Simple Theoretical Model May Explain Apparent Climate Instability

STEFAN RAHMSTORF AND ANDREY GANOPOLSKI

Potsdam Institute for Climate Impact Research, Potsdam, Germany

24 November 1997 and 23 July 1998

ABSTRACT

The authors propose a simple theory to explain a climatic drift previously found in a series of coupled general circulation model (GCM) experiments. Their theory places the GCM results on a simple stability diagram for the thermohaline circulation in the Atlantic. It is argued that the drift is due to Stommel's salt advection feedback and arises because the particular spinup procedure of the coupled model has in some cases resulted in an initial state on an unstable solution branch. Such drift may result from model initialization with unsuitable relaxation boundary conditions, rather than being a climatic transition that could occur in the real world.

1. Introduction

The stability of our present climatic state, and the possibility that it might undergo a strongly nonlinear transition in response to increasing greenhouse gas forcing, has become an important research topic in recent years. The latest report of the Intergovernmental Panel on Climate Change (IPCC; Houghton et al. 1995) warns that ''Future climatic changes may also involve 'surprises'. . . . Examples of such nonlinear behaviour include rapid circulation changes in the North Atlantic.'' Global warming scenarios generally predict a significant weakening of the thermohaline circulation and in some cases even a permanent shut down (for a recent discussion, see Rahmstorf 1997). Before we are able to quantify the risk of major circulation changes, we need to gain a better understanding of the mechanisms of climatic instability and state transitions.

The purpose of this note is to propose a physical mechanism for the apparent climatic instabilities found in a series of coupled ocean–atmosphere model experiments by Tziperman (1997, T97 hereafter). He prepared various initial states of the ocean component of the model by spinning it up with a range of different surface salinities in the North Atlantic. He found that initial states with a relatively fresh North Atlantic and weak thermohaline circulation led to a marked increase in thermohaline overturning and strong oscillations during the coupled model integration, and he attributed this to an inherent instability in the climate system. He further found that some initial states were more susceptible to a circulation collapse triggered by a brief freshwater anomaly. Here we propose a simple theoretical explanation for the behavior of the coupled model.

2. A simple conceptual model

We use Stommel's (1961) classic conceptual model of the thermohaline circulation, in a four-box version (Rahmstorf 1996, R96 hereafter) suitable for the interhemispheric Atlantic circulation. The model configuration is shown in Fig. 1. The model is kept as simple as possible by considering only one ''conveyor belt''–style loop of flow through the Atlantic, ignoring more localized cells and wind-driven circulation. A justification for this is provided by comparisons with general circulation model results (R96). In this four-box model, the meridional mass transport (or overturning) m in equilibrium is approximately described by the quadratic equation

$$\left[1 + \frac{k\alpha}{\gamma}(T_1^* + T_3^* - 2T_2^*)\right]m^2 + k\alpha(T_2^* - T_1^*)m + k\beta S_0 F_1 = 0 \qquad (1)$$

[see appendix of R96, Eq. (13)]. Here α, β are thermal and haline expansion coefficients, γ is a thermal coupling constant, and T_i^* are the temperatures that the southern, northern and tropical Atlantic are relaxed towards. Also, F_1 is the freshwater forcing (multiplied by a reference salinity, S_0, for conversion to a salt flux) that determines the rate of overturning in the Atlantic, while F_2 (see Fig. 1) determines the relative distribution of the freshwater forcing between the tropical and the northern Atlantic box. Note F_2 does not enter Eq. (1) and thus does not influence the equilibrium flow rate,

Corresponding author address: Stefan Rahmstorf, Potsdam Institute for Climate Impact Research, P.O. Box 60 12 03, 14412 Potsdam, Germany.
E-mail: rahmstorf@pik-potsdam.de

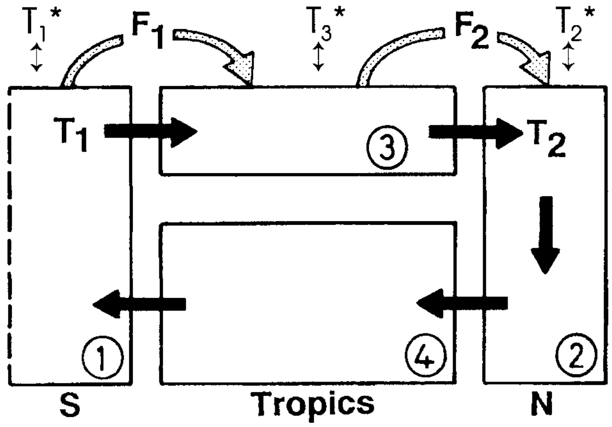

FIG. 1. Schematic of the four-box conceptual model of the thermohaline circulation in the Atlantic. The temperatures of boxes 1, 2, and 3 are relaxed toward the values T_1^*, T_2^*, and T_3^*, respectively, and salinities are forced by the freshwater fluxes F_1 and F_2 (there are only two independent freshwater fluxes so that total salinity in the system is conserved). The meridional flow (black arrows) is proportional to the density difference between boxes 2 and 1.

because in equilibrium it makes no difference whether freshwater enters box 2 directly at the surface or indirectly via box 3. The flux F_2 determines the relative salinities of tropical and northern boxes. This affects the transient behavior of the model and the stability of the steady solutions, as it determines the feedback by salt advection from lower latitudes. Detailed discussion of the roles of F_1 and F_2 is given in R96. The parameter k is an empirical constant linking the intensity of overturning to the north–south density gradient.

