to the depth corresponding to crests of topographic ridges that cross these latitudes. A siphon pump, representing Ekman transport, draws water from the surface of the Antarctic basin and deposits it onto the surface of subantarctic basin. The subantarctic basin is connected to the North Atlantic basin by a shallow tube, with a finite flow impedance, which represents frictional boundary flow. The major water masses (separated by dashed lines) are assumed to be well mixed, with no subsurface mixing between types and no impedance to lateral flow, except through the tube. Water at the surface of the subantarctic basin is assumed to represent newly formed AAIW, i.e., waters which sink near the APF. Water which forms at the surface of the Antarctic and North Atlantic basins are designated as Antarctic Surface Water (AASW) and North Atlantic Surface Water (NASW). When these same water types penetrate into the ocean interior they are assumed to represent AABW and NADW.

We first consider the behavior of the model when the water density is prescribed at the surface of each basin. Depending on the relative density of the water types, the model can operate under three possible conditions, analogous to the three overturning modes (modern, glacial "on," and glacial "off") shown previously in Figure 2. The modern overturning mode is obtained when $\rho_{NASW} > \rho_{AASW}$ and $\rho_{NASW} > \rho_{AAIW}$. In this case, NASW (as NADW) fills the deep interior. The action of the pump drives an upwelling of these waters in the Antarctic basin, and upon exposure to the Antarctic surface, these waters are modified to form a thin, buoyant surface layer. The pump skims off this layer, depositing it in the subantarctic basin as AAIW. The pump inflow to the subantactic basin is counteracted by an outflow of AAIW through the tube into the northern basin. The outflow depends on the hydrostatic pressure difference across the ends of the tube, which depends in turn on the density difference between northern and subantarctic waters and on the depth of the fluid interface, or "pycnocline," where northern and southern waters make contact below the surface in the subantarctic basin. If a balance between inflow and outflow in the subantarctic basin is not initially achieved, the pycnocline will shoal or deepen

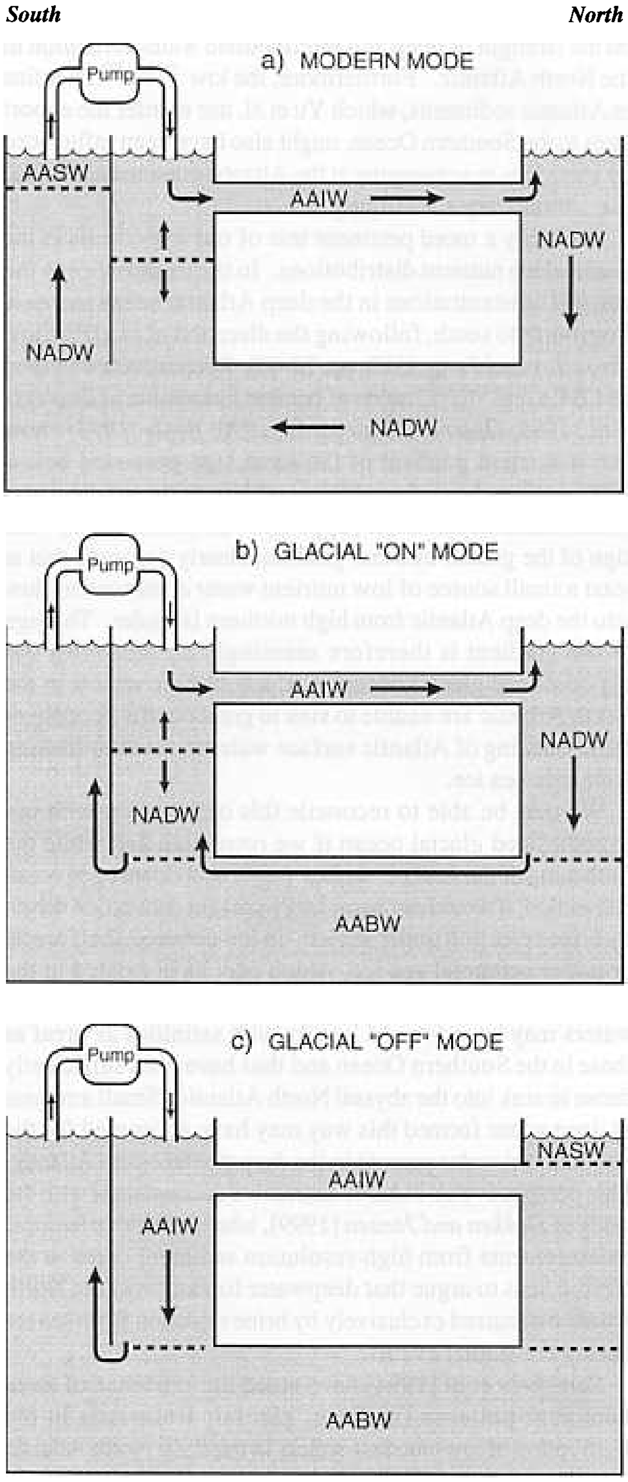

Figure 3. Hydraulic box model of the reconfigured conveyor, emphasizing the role the northward Ekman drift across the latitudes of Drake Passage, designated by the pump, which builds up a pool of low-density water Antarctic Intermediate Water (AAIW) north of Drake Passage. This in turn drives a northward pressure-driven flow through the tube, which is compensated by southward flow below. Water types are separated by dashed lines. Three qualitatively different overturning patterns Figures 3a-3c are possible, depending on the relative densities in the different surface source regions (see text). The model is inspired by the scale analysis of *Gnanadesikan* (1999).

until a balance is achieved. This balance has been discussed also by *Gnanadesikan* [1999].

