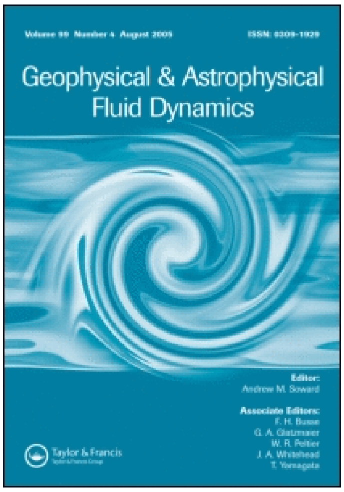

Geophysical & Astrophysical Fluid Dynamics
Publication details, including instructions for authors and subscription information:
http://www.informaworld.com/smpp/title~content=t713642804

Thoughts on a variable meridional overturning cell and a variable heat-flux to the atmosphere

Doron Nof[a,b]; Stephen Van Gorder[a]; Lisan Yu[c]
[a] Department of Oceanography, The Florida State University, Tallahassee, FL 32303, USA [b] Geophysical Fluid Dynamics Institute, The Florida State University, Tallahassee, FL 32303, USA [c] Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, MA, USA

First published on: 29 July 2010

To cite this Article Nof, Doron , Gorder, Stephen Van and Yu, Lisan(2010) 'Thoughts on a variable meridional overturning cell and a variable heat-flux to the atmosphere', Geophysical & Astrophysical Fluid Dynamics,, First published on: 29 July 2010 (iFirst)

To link to this Article: DOI: 10.1080/03091929.2010.481383

URL: http://dx.doi.org/10.1080/03091929.2010.481383

Geophysical and Astrophysical Fluid Dynamics
2010, 1–22, iFirst

Thoughts on a variable meridional overturning cell and a variable heat-flux to the atmosphere

DORON NOF*†‡, STEPHEN VAN GORDER† and LISAN YU§

†Department of Oceanography, The Florida State University, Tallahassee, FL 32303, USA
‡Geophysical Fluid Dynamics Institute, The Florida State University, Tallahassee, FL 32303, USA
§Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, MA, USA

(*Received 15 January 2009; in final form 27 January 2010*)

The atmospheric response to a potential slowdown of the Atlantic meridional overturning cell (AMOC) is examined using the *nonlinear* analytical approach (for Heinrich events) introduced by Sandal and Nof (Sandal, C. and Nof, D., A new analytical model for Heinrich events and climate instability. *J. Phys. Oceanogr*. 2008, **38**, 451–466; SN, hereafter). Most numerical global climate models predict that the atmosphere should *cool* in response to the increased freshwater-fluxes (''hosing'') that slow the AMOC down and significantly reduce the heat-flux to the atmosphere. Our application of SN to the modern day climate suggests that the answer to the question of how the atmosphere responds to a slowing AMOC is not that simple. Within the (admittedly limited) dynamics which SN invoke, we find that, as the global numerical climate models predict, a slowdown of the AMOC will indeed cause the *mean* atmosphere of the *entire* Northern Hemisphere to cool. However, in contrast to the numerical predictions, our analytical approach suggests that a region in the immediate vicinity of the Atlantic convection (up to a distance of $\sim$O(1000 km)) may *warm up*, *not cool down* (roughly 3°C for 50% mass-transport reduction). For some extreme conditions of a constant atmospheric transport independent of the AMOC (which is not a part of the dynamics involved by SN), the atmosphere can indeed locally cool, but the cooling is minimal (less than 0.3°C for a 50% ocean mass transport reduction), and the associated reduction in heat flux from the ocean to the atmosphere is almost totally negligible. We also place the SN results on a somewhat firmer ground by examining in detail about its closure condition and the most critical assumption adapted by SN. The first has to do with the ratio of the atmospheric and oceanic mass transports (assumed unity in SN) and the second involves up-to-date maps of the ocean–atmosphere heat-fluxes. We show that the system of governing equations admits physically relevant solutions only for particular relationships between the atmospheric and ocean mass transports participating in the ocean–atmosphere heat exchange. Still, as the analytics misses critical atmospheric components such as moisture and variability in the heat exchange interaction area, our results can only serve as an indicator of the problem complexity.

Keywords: Atmosphere; MOC; Heat flux

*Corresponding author. Email: nof@ocean.fsu.edu

Geophysical and Astrophysical Fluid Dynamics
ISSN 0309-1929 print/ISSN 1029-0419 online © 2010 Taylor & Francis
DOI: 10.1080/03091929.2010.481383

1. Introduction

We are motivated by the recent Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment (2007), which concludes that it is "very likely" (>90%) that there will be a slowdown of the Atlantic meridional overturning cell (AMOC) during the twenty-first century. The 11 models used in the IPCC suggest that the AMOC reduction is due to the changes in surface heat-flux associated with "hosing" (i.e., adding freshwater to the surface of the North Atlantic) further arguing for a better understanding of the coupled air and sea temperature variability in this region. Observational attempts to determine whether the AMOC transport has actually been altered during the past 50 years have so far been inconclusive (Bryden *et al.* 2005, Cunningham *et al.* 2007, De Boer and Johnson 2007, Kanzow *et al.* 2007, Meinen and Baringer 2008). However, the IPCC slowdown predictions are conceptually consistent with recent observations of an increased freshwater-flux into the Atlantic (e.g., Curry and Mauritzen 2005). Such an increase in freshwater does imply a slowdown of the AMOC because of the reduced salinity and a reduced convective temperature. Most global climate models predict that Europe should *cool* in response to "hosing" which results in an AMOC reduction (cooling of 1°C in the British Isles and 2°C in Scandinavia for a 30% reduction in transport according to the IPCC). The explanation that is normally given is that hosing reduces the strength of AMOC, which in turn reduces the heat-flux to the atmosphere. This warms the atmosphere less and, therefore, causes cooling. Here, we shall not argue with the results that cooling shows up in the numerical runs, but rather we shall argue that this is due to a reduced heat-flux.

The main premise of the work presented here is that the heat-flux from the water to the air is proportional to the temperature *difference* between the (commonly warmer) ocean and the (cooler) atmosphere, and that the specific heat capacity of water (C_{pw}) is much greater than that of the air (C_{pa}). Also, the temperature difference between the ocean and atmosphere is usually ~O(10°C), much larger than the typical meridional overturning circulation (MOC) variability, which is an order of magnitude smaller. Noting that reducing the heat-flux to the atmosphere implies reducing this temperature difference between the warmer ocean and the cooler atmosphere, we suggest that it can be achieved in two ways:

1. The ocean cools and the atmosphere warms.
2. The ocean cools and the atmosphere cools less.

Because the specific heat capacity of water is so large relative to that of the air ($C_{pw}/C_{pa} \approx 4$), the ocean can only *slightly* cool, usually no more than ~O(1°C). This means that with option (2), where the atmosphere must cool less than the ocean, one can only achieve very small and insignificant reductions in the heat-flux (<10%). On the other hand, the temperature difference can be significantly (50%) reduced when the system followed option (1). So, when the heat-flux is reduced significantly, the system must somehow allow the atmosphere to warm, not cool.

In this context, it is convenient to think about the hypothetical limit of $(C_{pw}/C_{pa}) \to \infty$, representing (hypothetical) water with an *infinite* specific heat capacity. In this limit, which is later referred to as the "high specific heat capacity limit", the ocean cannot cool at all and the heat-flux can only be reduced when the atmosphere warms up. The above premise is independent of where the atmospheric jet is coming from and of the flow direction.

1.1. *Background*

Historically, progress on new problems that have been identified during the last 80 years has initially been made using simple analytical and numerical models (e.g., the general oceanic circulation, boundary currents, eddies, outflows, island circulations, and mixed layers). By ''simple models'', we mean here models with straightforward structures and well-defined aspects that can be solved either analytically or semianalytically. Perhaps counter-intuitively by ''simple models'', we do not necessarily imply that they are simple to understand, as some of them are actually quite difficult to follow.

Complex global and local numerical models entered the investigative process later on and, by including many processes together, shed more light on the issues at hand. Paradoxically, the opposite has happened with the AMOC. Its variability has been identified as an important problem several years ago and yet, aside from the original classical box models with or without convection (e.g., Colin de Verdiere 2007 and the references given therein), there are hardly any simple models that describe it. At the same time, the literature is filled with numerous global hosing experiments that attempt to understand and explain how the system works. It is often very difficult to identify important processes without isolating them first, and there is a need for simple models that will bridge in the gaps in our understanding (see, e.g., Nof 2008). This is what we attempt to do here.

From a theoretical and physical point of view, it is very difficult to say upfront what will happen if the AMOC were to slow down because *nonlinear* heat exchange processes are both very hard to solve for and counter-intuitive. Here, we take the Sandal and Nof (2008; SN, hereafter) analytical model, which was originally developed for the analysis of Heinrich events, and apply it to the present-day climate. We provide additional justifications to the SN model by analyzing their closure condition associated with the atmosphere and ocean mass-flux ratio. To avoid duplication, this article starts where SN ends implying that this article is *not self-contained*. The reader who wishes to understand the details of this article and the reasons behind the various approximations is advised to go to SN first and understand the main issues discussed there before attempting to go through this work. On the other hand, the reader who is primarily interested in the results and is not particularly concerned with the logic leading to our various assumptions can get the desired information from this article. Both readers are warned in advance that, because of the neglect of moisture and other atmospheric processes, our model cannot be directly applied to the atmosphere. It can only serve as a tool suggesting that a further examination of the AMOC slowdown is warranted.

