

# 15 Decadal Variability of the Thermohaline Ocean Circulation

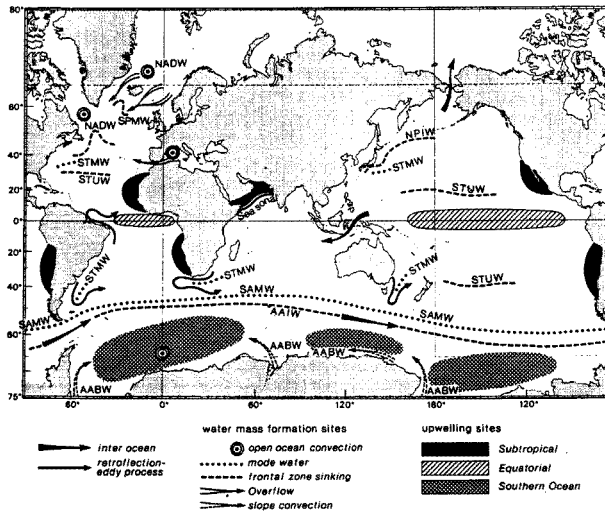
STEFAN RAHMSTORF

*Potsdam Institute for Climate Impact Research,  
Potsdam, Germany*

## 15.1 The thermohaline circulation as component of the climate system

In contrast to the wind-driven part of the ocean circulation, the thermohaline circulation is driven by density gradients and extends from the surface to the abyssal ocean. It is organized in basin-scale coherent current systems, which are globally connected through the Southern Ocean (see e.g. Schmitz 1995 for a review and Macdonald and Wunsch 1996 for a recent quantitative estimate). The flow is driven by surface heat and freshwater (i.e. buoyancy) fluxes. The heat fluxes dominate in that the densest surface waters are found in cold high-latitude regions, not in the salty subtropics; the flow is thus thermally direct. It is the downward penetration of buoyancy through diffusion which in the long run ultimately engages all ocean depths in the circulation, a fact first recognized by Sandström (1908). On shorter time scales, intense buoyancy loss (due to heat loss) in a few localized areas of deep convection has a controlling influence on the thermohaline ocean circulation: if all deep convection is interrupted in models (e.g. through a surface freshwater cap) the thermohaline circulation breaks down in a matter of years. On a time scale of millennia, diffusion then erodes the stratification until convection restarts and a long-term heat balance between downward diffusion and convection is re-established. The role of the thermohaline circulation in this balance is to transport heat horizontally towards the convection sites. The deep flow is constrained by continental boundaries and the bottom topography, so that the flow through deep channels and over sills plays an important role in the spreading of deep waters. Due to the Earth's rotation the flow is strongly concentrated in boundary currents attached to continental slopes or mid-ocean ridges. An overview of key localities and mechanisms for the thermohaline circulation is given in Fig. 15.1. Note that deep water renewal of the world ocean through convection occurs only in a few locations: in the North Atlantic in the Greenland and Labrador Seas, and in the Southern Ocean near the Antarctic continent.

The thermohaline circulation plays a crucial role in the climate system because of its large meridional heat transport. It roughly equals that of the atmosphere but is much more localized: in the present climate, about 1.2 PW (1 PW =  $10^{15}$  W) is transported towards the North Atlantic across  $24^{\circ}$ N by the ocean (Roemmich and Wunsch, 1985). A simple calculation illustrates this: about 17 Sv (1 Sv =  $10^6$  m<sup>3</sup>/s)



**Fig. 15.1** Location of generation and transformation areas of the major water masses: North Atlantic Deep Water (NADW), Antarctic Bottom Water (AABW), Subtropical Mode Water (SAMW), Subantarctic Mode Water (STMW), Subtropical Underwater (STUW), Subpolar Mode Water (SPMW) and North Pacific Intermediate Water (NPIW). Also indicated are interocean and intergyre exchange and retroreflection processes. (From CLIVAR Scientific Steering Group 1995, p. 84.)

of North Atlantic Deep Water (NADW) flows southward across  $24^{\circ}\text{N}$  at a temperature of  $\sim 2.5^{\circ}\text{C}$ , and is replaced by northward flow of near-surface waters which are some 16 degrees warmer; this implies a heat transport of 1.1 PW. In contrast, the wind-driven cell of about 23 Sv Florida current and 6 Sv Ekman transport recirculates towards the south at barely lower temperature, giving a heat transport of only 0.1 PW (Roemmich and Wunsch 1985). The wind-driven gyres are furthermore of limited latitudinal extent, in contrast to the deep boundary currents which provide an almost continuous flow from the Arctic to the Southern Ocean. The climatic effect of this thermohaline flow pattern is often illustrated by comparing the surface temperatures of the northern Atlantic with comparable latitudes of the Pacific; the former are  $4\text{--}5^{\circ}\text{C}$  warmer. It is thus plausible that variations of the thermohaline circulation could lead to multi-year sea surface temperature (SST) anomalies in the subpolar North Atlantic; this was first suggested in the 1960's by Bjerknes (1964).

The climatic effects are not limited to the North Atlantic, however. Model studies suggest that a climatic equilibrium without NADW formation would be significantly warmer in most ocean regions outside the North Atlantic (Manabe and Stouffer 1988; Rahmstorf and Willebrand 1995), as the ocean circulation moves less heat from other regions into the Atlantic. A recent coupled simulation of glacial climate (Ganopolski et al. 1998), however, shows that a change in the latitude

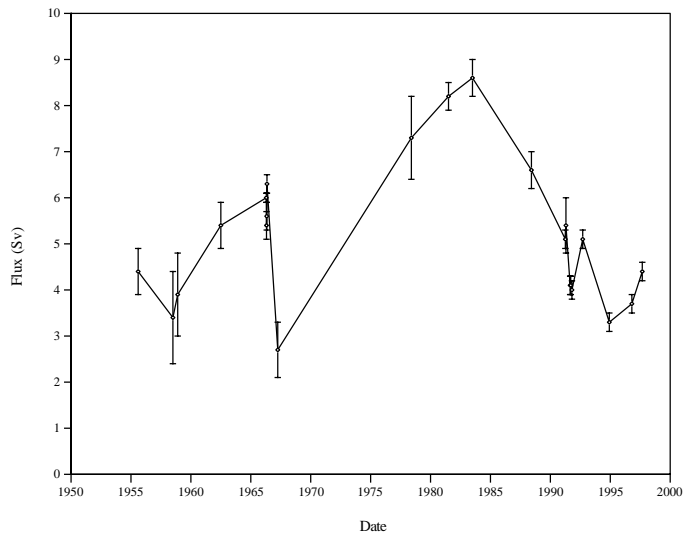
of NADW formation may also reduce the *global mean* temperature, as sea-ice advances southward in the Atlantic and the planetary albedo is increased. After a shutdown of deep convection the ocean can temporarily act as a global heat sink while the deep ocean warms. This can cause a transient global cooling as in a simulation of Fanning and Weaver (1997a; perhaps aided by unrealistically strong thermal coupling of northern and southern hemispheres in their diffusive energy balance model).

The thermohaline circulation also plays an important part in the global carbon cycle, by coupling the atmosphere to the huge oceanic carbon reservoir and by providing a connection to the carbon sink in the ocean sediments. Recent model simulations (Mikolajewicz 1997) show that a breakdown of NADW formation leads to a large transient increase in atmospheric CO<sub>2</sub> content, as the oceanic uptake of CO<sub>2</sub> in high latitudes and the outgassing in low latitudes get out of balance. A simulation of anthropogenic climate change which involves a reduction of NADW formation also shows that this leads to reduced oceanic carbon uptake and increased atmospheric CO<sub>2</sub> levels, providing a positive feedback which amplifies the anthropogenic global warming (Sarmiento and Le Quéré, 1996).