Our postulated glacial "on" mode is obtained when $\rho_{NASW} < \rho_{AASW}$ and $\rho_{NASW} > \rho_{AAIW}$. In this case, AASW (as AABW) fills the deep interior below the depth of the weir. Above this depth an overturning pattern persists with NADW sinking in the north and AAIW flowing through the tube. The only difference from the modern mode is that the flow of NADW into the Antarctic basin is now restricted to a thin sheet, which "spills" upward under the base of the weir toward the Antarctic surface, where it is modified to Antarctic surface water. The filling of the deep ocean has no direct effect on the overturning strength, which is still tied to the pump flow via the same balance as the modern mode. If the hydraulic model were modified to incorporate a finite flow impedance in the deep waters, the glacial "on" mode would differ in that a slight deepening of NADW below the depth of the weir might be needed to drive the deep southward flow of NADW. This would be more compatible with the perspective in Figure 2.

Our postulated glacial "off" mode is obtained when $\rho_{NASW} < \rho_{AASW}$ and $\rho_{NASW} < \rho_{AAIW}$. This mode is similar to the glacial "on" mode, in that AABW fills the ocean interior below the weir. The mode differs from the glacial "on" mode, however, in that AAIW now floods the full water column above weir depth in the subantarctic and northern basins, except for a thin inactive layer of NASW that floats at the surface in the North Atlantic basin. The northward transport by the pump is now balanced by southward flow of AAIW which spills under the weir.

The hydraulic model helps clarify several points pertinent to the earlier qualitative discussions in section 2: (1) The modern mode is associated with positive buoyancy forcing at the Antarctic surface, while the glacial modes are associated with negative buoyancy forcing, so the transition between the modern and glacial "on" modes occurs when the buoyancy forcing at the Antarctic surface changes sign. (2) The Antarctic surface is able to freeze over in the modern mode only if a layer of fresh water caps the Antarctic basin, thus insulating the ice from the warmer deep water. In this case, the water flowing through the pump must be fresher than NADW, so it is difficult to reverse the salinity difference between AAIW and NADW through sea ice processes in the modern mode. (3) The glacial "on" and "off" modes, in contrast, may allow relatively salty water to flow through the pump if the deep ocean is cooled sufficiently to permit ice cover in spite of Antarctic convection. The glacial "on" mode therefore permits (but does not require) AAIW to be saltier than NADW, as shown in Figure 2.

4.2 Multiple Steady States

A key element of our hypothesis is the assertion that both glacial modes can exist for certain ranges of external forcing, making abrupt transitions possible. To explore this possibility in the context of the hydraulic model of the previous section, it is necessary to consider separately influences of temperature and salinity on water density. Rather than simply adding freshwater forcing to the model in Figure 3, we first modify the model by adding an Indo-Pacific surface basin which intermixes at prescribed rates M_1 and M_2 with the subantarctic and North Atlantic basins, as shown in Figure 4. This modification allows for a rudimentary description of several pertinent processes, including atmospheric moisture transport, surface-water mixing, and overturning circulation, that control

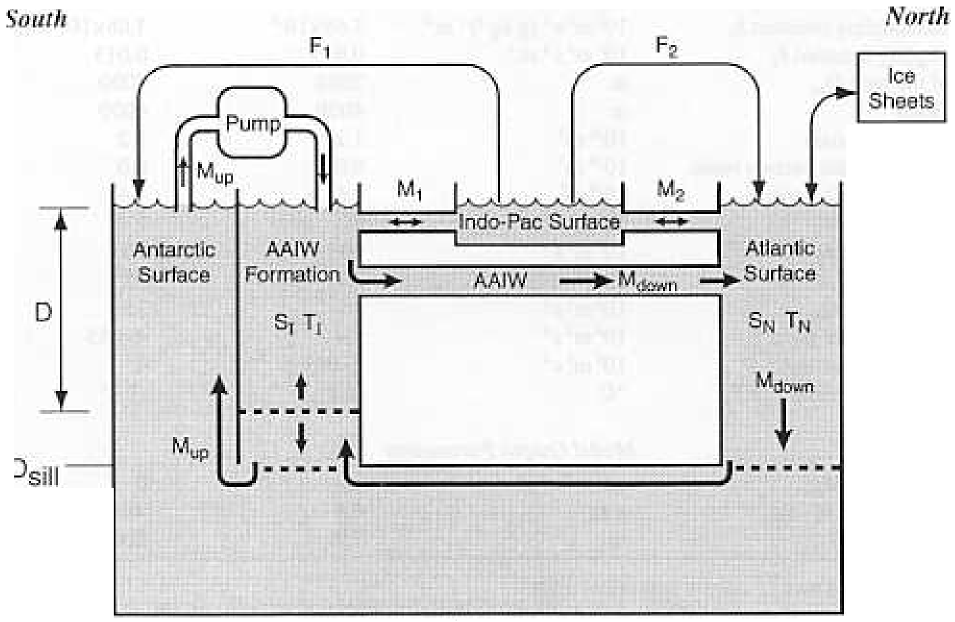

Figure 4. Modified hydraulic box model of Atlantic's thermohaline overturning, here with prescribed freshwater fluxes F_1 and F_2 and prescribed surface temperatures T_I and T_N.

the salinity differences between North Atlantic and subantarctic waters. Although we recognize that there is an element of arbitrariness to the surface divisions in this modified model, we suggest that the divide between the Indo-Pacific and subantarctic basins is best viewed as being aligned with the frontal zones of the ACC, while the divide between the Indo-Pacific and North Atlantic basins is best viewed as being located somewhere in the South Atlantic between the Equator in southern tip of Africa. With this interpretation, the North Atlantic basin includes the water column in the North Atlantic as well as Atlantic surface waters extending south of the Equator.

The model employs "mixed" boundary conditions consisting of prescribed freshwater transports F_1 and F_2, and prescribed surface temperatures. We assume that the surface outcrop of the subantarctic waters near the APF has negligible surface area, so these waters do not receive freshwater transport directly. In principle, the surface temperatures could be prescribed in all basins, although only the subantarctic and Atlantic temperatures are dynamically relevant. The temperature of the Antarctic surface waters is implicitly relevant, however, in two ways: (1) by influencing the magnitude of prescribed Antarctic freshwater flux F_1 and (2) by influencing the Antarctic surface buoyancy forcing and hence the distinction between modern and glacial "on" modes. In fact, the later distinction is relevant dynamically only in terms of the fate of F_1: in the modern mode the Antarctic freshwater flux is skimmed off the Antarctic surface and deposited in the subantarctic basin, whereas in the glacial modes the flux is mixed downward, consistent with negative surface buoyancy forcing.

