THE THERMOHALINE OCEAN CIRCULATION: A SYSTEM WITH DANGEROUS THRESHOLDS?

*An Editorial Comment**

Threatening scenarios of a breakdown of the Atlantic thermohaline circulation (Figure 1), a collapse of northern European agriculture and fisheries, and of glaciers advancing on Scandinavia and Scotland have captured the popular imagination in recent years, with a number of newspaper reports, magazine articles and television documentaries covering this topic with a widely varying degree of accuracy. The risk of critical thresholds in the climate system being crossed where some irreversible qualitative change sets in (such as a major ocean circulation change) is taken increasingly seriously in the discussion on anthropogenic climate change. While the 1995 IPCC report (Houghton et al., 1995) only mentioned it in passing, the upcoming third IPCC assessment will devote substantial space to this issue. So where do we stand scientifically?

Looking at the growing number of publications on thermohaline circulation stability (reviewed e.g. by Weaver and Hughes, 1992; Rahmstorf et al., 1996) one would be forgiven to be confused, finding the model results contradictory and the data inconclusive. Nevertheless I believe that a fairly consistent picture has emerged during the past years.

The first robust conclusion - taken for granted now but unknown only two-and-a-half decades ago (Oort and Stommel, 1976) - is that the thermohaline circulation makes a major contribution to the heat budget of the North Atlantic region (Roemmich and Wunsch, 1985), warming annual-mean surface temperatures locally by up to $\sim 10^{\circ}$C (Manabe and Stouffer, 1988). It is still a matter of debate how far the warm anomaly created by the oceanic heat transport extends into the European continent or affects the North American seaboard, but it is likely that at least the northwestern European countries (Great Britain and Ireland, Iceland, Scandinavia, Holland, Belgium) are warmed by several degrees, with the largest effect in winter.

The second robust conclusion is that the system is sensitive to the amount of freshwater entering the North Atlantic. This freshwater sensitivity has been the subject of numerous modeling studies and can best be summarised in a simple schematic stability diagram (Figure 2). This shows how under present-day climatic conditions the *equilibrium* rate of Atlantic overturning changes if the freshwater budget (precipitation and runoff minus evaporation) of the northern North Atlantic is altered.

* Editors' Note: This editorial was invited to provide a broad perspective on the emerging scientific debate about the likelihood and implications of altered thermohaline circulation as a result of human modifications to the atmosphere.

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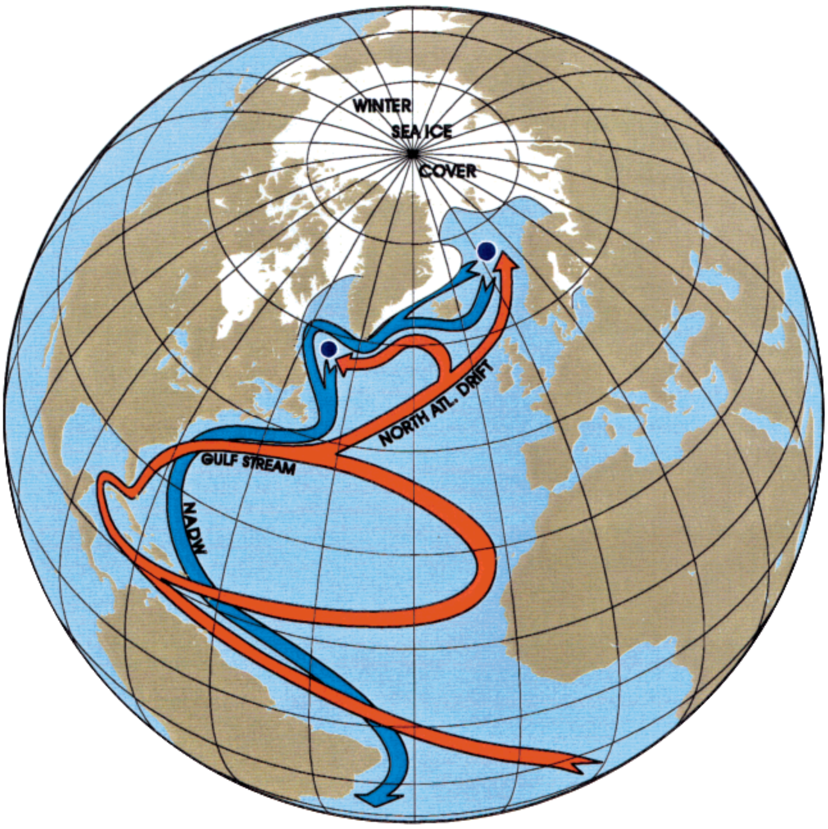

