The Bering Strait's Grip on the World Climate

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The Holocene interglacial period of the last 10,000 years and the penultimate interglacial ∼125,000 years ago have been characterized by distinctly stable climates. An analytical-dynamical model, which includes both wind and thermohaline processes, is used here to show that, during interglacial periods, variability in North Atlantic Deep Water (NADW) formation are damped out because of an Arctic freshwater feedback. During glacial periods, the stabilizing feedback is prevented by the closure of the Bering Strait (BS). The open-strait thermohaline circulation recovers exponentially from an internal perturbation with a rapid e-folding time of 32 years. However, a prolonged external forcing anomaly, such as the suggested increase in precipitation due to global warming, may shut down NADW formation entirely.

The North Atlantic (NA) is an integral part of the Earth's heat engine. Here, warm surface water from the South is transformed to deep water by releasing great quantities of heat to the atmosphere. The return flow surfaces again at low latitudes (through diffusion) or in the Southern Ocean (through wind forcing). This cycle is known as the ocean's thermohaline or overturning circulation. Coupled ocean-atmosphere modeling studies suggest that global-warming-induced precipitation increase and glacier melt may inhibit NADW formation and thereby halt the overturning circulation (*1–3*). The reduced heat transport to northern high latitudes could initiate large ice-sheet growth and, combined with the ice-albedo feedback on incoming solar irradiation, throw the planet back into an early glacial period. This mechanism is believed to be responsible for multiple rapid climate fluctuations (Fig. 1) during the last glacial period (*4, 5*). Curiously enough, such high amplitude oscillations have been absent during the Holocene and evidence is mounting that the penultimate interglacial was equally stable (*6–8*). In order to have confidence in predictions of rapid climate change due to anthropogenic warming, we need to understand what makes interglacials so stable.

We propose here that the switch between stable and unstable climates is the onset and cessation of flow through the BS as modulated by sea level variations. Our argument rests on a simple negative feedback mechanism between NADW formation and the inflow of freshwater from the Arctic. Due to the island-like geometry of the Americas during interglacials, an increase in NADW will draw in more freshwater from the Arctic (Fig. 2). The inflow of salty water from the south will also increase, but the net result of the mix is a freshening of the North Atlantic (NA). Hence, the NADW formation rate is reduced to its former strength. On the other hand, while the strait is closed, the stabilizing Arctic inflow feedback is prevented.

Observations of the last 40 Ka (Kilo Annum) support the idea of a BS control on climate stability. Since the last glacial maximum, 19 Ka BP (Before Present), global sea level has risen by 130m (*10*). The Bering Strait, being presently only 45m deep opened up to flow around 10.4 Ka BP (*11*) which is coincident with the onset of the stable Holocene. Correlating the BS throughflow and climate stability for the period preceding 40 Ka BP is not feasible, because the sea level history at the strait is unknown. Uncertainties exist in both the eustatic, or global, component (*12, 13*) and the isostatic component, i.e., the local response of the mantle due to