## 15.2 Evidence for decadal variability of the thermohaline circulation

The previous section has highlighted the active role of the thermohaline ocean circulation in the climate system; this suggests that variations in the circulation could have important climatic repercussions. However, direct measurements of decadal or longer term variability of the deep ocean circulation are not available. No deep boundary currents were even discovered in the Pacific until 1967 and in the Indian Ocean until 1970 (B. Warren, personal communication). The ongoing World Ocean Circulation Experiment (WOCE) is expected to give new estimates of the deep flow at key locations, but at present large uncertainties surround even the present mean transport of the deep western boundary currents, let alone their long-term variability. If the variance spectrum is red, measuring a true mean may not be possible (Wunsch 1992). Because of the problem of defining a valid “level of no motion” even repeated hydrographic sections may not yield reliable information on deep current variability. Wunsch (1992) further highlights the difficulty of interpreting variability, even if we manage to observe it.

The situation is best in the North Atlantic. Several moored current meter arrays (typically lasting one year) have measured the overflow of NADW from Denmark Strait and found a strikingly constant transport, possibly due to hydraulic control of the flow over the sill (Dickson, Gmitrowicz, and Watson 1990; Dickson and Brown 1994). However, this does not rule out variations on longer time scales. Very recently, Bacon (1998) has compiled hydrographic sections off Greenland and has concluded that there is good evidence for decadal variations in overflow from the Nordic Seas (Fig. 15.2). He linked these overflow changes to the North Atlantic



**Fig. 15.2** Transport of the deep western boundary current flowing southward off south-east Greenland, derived from various hydrographic sections since 1955. (From Bacon 1998.)

#### Oscillation (NAO) index.

More information is available about convection variability. Convection not only plays a role in driving the deep flow but is also the prime mechanism for releasing heat from the deep ocean, so that convection variability has a direct influence on surface climate. Because deep convection events in the ocean occur only sporadically under harsh winter conditions, they have not been observed directly until recently. However, convection homogenizes temperature and salinity in the water column, so that much information about past convective activity can be deduced from vertical temperature and salinity profiles and other tracers. For the Greenland Sea, Schlosser et al. (1991) concluded from tracer measurements that deep convection had been reduced by 80% during the 1980's. Strong convective variability was also found in the Labrador Sea. Regular observations from Ocean Weather Station Bravo (Lazier 1980) show that deep convection occurred almost every winter until 1967, but then stopped for several years until convective activity resumed in 1972. The interruption of convection was a consequence of the so-called Great Salinity Anomaly (Dickson et al. 1988). A similar event occurred in the early eighties (Lazier 1995). Model studies suggest that a change in convection leads to a change in deep flow within a few years (Rahmstorf 1995b), but direct evidence is lacking. In the Mediterranean, the effects of an individual convection event have been observed in the deep current downstream (Send, Font, and Mertens 1996); a pulse of cold water was seen passing several weeks after a convection event, but did not lead to a pulse in deep current transport.

Further indirect evidence for thermohaline circulation variability is provided by observed changes in water mass characteristics. Hydrographic data from different periods of the past decades have been compared, using either comprehensive data compilations (Levitus 1989a; Levitus 1989b; Levitus and Antonov 1995), individual repeated sections (Roemmich and Wunsch 1984; Bryden et al. 1996) or long-term time series (e.g. the one maintained near Bermuda, Joyce and Robbins 1996). As the signals are small, the former pose a significant quality control challenge. To interpret hydrographic changes, it is useful to distinguish changes in isopycnal depth from temperature and salinity changes on a given isopycnal (Bindoff and McDougall 1994). Isopycnals may move up or down in response to basin-scale wind changes, while an increase in thickness of a particular density layer may indicate an increased production of the water mass concerned. Changes of the  $\theta$ - $S$  (potential temperature vs. salinity) signature on an isopycnal, on the other hand, reflect the buoyancy forcing in the source region and should change only on a slower (advective) time scale. Increasing salinity from 1957 to 1992 in the intermediate water at 24°N in the Atlantic has been tentatively interpreted as a reduction of Antarctic Intermediate Water inflow (Lavin et al. 1994; Bryden et al. 1996). Deep water changes at the same latitude from 1981 to 1992 may reflect reduced flow of Labrador Sea Water (possibly a consequence of the interruption of convection discussed above) and a freshening of deep waters formed in the Greenland-Iceland-Norwegian Sea (Bryden et al. 1996). Further north, between 40°N and 50°N, repeat sections have revealed a significant cooling of Labrador Sea water since 1988 and an eastward spreading of the colder water (Sy et al. 1997). Changes in the properties of the Upper North Atlantic Deep Water layer near Bermuda appear to be linked to changes in Labrador Sea convection six years earlier (Curry, McCartney, and Joyce 1998).

Finally, paleoclimatic proxy data show that profound changes in the thermohaline circulation must have occurred in the past, some of which apparently happened on decadal time scales. During the last glacial maximum, the rate of NADW export from the Atlantic was probably similar to today's (Yu, Francois, and Bacon 1996), but convection sites were shifted to the south and NADW flow was shallower, while Antarctic Bottom Water (AABW) penetrated further northward (Sarnthein et al. 1994; Sarnthein et al. 1995). This picture is confirmed by the so far only coupled ocean-atmosphere simulation of glacial climate (Ganopolski et al. 1998). Several times during the glacial period the NADW circulation appears to have collapsed rapidly (within about a decade), probably due to an inflow of meltwater or a surge of icebergs (Heinrich event) into the North Atlantic (MacAyeal 1993; Sarnthein et al. 1994; Sarnthein et al. 1995; Björck and al. 1996).

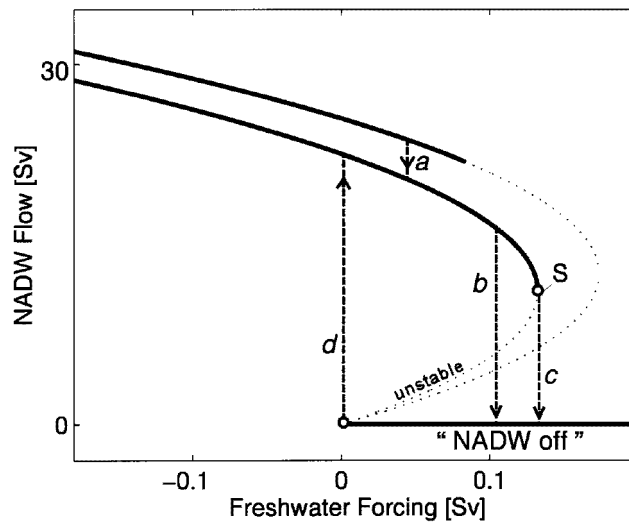
### 15.3 Physics of the thermohaline circulation

Before we discuss variability mechanisms of the thermohaline circulation, it is useful to recall some of its basic physical characteristics (see also the review papers

of Weaver and Hughes 1992 and Rahmstorf, Marotzke, and Willebrand 1996). As stated above, the flow is driven by thermally dominated density gradients which are ultimately a result of the uneven heating of our planet by the sun. It would thus be natural to conclude that the thermohaline circulation is driven by the density gradients between low and high latitudes, and that it acts to transport heat from low to high latitudes. This is indeed the picture chosen by Stommel (1961) in his famous conceptual model of a one-basin, one-hemisphere circulation, which was designed to demonstrate the salinity feedback (discussed below). However, the complex geography of the world ocean and the interaction and vertical layering of deep waters of North Atlantic and Antarctic origin make the situation more complex. For the Atlantic thermohaline circulation, Rahmstorf (1996b) has recently argued that it behaves like a true interhemispheric flow system driven by density gradients between the northern deep convection regions and the southern end of the enclosed Atlantic (south of 30°S), rather than the tropics. In this view, which is based on GCM experiments, the thermohaline circulation primarily transports heat from the South Atlantic to the North Atlantic, rather than from low to high latitudes. When NADW formation is shut down, the entire North Atlantic surface cools and the South Atlantic surface warms, with the zero line near the equator. Such a cross-hemispheric flow differs in its stability from Stommel's one-hemisphere circulation. For example, inflow of freshwater in the tropics, which in Stommel's model would decrease tropical density and thus enhance NADW flow, was found to weaken NADW flow in the GCM experiments. And as we will see below, the magnitude of the temperature difference driving the flow is an important parameter for the stability of the system.