1.2. *Present approach*

The issue of how the results of our simple analytical model should be compared to those given by the global numerical models is not a trivial one. Some feel that the resolution of modern state-of-the-art global climate models is so high that – when compared to simple models – their results represent the ''absolute truth''. Their argument is that global numerical models include many processes absent from the analytics and, therefore, should be closer to reality. While it is true that the global numerical models include processes absent from the simple models, some of their dynamics are not well understood and, as a result, it is often very difficult to say what the differences between the models imply. This comes about in several ways. First, there is the familiar diffusivity issue, which occasionally masks the reality and will be elaborated later (section 9). Second, there is the issue that one may not know *why* the global numerical models show what they do. For example, as mentioned earlier, the global numerical models show that when the North Atlantic is ‘‘hosed’’, the MOC transport reduces and Europe cools (see, e.g., Brayshaw *et al.* 2009 and the references given therein). But, there are processes besides the MOC’s variable transport that also affects Europe’s climate and these processes are also sensitive to hosing. For example, the hosing also affects storm tracks significantly (Brayshaw *et al.* 2009) and it may be that Europe cools in these models because of altered storm tracks, not because of a reduced MOC. Sea ice distribution issues may also contribute to the complexity.

Furthermore, it should be kept in mind that, no matter how high the resolution of the global numerical models is, the outcome of these global models is still an outcome from calculations. These are not observational data, and therefore should be regarded as questionable predictive results requiring verifications. (This, perhaps obvious, point has been mentioned on many occasions, but see, e.g., Nof 2008.) Consequently, we take here the approach that when simple and more complex models appear to disagree, one needs to further investigate the issues at hand keeping all the results in mind until the issue is resolved. For the problem at hand, a resolution of the disparity will likely require massive global models runs involving ‘‘nesting’’, which is well beyond the scope of this study. Hence, we will present the results of our simple model (figure 1) and discuss the outcome in detail, but we will not be able to answer unequivocally all the questions that it raises.

We shall suggest that when the reduced heat-flux warming is mixed around in either the actual atmosphere or the numerically modeled atmosphere, the mean effect will indeed be overall cooling, not warming. Accordingly, we shall propose that either: (1) as suggested by SN, the analytically predicted local warming and reduced mass-flux is merely camouflaged in the numerical atmosphere by mixing the air participating in the heat-exchange process with adjacent air masses that do not participate in the exchange, or (2) the numerically observed cooling of Europe is due to other, yet unknown, processes such as changing storms track. For general aspects of the AMOC, the reader is referred here to Weijer *et al.* (2001), Johnson *et al.* (2007), and Polton and Marshall (2007).

We will begin by considering an ultra simple model that does not even involve convection (section 2) to illustrate that, usually, a reduction in heat-flux from the ocean to the atmosphere implies atmospheric warming, not cooling. As mentioned, the main premise is that the heat-flux is proportional to the temperature difference between the (commonly warmer) ocean and the (cooler) atmosphere and that the specific heat capacity of water (C_{pw}) is much greater than that of the air (C_{pa}). Because the specific heat capacity of water is so large ($C_{\mathrm{pw}}/C_{\mathrm{pa}} \approx 4$), the ocean can only slightly cool implying that the temperature difference can be significantly (50%) reduced *only* when the atmosphere warms. We know of no physical process implying a particular dependency of the AMOC on the density structure and therefore see no reason to invoke it. After presenting the one-dimensional model, we will address some salinity issues (section 3), which are then followed by taking the SN solution and applying it to

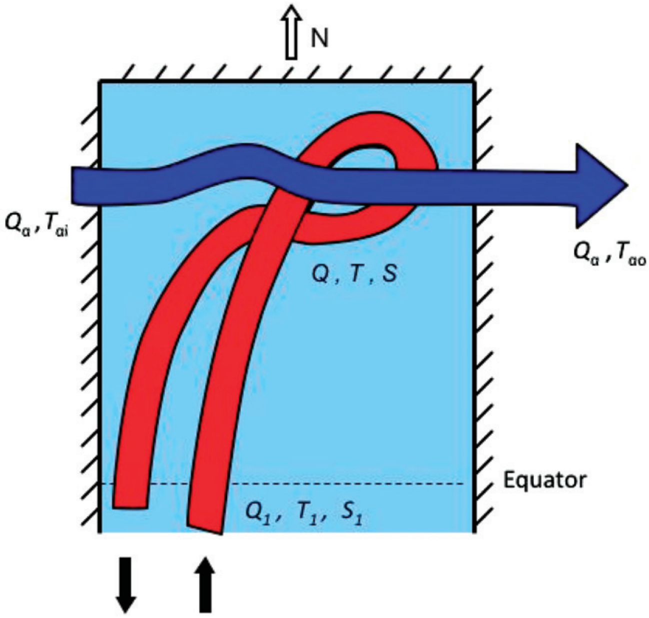

Figure 1. Schematic diagram of the present heat-exchange model.
Notes: The upstream upper limb of the AMOC has an unknown transport Q_1, but known temperature and salinity T_1, S_1. Due to the heat exchange with the atmosphere, which varies with the addition of a freshwater-flux F_F (not shown), it cools to a convective temperature and salinity (T, S). Just as the heat exchange draws the AMOC to the convection zone, it also draws an atmospheric jet with an unknown transport Q_a, but known temperature T_{ai}, into the region. The zonal jet, which should not be confused with the general atmospheric flow from the west to the east, is warmed and exits the area with a temperature T_{ao}.

present day climate (section 4). We then examine the possible transports variability range (section 5), the special case of constant atmospheric transport (section 6) and the collapsed MOC state (section 7). This is followed by an examination of diffusion issues (section 8) and the validity of our assumption regarding radiation (section 9). The results are summarized in section 10.

2. Ultra simple one-dimensional model

Consider first the ultra simple one-dimensional geostrophic jet with speed U and width L shown in figure 2. (SN did not consider this case.) There is an atmospheric Ekman layer below the inviscid jet and an oceanic Ekman layer further below. The two Ekman layers flux an equal and opposite mass-fluxes to the left and right (looking downstream) and are insulated from each other outside the jet's projection on the plane. The layers exchange heat as the ocean is warmer than the atmosphere (see, e.g., Tomczak and Godfrey 1994), but due to stratification (between the homogenous Ekman layers and the fluids above and below them) and relative weak heating, there is no convection. The oceanic Ekman layer enters (exits) the box with a known (and unknown) temperature $T_{wi}(T_{wo})$ whereas the atmospheric Ekman layer enters (exits) with a known

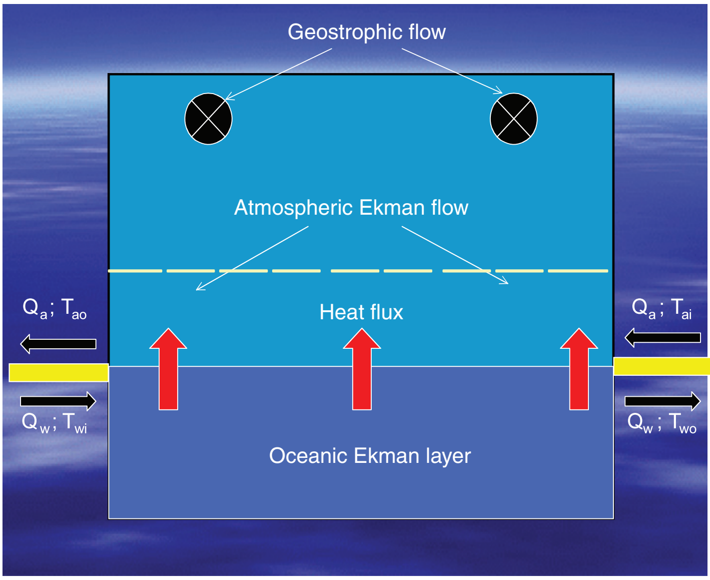

Figure 2. The ultra simple one-dimensional heat exchange problem.
Notes: An atmospheric jet with a geostrophic speed U and width L forces an atmospheric and oceanic Ekman layers, whose mass transports ($\rho_w Q_w$, $\rho_a Q_a$) are equal and opposite to each other, to exchange heat (in the direction perpendicular to U). The ocean is warmer than the atmosphere and the two are insulated from each other in the regions to the right and left of the jet (looking downstream). The oceanic (atmospheric) Ekman layer enters the exchange region with temperature $T_{wi}(T_{ai})$ and exits with the cooler (warmer) temperature $T_{wo}(T_{ao})$. There is no convection in either the ocean or the air. This straightforward model shows that, upon decreasing U, both the mass transports and the heat-flux reduce, but the exiting atmospheric temperature increases (figure 3).