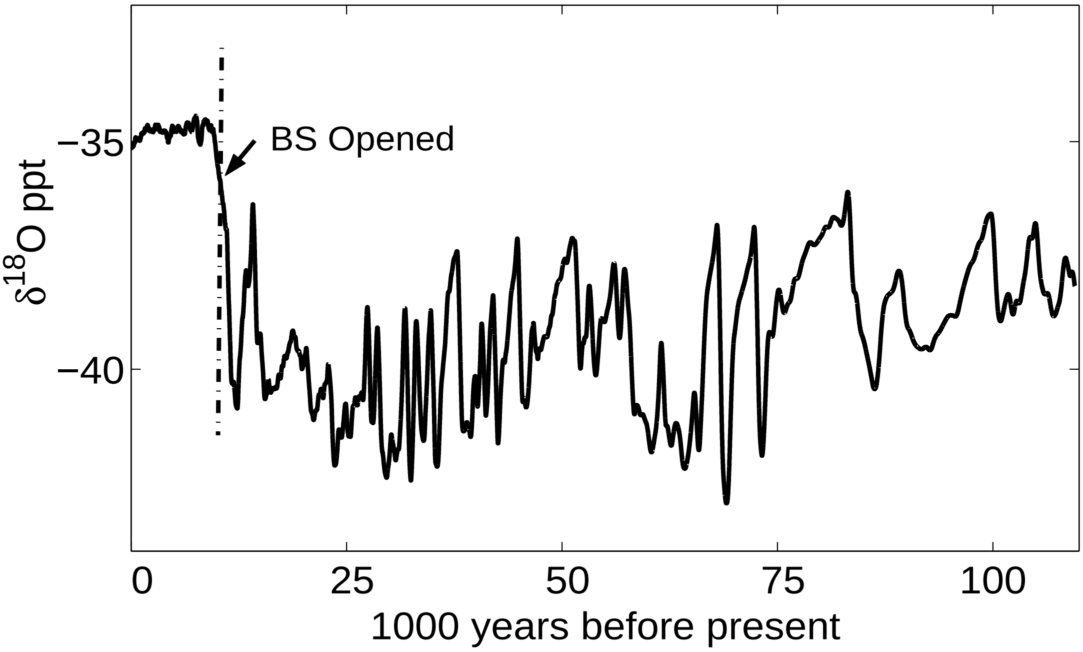

Figure 1: Oxygen isotope record from the Greenland GIPS2 ice core. Less negative values of $\delta^{18}O$ indicate higher atmospheric temperatures. The occurrence of high amplitude rapid oscillations decreased notably after the opening of the BS, 10.4 Ka BP. [GISP2 data obtained from the National Geophysical Data Center web site at http://www.ngdc.noaa.gov/paleo/icecore/greenland (*9*)]