Stommel (1961) was the first to recognize the fundamental salinity feedback which is one factor responsible for the non-linear behavior of the thermohaline circulation. This positive feedback is easy to understand: the circulation transports salt towards the deep water formation region, thereby increasing the salinity and density there and enhancing the circulation. It is again arguable whether this salt transport is essentially from the tropics to the high latitudes as in Stommel's simple model, or whether salt is transported from the South Atlantic (or even from outside the Atlantic) to the North Atlantic - for equilibrium conditions, the GCM results show a salinity reduction even in the Atlantic tropics in the absence of NADW flow. In either case, Stommel's salinity feedback makes the thermohaline circulation a partly self-sustaining system which can have multiple equilibria.

A schematic stability diagram (Fig. 15.3) illustrates this. The parabolic curves are solutions of Stommel's conceptual model, or more precisely of Rahmstorf's (1996b) modified version for cross-hemispheric flow in the Atlantic (the solutions of the two differ only in the values and interpretation of their parameters). Note that in a certain parameter regime multiple equilibria exist: essentially equilibria with NADW "on" and "off", and a third unstable branch (dotted line) in between. For the equilibrium marked "NADW off", Stommel's model would predict a reverse, haline driven flow. In GCMs Antarctic Bottom Water (AABW) dominates the deep Atlantic after the cessation of NADW formation, but the dynamics governing



**Fig. 15.3** Schematic bifurcation diagram for North Atlantic Deep Water (NADW) flow, as derived from GCM results and theoretical models. Solid lines show stable equilibrium states; unstable states are dotted. The upper two solid lines correspond to equilibria with different NADW formation sites, the bottom one to an equilibrium without NADW formation. The uppermost stable branch is not drawn out to its saddle-node bifurcation, indicating that it becomes convectively unstable before reaching the advective stability limit. Examples of the four basic transition mechanisms are shown by dashed lines: **(a)** local convective instability, a rapid shutdown or startup of a convection site, **(b)** polar halocline catastrophe, a total rapid shutdown of North Atlantic convection, **(c)** advective spindown, a slow spindown process triggered when the large-scale freshwater forcing exceeds the critical value at Stommel's saddle-node bifurcation *S* while convection initially continues, **(d)** startup of convection in the North Atlantic from a "NADW off" state. The convective transitions *a*, *b*, *d* can be triggered either when a gradual forcing change pushes the system to the end of a stable branch or by a brief but sufficiently strong anomaly in the forcing.

AABW flow into the Atlantic are not well understood and it is not included in the diagram. The origin of the freshwater axis divides a flow regime where both salinity and temperature gradients drive the flow (on the left), from a regime where temperature drives and salinity inhibits the flow (on the right). Only in the latter the salinity feedback is positive and the two basic equilibria exist. Beyond Stommel's saddle-node bifurcation *S* the freshwater input is so strong that no NADW formation can be sustained.

For a GCM the meaning of "zero freshwater input" is not obvious - the experiments were performed with a fully two-dimensional surface freshwater flux field (fixed in time), with anomalies added in certain regions. Zero freshwater input is in this case defined as the point where the regime of the "no NADW" equilibrium begins - i.e. the point where freshwater input starts to inhibit NADW flow, and

where the outflow of NADW becomes fresher than the northward near-surface return flow. The theory predicts that at this point the NADW flow for the “on” state is twice the flow at Stommel’s bifurcation, which is the minimum sustainable NADW flow. This is confirmed in the GCM experiments. The GCM results and the fitted conceptual model further show a ratio of NADW flow to freshwater input at Stommel’s bifurcation which is equal to  $m_{\text{crit}}/F_{\text{crit}} = 53$ . This dimensionless number depends only on the thermal and haline expansion coefficients and the temperature difference driving the flow; the latter was found to be  $5.9^{\circ}\text{C}$  (Rahmstorf 1996b). This is clearly not compatible with the idea that it is the equator-to-pole temperature difference which drives NADW flow.

Another consequence of the stability diagram is that the stability of the circulation will depend strongly on the mean NADW formation rate; the location of a model on the stability map is crucial for understanding its stability behavior. A model with very strong NADW formation (more than 25 Sv, say) will be situated in the left half of the diagram, where no ‘NADW off’ state exists, and it will require a very large freshwater anomaly to suppress NADW formation (in excess of 0.2 Sv). This situation is found e.g. in the ECHAM3/LSG model (Schiller, Mikolajewicz, and Voss 1997), which forms 26 Sv of NADW in the control run.

In the North Pacific, the net freshwater input is so large that it is most likely beyond the limit up to where deep water formation could be sustained, even with the help of the salinity increase that such a flow would bring. For a recent review of reasons for the difference between Atlantic and Pacific in this respect, see Rahmstorf, Marotzke and Willebrand (1996).

A second factor responsible for the non-linear behavior of the thermohaline circulation is convection. The integral effect of convection is essentially a vertical homogenization of the water column with little net vertical transport (Send and Marshall 1995). The net sinking motion associated with deep water formation is a secondary effect, which does not necessarily have to all occur in the same location as convection (see Ernst 1995 for Lagrangian deep water trajectories from an eddy-resolving model). Even in the absence of feedback, convection is strongly non-linear in its behavior: if the surface buoyancy flux is gradually varied, convection is turned ‘on’ or ‘off’ as the buoyancy flux changes sign. In addition there is a positive feedback, which can be described as follows: reduced surface density reduces vertical convection, which can lead to an accumulation of freshwater near the surface, which reduces surface density even further. This feedback can be illustrated with a simple conceptual model of two vertically stacked boxes (Welander 1982; Lenderink and Haarsma 1994). Like Stommel’s salt advection feedback, the convective feedback can lead to multiple equilibria if temperature and salinity are coupled at different time scales. An example of such coupling are mixed boundary conditions, where surface temperature is relaxed towards some chosen value and a fixed surface salt (or freshwater) flux is prescribed. Assume a convection patch where cooling exceeds the input of freshwater in its effect on buoyancy, keeping convection going. If convection is interrupted, the surface cools to the prescribed relaxation temperature; after that there is no more heat loss. The freshwater input



continues, however, assuring that convection remains permanently switched off. Convection thus behaves like a 'flip-flop' - in a certain parameter regime it can stabilize itself as being either 'on' or 'off'. In a three-dimensional ocean model, this can lead to the existence of many different, more or less stable convection patterns (Lenderink and Haarsma 1994; Rahmstorf 1995b) and transitions between them (indicated on Fig. 15.3 by the two upper branches connected with a convective transition, *a*). Such a convective state transition has recently also been found in the Hadley Centre coupled model, where Labrador Sea convection started up after 700 years of the control run, causing a sudden increase in NADW flow (Tett, Johns, and Mitchell 1997).