Figure 1. Highly simplified cartoon of the Atlantic circulation: surface currents are shown in red, the flow of North Atlantic Deep Water in blue. The two main areas of deep water formation are indicated by blue dots. The warm North Atlantic Drift heats north-western Europe, pushing back the winter sea ice margin.

The fascinating aspect of this stability diagram is its non-linearity. A simple positive feedback (first described by Stommel, 1961) - overturning enhances salinity due to salt transport from the south, and high salinity in turn enhances overturning - makes the governing equation quadratic rather than linear, hence the parabolic curves in Figure 2. This means that in a certain range of freshwater input (between zero and about 0.15 Sv; 1 Sv = 10^6 m^3s^{-1}) two fundamentally different climates can be stable: with and without the characteristic 'conveyor belt' style circulation in the Atlantic (Manabe and Stouffer, 1988). There is a well-defined threshold point *S* (mathematically speaking a saddle-node bifurcation) beyond which the circulation breaks down. The figure contains two parabolae computed for different temperatures (i.e., locations) of deep water formation, to indicate that the ocean can switch between different convection sites (Lenderink and Haarsma, 1994; Rahm-

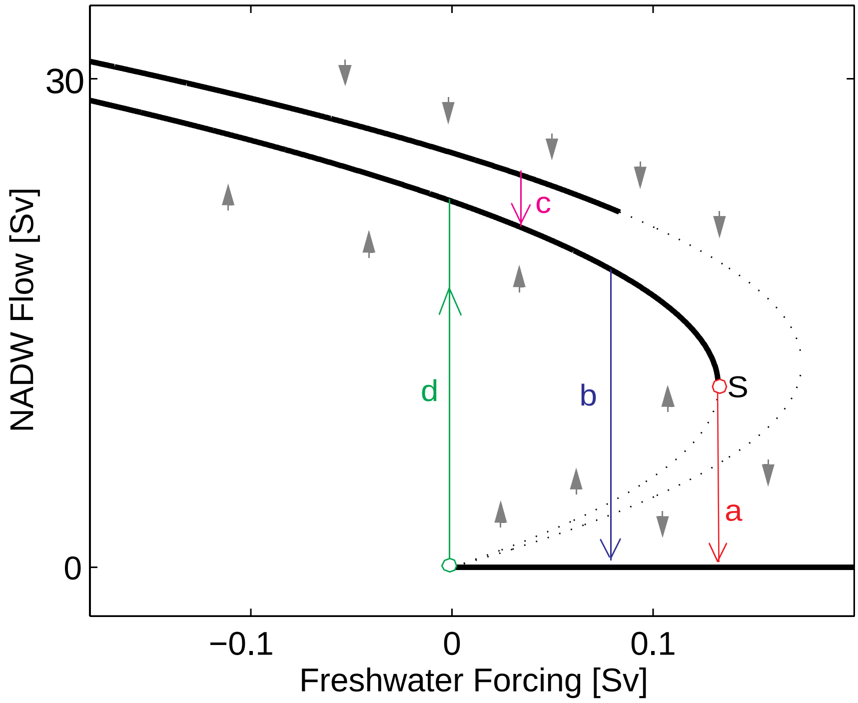

Figure 2. Schematic stability diagram of the Atlantic thermohaline circulation. The two upper heavy branches indicate the possibility of multiple states with different convection sites. Possible transitions indicated are: (a) advective spindown when reaching the bifurcation point *S*, (b) polar halocline catastrophe when convection is locally suppressed by a freshwater lid, (c) change in convection location, (d) start-up of NADW formation. The lower unstable branch (dotted) separates the basins of attraction of 'on' and 'off' modes; small grey arrows indicate the direction of movement in phase space. The point of zero freshwater forcing is well defined in a conceptual model but not easy to establish in a more complex model; it can be understood as the point where the thermohaline circulation changes from importing freshwater to the Atlantic (negative freshwater forcing) to exporting freshwater (see Rahmstorf, 1996 for a full discussion). Figure adapted from Rahmstorf (1999a); for an early version of the stability diagram see Stocker and Wright (1991).

storf, 1994) due to a second positive feedback (see Figure 1 for the two main convection locations).