The density and depth dependence of the overturning flow are taken from the scale analysis of *Gnanadesikan* [1999]. Water sinks in the Atlantic basin at a flow rate

$$M_{\text{down}} = k_1 \, D^2 [\, \beta(S_N - S_I) - \alpha(T_N - T_I) \,] \qquad (1)$$

where k_1 is a constant related to efficiency of boundary flow, D is the depth of the internal fluid boundary, or pycnocline, and β and α are linearized density coefficients. Water upwells in the Antarctic basin at a rate

$$M_{\text{up}} = M_{\text{Ek}} - k_2 D \qquad (2)$$

where M_{Ek} is the northward Ekman transport across the latitudes of Drake Passage, and k_2 is a constant related to eddy-induced transport across the circumpolar current. Conservation of mass of the subantarctic waters requires that

$$A \, dD / dt = M_{\text{up}} - M_{\text{down}} \qquad (3)$$

where A is the (sub)surface area of the subantarctic waters. The only other relations are conservation of salt in the

Table 1. Hydraulic Model Parameters

	Units	Modern Value	Glacial Value
	Model Input Parameters		
Density temperature coefficient α	g kg^{-1} °C^{-1}	0.15	0.15
Density salinity coefficient β	dimensionless[a]	1	
Hydraulic coupling constant k_1	10^6 m^3 s^{-1} (g kg^{-1})$^{-1}$ m^{-2}	3.68×10^{-5}	3.68×10^{-5}
Eddy coupling constant k_2	10^6 m^3 s^{-1} m^{-1}	0.013	0.013
Depth of the weir D_{sill}	m	2000	2000
Ocean depth	m	4000	4000
Volume Atlantic basin	10^{16} m^3	1.2	1.2
Volume Indo-Pacific surface basin	10^{16} m^3	6.0	6.0
Area of Subantarctic basin	10^{14} m^2	3.6	3.6
Interbasin mixing M_1	10^6 m^3 s^{-1}	7	7
Interbasin mixing M_2	10^6 m^3 s^{-1}	10	10
Average ocean salinity S_0	g kg^{-1}	35	35
"Pump" transport M_{Ek}	10^6 m^3 s^{-1}	25	35
Southern freshwater input F_1	10^6 m^3 s^{-1}	0.4	-0.065
Northern freshwater input F_2	10^6 m^3 s^{-1}	-0.092	0
Temperature difference $T_N - T_I$	°C	-1.5°	.0°
	Model Output Parameters		
Overturning flow M_{down}	10^6 m^3 s^{-1}	15	0
Salinity difference $S_N - S_I$	g kg^{-1}	0.5	-0.33
Pycnocline depth D	m	750	2000

[a]Density is treated here in salinity equivalent units.

[b]For the simulations of short-lived interstadial events, as shown in Figure 8, the Indo-Pacific surface basin volume was reduced to 1.2×10^{16} m^3 (see text). The other simulations are not sensitive to the choice of this volume.

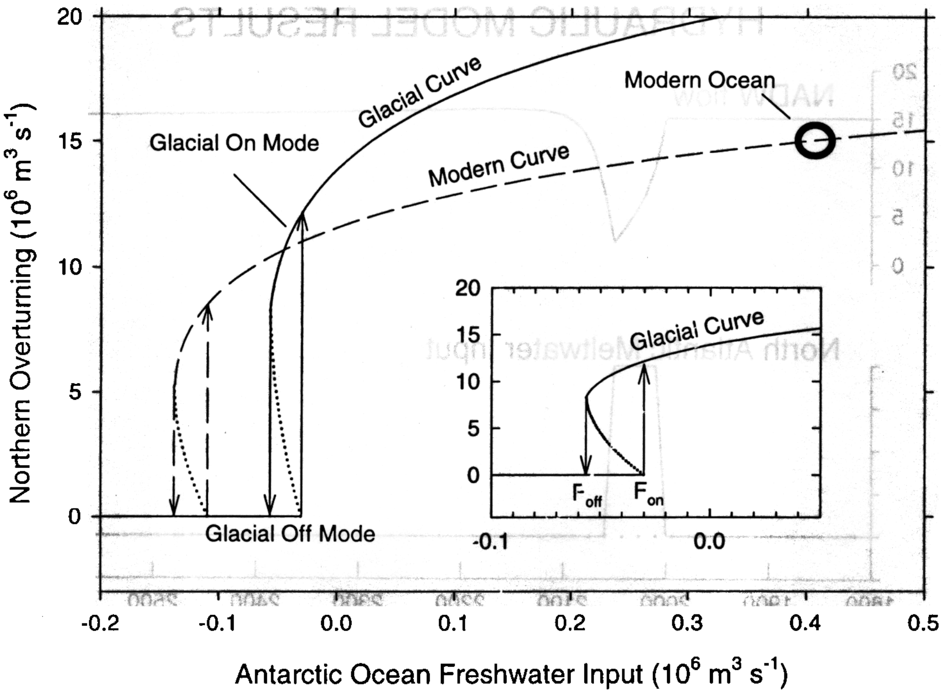

Figure 5. Steady state overturning of the hydraulic model as a function of the Antarctic Ocean freshwater input F_1. Curves are shown for both modern and glacial parameters. The curves are generated by keeping all parameters fixed except F_1 (see Table 1). The insert shows an expanded portion of the glacial curve. The "kink" in the glacial curve at a northern overturning rate of ~8.3×10^6 m^3 s^{-1} arises because the pycnocline depth D has deepened to the sill depth D_{sill}. The dotted portions of the curves are unstable.

individual water masses. Solutions are sought subject to the constraints that $M_{down} > 0$, $M_{up} > 0$, and $D < D_{sill}$, where D_{sill} is the weir depth (roughly the depth of the topographic ridges in the latitudes of Drake Passage). If the pycnocline D deepens to D_{sill}, then AAIW is allowed to recirculate under the wier to the Antarctic surface at a rate sufficient to prevent further deepening.