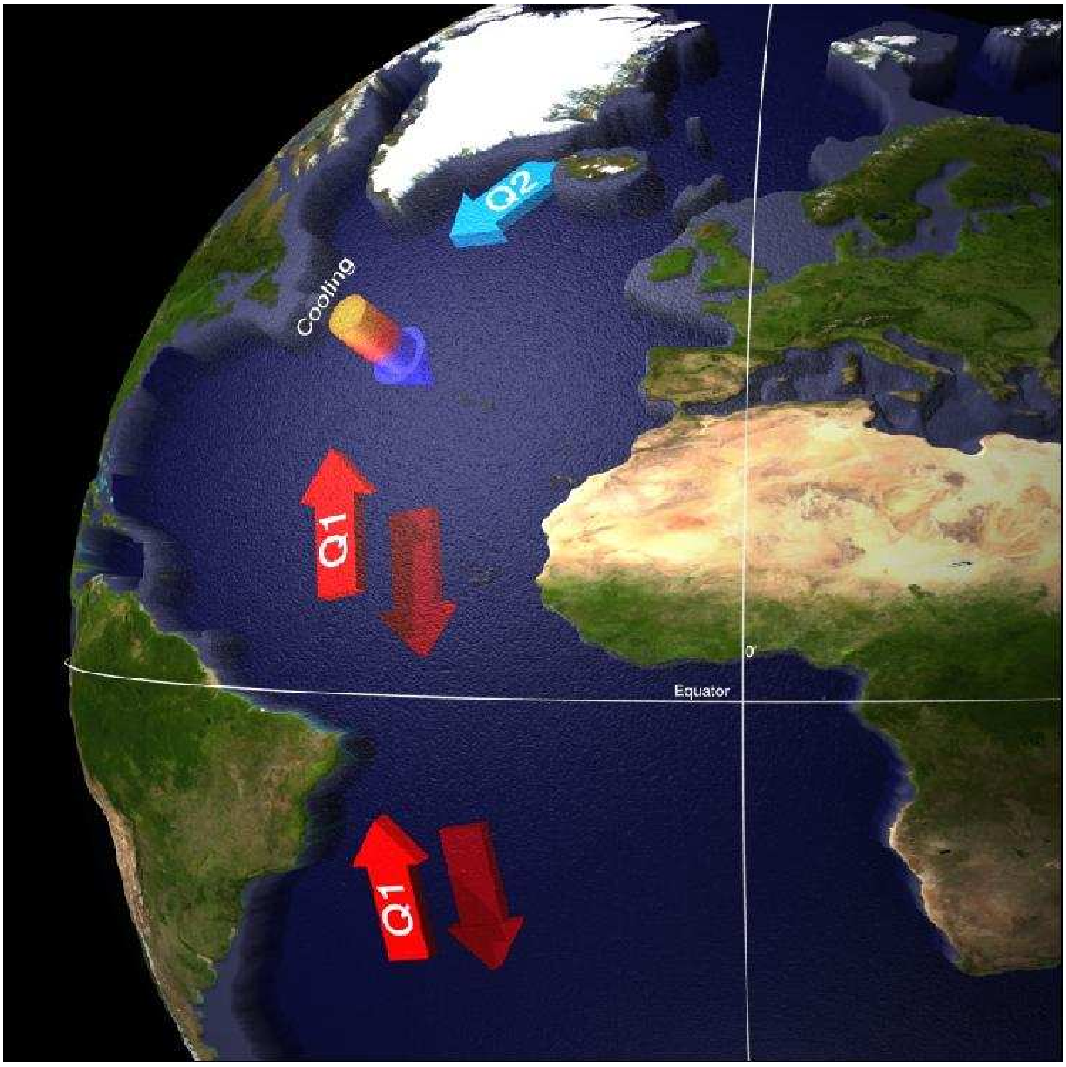

Figure 2: A 3D view of the study area. The blue oceanic surface is the main thermocline. Salty water from the South, Q_1 (red arrows), and fresh water from the North, Q_2 (blue arrows, originating in the BS), enter the North Atlantic convection region where it sinks below the thermocline (yellow-to-blue arrow) to return southward (dark red arrows). During interglacials, an increase in NADW formation draws in a mix of Arctic and SA water that has a net freshening effect on the NA. Consequently, the NADW formation is reduced back to its former strength.

shifts in surface weight (*14*). However, it is reasonable to assume that the BS was open during the height of the last interglacial, $\sim$125 Ka BP, when the global level was 7m higher than today (*15*).

The zonal asymmetry in the global precipitation and evaporation fields renders the NA saltier than the North Pacific (*16*). A steady state is maiinterglacialntained by the return of freshwater from the Pacific to the Atlantic through the BS (*17*). Recently, studies using ocean general circulation models (*18, 19*), 2-Dimensional models (*20*) and box models (*21*) indicate that NADW formation is sensitive to the BS throughflow. Our results agree with the conclusion that the BS flow has a weakening effect on the thermohaline circulation, but we include an important negative feedback mechanism that acts to stabilize, rather than destabilize, the overturning. Unlike these previous studies, that use fixed or parameterized BS throughflow, we derive the flow through the strait analytically from the global wind field and sinking in the NA. The mutual dependence of the freshwater influx into the North Atlantic and the deep water formation itself acts to smooth out instabilities in the thermohaline circulation.

We shall consider two types of problems. The first is the pure stability problem of what happens to a steady state ocean when it is perturbed. In this case, we are not interested in the external forcing (e.g., an ice sheet discharge) that caused the anomalous state, but only in the ocean's response. We shall call this an ”internal perturbation” and we shall see that the ocean always recovers from it and returns to its initial state. This will be the main result of our study. But it raises some questions. How does a -glacial transition occur in such a stable system? Does this imply that the overturning is robust to any changes that global warming may induce? These questions lead us to the next type of problem, i.e, what happens to the overturning when the external forcing is changed, not merely for a short time, but for an extended period. We shall consider the effect of a quasi-permanent increase in precipitation over the NA, as predicted by models of global warming. This is our ”external perturbation” case and we shall see that the thermohaline may collapse if the perturbation is sufficiently large.

Our analytical derivation consist of three stages: First we derive the dependence of Arctic and South Atlantic (SA) inflow on the NADW formation rate. Next, we express the salinity of the NA in terms of the inflows and finally we close the problem by relating the deep water formation rate to the salinity.

During interglacials the continents of the Americas form an island in the sense that they are entirely surrounded by water and, hence, we can apply Godfrey's Island Rule (IR) to it (*22*). The essence of the IR is an integration of the horizontal momentum equations around an island along a closed path that avoids the frictional Western boundary current region (Fig. 3). The assumption that mass is conserved above a level-of-no-motion to the east of the island then leads to an expression for the transport around the island in terms of the wind field along the integration path. We will apply the IR to the continents of the Americas in a similar fashion to Nof (*23*). The level-of-no-motion assumption is made here again merely for convenience. The results remain the same without this assumption, but the analysis is considerably more complicated and will, therefore, be reported elsewhere. In addition, we take not only the sea level to be continuous across the BS sill, but the entire vertically integrated pressure is also taken to be continuous. This assumption can also be relaxed without altering the fundamental result (*24*) but the procedure is too involved for presentation here. Our analysis differs from the classic IR in that we include thermodynamics by allowing sinking through the upper layer east of the island. Namely, the southern flow into the Atlantic, Q_1, differs from the Arctic inflow, Q_2, by the amount of sinking within the Atlantic, D. In terms of the wind field these are,

$$Q_1 = \frac{\oint \tau^r dr + f_2 D}{(f_2 - f_1)}, \qquad Q_2 = -\frac{\oint \tau^r dr + f_1 D}{(f_2 - f_1)} \tag{1}$$

where $\oint \tau^r dr$ is the clockwise integrated windstress along the path shown in (Fig. 3) and $f_{1,2}$ are the average Coriolis parameters at the Southern and Northern ends of the Atlantic respectively