This stability diagram can be analytically computed from simple conceptual models and traced with a surprising degree of agreement with sophisticated general circulation models (Rahmstorf, 1996). Transitions of the thermohaline circulation can be understood in terms of the transition mechanisms indicated on Figure 2. It is my contention that most if not all published model results are basically consistent with this stability diagram, even in a rough quantitative sense. Figure 2 could be taken as a null hypothesis for a general stability diagram until more is learned by further experiments. The stability diagram helps to interpret many model experiments, including the causes for differences or apparent contradictions between models.

Many early model studies were performed with ocean-only models with nearly fixed sea surface temperature (SST), neglecting the feedback of the ocean circulation on SST. This distorts the parabola in Figure 2 in the sense that the bifurcation point *S* shifts to the left, narrowing the bistable regime and bringing the critical threshold closer to the present climate (see Figure 2 in Rahmstorf and Ganopolski, 1998). This is why these ocean-only models were more sensitive to freshwater than coupled climate models.

As far as coupled models are concerned, the stability diagram has so far only been calculated for hybrid (e.g., Rahmstorf, 1995) or intermediate complexity (e.g., Ganopolski et al. 2000) coupled models but due to the high computational cost not for full coupled GCMs, so that for the latter we can only draw conclusions from various transient experiments. Nevertheless, taking all evidence from different model types together, it appears that the basic shape and size of the stability diagram is similar in the models, and that the major difference between models is in *where they locate the present climate* on the diagram.

Almost all models put the present climate somewhere in the bistable regime (between 0 and 0.15 Sv freshwater input in Figure 2); the classic example is Manabe and Stouffer (1988). This means that if a sufficiently large temporary freshwater input is applied, the climate moves to the right and via transition *a* onto the lower branch, where it remains even after the perturbation ends (shifting back to the left). There is observational evidence that the present climate is indeed on the right-hand side of Figure 2 (Weijer et al., 1999), even though the Atlantic is a net evaporative basin (Baumgartner and Reichel, 1975). The reason is that the Atlantic freshwater budget has three main components: surface exchange, wind-driven ocean currents and thermohaline circulation, so that net evaporation does not imply a negative freshwater forcing on the thermohaline circulation.

An exception is the model of Cubasch et al. (this issue), which locates the present climate in the left half of Figure 2 in a regime where *only* the 'conveyor belt' solution is stable. This is consistent with the large overturning rate of this model (26 Sv) and the large freshwater perturbation (ca. 0.3 Sv) required to shut down the circulation. After the addition of freshwater is ended, this model moves back to the left and via transition *d* back to the upper branch, i.e., the Atlantic circulation recovers. This behaviour has been reproduced both in the GFDL coupled model (Manabe and Stouffer, 1999) and the CLIMBER-2 model (Ganopolski et al. 2000) by increasing vertical diffusion in the ocean up to the high level found (for numerical reasons) in the ocean model of Cubasch et al. This is reassuring, since it shows that consistent results are obtained when consistent parameter values are used, and that the reasons for (at first sight) contradictory model results can be understood.

Further experiments (unpublished, in cooperation with R. Stouffer) have shown that changing from horizontal to isopycnal diffusion in the ocean component also shifts a model along the stability diagram, otherwise leaving the diagram largely the same. In addition to the level of diffusion, the surface forcing used during spinup and the flux adjustment procedure (if any is used) can affect the position of the model climate in Figure 2.