With these parameterizations the modified model takes account of wind-driven Antarctic upwelling, eddy transports and mixing within the ACC, changes in pycnocline depth, the role of temperature forcing, and advective feedbacks on salinity. The model has been "tuned" to crudely match modern estimates of the Atlantic overturning rate, to match the observed differences in salinity and temperature between AAIW and NADW, to match the Southern Ocean freshwater input, and to match estimates of southern Ekman transports. We apply the model alternatively to modern conditions using these parameter values or to glacial conditions using slightly different parameters, as summarized in Table 1.

The model exhibits stable steady state solutions, which are easily derived analytically. Figure 5 shows a plot of the steady state overturning rate M_{down} versus freshwater input into the southern ocean F_1. The functional form is qualitatively similar to that of Stommel's model, and admits a region of multiple steady states, with an "on" state, and "off" state, and an intermediate state, corresponding to the "overhang" (dotted curve). As for *Stommel*'s [1961] model, the overhanging solution is an unstable saddle point, and bifurcations occur where this solution merges with the "on" and "off" states. The saddle point bifurcations can be characterized by two critical values of the freshwater flux $F_1 = F_{off}$, and $F_1 = F_{on}$. Multiple steady states are possible only over a narrow range between F_{on} and F_{off}. More detailed analysis [*Keeling*, 2000] confirms that this region of multiple steady states requires that AAIW be saltier than NADW, in agreement with the mechanism proposed in Figure 2.

The glacial model parameters differ in three ways from modern parameters. First, the temperature difference between NADW and AAIW is reduced, consistent with both water masses approaching but not reaching the freezing point. This change reduces the separation between F_{on} and F_{off}. Second, the net freshwater loss from the Atlantic basin ($-F_2$) is reduced (here to zero), consistent with less low-latitude evaporation in a colder climate or with a direct transport of freshwater from the melting of Antarctic sea ice into the Atlantic. This has the effect of translating the curve to the right in Figure 5, thereby shifting both F_{off} and F_{on} toward zero. Third, the strength of the southern Ekman transport is increased, consistent with higher winds in a colder climate [*Petit et al.*, 1981]. This has the effect of increasing the magnitude of the overturning for fixed freshwater forcing.

