Sensitivity of the Atlantic thermohaline circulation and its stability to basin-scale variations in vertical mixing

Willem P. Sijp* and Matthew H. England.

Centre for Environmental Modelling and Prediction, School of Mathematics, University of New South Wales, Sydney, New South Wales, Australia

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*Corresponding author address: Willem P. Sijp, School of Mathematics, University of New South Wales, Sydney, NSW 2052, Australia. E-mail: wsijp@maths.unsw.edu.au
We show that a reduction in vertical mixing applied inside the Atlantic basin can drastically increase North Antarctic Deep Water (NADW) stability with respect to fresh water perturbations applied to the North Atlantic. This is contrary to the notion that the stability of the ocean’s thermohaline circulation simply scales with vertical mixing rates. An Antarctic Intermediate Water (AAIW) reverse cell, reliant upon upwelling of cold AAIW into the Atlantic thermocline, is found to be associated with stable states where NADW is collapsed. Transitions between NADW “on” and “off” states are characterised by interhemispheric competition between this AAIW cell and the NADW cell. In contrast to the AAIW reverse cell, NADW eventually upwells outside the Atlantic basin, and is thus not as sensitive to changes in vertical mixing within the Atlantic. A reduction in vertical mixing in the Atlantic weakens the AAIW reverse cell, resulting in an enhanced stability of NADW formation. Our results also suggest that the AAIW reverse cell is responsible for the stability of NADW collapsed states, and thereby plays a key role in maintaining multiple equilibria in the climate system. A global increase of vertical mixing in our model results in significantly enhanced NADW stability, as found in previous studies. However, an enhancement of vertical mixing applied only inside the Atlantic Ocean results in a reduction of NADW stability. We conclude that the stability of NADW formation to freshwater perturbations depends critically on the basin-scale distribution of vertical mixing in the world’s oceans.
1. Introduction

Stommel (1961) first proposed the possibility that the ocean’s thermohaline circulation could have two stable regimes of flow. Bryan (1986) demonstrated how the operation of a positive salt feedback could bring about two asymmetric overturning states under symmetric surface flux conditions in a rectangular basin geometry. In his idealised model, there is competition for interhemispheric fresh water (FW) export between two overturning cells involving the sinking of water originating at the antipodean polar surface of the basin. The real ocean, however, contains an obstruction to southward geostrophic flow across the latitudes of Drake Passage, thus inhibiting the formation of the vigorous Southern Hemisphere (SH) cell observed by Bryan (1986). Instead, studies employing a more realistic geometry (e.g. Manabe and Stouffer 1988, 1999; Saenko et al. 2003; Gregory et al. 2003; Rahmstorf 1996; Sijp and England 2005) find that SH overturning states exhibit a shallower Antarctic Intermediate Water (AAIW) reverse cell inside the Atlantic basin. Manabe and Stouffer (1999, hereafter MS99) point out that this small reverse cell is associated with a stable NADW “off” state. In contrast to the SH cell of Bryan (1986) and other studies employing a Drake Passage closed geometry (e.g. Mikolajewicz et al. 1993; Sijp and England 2004), this AAIW reverse cell originates at the AAIW formation regions. Saenko et al. (2003) use a model to suggest that due to the existence of this cell in a North Atlantic Deep Water (NADW) “off” state, the density difference between surface waters at the formation regions for AAIW and NADW determines the strength and polarity of the meridional overturning circulation (MOC), at least in their model. Sijp and England (2005) show that the nature of this relationship depends on
the depth of the Drake Passage sill, with Antarctic Bottom Water (AABW) playing a reduced role as the DP deepens. Furthermore, Saenko et al. (2003) and Gregory et al. (2003) suggest that this AAIW reverse cell can contribute to maintaining a stable NADW “off” state by importing FW into the Atlantic basin, and that it cannot co-exist with the NADW formation cell.