We consider two versions of this model: Version 1 corresponds to an ocean-only spinup driven by traditional relaxation boundary conditions for temperature and salinity, and version 2 mimics a coupled climate model. The latter is driven by a fixed freshwater flux and a weak thermal coupling of 10 W m^{-2} K^{-1} (see Rahmstorf and Willebrand 1995 for a justification of this approach).

The equilibrium solutions of thermohaline mass transport m as function of the freshwater flux F_1 are shown in Fig. 2 for both model versions. Being almost quadratic [or in the approximate Eq. (1) exactly quadratic], the shape of $m(F)$ is essentially determined by two independent parameters. It is practical to use the two unknown quantities k and $(T_2^* - T_1^*)$ as tunable parameters, and to fix a priori the other parameter values that are well constrained by observations. This includes the freshwater flux F_2, which is found by requiring a realistic salinity gradient between the tropical and northern boxes, and the value of $(T_2^* - T_3^*)$, which sets the corresponding thermal gradient. First, we determine the two tunable parameters for the "coupled" version 2, by fitting the model to a hysteresis experiment with an ocean circulation model coupled to an atmospheric energy balance model (as described in R96). The fit obtained is shown in Fig. 3. The difference to the fit in

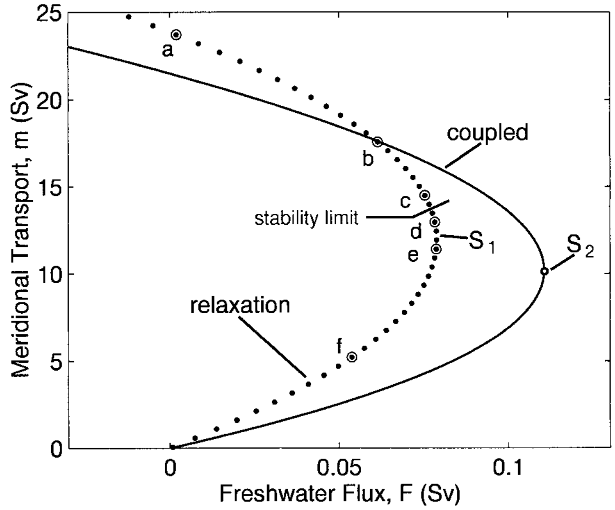

FIG. 2. Equilibrium solutions of the conceptual thermohaline circulation model: the dotted line shows solutions of version 1 with relaxation boundary conditions, the solid line of the coupled version 2 [shown are the full solutions, not the quadratic approximation given in Eq. (1)]. Points a–f correspond to the initial states of Tziperman's circulation model experiments. Note that the bifurcation point S_2 of the coupled system differs from that obtained by experiments initialized with relaxation boundary conditions, S_1.

R96 arises because here we fit the conceptual model to the OGCM with isopycnal diffusion (in R96 horizontal diffusion was used), that is, to an ocean model identical to the one used by T97.

For the "relaxation" version 1 we use a stronger thermal coupling of 75 W m^{-2} K^{-1} to mimic an ocean-only spinup with relaxation boundary conditions, but we retained the same flow constant k (which obviously should not depend on the surface boundary condition used).

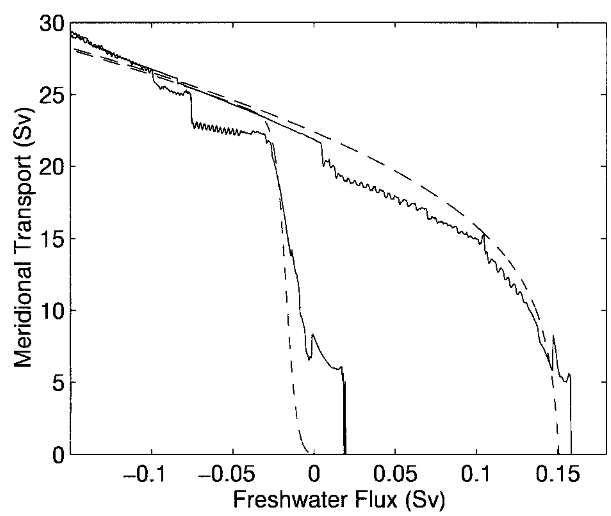

FIG. 3. Hysteresis curve of the box model (dashed) and a general circulation model (solid). These curves are traced by slowly increasing and decreasing the freshwater flux into the North Atlantic, at a rate of 0.05 Sv kyr^{-1}.