Importantly, all three of these changes, which are reasonable

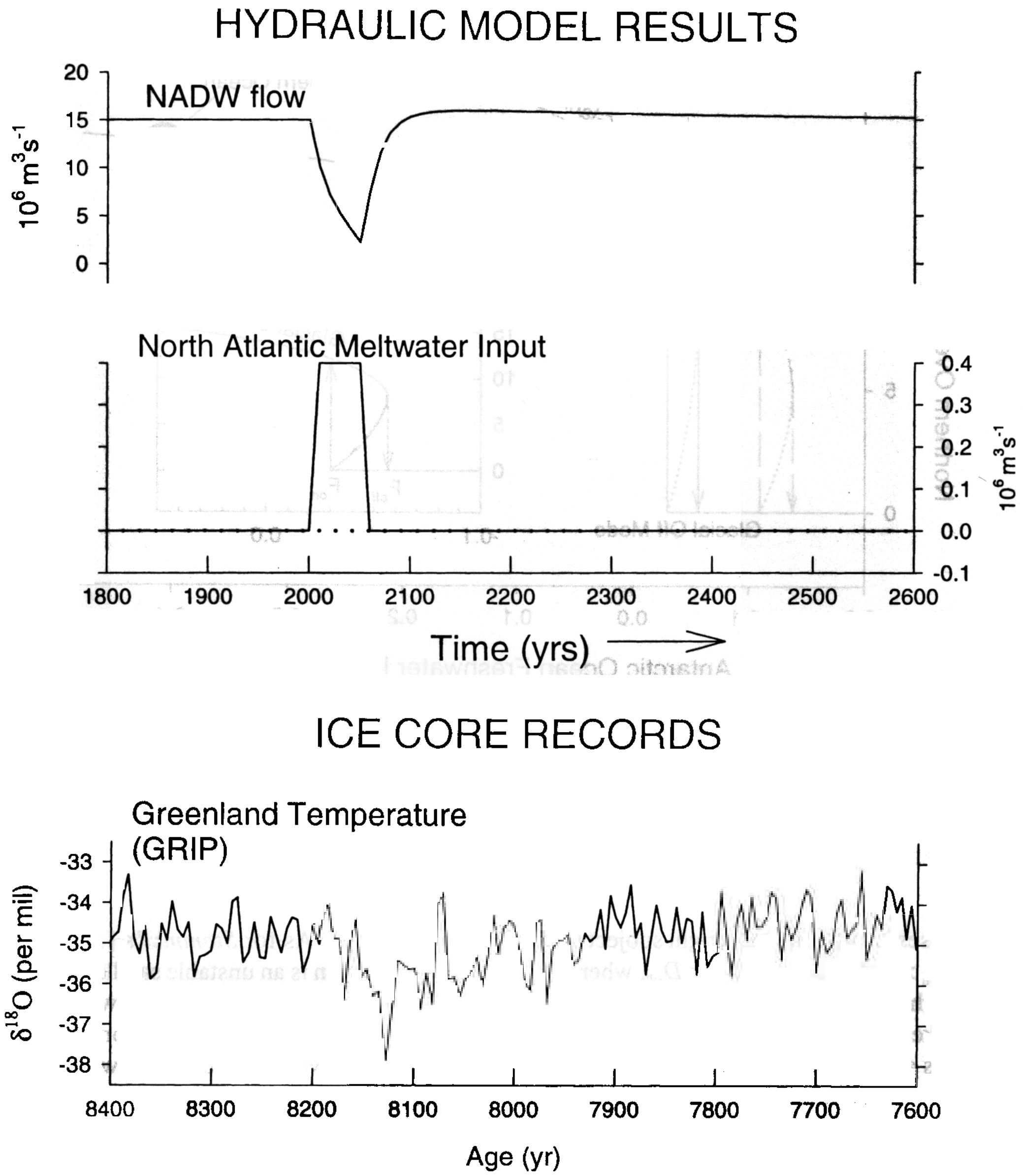

Figure 6. Response of hydraulic model with modern parameters (see Table 1) to transient freshwater input to the Atlantic basin. Also shown for comparison is the Greenland Ice Core Project (GRIP) ice core record [*Dansgaard et al.*, 1993] of the 8 kyr B.P. cold event, showing the change in Greenland climate, possibly related to freshwater input from the draining of Lake Agassiz.

to postulate for a colder climate, shift the steady state solutions so as to allow multiple steady states when the Antarctic freshwater balance F_1 is near zero. Although the exact range depends on parameters (F_2, M_2, and M_1) which are poorly defined, no reasonable choice of these parameters allows multiple steady states for F_1 anywhere near modern values. A regime shift in Antarctic freshwater balance is clearly required. Our earlier discussions and the hydraulic system in Figure 3 offer a simple suggestion for how such a regime shift may have occurred, involving the exhaustion of the sensible heat reserve of the deep ocean and associated changes in Antarctic sea ice formation and in the salinity of Antarctic surface waters. The model thus helps to illustrate how the origin of multiple steady states may be tied to a regime shift in Antarctic surface waters. Nevertheless, because the model does not explicitly treat the thermodynamic controls on sea ice formation and deep ocean temperature, its application is restricted to scenarios involving either glacial or modern (interglacial) conditions but not glacial-interglacial transitions.

4.3. Holocene Cold Event

To illustrate the stabilization of the model caused by Antarctic freshening, we run the model for modern conditions

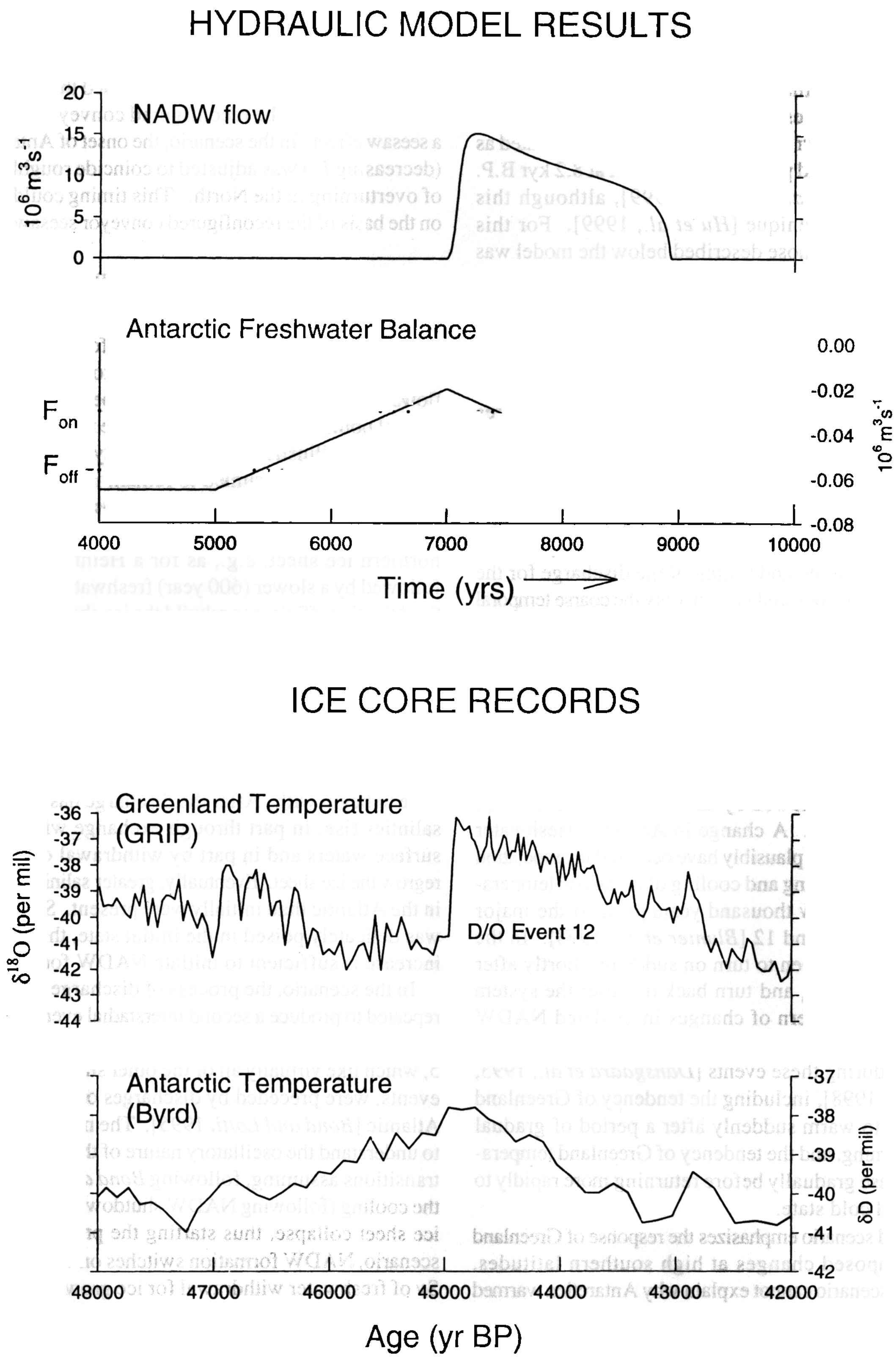

Figure 7. Response of hydraulic model with glacial parameters (see Table 1) to transient changes in Antarctic freshwater budget. Also shown for comparison are ice core records [*Johnsen and Dansgaard*, 1972; *Dansgaard et al.*, 1993], as synchronized by *Blunier et al.* [1998], for the period of Dansgaard/Oeschger (D/O) event 12, in which antecedent Antarctic warming (possibly linked to changes in sea ice coverage) appears to have triggered sudden warming in Greenland and in which transient Antarctic cooling anticipated cooling in Greenland.