A number of further feedbacks affect the thermohaline circulation. The most important is the temperature advection feedback, which is analogous to the salinity advection feedback discussed above, except that it is negative: the advection of warm water from the south by the thermohaline circulation reduces density in the convection region, thereby weakening the flow. Both simple estimates and GCM experiments (Rahmstorf 1995b; Rahmstorf and Willebrand 1995; Rahmstorf, Marotzke, and Willebrand 1996) demonstrate that this feedback is powerful enough to have a strongly stabilizing effect on the flow rate of NADW. The traditional restoring boundary condition on temperature (Haney 1971) largely suppresses this important feedback, and if combined with fixed freshwater fluxes (allowing for the positive salt advection feedback) leads to a much too unstable thermohaline circulation in models.

Other feedbacks operate through the atmosphere. The first of these is intimately connected to the temperature feedback discussed above and limits it: if the surface of the ocean is warmed through increased oceanic advection of heat, the atmosphere will absorb some of this extra heat and both transport it away and radiate it to space. Atmospheric transport is much more effective for small scale temperature anomalies (Bretherton 1982). On the smallest scales (e.g. those of convective heat release by the ocean) the damping of SST anomalies is limited only by the efficiency of local air-sea coupling (typically  $\sim 50 \text{ W/m}^2\text{K}$  but strongly dependent on atmospheric conditions - it can be especially large during the winter-time convection season). A large-scale warming of North Atlantic SST due to heat advection in the thermohaline circulation, however, is damped by the atmosphere at only about  $10 \text{ W/m}^2\text{K}$ : at this rate, a heat transport of 1.2 PW across  $24^\circ\text{N}$  leads to an average SST warming of  $4^\circ\text{C}$  north of this latitude. Simple energy balance models of the atmosphere can be used to account for this feedback (e.g. Power et al. 1995; Rahmstorf and Willebrand 1995), and recent comparisons have shown that these compare well with coupled models in studies of thermohaline circulation sensitivity (Capotondi and Saravanan 1997; Weber 1997).

Another feedback is the change in atmospheric vapor transport in response to thermohaline circulation (i.e. SST) changes (Nakamura, Stone, and Marotzke 1994). However, this feedback is probably weak and not even its sign is known with certainty (Tang and Weaver 1995; Hughes and Weaver 1996; Weber 1997). Finally, there is the possibility of wind feedback (Fanning and Weaver 1997a;

Mikolajewicz 1997; Schiller, Mikolajewicz, and Voss 1997) which deserves further study. If this feedback primarily acts on small scale convection regions, the resolution and realism of present models is probably not adequate to draw firm conclusions.

The same is true for all feedbacks acting locally on convection, rather than on the large-scale forcing of the thermohaline circulation. Atmospheric models do not yet reliably forecast regional precipitation changes. Another problem is that present global ocean models are unable to simulate the physics of the overflow of NADW over the Greenland-Iceland-Scotland sill, and usually form most of their NADW through convection south of Greenland or Iceland. NADW formation in these models is well removed from the sea ice edge, reducing interaction with ice and ice-albedo-feedback (Ganopolski et al. 1998). This could be one of the reasons why the significant reduction of NADW formation found in most greenhouse scenario experiments (e.g. Manabe and Stouffer 1994; Cubasch et al. 1995) only leads to a regional SST cooling south of Greenland, but has surprisingly little effect on Northern European air temperature (Rahmstorf 1997).

## 15.4 Basic variability mechanisms

This section will provide a brief general discussion of possible mechanisms of thermohaline circulation variability. More specific examples of variability modes found in circulation models will follow in later sections. One may group the variability mechanisms into the following five categories:

- Hasselmann's mechanism: integration of white noise forcing
- stochastic excitation of damped oceanic internal modes
- self-sustained internal oscillations
- externally forced variability (including state transitions)
- coupled modes.

Other ways to classify variability are of course equally possible - see e.g. Stocker (1996) and Sarachik et al. (1996) for useful reviews. We will now give a brief discussion of some characteristics of the variability types.

### 15.4.1 Hasselmann's mechanism

In this mechanism the ocean plays but a passive role, integrating the effect of weather fluctuations in the overlying atmosphere, which essentially act like a random forcing on the ocean. Consider a temperature response of the form

$$\frac{dT}{dt} = Q - \lambda T \quad (15.1)$$

where  $Q$  is a stochastic thermal forcing and  $-\lambda T$  a damping term, then a Fourier transformation of this equation leads to the temperature variance spectrum

$$S_T(\omega) = \frac{S_Q(\omega)}{\omega^2 + \lambda^2} \quad (15.2)$$

where  $S_Q(\omega)$  is the heat flux variance. For a white noise forcing ( $S_Q = \text{const.}$ ) the response is red roughly up to the damping frequency  $\lambda$  of temperature anomalies; for longer frequencies the spectrum becomes flat. If the memory of the ocean ( $1/\lambda$ ) is long enough, this mechanism can lead to climate variability at decadal time scales. This mechanism does not directly involve the thermohaline circulation, but could force thermohaline circulation variability as a result of persistent density anomalies in the upper ocean (as in the model of Weisse, Mikolajewicz, and Maier-Reimer 1994).

#### 15.4.2 Internal variability modes of the ocean

Internal variability modes come in two flavors: modes that are self-sustaining due to non-linear feedback and modes which are damped, so that they would die out if not excited by some forcing. Simple conceptual models are useful to demonstrate how thermohaline feedbacks can lead to internal oscillations; an instructive “gallery” of such oscillators was presented by Welander (1986). A prototype of advective oscillations is the Howard-Malkus loop oscillator (Welander 1967), where flow in a differentially heated ring of fluid is driven by temperature differences. In this flow loop, a warm anomaly can be sustained if it passes through the region of cooling more quickly than through the heating region. If the flow responded instantaneously to the temperature changes, this would not be the case - the key to make the oscillation self-sustaining is a delay in the flow response. The oscillation arises because the same anomaly alternates between decelerating or accelerating the flow in different phases of the oscillation cycle; the period of the oscillation is thus set by the advection time around the loop. If we follow the anomaly around the loop, it experiences a positive feedback which maintains it, but if we look at one particular place we see a delayed negative feedback: the warm anomaly creates a current which removes it. Because of the delayed response, the reaction overshoots and a cold anomaly is generated instead of the previous warm anomaly, and an oscillation starts. This is a simple thermal/advective oscillation, characterized by feedback between heating and advection. A linear box model where circulation anomalies are related out of phase to temperature anomalies, and which therefore can oscillate, has been suggested by Huck, Colin de Verdière and Weaver (1997). In this model the feedback is not dominated by the advection of a thermal anomaly in the mean flow ( $\bar{v}T'$ ) as in Welander’s loop oscillation, but by the advection of the mean thermal gradient due to a flow anomaly ( $v'\bar{T}$ )

Other oscillations are truly thermohaline, i.e. depend on the different response of temperature and salinity; anomalies of temperature and salinity are damped at different time scales. Welander (1986) discusses a thermohaline/advective version of the loop oscillator which does not require any delay in the dynamic response in order to oscillate. A direct oceanic analogy to the loop oscillation would be a density anomaly travelling with the thermohaline overturning in a loop around the

North and South Atlantic; indeed such oscillations are found in ocean circulation models on centennial time scales. However, more localized decadal advective oscillations, in which density anomalies are not advected around the whole loop but arise from flow anomalies ( $v'\bar{T}$ ,  $v'\bar{S}$ ) are more common in circulation models. A simple linear thermohaline oscillator model has been proposed by Griffies and Tziperman (1995); this shows damped oscillations where temperature anomalies in the sinking region (providing negative feedback) lag behind salinity anomalies (providing positive feedback). This characteristic phase relation is similar to that found in the coupled model of Delworth, Manabe and Stouffer (1993).