(and unknown) temperature $T_{ai}(T_{ao})$. The system is governed by two equations, the bulk formula for heat exchange and the condition that the atmosphere accepts all the heat released by the ocean. The bulk formula for the total heat-flux F_H is

$$F_H = A(F_S + F_L), \tag{1}$$

where F_S, F_L are the sensible and latent heat-fluxes given by

$$F_S = \rho_a C_{pa} C_S U_{10}(T_{w\,mean} - T_{a\,mean}), \tag{2}$$

$$F_L = \rho_a L_e C_L U_{10} q^*(1 - R_H) + \rho_a C_L R_H \frac{C_{pa}}{Be} U_{10}(T_{w\,mean} - T_{a\,mean}), \tag{3}$$

and

$$T_{a\,mean} = \frac{T_{ai} + T_{ao}}{2}, \; T_{w\,mean} = \frac{T_{wi} + T_{wo}}{2}. \tag{4}$$

Here, A is the box interaction area, F_S and F_L (Wm^{-2}) the sensible and latent heat-fluxes, ρ_a and ρ_w the air and water densities, C_S and C_L constants, U_{10} (m s^{-1}) the wind speed at 10 m above the surface, q^* (g kg^{-1}) the saturation specific humidity of the air, L_e (J kg^{-1}) the latent heat of evaporation, R_H the relative humidity of the air, *Be* the *equilibrium* Bowen ratio, $T_{w\,mean}$ the mean upper oceanic temperature of the box, and

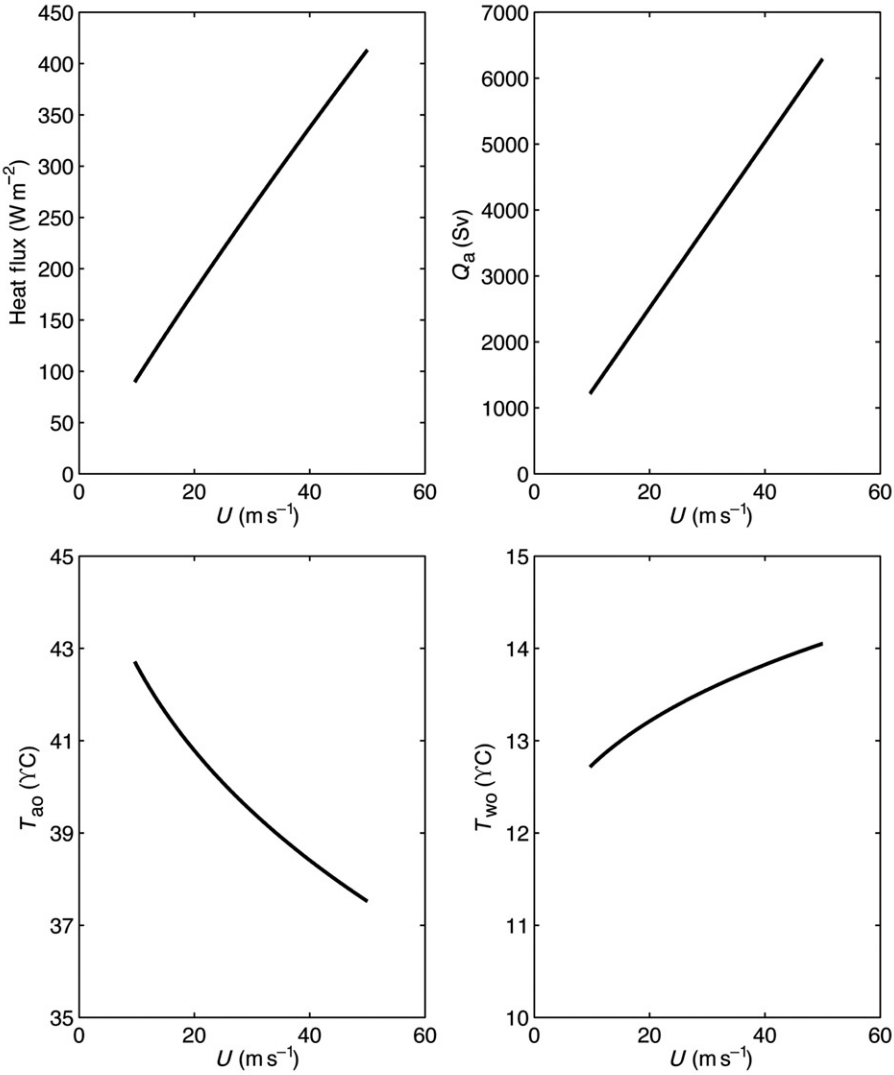

Figure 3. Analytical solution for the one-dimensional jet problem.
Notes: The heat-flux (upper left), atmospheric Ekman layer volume-flux Q_a (upper right), exiting air temperature T_{ao} (lower left), and exiting water temperature T_{wo} (lower right), are all given as a function of the air speed aloft U. The surface area is $10^{12}\,m^2$, the oceanic incoming temperature, 15°C, the incoming air temperature, 5°C, α, the thermal expansion coefficient for water, is $1.5 \times 10^{-4}\,K^{-1}$, β, the salinity expansion coefficient for water, is 8×10^{-4} (practical salinity unit; *PSU* to the −1), *Be*, the equilibrium Bowen Ratio, 0.65, q_s, the surface saturation specific humidity, 10^{-3} (g kg^{-1}), relative humidity, 0.76, U_{10} is calculated from U, water density is 1000 kg m^{-3}, air density 1.5 kg m^{-3}, specific heat of water and air 4000 J (kg K)$^{-1}$ and 1030 J (kg K)$^{-1}$, non-dimensional aerodynamic transfer coefficient for heat (Stanton Number, Cs), 0.0009, aerodynamic transfer coefficient for evaporation (Dalton number), 0.00135, and the latent heat of evaporation (J kg^{-1}) is 2.5×10^6.

$T_{a\,mean}$ the mean temperature of the air in the atmospheric Ekman layer which gets in at the known temperature T_{ai} and leaves at T_{ao}, the (unknown) outgoing atmospheric temperature. The above heat-flux parameterization follows the approach taken by Hartmann (1994) and is adequate for our analysis. Also, q^* and *Be* are dependent on the mean air temperature in the immediate vicinity of the ocean surface in a nonlinear fashion. However, since we would like to use a linear equation in T, we have taken these parameters to be constants. After completing our calculations, we verified that this simplification introduces only minor and negligible changes in the two unknowns. This is not surprising as the 1 : 4 ratio of the air and water specific heat capacity implies small variations in the ocean temperature compared to the atmospheric temperature. U_{10} is calculated by equating the stress exerted by the atmospheric Ekman layer on the surface of the ocean (expressed as a function of U, the geostrophic flow above) to $\rho_a C_d U_{10}^2$.

The second governing equation is

$$\rho_a C_{pa} Q_a (T_{ai} - T_{ao}) = \rho_w C_{pw} Q_w (T_{wi} - T_{wo}) = F_H, \tag{5}$$

where Q_a and Q_w are the known volume transports of the atmospheric and oceanic Ekman flows participating in the heat exchange. The system (1–5) includes two equations with two unknowns, the exiting temperatures of the air and water, T_{ao} and T_{wo}. Its solution is straightforward and unequivocally demonstrates that, upon reducing the geostrophic flow aloft U, the heat-flux from the ocean to the atmosphere *reduces* (figure 3, upper left panel), the Ekman transports also reduce (figure 3, upper right), the ocean slightly *cools* (figure 3, lower right), and most importantly, the atmosphere *warms* (figure 3, lower left). This ultra-simple model illustrates the main point of this article – a significantly reduced heat-flux to the atmosphere should involve atmospheric warming, not cooling. It does so without introducing salinities and convection issues that (for completeness) will be addressed in the following sections.

3. Salinity issues

We now return to our more general SN model shown in figure 1, where a northward flowing ocean current with a (unknown) volume-flux Q_1, (known) upstream temperature T_1 and (known) upstream salinity S_1 is drawn into a "convection box" representing the North Atlantic. Within the box, the current cools to the point that it convects and sinks into the abyss. The temperature S and salinity T at which convection occurs are both unknowns. That current is the modeled AMOC. Just as the AMOC is drawn by the convection into the box, a fraction of the atmosphere is also drawn into the region directly above the oceanic box. That fraction forms a zonal atmospheric jet with an unknown volume-flux Q_a (and a known initial upstream temperature T_{ai}) and needs to be distinguished from the familiar general eastward atmospheric flow. This general eastward flow depends on the equator-to-pole density gradient that is unrelated to the atmospheric jet drawn by the convection, which takes place in both the ocean and the atmosphere. In a way, one can think of this neglected atmospheric flow as analogous to the wind-driven Gulf Stream, which is typically neglected in idealized models of the AMOC.

The warm ocean below warms the atmospheric jet to T_{ao} and a reduction in the AMOC volume-flux is achieved (in the model) by introducing a salinity-reducing freshwater-flux (F_F). This reduction implies a reduced convective temperature, a reduced AMOC transport and a reduced northward heat-flux. The coupled model is governed by the heat exchange equations, bulk heat-flux formulas similar to those earlier introduced, and a convection condition.