and subject the Atlantic basin to a transient discharge of freshwater sufficient to raise global sea level by 1.5 m over a period of 50 years, as shown in Figure 6. A discharge of slightly smaller magnitude likely occurred from the sudden draining of glacial lakes Agassiz and Ojibway some time between 8.2 and 8.7 kyr B.P., which has been implicated as a cause of the brief cold period that occurred at 8.2 kyr B.P. [*Alley et al.*, 1997; *Barber et al.*, 1999], although this interpretation is not unique [*Hu et al.*, 1999]. For this simulation as well as those described below the model was run by forward time stepping the governing equations.

It is seen that the overturning rate slows markedly in response to the freshwater discharge but recovers quickly (with a slight overshoot) after the discharge ends. The simulated changes in overturning are qualitatively similar to the ice core proxy record of Greenland temperatures associated with the 8.2 kyr B.P. cold event. Details of the simulation depend on the temporal pattern of discharge as well as on the effective volume of the North Atlantic basin into which the discharge is promptly mixed in the model. Because these parameters are somewhat arbitrarily chosen and because magnitude and timing of the discharge for the 8.2 kyr cold event are not well known, only the coarse temporal pattern is significant.

4.4. Long-Lived Interstadial Events

Next, we switch to the glacial conditions, and explore the response of the model to an imposed sequence of changes in Antarctic freshwater balance consisting of an upward ramp across F_{on} and F_{off}, followed by an equal downward ramp, as shown in Figure 7. A change in Antarctic freshwater balance of this sort may plausibly have occurred in association with the gradual warming and cooling of Antarctic temperatures that began a few thousand years prior to the major interstadial events 8 and 12 [*Blunier et al.*, 1998]. In the scenario, NADW is seen to turn on suddenly shortly after the system crosses F_{on} and turn back off after the system crosses F_{off}. The pattern of changes in modeled NADW formation rate matches the general pattern seen in Greenland temperatures during these events [*Dansgaard et al.*, 1993, *Blunier et al.*, 1998], including the tendency of Greenland temperatures to warm suddenly after a period of gradual Antarctic warming, and the tendency of Greenland temperatures to decrease gradually before returning more rapidly to the full glacial cold state.

This second scenario emphasizes the response of Greenland climate to imposed changes at high southern latitudes. Although the scenario cannot explain why Antarctica warmed in the first place prior to interstadials 8 and 12, it may offer explanations for why Antarctica subsequently cooled back down. First, the onset of the conveyor circulation can be expected to cool the southern high latitudes via the heat transported in the South Atlantic by the overturning flow (i.e., $T_l > T_N$). Second, the model also predicts that the onset of overturning is associated with a shoaling of the pycnocline (D), which would be expected to cool high southern latitudes via reduced southward eddy heat transport. Both processes have analogues in the normal seesaw mechanism [*Stocker*, 1998]. Thus, in a model which included thermal feedbacks, we would expect the reconfigured conveyor to also exhibit a seesaw effect. In the scenario, the onset of Antarctic cooling (decreasing F_1) was adjusted to coincide roughly with onset of overturning in the North. This timing could be justified on the basis of the reconfigured conveyor seesaw mechanism.

4.5. Short-Lived Interstadial Events

Although the model emphasizes Antarctic freshwater balance as the ultimate control point for NADW formation, the model can also be very sensitive to transient freshwater flows in the Northern Hemisphere, especially if the Antarctic control point is delicately poised. The third scenario, shown in Figure 8, illustrates this possibility.

The system is assumed to remain in a state with F_1 just smaller than F_{off}. The Atlantic is exposed to a brief (50 year) freshwater pulse to simulate a sudden ice discharge from the northern ice sheet, e.g., as for a Heinrich event. This is followed by a slower (600 year) freshwater withdrawal from the Atlantic sufficient to rebuild the ice sheet. The overturning responds by NADW flickering on, as for a short-lived interstadial event, near the end of the regrowth stage. The response can be understood by considering the evolution of Atlantic surface salinity: The Atlantic salinity is initially driven lower due to the meltwater input. This freshening has no influence on NADW formation, which was in the "off" state to begin with. After the discharge has ended, Atlantic salinties rise, in part through exchange with Indo-Pacific surface waters and in part by withdrawal of freshwater to regrow the ice sheet. Eventually, greater salinities are achieved in the Atlantic than initially were present. Since the system was delicately poised in the initial state, this small salinity increase is sufficient to initiate NADW formation.

In the scenario, the process of discharge and regrowth is repeated to produce a second interstadial event. The sequence can be compared to the short-lived interstadial events 6 and 5, which like virtually all of the other short-lived interstadial events, were preceded by discharges of glacial ice into the Atlantic [*Bond and Lotti*, 1995]. The model thus allows us to understand the oscillatory nature of the interstadial/stadial transitions assuming, following *Bond and Lotti* [1995], that the cooling (following NADW shutdown) somehow triggers ice sheet collapse, thus starting the process anew. In the scenario, NADW formation switches on with as little as 0.03 Sv of freshwater withdrawal for ice growth, a flux which is the same order of magnitude as postulated regrowth rates for wasted ice sheets [*MacAyeal*, 1993].