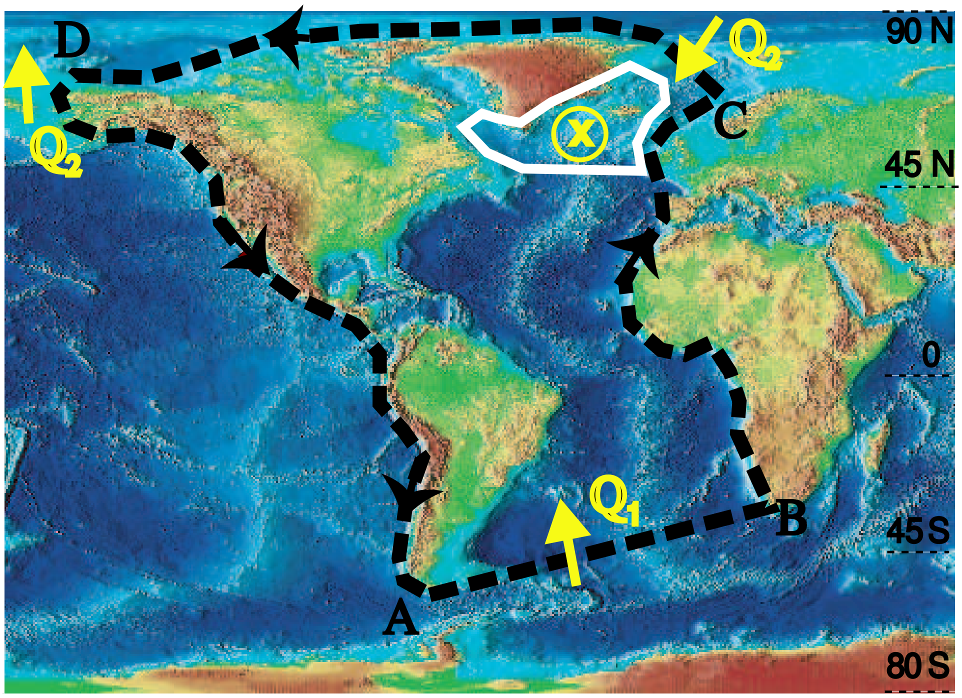

Figure 3: Integration path of the wind around the Americas and Atlantic (black dashed line). $Q_{1,2}$ are the transports into the Atlantic from the south and north respectively. The convection region in the NA (whose upper volume is V) is outlined in white and downwelling is indicated by $\otimes$.

(i.e., sections AB and CD). Both Q's are positive towards the Atlantic and represent water from the surface to a depth of, say, 1000m. In what follows, we will assume that all downwelling from the upper layer occurs (through deep convection) north of 50N (Fig. 3).

It is informative at this point to derive the present day downwelling in the NA and the transports into the Atlantic from eqs. 1. Integration of 40 years of NCEP annual winds (*25*) along the path shown in Fig. 3 gives 4 Sv (1Sv= $10^6 m^3/s$) for the first term in eqs. 1 [i.e., $\oint \tau^r dr/(f_2 - f_1)$, where f_1 and f_2 are the Coriolis parameters at $45^o S$ and $75^o N$ respectively (*26*)]. Observations (*27*) indicate that Q_2 is almost 1 Sv so that D=11 Sv and Q_1=10 Sv. Note that D is only the net downward flux and does not negate a superimposed internal overturning circulation.

The salinity in the convection region, S, is determined by the equation of conservation of salt,

$$\frac{\partial(S)}{\partial t} = \frac{1}{V}\left[(S_2 - S)Q_2 + (S_1 - S)Q_1 - SP\right] \tag{2}$$

where $S_{1,2}$ are the salinities of the incoming water $Q_{1,2}$, V is the volume of the region in which the downwelling occurs (Fig. 3) and P is the surface freshwater flux (precipitation - evaporation + runoff) into the region.