For convenience, we now rewrite the water–air heat exchange equation in the form

$$\gamma C_{\mathrm{pw}}(T_1 - T) = C_{\mathrm{pa}}(T_{\mathrm{ao}} - T_{\mathrm{ai}}), \tag{6}$$

where γ measures the (unknown) ratio of the oceanic and atmospheric mass transports, $\rho_{\mathrm{w}}Q_1/\rho_{\mathrm{a}}Q_{\mathrm{a}}$, which, by definition, is always positive. We shall see that $\gamma \sim \mathrm{O}(1)$. Equation (6) states that the atmospheric jet accepts all the heat released by the ocean. The ocean is subject to a convection condition (originating in the linearized equation of state) assuring that the density of the surface water is as large as that of the deep water, but not more:

$$S = S_{\mathrm{D}} + \frac{\alpha}{\beta}(T - T_{\mathrm{D}}), \tag{7}$$

where S is the (unknown) salinity in the upper North Atlantic convection box, α and β the familiar temperature and salinity expansion coefficients, respectively, and T_{D} and S_{D} the (known) temperature and salinity of the deep layer (below the thermocline). Note that the upper layer temperature cannot be reduced below that given by equation (7) because the surface water sinks immediately when the temperature given by equation (7) is reached. Substitution of equation (6) into equation (7) gives

$$S = S_{\mathrm{D}} + \frac{\alpha}{\beta}\left[T_1 - T_{\mathrm{D}} + \frac{C_{\mathrm{pa}}}{C_{\mathrm{pw}}}\left(\frac{T_{\mathrm{ai}}}{\gamma}\right)\right] - \frac{\alpha}{\beta}\left(\frac{C_{\mathrm{pa}}}{C_{\mathrm{pw}}}\right)\left(\frac{T_{\mathrm{ao}}}{\gamma}\right). \tag{8}$$

Assuming, temporarily, that γ is not a function of the freshwater-flux (F_{f}), one immediately finds

$$\gamma\frac{\partial S}{\partial F_{\mathrm{f}}} = -\frac{\alpha}{\beta}\left(\frac{C_{\mathrm{pa}}}{C_{\mathrm{pw}}}\right)\frac{\partial T_{\mathrm{ao}}}{\partial F_{\mathrm{f}}}, \tag{9}$$

which has very important implications. Since adding freshwater (usually) lowers the salinity (i.e., $\partial S/\partial F_{\mathrm{f}} \leq 0$), equation (9) shows the same counter-intuitive result that the one-dimensional model showed before – the exiting air temperature T_{ao} must *increase* with increasing freshwater-flux (i.e., $\partial T_{\mathrm{ao}}/\partial F_{\mathrm{f}} \geq 0$) again implying a *warming* atmosphere in response to a slowing AMOC. We shall see later that this is also the case for almost any choice of γ (i.e., even when it *is* a function of F_{f}). An exception is the case of a constant atmospheric flow, where in the high specific heat capacity limit, $(C_{\mathrm{pw}}/C_{\mathrm{pa}}) \to \infty$, the system has no solution and, in the realistic 4 : 1 ratio case, the atmosphere very slightly cools and the heat-flux to the atmosphere hardly changes (section 6). We shall argue that this exceptional case is probably irrelevant to the ocean–atmosphere system because the oceanic deepwater formation induces a convection in the atmosphere and so generates its own atmospheric flow.

4. The hybrid dynamical-box model

We shall now apply the SN model to present-day conditions and strengthen the justification for some of the assumptions originally made in SN by analyzing more carefully the equations involved in the heat-exchange process. However, we will not be able to unequivocally support all the assumptions that were made and so the results of this study should be interpreted with caution. Recall that Heinrich events are paleoceanographic collapses of the AMOC during the last glaciation. The equations and solutions for present-day AMOC slowdowns are the same as those for glacial times except that the numerical values of the parameters involved must be chosen differently. The model (figure 1) is a combination of a dynamical aspect and a conventional Atlantic box model situated between 50°N and 70°N. As mentioned, a detailed explanation of the SN model is not given here – only the general equations needed for the present ocean and atmosphere mass-flux ratio analysis are presented here. These equations correspond to conservation of volume, salt, heat, a convection condition, and a closure condition relating the ocean and atmosphere mass transports ratio, γ. They were also used in SN.

4.1. *Volume, salt, and heat*

For steady flow, volume-flux conservation within the modeled box is given as

$$W = Q_1 + F_F, \tag{10}$$

where W is the (unknown) volume-flux of water that sinks from the upper layers to the deep layer, Q_1 the (unknown) convection-induced upper limb transport of water from the Southern Ocean and South Atlantic into the North Atlantic box, and F_F the (known) freshwater-flux into the upper North Atlantic box. The conservation of salt is

$$(S_1 - S)Q_1 = SF_F, \tag{11}$$

where S is the (unknown) salinity in the upper North Atlantic convection box and S_1 the (known) salinity of the entering water from the south, Q_1. North Atlantic deep water (NADW) is formed in the box through cooling of all upper waters

$$Q_1(T_1 - T) = \frac{A}{\rho_w C_{pw}}(F_S + F_L), \tag{12}$$

where the heat-fluxes are the same as those introduced by equations (4) and (5) with T_1 and T replacing T_{wi} and T_{wo}.

As mentioned previously, q^* and Be are dependent on the mean air temperature in the immediate vicinity of the ocean surface in a nonlinear fashion. However, since we would like to use a linear equation in T, we have taken these parameters to be constants. After completing our calculations, we verified that this simplification introduces only minor and negligible changes in the five unknowns (Q_1, W, S, T, and T_{ao}). This is not surprising as the 1 : 4 ratio of the air and water specific heat capacity implies small variations in the ocean temperature compared to the atmospheric temperature. Note that the freshwater-flux has been neglected in equation (12), as it is much smaller than Q_1. Even so, the equation is still nonlinear because both Q_1 and T are unknowns. Note that the right-hand side of equation (12) shows through equations (4) and (5) that, in the limit of U_{10} going to zero, there can be no convection because no heat can be removed from the ocean to the atmosphere.

4.2. *Convection and closure conditions*

To allow water from the box to sink into the deep layer, SN invoked the linearized *convection* condition analogous to equation (7):

$$T = T_\mathrm{D} + \frac{\beta}{\alpha}(S - S_\mathrm{D}). \tag{13}$$

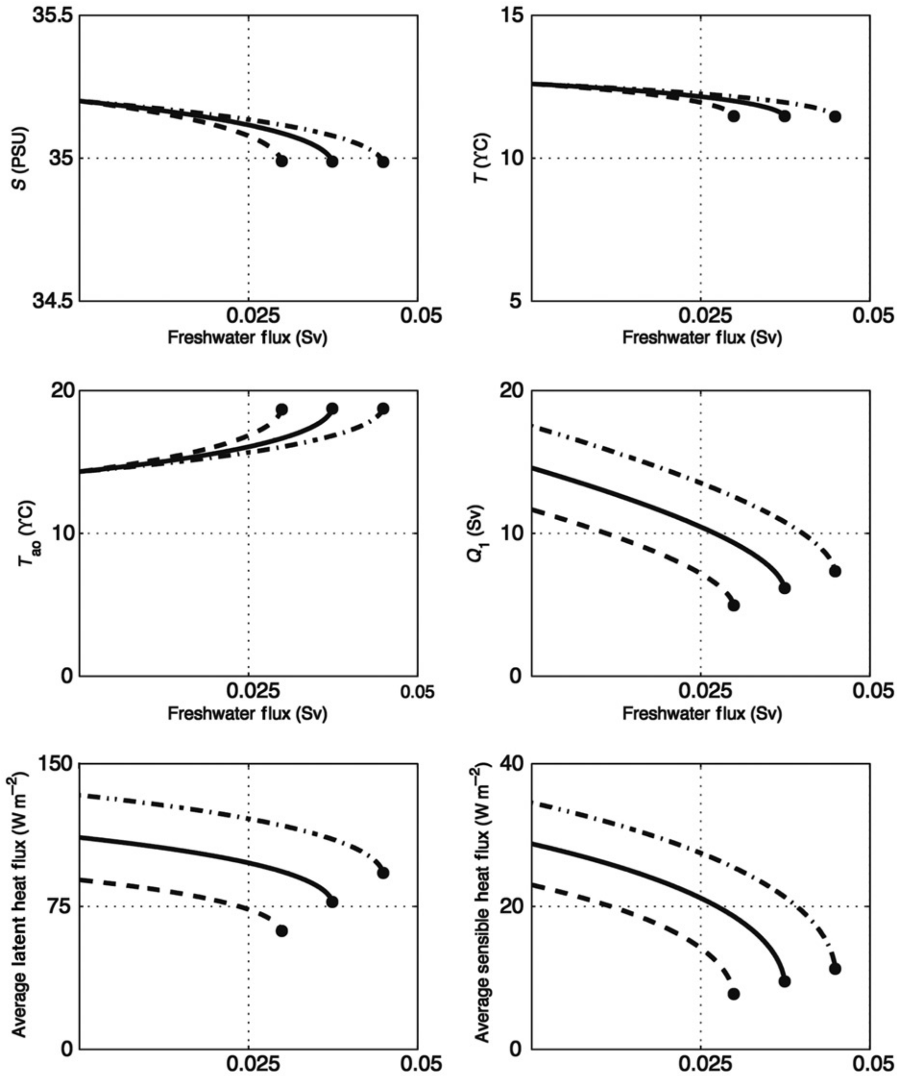

Figure 4. Analytical solution for the general modern day AMOC problem with equal mass transports in the ocean and atmosphere ($\gamma = 1$).
Notes: The AMOC for modern day oceanic salinity at the convection site (upper left), temperature at the convection site (upper right), outgoing air temperature (middle left), AMOC transport (middle right), latent heat-flux (lower left) and sensible heat-flux (lower right) as a function of the freshwater-flux, F_F [for three different values of U_{10}, $4\,\mathrm{m\,s^{-1}}$ (dashed line), $5\,\mathrm{ms^{-1}}$ (solid line) and $6\,\mathrm{ms^{-1}}$ (dashed-dotted line)]. The termination of the curves corresponds to the critical freshwater-flux where the solution breaks down (solid dots). When the critical points are reached, the AMOC collapses and warm equatorial water no longer reaches the high latitudes so that all temperatures and heat-fluxes plunge to extremely low values (figure 6).