Yet other oscillations involve oceanic convection instead of advection, e.g. the thermohaline/convective oscillator of Welander (1982). Here, convection is periodically switched on and off. When it is off, salt slowly accumulates at the surface until convection starts. However, once the salt has been removed by convection, strong surface heating stops convection until the slower, but ultimately dominant salt accumulation triggers another convection event. Note that this particular oscillation works only for haline (not thermally) driven convection. However, a similar, purely thermal convective oscillation can be obtained due to the temperature non-linearity in the equation of state (Pierce, Barnett, and Mikolajewicz 1995; Cai and Chu 1997).

Finally, diffusive 'flushes' are another type of internal thermohaline variability (reviewed e.g. by Weaver 1995), but their long time scales are beyond the scope of this paper.

To understand a particular internal oceanic oscillation, we thus have to ask which is the essential feedback loop which causes the oscillation and sets its time scale, and which feedbacks act to dampen or maintain the oscillation. The basic ingredients of internal decadal oscillations will be temperature and salinity anomalies, advection, convection and wave propagation, and in some cases interaction with sea ice (Yang and Neelin 1993).

#### 15.4.3 Forced variability

As forced variability we consider such variations in thermohaline circulation which are directly forced by the atmosphere or from the land. For example, if there are internal decadal variations in atmospheric circulation, these could force decadal variations in thermohaline circulation. The ocean would simply be a slave to periodic variations in surface conditions.

An interesting case of forced variability are state transitions in the thermohaline circulation. Due to the advective and convective feedbacks discussed earlier, the thermohaline circulation may have several different equilibrium states, and changes in surface conditions may lead to a highly non-linear response in the ocean, such as a shutdown of NADW formation or a shift in convection sites (Rahmstorf 1995a). An example is the spectacular collapse of NADW flow during the so-called Heinrich Events (MacAyeal 1993; Sarnthein et al. 1994) mentioned above. Simulations of anthropogenic climate change in the next century show a

reduction or in some cases complete cessation of NADW formation (Cubasch et al. 1992; Manabe and Stouffer 1994).

Both types of feedback - advective and convective - have their own critical thresholds and time scales for state transitions (Rahmstorf 1995b; Rahmstorf 1995a; Rahmstorf 1996a; Rahmstorf, Marotzke, and Willebrand 1996). The advective spindown mechanism has a time scale of several centuries; its critical threshold is Stommel's bifurcation shown in Fig. 15.3. Convective transitions are much faster and can occur within a few years; they are also more complex and difficult to predict as convection is a very localized process depending on regional details, which are not usually well represented in models. Convective transitions include regional shifts in convection sites, but also the so-called "polar halocline catastrophe" (Bryan 1986), i.e. a rapid and total shutdown of convection in the high latitudes of the North Atlantic.

#### 15.4.4 Coupled modes

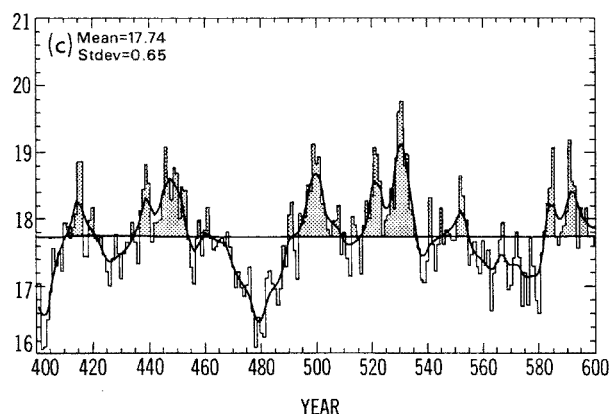
There is no universally agreed definition of what constitutes a coupled ocean-atmosphere mode. It is clear that oceanic variability arising in an ocean model driven by fixed surface fluxes is not a coupled mode. Likewise, stochastic surface forcing does not make a mode coupled, as it does not respond to oceanic changes. Variability of an atmosphere model driven by prescribed SST at its lower boundary could also not be called coupled. The equivalent boundary condition for the ocean is to prescribe fixed atmospheric conditions (temperature and other variables such as wind speed), which leads to a restoring boundary condition (Haney 1971; Rahmstorf and Willebrand 1995); ocean variability arising under such a completely unchanging atmosphere could hardly be called a coupled mode, even though surface fluxes are not fixed. We will therefore consider an oceanic mode uncoupled if it arises either with prescribed fluxes or under a restoring boundary condition (i.e. a purely local damping of anomalies), or a combination of the two (mixed boundary conditions). This is in line with common terminology calling a model driven by any of these boundary conditions an ocean-only model, not a coupled model. We will speak of coupled modes only if a *non-local feedback* via the atmosphere is important in generating the variability, as in simple ENSO models (Cane and Zebiak 1985) for wind feedback or in diffusive or advective energy balance models (Power et al. 1995; Rahmstorf and Willebrand 1995), which simulate atmospheric heat transport changes. Conversely, with this definition atmospheric models with a local swamp or slab ocean boundary are uncoupled atmosphere-only models, and variability arising in these models is internal atmospheric variability.

The classic example of a coupled mode is ENSO, where changes in atmospheric winds trigger a dynamical response of the oceanic mixed layer and equatorial waves, which ultimately lead to a delayed negative feedback on the initial anomaly (Suarez and Schopf 1988; Enfield 1989; Cane 1992). Another coupled mode has recently been suggested for the mid-latitude North Pacific (Latif and Barnett 1994). This involves the wind-driven gyre circulation. The best place to look for a coupled mode involving the thermohaline circulation is the North Atlantic; as early as 1964

Bjerknes proposed that variations in the ocean's meridional heat transport could explain decadal climate variations there. This hypothesis is finding increasing support from both data analysis and models (Bryan and Stouffer 1991; Kushnir 1994; Hall and Manabe 1997). The link between the North Atlantic Oscillation (NAO) pattern in the atmosphere and the ocean circulation has become an important research topic in recent years. Decadal temperature anomalies can be followed along the North Atlantic Current in step with the NAO (McCartney, Curry, and Bezdek 1996). Likewise, the latitude of the Gulf Stream correlates with the NAO index (Taylor 1996). Wohllleben and Weaver (1995) have analysed observed data and proposed a coupled subpolar North Atlantic mode which involves Labrador Sea convection. They suggest the following feedback loop: high Labrador SST → high Greenland sea level pressure → transport of ice/freshwater through Fram Strait and Denmark Strait into Labrador Sea → reduced convection → cooling of Labrador Sea. Given the delays in the system this leads to a 20-year oscillation. While the existence of a coupled decadal climate mode involving the thermohaline circulation is thus not yet firmly established, there are some interesting results pointing in this direction. In the remainder of this paper we will discuss decadal variability of the thermohaline circulation as it is found in models.

## 15.5 Models with realistic geography

The first interdecadal oscillation of the thermohaline circulation in an ocean-atmosphere GCM was described by Delworth, Manabe and Stouffer (1993, hereafter DMS93) in the GFDL coupled model. A time series of the maximum of the meridional transport stream function in the Atlantic is reproduced in Fig. 15.4. Spectral analysis of this oscillation revealed a fairly broad peak centered at a period of ca. 50 years. DMS93 point out similarities between the associated SST pattern and the observed interdecadal SST variability in the Atlantic (Kushnir 1994). The current and dynamic height anomaly pattern associated with this oscillation is shown in Fig. 15.5. The oscillation is centered on the regions to the east and south of Newfoundland in a dipole pattern. The mechanism of the oscillation was described essentially as a delayed negative feedback between oceanic advection and density. The feedback loop has an additional twist, though, involving a horizontal gyre circulation in quadrature with the vertical overturning. Consider the weak phase of the thermohaline flow: northward heat transport in the Atlantic is reduced, and a pool of cold dense water accumulates in the central North Atlantic. This then causes an anomalous cyclonic flow which transports salty water into the deep water formation region of the model, enhancing the thermohaline circulation and starting the warm phase of the oscillation. The atmosphere appears to play little role in this oscillation, except perhaps by exciting it and making it irregular due to its stochastic buoyancy forcing (DMS93). Subsequent work has linked this variability with salinity and temperature variability in the Greenland Sea and fluctuations in the East Greenland current (Delworth, Manabe, and Stouffer 1997). These