Another possible source of confusion when looking at model results is the *duration* and *location* of freshwater input in sensitivity experiments. Duration is crucial because the advective spindown process (transition *a*) can take centuries - the less the amount of freshwater exceeds the critical threshold value, the longer it takes (Cessi, 1994). If the circulation is not completely shut down when the additional freshwater input ends, the model climate may end up in the basin of attraction of the upper branch and the 'conveyor belt' can recover even from a very low overturning rate. For example, in the GFDL coupled model an addition of freshwater at the rate of 1 Sv for 10 years reduces the circulation to 7 Sv (Manabe and Stouffer, 1995) while adding 0.1 Sv for 500 years reduces it to 4 Sv (Manabe and Stouffer, 1997) before it recovers. The latter freshwater addition (0.1 Sv) is obviously just beyond the bifurcation *S* in this model, as the circulation spins down only slowly over 500 years, not quite reaching the stable state on the 'off' branch. Adding 1 Sv for 100 years makes the model climate reach this state (Manabe and Stouffer, 1999).

The location of the freshwater perturbation is also important - a rule of thumb is: the closer to the deep water formation regions, the more effective it is (Rahmstorf, 1996; Manabe and Stouffer, 1997). This is plausible, as the deep water formation regions represent the 'achilles heel' of the conveyor belt (W. Broecker), and adding freshwater further south (say, in the tropical Atlantic) means that only part of it will be advected north to affect deep water formation, and that only with delay. This delay becomes important in transient problems such as greenhouse warming, and the location of the freshwater source is more important for the transient than the equilibrium response.

An important reason for studying the freshwater sensitivity of the Atlantic thermohaline circulation is paleoclimatic data showing major and rapid ocean circulation changes in the past. Past water mass characteristics and currents can be reconstructed from ocean sediments. Three main modes of Atlantic circulation can be identified in the data (Alley et al., 1999): (i) a warm or interglacial mode with deep water forming in the Nordic Seas, (ii) a cold or glacial mode with deep water forming south of the shallow sill between Greenland, Iceland and Scotland, and (iii) a 'switched off' mode with practically no deep water formation in the North Atlantic. Transitions between these modes are associated with dramatic changes in surface climate known as Dansgaard-Oeschger events, which are centered on the North Atlantic region and register for example in the Greenland ice cores but in some cases as far afield as New Zealand. The trigger mechanism for these past mode switches in climate is unknown, but the regularity of their occurrence points at an underlying mystery climate cycle with a period of about 1500 years (Bond et al., 1997) which pervades both the last Glacial and the Holocene. The stability properties of the Atlantic circulation could be such that the cycle triggers mode switches only in the cold glacial mode, while the Holocene ‘warm mode’ is more stable with respect to this particular forcing.

Not surprisingly, the existence of abrupt past climate changes has fuelled concern over the possibility of setting off similar changes in future, first voiced strongly by Broecker (1987). When the Earth heats up, two factors affect the density of ocean waters and thereby the thermohaline circulation: the temperature increase and the change in freshwater budget. This makes the problem more complex than the issue of pure freshwater sensitivity discussed above and compounds the uncertainty.

First, the *initial position* of the pre-industrial climate on the stability diagram will determine the sensitivity of the Atlantic circulation for global warming scenarios. The model’s initial rate of Atlantic overturning is a good indication for this. Second, as it is not the greenhouse gases per se but the surface warming which affects the circulation, the traditional *climate sensitivity* (surface temperature rise for CO_2 doubling) is important. Third, the change in the Atlantic freshwater budget for a given surface temperature change factors in. This can be described by an Atlantic *hydrological sensitivity* parameter (defined in Rahmstorf and Ganopolski, 1999).