For purely thermally driven sinking, a surface heat flux, H, must change the upper layer temperature by an amount equivalent to the density difference between the upper and lower layer. Conservation of heat determines the amount of downwelling that H can support,

$$D = \frac{-H}{\rho C_p (T_l - T - \frac{\beta}{\alpha}(S_l - S))} \tag{3}$$

where ρ is the upper water density, C_p the heat capacity of sea water and T is the upper layer temperature. The subscript l denotes lower layer quantities and α and β are the expansion coefficients for temperature and salinity, respectively.

The four equations 1, 2 and 3 fully describe our system for the four unknowns $Q_{1,2}$, D and S. The dependence of the wind field and precipitation on the overturning rate is not straight forward. Hence, for the purpose of this study, we shall regard them as external variables and assume that their response to NADW formation perturbations is negligible. For the stability analysis of the internal perturbation case, we shall perturb the four unknowns around a steady mean state. When considering the case of a prolonged external forcing, we shall, in addition, perturb the freshwater P. The analysis that follows is only valid for small perturbations, but it is easy to see how the ocean will also rapidly recover from large anomalous forcings. Consider, for instance, a freshwater input into the NA that is large enough to shut down the overturning altogether. With the aid of eqs. 1, one can show that, due to the strong southern winds, 4 Sv enters the SA (from the Southern Ocean) and, without any sinking, it will exit the NA through the BS. This means that the anomalous freshwater is flushed out of the NA by the winds. Once the NA regains its former salinity, the overturning will resume. The closure of the BS during glacial periods will trap the freshwater anomaly and prevent a thermohaline recovery.

We now proceed with our analysis for small deviations from the mean state. The perturbation of eqs. 1 and 2 is straightforward. For eq. 3 it is necessary to expand and linearize around a mean steady state. This yields,

$$D' = \frac{-H\rho\beta\alpha S'}{C_p\overline{\Delta_l\rho}^2} \tag{4}$$

where $\overline{\Delta_l\rho}$ is the steady state difference between the lower and upper layer density, the primed variables denote small perturbations and variables with a bar (-) denote a mean steady state. The linearization is valid for $\beta S' << \overline{\Delta_l\rho}/\rho$.

We can now immediately solve for D' from a perturbation of eqs. 1 and 2 and from eq. 4. For the case of a short-lived external forcing, we find,

$$D' = D'_0 e^{-\lambda\sigma t} \tag{5}$$

$$\text{where,} \quad \lambda = \frac{-H\rho\beta\alpha}{VC_p\overline{\Delta_l\rho}^2} \qquad \sigma = \frac{f_1(S_2 - \overline{S}) - f_2(S_1 - \overline{S})}{(f_2 - f_1)}$$

and D'_0 is the initial perturbation. The variables $Q'_{1,2}$ and S' have a similar decaying solution. Reasonable parameter choices (*28*) yield an e-folding timescale, $\lambda\sigma$, of 32 years (Fig. 4, blue curve). Note, however, that arguments can be made for a wide range of parameter sets that would lead to e-folding scales ranging anywhere from 10 to 600 years. Qualitatively our result remains the same. We see that, while the BS is open, the ocean is able to smooth out any internal perturbations to its overturning circulation.

We shall now address the question of how the ocean will respond if the perturbation is external and is sustained for a long period. For brevity, we shall consider here only the case of a quasi-permanent freshwater flux anomaly. Changes to the windfield give similar results. The inclusion of a freshwater anomaly, P', changes our solution as follows,

$$D' = -\frac{\overline{S}P'}{\sigma} + \left[D'_0 + \frac{\overline{S}P'}{\sigma}\right] e^{-\lambda\sigma t} \tag{6}$$

Here, the overturning will decay to a new stable state reduced by $\overline{S}P'/\sigma$ and the salinity will adjust according to eq. 4 (Fig. 4, green curve). Even though the extra precipitation or melt-water freshens the NA, a salinity balance is maintained by the reduction of fresh Arctic inflow. However, the Arctic inflow can only be reduced by 1 Sv, the current flow through the BS. An additional increase in precipitation will reverse the flow through the strait and thereby disable this stabilizing negative feedback. In other words, if the NADW formation rate is decreased beyond this 'reversal threshold' (Fig. 4, dashed pink line), less high salinity water from the South will enter the convection region, but now without the accompanying reduction of Arctic freshwater inflow. Hence, that NA becomes fresher and fresher and the deep water formation