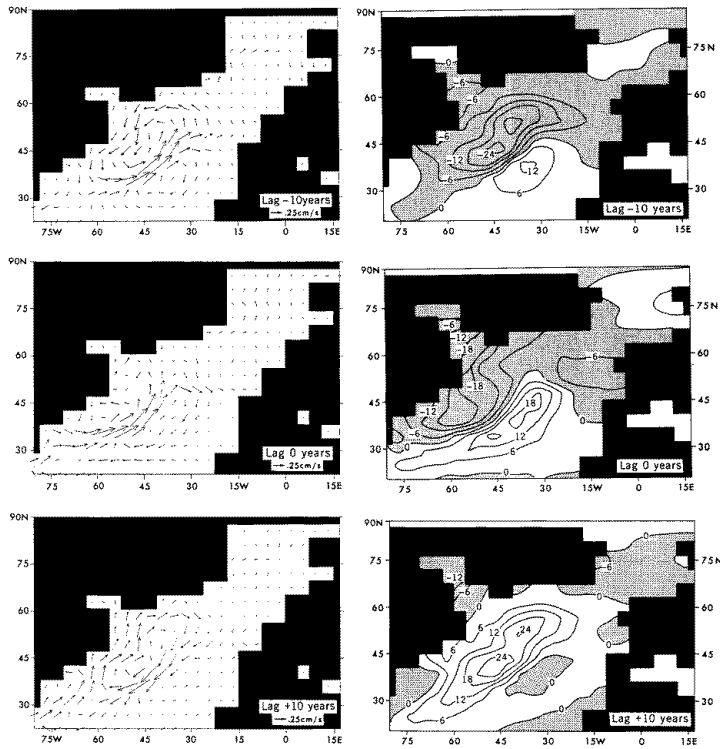


**Fig. 15.4** Time series of 200 years of the Atlantic meridional overturning in the coupled model of Delworth, Manabe, and Stouffer 1993.

results have spawned a number of further studies attempting to reproduce the oscillations of the coupled model with simplified models, some of which will be discussed below. Griffies and Tziperman (1995) have proposed a linear damped oscillator in the form of a simple box model which reproduces aspects of the DMS93 oscillation, in particular the phase relations between salinity, temperature and flow anomalies.

A thousand-year integration of the Hamburg ECHAM1/LSG coupled model has also revealed thermohaline oscillations (von Storch et al. 1997). The first EOF of the variability of the Atlantic meridional overturning stream function is shown in Fig. 15.6 (lower panel). It shows a pattern of anomalous downwelling north of  $25^{\circ}\text{N}$  accompanied by anomalous upwelling south of this latitude, with little connection to the outflow across  $30^{\circ}\text{S}$ . This pattern is different from the mean flow pattern, so it is not a loop oscillation with the whole Atlantic “conveyor belt” accelerating and decelerating, but rather a more localized advective mode. Consistent with this is the period of the oscillation of broadly 50 years. The overturning pattern associated with the mode of DMS93 is strikingly similar (Fig. 15.6, upper panel). A uniform upwelling anomaly of 1 Sv across the Atlantic between the equator and  $25^{\circ}\text{N}$  would produce a vertical excursion of isopycnals by about 15 m if maintained for 10 years, so that isopycnal movements associated with decadal variability would be quite small unless the upwelling was highly localized.

In a periodically coupled 700-year simulation of the ECHAM3/LSG model (a successor of the above), Timmermann et al. (1998) found a 35-year oscillation of the Atlantic thermohaline circulation. It resembles the DSM93 mode in the ocean component, but in contrast to DSM93 it was identified as a coupled ocean-atmosphere mode in which the atmospheric response to SST anomalies provides the negative feedback on the thermohaline overturning. When the thermohaline circu-



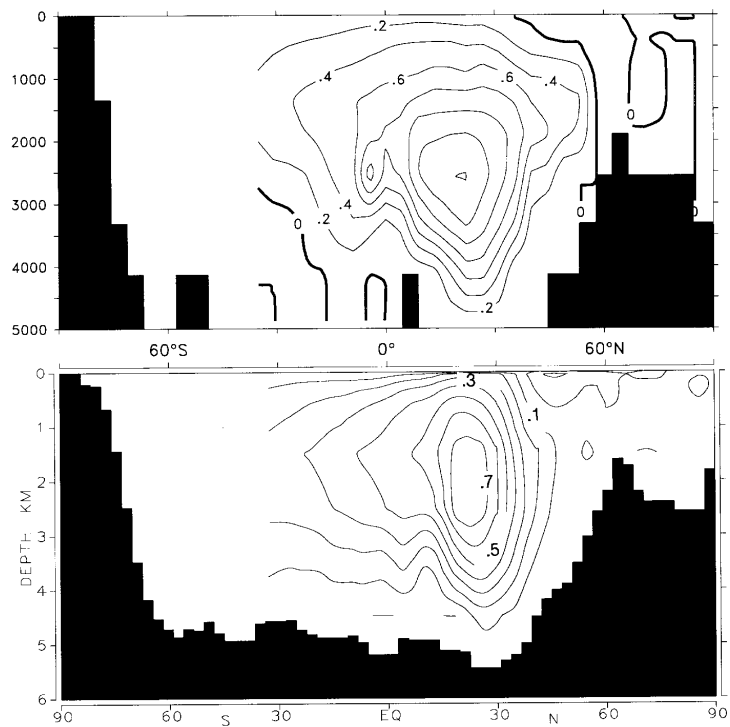
**Fig. 15.5** Regression between the Atlantic overturning intensity (see Fig. 15.4) and the currents at 170 m depth (left panel), and the dynamic topography relative to 915 dbar (in  $10^2 \text{ cm}^2 \text{ s}^{-2}$ , right panel). (From Delworth, Manabe, and Stouffer 1993.)

lation is strong, a warm SST anomaly develops in the subpolar North Atlantic which leads to a high NAO index in the atmosphere. This in turn leads to a positive surface freshwater input, lowering surface salinity and density and weakening the thermohaline circulation.

In the coupled CSM model of NCAR, Capotondi and Holland (1998) have found a rich decadal climate variability. This model is distinguished by a higher resolution than the above models (T42 atmosphere with 18 levels,  $2.4^\circ$  ocean resolution with 45 levels) and was consequently integrated for only 300 years. The model uses no flux adjustments. A 50-year mode of the meridional overturning stands out, which is similar to a mode found in an ocean-only version of the model. The authors therefore interpret the mode as a stochastically forced internal ocean mode, similar to DSM93.

Decadal oscillations of the thermohaline circulation were also found in Rahmstorf's (1995a) hybrid coupled model, using an ocean model configuration





**Fig. 15.6** First EOF (empirical orthogonal function) of the meridional mass transport stream function variability in the Atlantic for the two coupled model experiments of Delworth, Manabe, and Stouffer 1993 (top) and von Storch et al. 1997 (bottom).