The system of five equations (6), (10), (11)–(13) has six unknowns ($Q_1, Q_a, S, T, W, T_{ao}$), respectively, and, consequently, a *closure* condition is necessary. This condition will be specified through γ, which, as previously mentioned, measures the ratio of the oceanic and atmospheric mass transports, $\rho_w Q_1/\rho_a Q_a$ and, by definition, is always positive. Recall that SN invoked the plausible assumption that the atmospheric and oceanic mass-fluxes participating in the heat exchange are the same (i.e., $\gamma = 1$). In section 5, we shall relax this SN constraint and examine our solution for various values of γ, including the case when it is a function of F_F, as well as the case where the atmospheric transport is constant. This relaxation is one of the main contributions of this article.

To obtain the solution, we now take the Southern Ocean, deep Atlantic, and incoming air masses from the North American continent to be very large, so that T_1, T_D, T_{ai}, S_1, and S_D can all be taken as known. Following SN, we temporarily choose γ to be a constant (i.e., not a function of F_F) equal to unity, so that the five unknowns are then T_{ao}, T, S, W, and Q_1 and, as should be the case, we have five equations (6), (10), (11)–(13). The solution to this system of nonlinear algebraic equations was found analytically (SN) and need not be repeated here. The results for *present-day* conditions (instead of the glacial conditions presented by SN) are shown in figure 4. It illustrates that increased freshwater-flux reduces the AMOC volume-flux, decreases both the salinity and temperature at the convection site, but *increases* the outgoing air temperature, in line with our earlier statements. Recall that, since, here, the atmospheric jet participating in the heat-exchange process is in concert with the AMOC (in the sense that, regardless of the freshwater-fluxes, they carry identical mass-fluxes), a reduced AMOC also implies a reduced mass-flux of the atmospheric flow. The numerical values that we used for the various parameters are, by and large, the same as those of SN save that the present-day ocean is warmer and fresher so that we took $S_1 = 35.2\,\text{PSU}$, $S_D = 33.4\,\text{PSU}$, $T_1 = 15°\text{C}$, $T_D = 3°\text{C}$, $T_{ai} = 5°\text{C}$, $B_e = 0.4$, $q^* = 1.4\,\text{g}\,\text{kg}^{-1}$, and we took the expansion coefficient α to be three times its glacial value.

5. How does γ depend on increasing F_F?

We shall now relax our choice of γ to be a constant equal to unity and take it to be some unknown function that depends on F_F. The main issue that will be pointed out is that the system has physically relevant solutions only for some particular (and limited) choices of γ. We will show that, with the exception of some limited circumstances where there are hardly any changes in the heat-flux when freshwater is added (e.g., the case of constant atmospheric-flux) and, regardless of the choice of γ, the atmosphere can only *warm* when the AMOC transport is reduced. Otherwise, the system of equations does not have a physically relevant solution. We begin by inserting γ into equation (6) and using the chain rule $[\Delta T_{ao} = (\partial T_{ao}/\partial T)\Delta T + (\partial T_{ao}/\partial \gamma)\Delta\gamma]$ to examine how variations in the exiting atmospheric temperature ΔT_{ao} (due to increased freshwater-flux) are related to variations in the oceanic convective temperatures ΔT and $\Delta\gamma$. We get

$$\Delta T_{ao} = \frac{C_{pw}}{C_{pa}}[-\gamma\Delta T + (T_1 - T)\Delta\gamma], \tag{14}$$

where the first term in the brackets is always positive for an ocean subject to an increased freshwater-flux (F_F) that reduces the salinity. This is because the reduced salinity requires cooler temperature for convection to be maintained ($\Delta T < 0$). The second term sign depends only on the sign of $\Delta\gamma$, because the ocean always cools at the convection site ($T_1 - T > 0$). Note that $\Delta\gamma$ is positive when the atmospheric transport decreases (relative to the oceanic transport) due to increased freshwater-flux and negative for increasing atmospheric transport. Next, we define the total heat-flux F_H to be the sum of the latent heat-flux F_L, and the sensible heat-flux F_S (given by equation (12)) and again use the chain rule $[\Delta F_H = (\partial F_H/\partial T)\Delta T + (\partial F_H/\partial T_{ao})\Delta T_{ao}]$ to get

$$\Delta F_H = \frac{1}{2}\rho_a C_{pa} U_{10}\left(C_S + C_L\frac{R_H}{Be}\right)(\Delta T - \Delta T_{ao}). \tag{15}$$

Inserting equation (14) into equation (15) one finds

$$\Delta F_H = \frac{1}{2}\rho_a C_{pa} U_{10}\left(C_S + C_L\frac{R_H}{Be}\right)\left\{\left(1 + \gamma\frac{C_{pw}}{C_{pa}}\right)\Delta T - \frac{C_{pw}}{C_{pa}}(T_1 - T)\Delta\gamma\right\}. \tag{16}$$

Here, the first term in the square brackets is always negative (for negative ΔT) because, by definition, γ is always positive. The sign of the second term depends on the sign of $\Delta\gamma$; it is positive for positive $\Delta\gamma$ (decreasing atmospheric transport with increasing freshwater-flux) and negative for negative $\Delta\gamma$ (increasing atmospheric transport for increasing freshwater-flux).

Equations (14) and (16) will now be used to determine what behavior γ can have in order for the system to have a physically relevant solution. In particular, this will allow us to assess whether the atmosphere will cool or warm in response to an increased freshwater-flux. For clarity, we shall examine two different scenarios.

5.1. *Positive ΔT_{ao} and negative ΔF_H, i.e., warming atmosphere and a reduced heat-flux*

This is the scenario that, in our view, is the most common. For both of the above conditions to hold simultaneously, equations (14) and (16) imply

$$(T_1 - T)\Delta\gamma > \gamma\Delta T,$$

$$(T_1 - T)\Delta\gamma > \left(1 + \gamma\frac{C_{pw}}{C_{pa}}\right)\Delta T.$$

Since ΔT is negative (for increasing F_F) and $(T_1 - T)$ is positive, and since γ is, by definition, positive, the second condition is always satisfied if the first one is. So, the heat-flux reduces and the atmosphere *warms* (in response to an increase freshwater-flux) whenever

$$\frac{\Delta\gamma}{\gamma} > \frac{\Delta T}{(T_1 - T)}. \tag{17}$$

This condition is satisfied under most circumstances that we can envision. First, we immediately see that, whenever the ratio of ocean/atmosphere mass-fluxes is *constant* (i.e., $\Delta\gamma = 0$), the condition is satisfied regardless of which medium (ocean or atmosphere) has larger transport and regardless of the freshwater-flux. Second, the condition is also satisfied when the ratio of γ increases (i.e., the atmospheric transport decreases relative to the oceanic) or decreases (i.e., the atmospheric transport increases relative to the oceanic) as a function of F_F provided that the decrease is not more severe than the reduction in the oceanic temperature. Under all of these circumstances, the heat-flux *decreases* and the atmosphere *warms* with increasing F_F. Can the atmosphere cool while the heat-flux decreases, as suggested by most of the global climate models? This point is examined below where it is argued that, within the limited dynamics involved here and in SN, such a scenario is not really feasible.

5.2. *Negative ΔT_{ao} and negative ΔF_H, i.e., cooling atmosphere and a reduced heat-flux*

These are the conditions stipulated by the numerical models to be the norm. For these conditions to hold simultaneously, equations (14) and (16) imply

$$(T_1 - T)\Delta\gamma < \gamma\Delta T;$$
$$(T_1 - T)\Delta\gamma > \left(\frac{C_{pa}}{C_{pw}} + \gamma\right)\Delta T.$$

First, one immediately sees that in the large specific heat capacity limit, $(C_{pw}/C_{pa}) \to \infty$, the two equations are contradictory so the system has no solution. Taking the ocean and atmosphere heat capacity ratio (C_{pw}/C_{pa}) to be four (actual value) rather than infinity, the above can be combined to

$$\frac{5\Delta T}{4(T_1 - T)} < \frac{\Delta\gamma}{\gamma} < \frac{\Delta T}{(T_1 - T)}. \tag{18}$$

Since ΔT is negative, it is technically possible to find a very narrow range of parameters for which a decrease in γ will satisfy equation (18). We did find some extreme conditions (consistent with equation (18)) under which this happened, but for these conditions, the atmospheric temperature changes were minute and insignificant. The constant atmospheric flow discussed in section 6, which we will argue is probably irrelevant to nature, is one such case.