A large uncertainty is associated with the latter, for several reasons. Global mean evaporation and precipitation changes for a given warming differ by up to a factor of three between different atmospheric models. Regional precipitation and runoff changes over the North Atlantic will differ even more. In the present climate, the freshwater export from the Atlantic is larger by about 0.1 Sv in El Niño compared to La Niña conditions, so that a long-lasting shift in the prevalence of El Niño could have a significant effect on the Atlantic (Schmittner et al., 2000). In one model, increased El Niño frequency resulting from global warming draws enough water vapor from the subtropical Atlantic across into the Pacific to cancel out the weakening effects on the thermohaline circulation (Latif et al., 2000). Though other models so far do not show this effect, this is an important issue requiring further study. Estimates for runoff from the Greenland ice sheet and other melting glaciers in the North Atlantic catchment are also highly uncertain (Houghton et al., 1995), and many models do not include this potentially important source of freshwater at all. Melting sea ice can make a small (but not negligible) further contribution close to the sensitive deep water formation regions (Manabe and Stouffer, 1994). The observed loss of perennial Arctic sea ice over the past decades (Johannessen et al., 1999; Rothrock et al., 1999) would be equivalent to a freshwater flux of about 0.01 Sv (M. Morales Maqueda, personal communication).

The relative importance of temperature increase and freshwater input in weakening the Atlantic conveyor belt differs between different models (Mikolajewicz and Voss, 1998; Dixon et al., 1999; Rahmstorf and Ganopolski, 1999). However, it appears that warming alone cannot close down the circulation; ultimately a substantial amount of freshwater is required to push the conveyor belt to extinction.

The fact that global warming happens on a time scale close to the response time of the ocean circulation complicates matters further. The circulation appears to be more sensitive to a given warming if it occurs more rapidly (Stocker and Schmittner, 1997; Stouffer and Manabe, 1999), so that slowing down the *rate* of greenhouse gas increase appears to be a policy that can buy greater security even if the same level is reached in the end.

Having said all this, what does the "murky crystal ball" of climate models tell us about the Atlantic ocean in a global warming future? Most models show hardly any greenhouse-induced circulation change up to the present day (neither is there any clear observational evidence for such a change), but they do show a decline of the Atlantic overturning rate by 20-50% by the end of the 21^{st} century (Rahmstorf, 1999b). No model reaches the threshold for a complete collapse by this time, and few longer-term scenarios are available. A complete shutdown was simulated by Manabe and Stouffer (1993, 1994) for a quadrupling of atmospheric CO_2 and by Rahmstorf and Ganopolski (1999) for a transient peak in CO_2 content. These studies suggest that the risk of a shutdown arises after a global warming of 4-5°C in a century if the Atlantic hydrological sensitivity is relatively high. It takes several centuries until the circulation is completely shut down in both cases. The simulations of Manabe and Stouffer (1993) and Hirst (1999) further show the possibility of a shut-down of the formation of Antarctic Bottom Water, which is the second major deep water source of the world ocean.

The models generally agree that during the phase of greenhouse gas increase a weakening or even collapse of the conveyor belt does not lead to a surface cooling below pre-industrial levels. A serious cooling of the North Atlantic region (including northwestern Europe) results only in the longer term, when greenhouse gases decline again and the circulation remains in the 'off' mode. In the worst case scenario of Rahmstorf and Ganopolski (1999), regional surface temperature increases by around 3°C during the coming hundred years, then dramatically drops back to preindustrial levels in the first decades of the 22^{nd} century, declining more gradually thereafter. Among the global impacts of a circulation shutdown are an increased rate of sea level rise (Knutti and Stocker, 2000) and a reduced ability of the ocean to take up CO_2 from the atmosphere (Sarmiento and Le Quéré, 1996). It should be noted that once the Atlantic circulation has collapsed, it is likely to remain off for many centuries.

Finally, the possibility of a regional circulation change, rather than a full collapse, should not be forgotten. In the British Hadley Center model (Wood et al., 1999), convection in the Labrador Sea (Figure 1) shuts down early in the new century. This is a convective mode transition of type *c* in Figure 2. While not as dramatic as a complete shutdown of deep water formation, it is still an important qualitative change in the climate system.

A recent workshop* on the thermohaline circulation stability problem concluded that a major ocean circulation change should be considered a 'low probability - high impact' risk, and emphasized that proper risk analysis is crucial for this type of non-linear climatic change. Performing single 'best guess'-style greenhouse scenario simulations has only limited value for capturing such climatic risks or evaluating their probability. This presents a major challenge to the modeling community, especially since society and policy-makers have a great interest in understanding this type of risk.

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