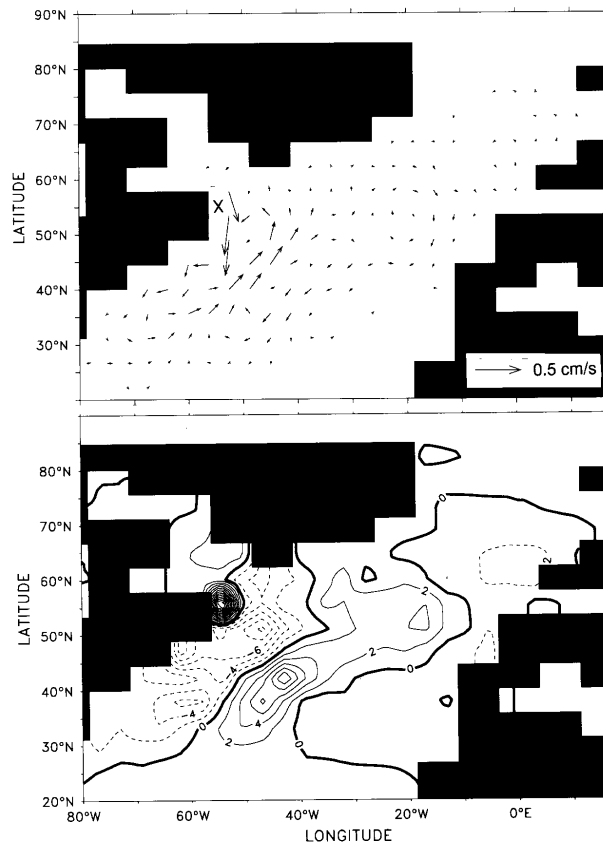
similar to that of DMS93 (the main difference being the use of isopycnal diffusion in the latter), combined with a simple energy balance model and fixed freshwater fluxes. This model produces a steady circulation when spun up for present-day climatological conditions, but a minor increase in precipitation in the high-latitude North Atlantic leads to a transition to an oscillatory regime through a Hopf bifurcation (without stochastic forcing). Only a brief discussion of this oscillation was included in Rahmstorf (1995a); results of a further analysis are summarized here. The oscillation shows many similarities to the one of DMS93, e.g. in the phase relations: sea surface salinity (averaged over the northern North Atlantic) leads the overturning peak by a year or two while SST lags it by a similar amount. For depth-averaged quantities, density leads (as in Figure 8 of DMS93) while salinity and temperature both lag (as in their Figure 9). Dynamic height and current anomalies (Fig. 15.7, compare to Fig. 15.5) show two distinct patterns: a large-scale flow pattern in the form of a band of northeastward flow across the interior Atlantic, very

similar to that of DMS93, and a strong anomaly at the convecting grid point off Newfoundland which is in opposite phase and dominates the vertical overturning response. The large-scale dipole pattern in dynamic topography is dominated by temperature anomalies just as in DMS93 (their Figures 18 and 19). We suggest that a non-linear self-sustained oscillation localized at the convection site off Newfoundland is what drives this oscillation; the moment that this convection is stopped (through a further prescribed increase in precipitation) the oscillation also ends. The dynamics of a single convection point next to a land boundary are unrealistic (as analysed in Rahmstorf 1995b); this oscillation probably behaves like a low-order box model involving a few grid cells (this idea is consistent with the clean Hopf bifurcation and the phase relationships). This local oscillator then forces the large-scale flow variations just as stochastic noise excites them in DMS93, and in a very similar pattern. In this interpretation, the large-scale oscillation would be a robust internal ocean mode which is essentially the same as in DMS93, except that it is excited by a local grid-cell instability instead of stochastic surface forcing.

It is a major advantage of hybrid coupled models that they can include specified climatic feedbacks without internal variability of the atmosphere; they can produce a 'pure' form of thermohaline oscillations (almost sinusoidal in Rahmstorf, 1995a) which does not need to be extracted from noise by statistical techniques and is thus easier to analyse. To make full use of this advantage, hybrid models should be constructed which exactly match coupled GCM's in their ocean component and their climatological surface fluxes, in order to emulate the variability found in the GCM. This would not only help to isolate the oscillation mechanism, but would also allow sensitivity studies, such as testing where stochastic noise is necessary to excite the oscillation, or how changes in surface climatology alter it. At present many modelers have used their rather different models to produce oscillations and then speculate how similar to DMS93 they are. A perhaps more productive approach would be to start with the very model of DMS93 and progressively 'strip it down' to establish the essence of the oscillation. A good step in this direction has been taken by Weaver and Valcke (1997), albeit using fixed surface fluxes rather than a feedback model. They showed that with fixed climatological surface fluxes taken from DMS93, with or without stochastic component, the oscillation of DMS93 cannot be reproduced.

Weisse et al. (1994) describe decadal variability in the North Atlantic in a stochastically forced experiment with the LSG model. They show that this mode is due to the formation of salinity anomalies in the Labrador Sea and their subsequent advection into the deep water formation sites of the model south of Greenland. The salinity anomalies are generated by Hasselmann's mechanism (see section 15.4.1) and arise because of the absence of convection in the Labrador Sea in this model. This mechanism is thus rather different from the advective feedback described in DMS93, and the spectrum is red rather than showing a decadal spectral peak.

The same model also shows a thermohaline oscillation with a period of 320 years, which at first sight looked similar to Welander's loop oscillation (Mikolajew-



**Fig. 15.7** Current anomalies at 170 m (top) and dynamic height anomalies (as in Fig. 15.5, bottom) at the time of minimum overturning in the decadal oscillation of Rahmstorf 1995a. The deep convection point off Newfoundland is marked by 'X'.

icz and Maier-Reimer 1990). Subsequent work (Mikolajewicz and Maier-Reimer 1994) showed that with a more realistic, weaker thermal damping (and thus a more effective negative temperature feedback) the oscillation disappears unless atmospheric noise is increased. Pierce et al. (1995) and Osborn (1997) have shown that this oscillation is caused by a convective flip-flop oscillator in the Southern Ocean; it may be a model artifact related to the spurious widespread convection which most models display in this region (an exception are models with sophisticated eddy mixing parameterisations, e.g. Danabasoglu, McWilliams, and Gent 1994). Pierce et al. (1996) have demonstrated that the oscillation can also occur in a hybrid coupled model (with atmospheric energy balance model), unless suppressed by a flux correction. Such century-scale overturning variability is however not the main theme of this review; for further discussion see e.g. Weaver (1995).

A common feature of the variability modes described in this section is that they are centered on the regions of convection and deep water formation. It is therefore of some concern that in these models these convection regions are not well resolved, nor are they necessarily in the right places (see Fig. 15.1). Typically the models form their NADW south-east of the tip of Greenland, in the kind of shortcut thermohaline circulation that probably existed during the last glacial maximum 21,000 years ago. The formation of lower NADW by convection in the Greenland-Norwegian Seas and subsequent overflow over the Greenland-Scotland sill is not properly represented. It is therefore not yet clear how realistic the variability patterns and feedback mechanisms in these coarse models can be, and to what extent they can be compared with observed regional patterns.

## 15.6 Idealized models

A rapidly growing number of idealized modelling studies has been performed to pin down the essential ingredients of thermohaline oscillations and to investigate their parameter sensitivity. Oscillations have been found in idealized ocean models (usually square basins spanning one or both hemispheres) driven by fixed fluxes, restoring boundary conditions, mixed boundary conditions or atmospheric energy balance models, both with three-dimensional or zonally averaged circulations. It is difficult to gain an overview over the various oscillation modes and to determine to what extent they share essentially the same mechanisms and where they differ. However, most oscillations appear to be due to a delayed negative advective feedback and have a multi-decadal period.

At first sight one might expect fixed surface fluxes to be most conducive to variability, as they do not damp any anomalies. However, when both salinity and temperature effects are considered, their effect on density (and thus flow) is often partly cancelling. A warm anomaly (e.g. due to advection from the subtropics) is often accompanied by high salinity. In these cases mixed boundary conditions are more prone to variability, as they damp the temperature anomaly but leave the salinity anomaly intact. For the effect on thermohaline overturning this means that the negative thermal feedback is suppressed, allowing the positive haline feedback to dominate.