6. The constant atmospheric flow case

Here, we repeat the calculations presented in section 4, except that we drop the closure condition γ and, instead, introduce a fixed Q_a back into the equations and take it to be a known constant. This case is not as straightforward as it may initially seem. We present it here merely to show that even though the changes in its AMOC are as large as 50%, this scenario involves almost no changes in the heat-flux to the atmosphere and so it is not relevant to those cases where the GCMS predict a cooling Europe. Also, from a logical point of view, we do not believe that this is a realistic scenario because Q_a is not known in advance but rather should be determined as part of the air–sea interaction problem. The five equations that govern this case are now (6), (10), (11)–(13) with the five unknowns, Q_1, W, T, T_{ao}, and S, respectively.

It is instructive to first examine this case in the high specific heat capacity limit $(C_{pw}/C_{pa}) \to \infty$. In this limit, equation (6), which is now re-written as

$$\rho_w Q_1(T_1 - T) = \rho_a Q_a\left(\frac{C_{pa}}{C_{pw}}\right)(T_{ao} - T_{ai}),$$

shows that $T \rightarrow T_1$ implying that the ocean does not change its temperature no matter what happens, including the addition of freshwater. The bulk formula (1)–(4) then implies that the heat-flux can only be reduced if T_{ao} increases. However, since Q_a is now fixed and the heat-flux is also proportional to $Q_a(T_{ao} - T_{ai})$, increasing T_{ao} *increases* the heat-flux which is a contradiction. So, in the particular limit of $(C_{pw}/C_{pa}) \rightarrow \infty$ and Q_a a constant, there is *no solution* that allows changes in the heat-flux. Namely, there can

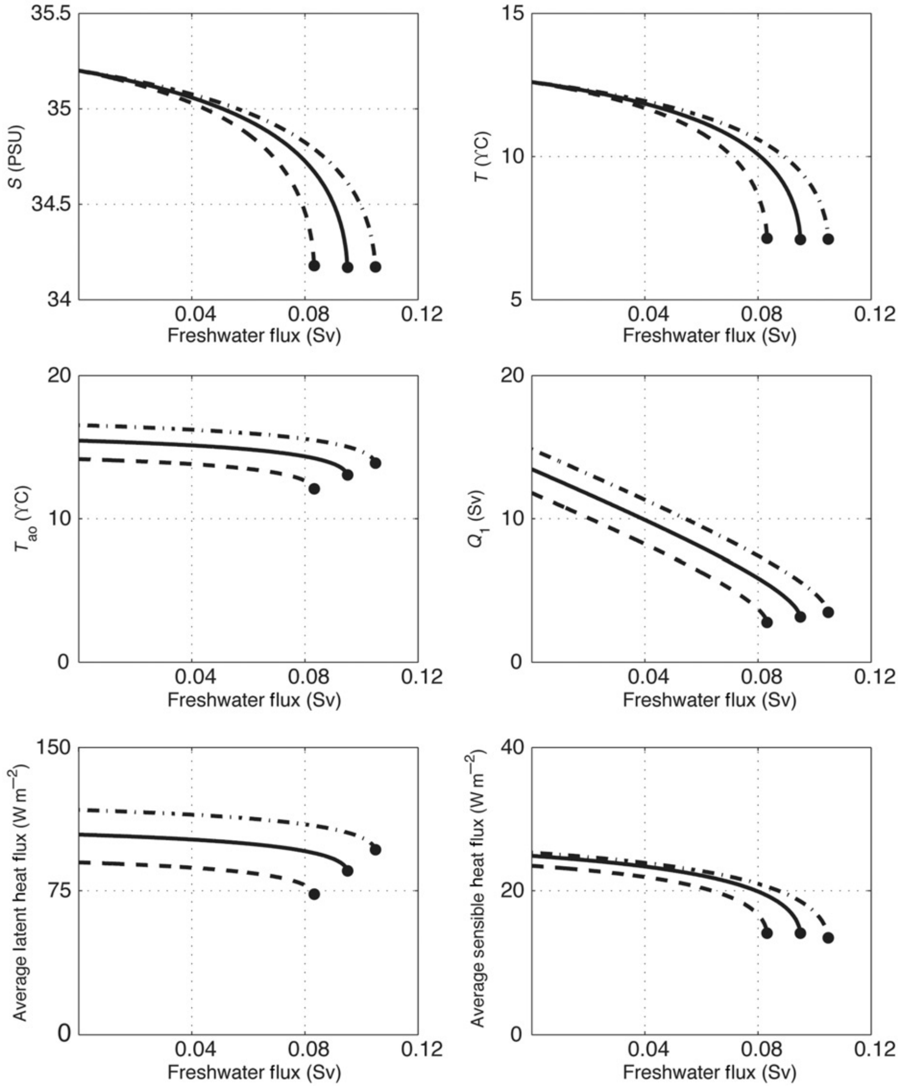

Figure 5. The constant atmospheric flow case (i.e., $Q_a =$ constant), which we do not believe is relevant to the ocean, but present it here for completion.
Notes: In this case, most of the changes take place in the ocean (upper panels, middle right panel) where the variability is unrealistically high, and, as shown in the middle left panel, very little happens in the atmosphere unless one approaches the collapsing points (solid dots). Furthermore, there are hardly any changes in the total heat-flux from the ocean to the atmosphere, which is dominated by the latent heat-flux (lower left).

be no changes in the heat-flux or the temperatures. When we drop the condition of fixed Q_a, but retain the limit $(C_{pw}/C_{pa}) \rightarrow \infty$, we recover the case discussed in section 4 with an ocean that does not change its temperature and a warming atmosphere in response to increased freshwater-flux.

In the more realistic specific heat capacity ratio of 4 : 1, the constant Q_a case does have a solution (figure 5). However, while this solution corresponds to a significant (and unrealistic) oceanic cooling (upper right panel) and, as shown in the middle right panel, a significantly reduced AMOC (Q_1), the changes in both the atmospheric temperature (middle left) and the heat-fluxes (lower panels) are *totally insignificant* unless one approaches the collapsing points (solid dots). Physically, this is because both the ocean and atmosphere cool, and so the temperature difference between the two can decrease only when the atmosphere cools less than the ocean does. Under these circumstances, the ocean cools more than it is usually expected to do and the atmosphere less than what it is usually expected to.

7. The collapsed AMOC

As shown schematically in figure 6 (based on SN analysis for Heinrich events), as long as the AMOC is operational, the outgoing air temperature increases with increasing freshwater-flux causing a decreasing mass transport. Under these conditions, the outgoing air is warmer but a smaller amount of air is being warmed so that, overall the heat-flux is smaller. The solution describing this behavior is valid all the way up to the collapsing point (solid dot). Beyond that point (dashed line), our solution is invalid and what is schematically shown is based on our understanding of the problem. When the AMOC completely collapses due to too much freshwater-flux (which prevents convection), warm equatorial water no-longer reaches the high-latitudes and so the atmosphere directly above the convection must, of course, cool. In this case, the heat exchange is not between the warm water of equatorial origins and a high-latitude cold atmosphere but rather between cold high-latitude water and a still colder high-latitude atmosphere. In other words, the process of reducing the AMOC mass-flux is nonlinear – small reductions in the AMOC mass-flux cause warming but beyond a critical point, the opposite takes place – the atmosphere cools. Note that as mentioned previously, when we speak here about the "atmosphere", we speak about the particular atmospheric jet directly participating in the convective heat-exchange process, not the entire atmosphere over the northern hemisphere which is governed by other processes.

8. How important is radiation?

Climatologists and global numerical modelers have come up with various qualitative arguments to explain the numerically observed atmospheric cooling with a decreasing AMOC. While these explanations may have merit, none of the three appears to be valid for the atmospheric jet participating directly in the convective heat-exchange process. The one that we will be addressing now is that the main heat loss from the ocean occurs via radiation, which is not strongly dependent on the atmospheric temperature so that a reduction in sea surface temperature (SST) implies a reduction in the atmospheric temperature. Using relatively old observations collected

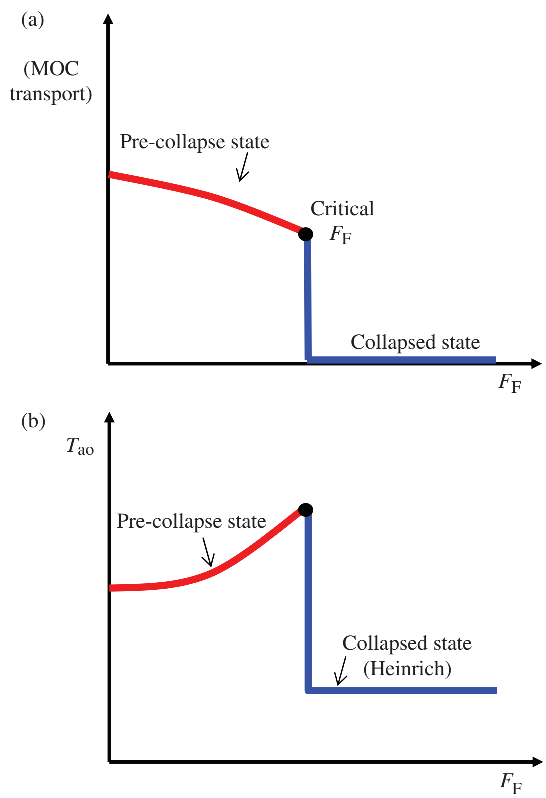

Figure 6. Schematic diagram of the AMOC transport (upper panel), and the outgoing atmospheric temperature (T_{ao}, lower panel), versus the freshwater-flux into the North Atlantic (F_F).
Notes: The solution of the pre-collapsed state (red line) was determined both here and in SN. The collapsed state (blue line) does not obey the equations presented here and its configuration is purely speculative. The transport dependency on the freshwater-flux (upper panel) is straightforward in the sense that the transport decreases until one gets to the collapse point where the convection condition can no longer be satisfied so that the whole AMOC system breaks down and the transport goes to zero. The outgoing air temperature dependency on the freshwater-flux (lower panel), on the other hand, is not at all standard. Due to the non-linearity of the problem, it first increases until the collapse point is reached. At that point, it drops dramatically because warm equatorial water is no longer imported to the high-latitudes.

in 1950s and 1960s and displayed by Hartmann (1994), we already showed in SN that radiation could not possibly be important for the high-amplitude variability of the AMOC.