TABLE 1. Parameters of the conceptual model. Note that the model is formulated independently of the box volume, hence many parameters are given in yr^{-1}. To convert to absolute values, box volumes of 10^{17} m^3 were assumed. The freshwater flux F_2 is not relevant for the relaxation version of the model.

Model parameters		
α	1.7×10^{-4} K^{-1}	Thermal expansion coeff.
β	8×10^{-4} psu^{-1}	Haline expansion coeff.
k	25 yr^{-1}	Empirical flow constant
Forcing for version 1: Relaxation conditions		
T_3^*–T_2^*	12.3 K	Relaxation temperatures
T_1^*–T_2^*	2.1 K	Relaxation temperatures
γ	0.3 yr^{-1}	Thermal coupling = 75 W m^{-2} K^{-1}
Forcing for version 2: Coupled		
T_3^*–T_2^*	15 K	Relaxation temperatures
T_1^*–T_2^*	3.8 K	Relaxation temperatures
γ	0.04 yr^{-1}	Thermal coupling = 10 W m^{-2} K^{-1}
F_2	7.9×10^{-5} yr^{-1}	Freshwater forcing = 0.25 Sv

The values of T_i^* were adjusted so that both model versions have the same surface heat fluxes for the present-day overturning rate of 18 Sv (Sverdrup = 10^6 m^3 s^{-1}); this makes the model states of versions 1 and 2 identical at point *b* in Fig. 2. This is mathematically fully equivalent to the flux adjustment procedure used in general circulation models to make ocean-only and coupled models coincide for the present-day climate. All parameter values of the conceptual models are thereby fixed; they are given in Table 1. It is important to note that our parameter fit for both model versions is independent of the results of T97 that we set out to explain.

Points a–f in Fig. 2 correspond directly to the initial states prepared by T97 and were obtained in the same way, using relaxation boundary conditions with different relaxation salinities and strong thermal coupling (version 1). They lie on the solution parabola for relaxation conditions, which differs from the coupled solution parabola (solid line) because of the stronger thermal coupling, which suppresses the negative thermal feedback on the overturning (weaker overturning leads to a colder northern box; see Rahmstorf and Willebrand 1995). The box-average salinity increments we used to obtain points a–f are one-quarter of the values reported by T97, to account for the limited spatial extent of the anomalies of T97 compared to our box size (the numbers given by T97 are actually maxima of regional Gaussian-shaped salinity anomalies). The conceptual model then reproduces the locations of T97's initial states in the (F_1, *m*) phase plane: the flow values agree well, and the freshwater fluxes of states c, d, and e are almost identical to each other while both the strong (a, b) and weak (f) overturning states correspond to a smaller freshwater flux (see Figs. 1c and 2 of T97).

The instability of some of the initial climate states found by T97 can now be understood with the help of the conceptual model. If we start a coupled integration from the initial states, the freshwater flux will remain constant in our simple model and approximately constant in the general circulation model (GCM). This is because flux adjustments make sure that it is initially the same, and the feedback of thermohaline circulation changes on the atmospheric hydrological cycle is weak (Hughes and Weaver 1996). Tziperman further uses flux adjustments (different for each initial state) on the heat flux to keep the coupled model close to each initial state; otherwise the model would drift toward the line of coupled equilibria (solid line). With these flux adjustments, the initial states a–c can indeed be stabilized in the coupled model. However, the *lower* solution branches (below the bifurcation points S_1 or S_2) in Stommel's box model are known to be unstable for fixed freshwater fluxes (see Weaver and Hughes 1992) and are only accessible if a salinity relaxation boundary condition is used. The reason for their instability is the positive salt advection feedback: if the circulation increases, more salt is advected to the northern box, increasing its density and increasing the circulation further. State *f* is clearly well inside this unstable regime.

States c–e are very close to the bifurcation S_1 and require further discussion. Because not only salinity feedback changes when the coupled integration starts, but also the thermal damping becomes weaker, states just *below* S_1 can be stabilized by thermal feedback. This has to do with the fact that the dotted line is not an equilibrium line of the coupled system. Each initial point is on its own coupled branch, which intersects the dotted line at that point. It differs from the coupled line shown (which intersects at b) by its different heat flux adjustment. Some of the intersection points (just below S_1) are between the lower relaxation branch and an upper coupled branch. On the other hand, states on an upper branch just above the bifurcation can become oscillatory unstable (Tziperman et al. 1994) due to salt advection from the tropical box in a four-box model (this oscillatory instability is not found in Stommel's original two-box model). Stability analysis of our conceptual model shows that when both these effects are taken into account, the stability threshold turns out to be at a flow of 12.9 Sv, between points c and d (see Fig. 2) exactly as in the climate simulations of T97.