Decadal or interdecadal oscillations under mixed boundary conditions have been found in many models (e.g. Weaver and Sarachik 1991; Weaver, Sarachik, and Marotzke 1991; Weaver et al. 1993; Chen and Ghil 1995; Yin and Sarachik 1995). These were mainly due to various advective mechanisms. The studies identified certain surface forcing conditions which are conducive to oscillations: strong fresh-water forcing (as compared to the thermal forcing) helps to destabilize the thermally direct thermohaline flow, and a minimum in P-E in mid-latitudes combined with high-latitude freshening supports decadal variability because it leads to a northward salt advection. Chen and Ghil (1996) obtained a 40-50 year oscillation in a hybrid model using an atmospheric energy balance model, with no salinity

included in the model. Greatbatch and Zhang (1995) describe a decadal oscillation driven only by constant surface heat flux. Again, the mechanism is due to a delayed negative advective feedback. High-latitude temperatures control density and overturning, and are in turn controlled by the intensity of advection from low latitudes. Several authors stress that surface fluxes derived from a spinup with restoring boundary conditions may be unique in giving the ocean the fluxes it needs for a steady solution, and thus tend to suppress variability (Cai 1995; Pierce, Barnett, and Mikolajewicz 1995). Even under a thermal restoring boundary condition (again without salt) oscillations can in some circumstances be found (Cai and Chu 1997). Capotondi and Holland (1997) performed a sensitivity study with respect to surface forcing, and found in their two-hemisphere square basin that decadal variability (period around 24 years) arose with weak thermal restoring, an energy balance model or fixed fluxes, but irregular 'flushes' were triggered when mixed boundary conditions with strong thermal restoring were used. They also investigated the role of stochastic forcing, and found essentially the same variability mode irrespective of the pattern of stochastic forcing, or whether the stochastic component was in the heat or freshwater fluxes. However, large scale stochastic forcing patterns were more effective in exciting the variability (leading to larger amplitude) than small scale forcing.

An interesting question is the potential role of waves in decadal variability. Greatbatch and Peterson (1996) propose that decadal oscillations can arise because the adjustment of the flow to density changes, occurring through Kelvin waves, is so slow that the north-south density gradient and overturning vary out of phase (overturning is in phase with east-west density gradient). The almost absent stratification at their convective polar model boundary slowed the Kelvin waves there enough to give a decadal oscillation, which thus would be determined by a wave propagation rather than advective time scale. A similar mechanism was invoked by Cai and Chu (1997). On the other hand, Huck, Colin de Verdière and Weaver (1997) performed a thorough sensitivity study of decadal variability in a square basin and found that it did not depend on boundary waves; the variability survived a removal of the boundaries (by extending the basin far away from the region of variability) except for the western boundary. They further confirmed Winton's (1996) earlier finding that the oscillations also occur in an  $f$ -plane model, eliminating Rossby waves from the possible mechanisms.

A weakness of idealized models is that they don't include topography, which appears to have an important influence on decadal thermohaline variability. Moore and Reason (1993) report oscillations in the North Atlantic in a global model associated with deep convection, which arose only with flat bottom. Likewise, Weaver et al. (1994) have found a convective-advective oscillation centered on periodic convection in the Labrador Sea in an Atlantic model, which was suppressed when topography was introduced into the model. Winton (1997) suggests that some of the variability found in flat-bottom models is caused by currents impinging on weakly stratified coasts (i.e. vertical walls), and shows that variability is suppressed when a bowl-shaped bottom topography is included. Another limitation of most

models is their coarse resolution; Fanning and Weaver (1997b) show that coarse resolution and high diffusion both tend to suppress variability.

Even two-dimensional models have shown advective decadal oscillations (Saranvanan and McWilliams 1995). Taking Welander's conceptual models again as guidance, we expect that a model will be more prone to advective oscillations if there is either a delayed response of the flow to density changes or different damping time scales for temperature and salinity perturbations. Indeed the model of Winton (1996) with an instantaneous diagnostic momentum balance and purely thermal forcing does not support decadal oscillations. However, such oscillations can be easily induced in a two-dimensional model if an artificial delay is added to the diagnostic momentum balance by restoring the velocities to their equilibrium values on a 10-year time scale (A. Ganopolski, pers. comm.).

## 15.7 Conclusions

Decadal variability of the thermohaline circulation probably exists. Both observational evidence and the prevalence of decadal oscillations in all types of ocean models point to this conclusion, and it would indeed be surprising if the complex turbulent flow in the ocean would not vary on decadal time scales (Wunsch 1992). There is as yet no consensus on the exact mechanism of this variability and models show a range of possibilities. The simplest mechanism, and the one which seems to play a role in most models, is an advective oscillation centered on a deep water formation site and the current feeding it. By necessity deep water formation sites are points of maximum surface density and water of lesser density is advected towards these regions near the surface. If deep water formation increases, more light water is advected into the convection site, tending to decrease deep water formation. This is a negative feedback. Add a delay in the response of the flow to the density change and the basic ingredients for an oscillation are there. This advective oscillator is captured in essence by the conceptual model of Huck, Colin de Verdière and Weaver (1997) or similar simple models and can be considered the null hypothesis for a decadal oscillation involving the ocean's deep circulation. It works for purely thermal forcing and does not require salinity effects. However there is also a different type of advective oscillation which depends on the different damping of temperature and salinity anomalies, but does not require a delay in the flow response. A non-linear, self-sustaining prototype has been described by Welander (1986), while Griffies and Tziperman (1995) have used a linear damped version. The oscillation of DMS93 shows the tell-tale phase relation between salinity and temperature anomalies (Griffies and Tziperman 1995), but also a 3-year lag in the flow response. A further indication that the fast damping of temperature (compared to salinity) anomalies may be important is the fact that the oscillation of DMS93 could not be reproduced with fixed fluxes (Weaver and Valcke 1997).

The details of the advective pathways, the interaction with the wind-driven circulation, the topography and the air-sea coupling differ between idealized and more

geographically explicit models, and therefore a detailed scenario that can be compared with observations can only come from the most realistic coupled models. An in-depth analysis and sensitivity study of a particular oscillation mode found in such a coupled model can probably best be achieved by fitting a simple atmospheric feedback model to the coupled model, with the same mean climate but controllable noise and feedback.

Some models suggest a possibility for other oscillation mechanisms which are fundamentally different from the simple advective oscillators described above, e.g. the convective oscillation of Weaver et al. (1994), where a time scale for cooling the Labrador Sea by convection is important in setting the period, or the Hasselmann mechanism in Weisse et al. (1994). An as yet controversial issue is the role of Kelvin waves in the oscillation mechanism (Greatbatch and Peterson 1996). Finally, there is also the possibility of a sudden shut-down (or start-up) of a particular convection site; this has been studied in idealized models (Lenderink and Haarsma 1994; Rahmstorf 1995b), a hybrid model (Rahmstorf 1995a) and was found to occur in a full coupled model (Tett, Johns, and Mitchell 1997). It is possible - indeed likely - that thermohaline variability in the real ocean may not be explained by one simple mechanism but by a combination of several factors.

Given the existence of thermohaline circulation oscillations, do they offer an opportunity for long-term prediction of climatic conditions over the Atlantic and Europe? Griffies and Bryan (1997) have analysed ensembles of coupled model integrations and found predictability up to ~20 years ahead, particularly during periods of strong circulation variability. The behavior of the model could be explained by damped harmonic oscillations of the thermohaline circulation excited by stochastic weather 'noise', with an oscillation period and damping time scale of around 40 years. These results suggest that there are real prospects for multi-year climate forecasts in the North Atlantic region. Modelling progress will probably continue apace in the next decade if computer resources continue to increase. A greater challenge will be to put a monitoring system in place to provide the necessary input data for forecast models.

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