The new heat-fluxes maps displayed for the first time in figure 7 (based on Gupta *et al*. 2006 and Yu and Weller 2007) repeatedly illustrate that the latent and sensible heat-fluxes for the Atlantic (where there is convection) and Pacific (where there is no convection) are dramatically different, indicating the critical role that both *latent* and *sensible* heat-fluxes play in the convective heat exchange process. We see here again that, as discussed in SN (who used old heat-flux maps, see their figure 3), the net *radiation* heat-fluxes in the two oceans are almost identical, indicating that despite the non-negligible size of the radiation terms, radiation clearly does not play a critical role in the high-amplitude variability of the AMOC. Namely, it makes no difference for the radiation field whether there is an AMOC or not and so it makes sense that the

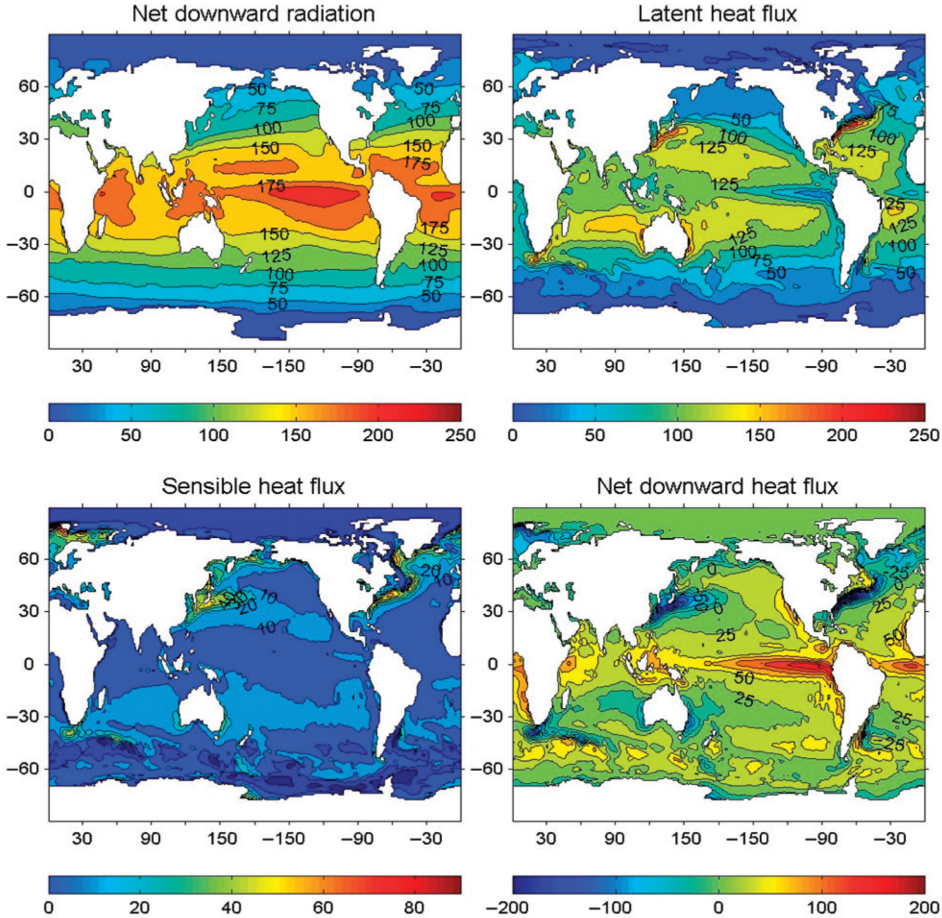

Figure 7. Updated annual average heat-flux (W m^{-2}) maps for the world ocean.
Notes: Both the latent and sensible heat-fluxes show a dramatic difference between the North Atlantic (where there is convection) and the North Pacific (no convection) indicating that both are important to NADW formation and the MOC (with latent being more important than sensible). And yet, by sharp contrast, the radiation maps show almost no difference between the two oceans indicating that it is not very important to the convection. This is because, even though the radiation terms are not small compared to latent and sensible heat-fluxes, they depend primarily on the SST, which, on a basin scale, does not vary much (compared to the atmosphere) between convective and non-convective states.

reverse is also true, i.e., turning the AMOC off-and-on is not directly affected by radiation at the ocean surface.

There is a good physical reason for this. Radiation is not so important to the convection (turning on-and-off) because it depends strongly only on the SST whereas the latent and sensible heat-fluxes are primarily a function of the temperature *difference* between the SST and statistic air temperature (SAT). The variability of the atmospheric temperature is much greater than that of the ocean (due to the much smaller specific heat capacity of air) making the difference between the convective latent and sensible heat-fluxes and non-convective-fluxes much more important than the variations in radiation. Also, radiation is independent of the salinity (Warren 1983), which again makes its distribution similar in the two oceans.

9. Is atmospheric diffusion the culprit?

As mentioned previously, most experiments with numerical models carried out to date are not in obvious agreement with our analytical results nor are they even in agreement with the very simple heat-flux considerations that we presented earlier. Global numerical modelers who sometimes view the results of high-resolution numerical models as the "absolute truth" would probably say that this is due to a problem with the analytics. This may certainly be the case because we have ignored many potentially important processes such as atmospheric moisture. However, we demonstrated in this article that the discrepancy is not due to what appear to be the most obvious weaknesses. Namely, we showed that it is not due to the old heat-flux maps that we used in SN, nor is it due to our mass transport ratio closure condition. All other aspects simply follow from first principles and should be valid in any model.

SN suggested that the discrepancy is perhaps due to the unavoidably high atmospheric diffusion that most of the numerical models employ directly or indirectly. (Some diffusive energy balanced models and intermediate complexity models such as the *Uvic* model have it as high as $10^6\,\mathrm{m^2\,s^{-1}}$ and even the atmosphere–ocean general circulation models (AOGCMs) runs which have much higher resolution of 50 km in both the atmosphere and the ocean do not resolve the convection, which is in the order of 1 km.) To see this idea clearly, consider two adjacent eastward-flowing atmospheric jets of identical mass-flux Q. One is flowing above the convection zone and exits with a (hypothetical) temperature anomaly $4\Delta T$ (above some mean) whereas the other is situated farther to the north (away from the convection) and exits with a smaller temperature anomaly of ΔT. Mixing the two jets together downstream yields a mean temperature anomaly of $2.5\Delta T$. Suppose now that the rate of freshwater-flux into the ocean increases so that the transport of the southern jet (i.e., the jet affected by the convection) decreases from Q to $0.25Q$ but its temperature anomaly increases to $5\Delta T$ because it warmed.

The northern jet outside the convection region remains the same as before, as it is not above the convection region. Mixing the two jets downstream now gives a mean temperature of $1.8\Delta T$, which is $0.7\Delta T$ cooler than before, even though the southern jet warmed up by one ΔT. Hence, because of horizontal mixing, the effect of a reduced heat-flux to the atmosphere appears in the numerics as a broad cooling even though the original process involved local warming. Note that the idea that mixing can completely mask the dynamics, even when the process in question is not of *sub-grid* scale such as an AMOC, is not new. Nof *et al.* (2007) suggested that the presence of the widely discussed AMOC hysteresis depends strongly on the mixing employed in the runs. They argue that the hysteresis goes to zero when the diffusivity goes to zero and questioned whether hysteresis really exists in the ocean. It seems that, even in the cases where the atmospheric resolution is satisfactory for atmospheric processes, it is still not satisfactory for ocean–atmosphere heat exchange processes because of the very large difference in the scales (~800 km in the atmosphere compared to ~30 km in the ocean). This implies that a *nested* atmosphere may be needed in order to resolve the issues under discussion.

Another possibility is that the observed hosing-induced cooling of Europe is not due to the slowing AMOC, but rather due to other processes such as the shifting storm tracks. Resolving the issues at hand (i.e., determining what is the discrepancy due to) is a major task involving complicated global runs (possibly with nesting) and is left as a subject of future investigation.