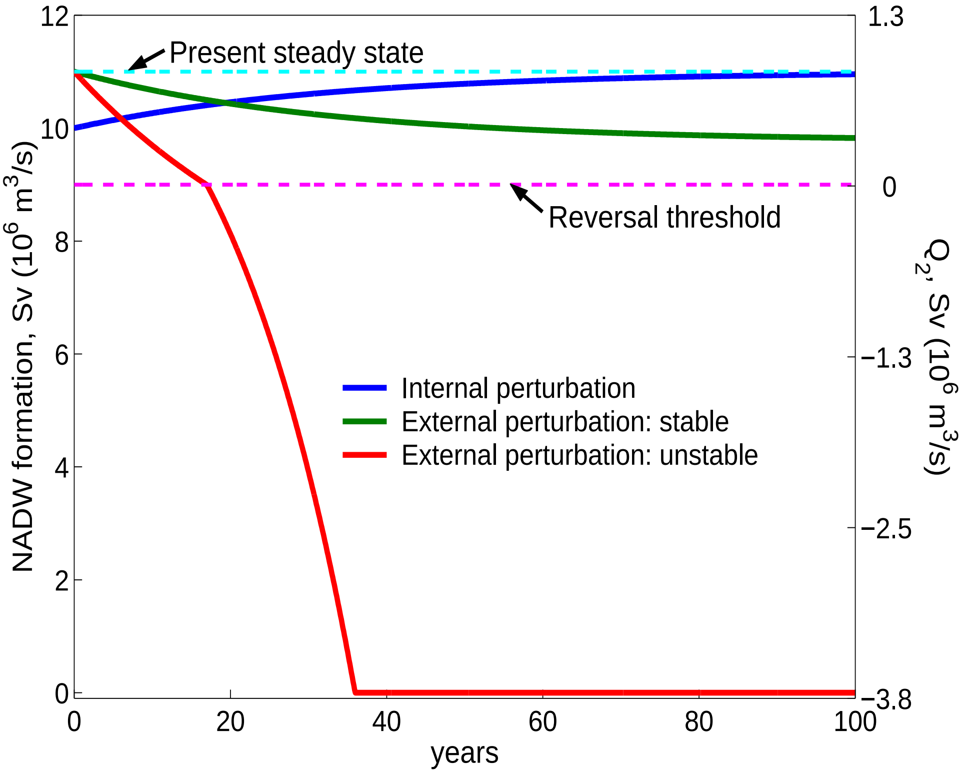

Figure 4: The solution (eq. 5) for an initial internal perturbation to the NADW formation of $D_0' =$ -1 Sv (blue curve) and two solutions (eq. 6) for a prolonged increase in precipitation of $P' =$ 0.01 Sv (green curve) and $P' =$ 0.04 Sv (red curve). The corresponding flow from the Arctic into the NA, Q_2, is indicated on the right axis. As long as the precipitation anomaly is less than the reversal threshold (beyond which the BS throughflow is reversed), the overturning will be stable. Any additional increase will arrest the formation of NADW.

will come to a rapid halt (Fig. 4, red curve). Mathematically, the solution changes to one of exponential growth, due to the sign change of σ, which becomes, $-f_2(S_1 - \overline{S})/(f_2 - f_1) < 0$. From eqs. 1, 4 and 6 we find that the precipitation increase necessary to reverse the BS flow (and, therefore, shut down the thermohaline circulation) is P'=0.02 Sv. This is a 13% increase to the current net surface freshwater flux into the NA (from 50N to 70N) which is 0.15 Sv (*29*).

In conclusion, an open BS allows the use of an integral constraint that links the Arctic inflow to deep water formation in the NA. The interplay between the incoming freshwater and the strength of the wind and thermohaline-induced overturning acts to stabilize (rather than destabilize) the climate. However, a prolonged increase in the external freshwater flux that suffices to reverse the Arctic inflow will arrest the overturning circulation. The reversal of the Arctic inflow (and, hence, the BS throughflow) as a precursor of an overturning collapse, points to the importance of monitoring the long term transport in the BS.

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28. H=-2e14 W, V=3.75e15 m^3, ρ=1000 Kgm^{-3}, C_p=4000 JKg$^{-1}K^{-1}$, β=8e-4 PSU^{-1}, α=1.5e-4 K^{-1}, $\overline{\Delta_l \rho}$=0.7 Kgm^{-3}, f_1=-1.0e-4 s^{-1}, f_2=1.4e-4 s^{-1}, $(S_2 - \overline{S})$=-1.9 PSU, $(S_1 - \overline{S})$=0.9 PSU, $\overline{S}$=34.9 PSU].

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