The duration of the simulated interstadial events depends on how long ice sheet growth continues following the onset of northern overturning. Several adjustments were needed here to produce short events. First, it was necessary to terminate the regrowth of the ice shortly after the onset of

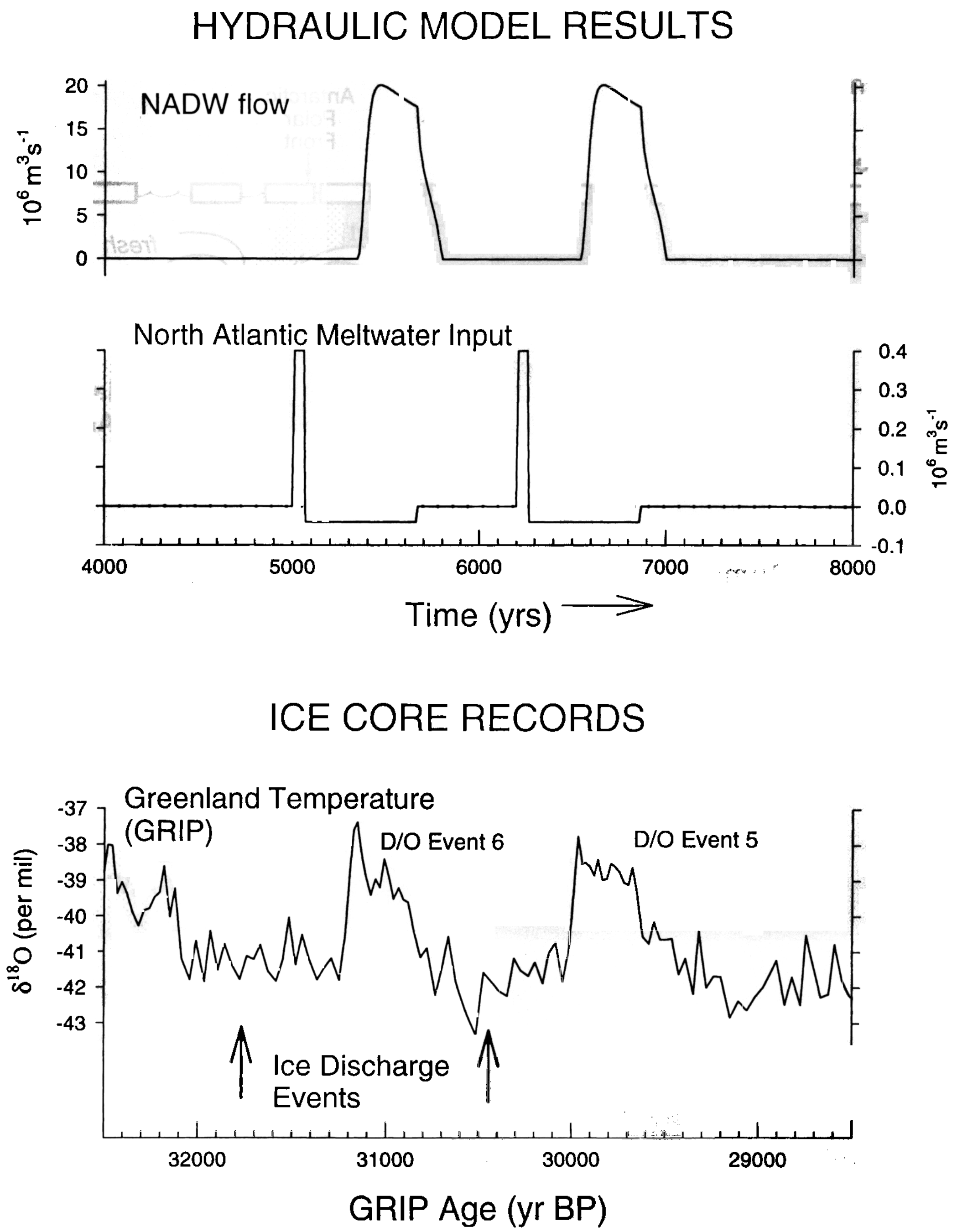

Figure 8. Response of hydraulic model with glacial parameters (see Table 1) to transient freshwater input and losses from the Atlantic basin, simulating thermohaline response to sudden glacial collapse and subsequent slower regrowth. Also shown for comparison is the GRIP ice core record spanning D/O events 5 and 6 [*Dansgaard et al.*, 1993] and approximate times of ice discharge events into the North Atlantic based on *Bond and Lotti* [1995]. Relative timing of the ice discharge events and Greenland temperature changes is not well constrained.

overturning, as seen in Figure 8. Second, it was necessary to reduce the volume of the Indo-Pacific surface basin to roughly the size of the Atlantic basin. Both are potentially justifiable adjustments, the first on the grounds that the onset of overturning might lead to enhanced atmospheric moisture transport and hence rapid "topping up" of the ice sheet, the second on the grounds that only a small portion of the Pacific and Indian surface oceans can intermix with the Atlantic on such short timescales.

Another explanation for the short duration of the events also suggests itself, however. As discussed previously, the onset of overturning can be expected to produce cooling at high southern latitudes via a seesaw effect. This cooling can be expected to enhance sea ice formation and hence tend to