10. Summary and discussion

We used two simple models to show that the answer to the question whether *local* atmosphere (i.e., with an atmospheric Rossby radius of the convection) should usually cool or warm in response to an AMOC slowdown is not so obvious. First, we examined the heat exchange in an ultra simple one-dimensional model (figures 2 and 3, section 2), and then we looked at the convection condition and its relationship to the salinity and heat exchange (section 3). In both, we note that according to the bulk formulas, the heat-flux from the ocean to the air is proportional to the temperature *difference* between the (warmer) ocean and the (cooler) atmosphere. Since the specific heat capacity of water is much larger than that of the air, the oceanic temperature can only be subject to small variations. Accordingly, the only way for the difference in the ocean/atmosphere temperature to decrease in a significant way is for the atmosphere to warm, not cool. All other variations, such as variations in transports, must be in concert with the implications of the above condition.

The atmospheric heat transport in our more complicated analytical-hybrid model (sections 4 and 5) still obeys this same principle. It incorporates both the sensible and latent heat-fluxes from the ocean to the atmosphere as well as the difference between the advection of heat into and out of the atmospheric box (figure 1). The ocean responds to an increased freshwater input by cooling (slightly) and reducing its northward heat and mass transports (figure 4). As a result, the heat-flux from the (warm) ocean to the (cool) atmosphere is also reduced, but despite this reduction, the atmosphere warms in the convection region. This is because the atmosphere also responds by reducing its own mass and heat transports so that the difference between the heat advected into and out of the atmospheric box is reduced to match the reduced heat-flux from the ocean to the atmosphere. The major contributor to this reduction in atmospheric heat advection is the reduction in the volume transport (i.e., a reduction in surface winds) rather than changes in the surface air temperature (figure 4). The atmosphere warms approximately 1.2°C for 20% of an AMOC reduction (requiring ≈ 0.02 Sv of freshwater which cause a volume transport change from 16.6 to 11.6 Sv). As mentioned previously, this warming required by the bulk formula is achieved via a reduction in the atmospheric mass transport, i.e., the reduced heat-flux from the ocean is heating an even smaller amount of air, and so the air warms.

For completeness, we also presented the solution for the case of constant atmospheric flow (section 6, figure 5), a case which is interesting from a theoretical point view, but is probably less relevant for the problem of interest here because it involves almost no change in the heat-flux to the atmosphere. Two more points should be made here. First, prescribing the inflowing atmospheric temperature in the manner that we did in section 4 is reasonable, but it is clear that it will decrease in the long run (in response to a weakening AMOC) because of lateral diffusion in the atmosphere. Second, our heat exchange slowest time scale is the oceanic advection, which in the North Atlantic is a few months. Even if we take the Rossby waves time scale for the Atlantic (decadal) as the problem time scale, we still get that the model is adequate for a variability of a time scale of 50–100 years, which is often the considered variability.

Global climate models are wonderful tools when one can confidently rely on what they predict. However, the convective AMOC representation in these models is far from certain. Even the AOGCMs with 50 km resolution in both the ocean and the atmosphere may not be enough to resolve for the oceanic convective scales which at times are as small as ~1 km, much smaller than the Rossby scale. We propose that the almost-uniform numerical prediction of cooling due to hosing and an AMOC slowdown is either due to: (1) high diffusivities and mismatch in the two fluid Rossby scales (~800 km in the atmosphere compared to ~30 km in the ocean) or due to (2) changes in processes other than the AMOC (e.g., storm tracks, sea ice cover, and associated albedo). Alternatively, there may be serious implications to the approximations involved in our analytics (e.g., the neglect of moisture and changing interaction area). Resolution of these issues requires massive global computations involving nesting and is left as a subject of future investigation.

Recall that, regardless of how high the resolution of numerical global climate models is, and regardless of what comparison one makes to other models, the predictive results of the global climate models should still be regarded as questionable and unverifiable modeling results. Also, the frequently held perception that global numerical models' results are always correct unless the process in question is a sub-grid process is very questionable. Since sub-grid processes, such as mixing, govern the entire flow field and get integrated over many grid points, there are counter-examples where the model dynamics are totally wrong even when it is "merely" the sub-grid processes that are wrong (see, e.g., Nof *et al.* 2007). The Reynolds number (UD/ν, where U is the speed, D the distance, and ν the viscosity) in a one-degree (or even 50 km resolution) global climate model is typically of order one [~O(1)], so it is no exaggeration to state that these models are representing most convective processes badly.

We originally (SN) obtained our main results under the assumption that the atmospheric and oceanic mass transport are the same, but we show here that any reasonable relationship of the two transports essentially produces the same result (though the actual values do, of course, vary). An important aspect that we have totally ignored here is that the interaction area A can vary with the freshwater-flux and that moisture could also be an important parameter for the problem. Also, as one of the reviewers eloquently pointed out, the results of the model might be different if radiation and turbulent heat transport in the atmosphere were included in a manner different from that which we chose. Specifically, she/he showed that if the atmospheric horizontal diffusion due to storm tracks is proportional to the temperature gradient to some power and the radiation at the top of the atmosphere is taken to be proportional to the convective temperature, then option (2) rather than option (1) (section 1) is the answer, i.e., a slowing AMOC corresponds to a cooling atmosphere, not warming. As mentioned earlier, however, this situation corresponds to a minimal change in the heat-flux even for large changes in the AMOC transport. Overall, the main message of this article is somewhat speculative – we wish to point out that the answer to the question of what will happen if the AMOC transport is reduced is still open.

Acknowledgments

This work is supported by NSF's Physical Oceanography program (OCE-0241036 and OCE-0545204) and the Office of Polar Program (ARC-0453846), as well as NASA

(NNX07AL97G) and BSF (2006296), and FSU. We thank an anonymous reviewer for an extremely thorough and penetrating review.

References

Brayshaw, D.J., Woollings, T. and Vellinga, M., Tropical and extratropical responses of the North Atlantic atmospheric circulation to a sustained weakening of the MOC. *J. Climate*. 2009, **22**, 3146–3155.

Bryden, H.L., Longworth, H.R. and Cunningham, S.A., Slowing of the Atlantic meridional overturning circulation at 25°N. *Nature* 2005, **438**, 655–657.

Colin de Verdiere, A., A simple model of millennial oscillations of the thermohaline circulation. *J. Phys. Oceanogr*. 2007, **37**, 1142–1155.

Cunningham, S.A., Kanzow, T., Rayner, D., Baringer, M.O., Johns, W.E., Marotzke, J., Longworth, H.R., Grant, E.M., Hirschi, J.J.-M., Beal, L.M., Meinen, C.S. and Bryden, L.H., Temporal variability of the Atlantic meridional overturning circulation at 26.5°N. *Science* 2007, **317**, 935–938.

Curry, R. and Mauritzen, C., Dilution of the Northern North Atlantic ocean in recent decades. *Science* 2005, **308**, 1772–1774.

De Boer, A.M. and Johnson, H.L., Inferring the zonal distribution of measured changes in the meridional overturning circulation. *J. Ocean Sci*. 2007, **3**, 55–57.

Gupta, S.K., Stackhouse Jr, P.W., Cox, S.J., Mikovitz, C.J., Zhang, T. and Hinkelman, L., The NASA/GEWEX surface radiation budget dataset, American Geophysical Union. 2006, 22–26, May, Baltimore, Md.

Hartmann, D., *Global Physical Climatology*, p. 411, 1994 (San Diego, CA: Academic Press).

Johnson, H.L., Marshall, D.P. and Sproson, D.A.J., Reconciling theories of a mechanically-driven overturning circulation with thermohaline forcing and multiple equilibria. *Clim. Dyn*. 2007, **29**, 821–836.

Kanzow, T., Cunningham, S.A., Rayner, D., Hirschi, J.J.-M., Johns, W.E., Baringer, M.O., Bryden, H.L., Beal, L.M., Meinen, C.S. and Marotzke, J., Observed flow compensation associated with the MOC at 26.5°N in the Atlantic. *Science* 2007, **317**, 938–941.

Meinen, C.S. and Baringer, M.O., Atlantic meridional overturning circulation shows significant changes in early data from international monitoring system at 26.5°N. *US CLIVAR Variations* 2008, **6**, 1–3.

Nof, D., Simple versus complex climate modeling. *EOS* 2008, **89**, 544.

Nof, D., Van Gorder, S. and De Boer, A.M., Does the Atlantic meridional overturning cell really have more than one stable steady state? *Deep Sea Res. Part I* 2007, **54**, 2005–2021.

Polton, J.A. and Marshall, D.P., Overturning cells in the southern ocean and subtropical gyres. *J. Ocean Sci*. 2007, **3**, 17–30.

Sandal, C. and Nof, D., A new analytical model for Heinrich events and climate instability. *J. Phys. Oceanogr*. 2008, **38**, 451–466.

Tomczak, M. and Godfrey, J.S., *Regional Oceanography: An Introduction*, 422 pp., 1994 (Oxford: Pergamon Publications).

Warren, B.A., Why there is no deep water formed in the North Pacific? *J. Mar. Res*. 1983, **41**, 327–347.

Weijer, W., De Ruijter, W.P.M and Dijkstra, H.A., Stability of the Atlantic overturning circulation: competition between Bering Strait freshwater flux and Agulhas heat and salt sources. *J. Phys. Oceanogr*. 2001, **31**, 2385–2402.

Yu, L. and Weller, R.A., Objectively analyzed air–sea heat fluxes for the global ice-free oceans (1981–2005). *Bull. Am. Meteorol. Soc*. 2007, **88**, 527–539.