The oscillations found by T97 are also consistent with our conceptual model, as it shows damped oscillations above and growing oscillations below the stability limit. Finally, our simple model also reproduces the coupled GCM response to an anomalous freshwater input of 1 Sv (10 yr)$^{-1}$. As in T97, the circulation recovers from this if initialized with the present climate (b), but not if initialized with one of the unstable initial states.

3. Discussion and conclusions

We have tried to construct the simplest possible model that could be used to reproduce Tziperman's GCM experiments. In essence it is a model of one simple loop of thermohaline flow passing through the Atlantic, driven by a density difference between the North and South Atlantic and subject to various feedbacks between salinity, temperature, and flow rate. The wind-driven circulation, all regional details, and possible changes in convection locations (Rahmstorf 1995) are ignored. In spite of its simplicity, the conceptual model reproduces all the major aspects of the coupled GCM simulations reported in T97. When equivalent experiments are performed with the conceptual model, the results are (even quantitatively) similar to the GCM results. If this agreement is not just fortuitous, then the simple oceanic feedbacks captured in our model may indeed dominate the response of the GCM and provide an explanation for its stability behaviour.

The drift of the coupled model of T97 away from initial states d–f would then essentially result from the fact that with relaxation boundary conditions any initial state can be obtained, even states that in a coupled system are destabilized by the positive thermohaline circulation–salinity feedback described by Stommel. A coupled model will drift from such initial states toward a stable state. The theoretical model underscores the basic conclusion of T97 and previous studies (reviewed in Rahmstorf et al. 1996) that states with weak thermohaline circulation tend to be unstable, but it also suggests that model experiments have to be designed and interpreted with caution.

The solid curve in Fig. 2 corresponds to the stability curve of the coupled climate system (as approximated by the simple model) and has a straightforward interpretation: it shows how the equilibrium ocean circulation would change if the freshwater input into the Atlantic were to change. The dotted curve, however, has no obvious application to the real world. For example, states c–f have a considerably weaker thermohaline circulation than is observed today, but they do not have a sea surface temperature (SST) field that is physically consistent with this. A weaker circulation would lead to a colder North Atlantic, but these states have been obtained by strongly restoring the model SST to present-day observed SST. In this sense these states can be considered physically implausible. Tziperman (1997) has investigated whether, by using flux adjustments, such initial states can be kept stable in a coupled model. This is an interesting technical question for initializing coupled models, but it is hard to relate to a climatic phenomenon in the real world. Our conceptual model suggests that the apparent stability limit found in these experiments (on the dotted line near S_1) could be much closer to the present-day climate than the real stability threshold of the coupled climate system (on the solid line near S_2).

Acknowledgments. This work was supported by the European Union's Environment and Climate programme.

REFERENCES

Houghton, J. T., L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell, Eds., 1995: *Climate Change 1995.* Cambridge University Press, 572 pp.

Hughes, T. M. C., and A. J. Weaver, 1996: Sea surface temperature–evaporation feedback and the ocean's thermohaline circulation. *J. Phys. Oceanogr.,* **26,** 644–654.

Rahmstorf, S., 1995: Multiple convection patterns and thermohaline flow in an idealized OGCM. *J. Climate,* **8,** 3028–3039.

——, 1996: On the freshwater forcing and transport of the Atlantic thermohaline circulation. *Climate Dyn.,* **12,** 799–811.

——, 1997: Risk of sea-change in the Atlantic. *Nature,* **388,** 825–826.

——, and J. Willebrand, 1995: The role of temperature feedback in stabilizing the thermohaline circulation. *J. Phys. Oceanogr.,* **25,** 787–805.

——, J. Marotzke, and J. Willebrand, 1996: Stability of the thermohaline circulation. *The Warm Water Sphere of the North Atlantic Ocean,* W. Krauss, Ed., Borntraeger, 129–158.

Stommel, H., 1961: Thermohaline convection with two stable regimes of flow. *Tellus,* **13,** 224–230.

Tziperman, E., 1997: Inherently unstable climate behaviour due to weak thermohaline ocean circulation. *Nature,* **386,** 592–595.

——, J. R. Toggweiler, Y. Feliks, and K. Bryan, 1994: Instability of the thermohaline circulation with respect to mixed boundary conditions: Is it really a problem for realistic models? *J. Phys. Oceanogr.,* **24,** 217–232.

Weaver, A. J., and T. M. C. Hughes, 1992: Stability and variability of the thermohaline circulation and its link to climate. *Trends in Physical Oceanography,* Council of Scientific Research Integration, Trivandrum, 15–70.