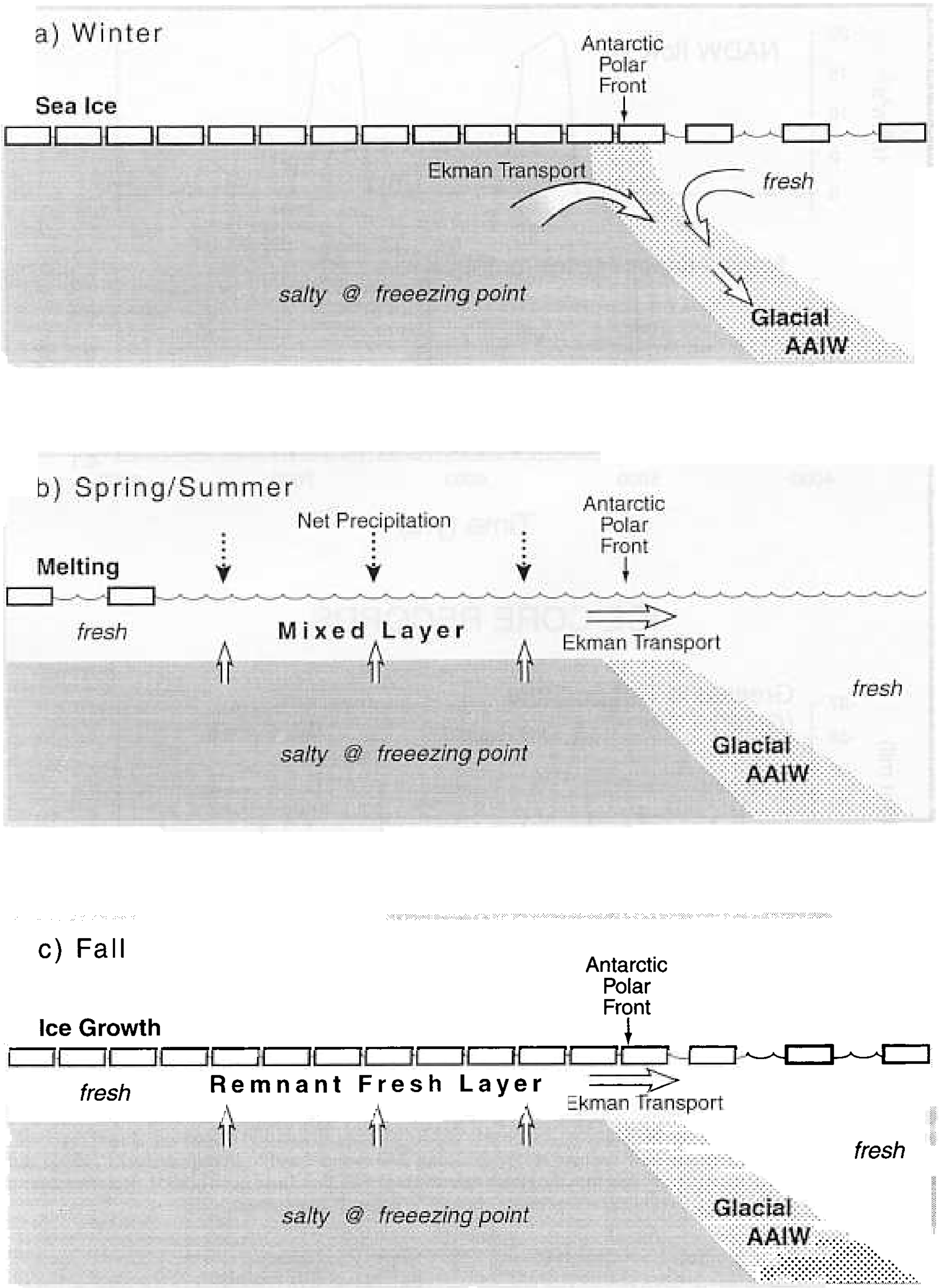

Figure A1. Seasonal controls on the freshwater excess of Antarctic surface waters. (a) Winter ice cover, which allows for formation of salty AAIW. (b) Spring/Summer melting of ice, which produces a freshwater excess in the surface mixed layer, and this excess evolves through the summer as determined by precipitation and entrainment of deeper waters balanced by northward Ekman drift. (c) Fall ice coverage, which shields water column from precipitation, while entrainment and surface Ekman drift continue.

shift the freshwater budget of the Antarctic Ocean in the direction of net loss. Under sufficiently delicate poising the onset of overturning may thereby sow the seeds of its own demise. With this explanation the duration of the short-lived interstadial events would be linked to the timescale for NADW to propagate from the North Atlantic to high southern latitudes. A simulation which accounts for this process would require extending the model to include thermal feedbacks of overturning on sea ice formation.

The simulations for both the short-lived interstadial events (this section) and the long-lived interstadial events (section 4.4) require that the Antarctic freshwater budget starts off delicately poised with the freshwater flux just smaller than F_{off}. Relative to the large change in the freshwater flux from modern conditions (see Figure 5), such fine control might seem improbable. However, once the Antarctic sea surface is effectively covered with ice, we can expect the freshwater budget of the water under to the ice to be much less sensitive to climate changes than the freshwater budget of open water. Hence it is reasonable to invoke much finer control during cold ice-covered periods. Feedbacks of overturning on sea ice formation might also be relevant for maintaining delicate poising, although an exploration of such feedbacks will require an examination of thermal feedbacks, as mentioned above.

5. Concluding Remarks

We have outlined a hypothesis for the origin of Pleistocene climate instability that builds on the work of *Stommel* [1961] showing that feedbacks involving salinity can potentially lead to multiple states of thermohaline overturning, and draws on work of *Toggweiler and Samuels* [1993a, 1993b,1995, 1998] showing that a "reconfigured conveyor" overturning pattern can be sustained by the action of the winds in the

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are therefore not a sensitive test of our hypothesis. With the seasonal perspective given in Figure A1, only the winter sea ice limits are critical.

The seasonal perspective shown in Figure A1 raises a significant new issue, however. The summer upwelling and northward surface drift results in a net conversion of deep water into lower latitude surface waters. To sustain a steady state, a compensating pathway for converting lower latitude surface waters into deep waters is required.

Suitable pathways are not hard to identify. In the glacial "on" state a possible pathway is the conversion of lower latitude surface waters to NADW, and the subsequent penetration of NADW into abyssal layers. In the glacial "off" state a possible pathway is the entrainment of lower latitude surface waters into AAIW and the subsequent penetration of a portion of the AAIW into abyssal layers. The diapycnal mixing needed to support downward penetration of NADW or AAIW could occur via several possible mechanisms, such as via eddy mixing within the ACC or turbulence generated over rough topography in the oceans interior [Ledwell *et al.*, 2000]. The need for this compensating pathway clarifies that diapycnal mixing actually promotes our hypothesized glacial overturning patterns, provided the mixing is concentrated at high southern latitudes or at depths below ~2500 m but not within the main pycnocline.

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