In ocean models the upwelling branches of the NADW cell and the AAIW reverse cell rely in part on vertical mixing rates at lower latitudes. This upwelling into the thermocline consists of a balance between downward diffusion of heat from the surface and upward advection of colder water from below the thermocline (e.g. Munk 1966; Munk and Wunsch 1998), the balance determining the global thermocline depth. Enhanced vertical mixing therefore increases transport in the upwelling branches of these cells. Indeed, Bryan (1987) found that larger vertical mixing rates lead to increased MOC in an idealised sector geometry model. There have been several other studies that investigate the role of vertical mixing in setting the thermohaline circulation (THC) strength in global ocean climate models. MS99 showed that the stability of their NADW “off” state depends on the value of vertical mixing. They also find enhanced NADW formation and greater NADW stability with respect to FW perturbations under a global increase in vertical mixing. Studies employing a zonally averaged model (Schmittner and Weaver 2001) and a three dimensional ocean model (Prange et al. 2003) use hysteresis experiments to show that the stability of the NADW “off” and “on” states depend on the rate of vertical mixing. For instance, in agreement with MS99, Schmittner and Weaver (2001) find enhanced NADW stability in their hysteresis experiments when vertical mixing is increased globally. These studies, however, employ horizontally uniform
vertical mixing modifications. In the present study, we will assess the stability of the North Atlantic MOC with respect to basin-wide changes in the vertical mixing coefficient ($K_v$).

The AAIW reverse cell of the NADW “off” state relies on upwelling inside the Atlantic basin, whereas in the NADW “on” state, NADW outflow leaves the Atlantic basin at 30°S to upwell predominantly outside this basin. Broecker (1991) highlights the role of the Indian and Pacific Oceans in the removal of NADW by vertical mixing at low latitudes in the classical “global ocean conveyor belt” schematic. Later studies (e.g. Toggweiler and Samuels 1995; Döös 1995) suggest instead that a significant portion of NADW resurfaces via wind-driven upwelling in the Southern Ocean, undergoing subsequent buoyancy changes due to surface fluxes. Model studies employing weak background vertical mixing also support this idea (e.g. Saenko and Merryfield 2005). Another example of how a reasonable overturning and a thermocline of realistic depth can be achieved without the need for strong vertical diffusivity is described in the theory of the ventilated thermocline (Luyten et al. 1983).

As the upwelling branches of the NADW cell and the AAIW reverse cell occur at different locations, the spatial distribution of vertical mixing in the world ocean will likely affect the relative potential of these cells for dominance of deep ocean ventilation. In particular, the Atlantic AAIW reverse cell relies upon the removal of intermediate water across the Atlantic thermocline via diapycnal mixing. Hence, a reduction in vertical mixing at low latitudes of the Atlantic should reduce the strength of the AAIW reverse cell and therefore its ability to compete with NADW. The NADW cell, on the other hand, does not rely on Atlantic upwelling and may therefore be less sensitive to a reduction of vertical mixing inside the
Atlantic. In this situation, a greater robustness of NADW overturning with respect to FW perturbations may ensue. In this study, we use an intermediate complexity coupled climate model to examine the stability of NADW formation with respect to FW perturbations under different spatial distributions of vertical mixing.

The remainder of this paper is divided as follows. Section 2 covers a description of the model and experimental design. We will consider experiments where $K_v$ is applied only inside the Atlantic Ocean to examine the role of the AAIW reverse cell in suppressing NADW formation. We will also assess experiments wherein a reduced $K_v$ is applied only inside the Indian and Pacific Oceans. To compare our findings to previous studies, we further present an experiment similar to MS99 where $K_v$ is increased globally, and an experiment where $K_v$ is increased by a similar magnitude, but only inside the Atlantic. In section 3 we first discuss the steady state fields under the different vertical mixing scenarios, and then describe their response to the application of external FW pulses of varying magnitude. Finally, section 4 covers a discussion and the conclusions.

2. Model and Numerical Experiments

We use the University of Victoria Intermediate Complexity Coupled Model described in detail in Weaver et al. (2001), which comprises a global ocean general circulation model (GFDL MOM Version 2.2, Pacanowski 1995) coupled to a simplified atmospheric model and a dynamic-thermodynamic sea-ice model. A global domain is used with horizontal res-
olution 3.6° longitude by 1.8° latitude in each model component. While air-sea heat and freshwater fluxes evolve freely in the model, a non-interactive wind field is employed. FW fluxes between the ocean and the atmosphere are determined by evaporation and precipitation, while river run-off and changes in sea-ice volume also affect oceanic salinity. Moisture transport in the atmosphere occurs by way of advection and diffusion. Precipitation occurs when relative humidity exceeds a threshold value. The wind forcing is taken from the NCEP/NCAR reanalysis fields (Kalnay et al. 1996), averaged over the period 1958-1997 to form a seasonal cycle from the monthly fields. Brine rejection during sea-ice formation is parametrised after Duffy and Caldeira (1997). The atmospheric model advects and diffuses moisture, and it diffuses heat. No flux corrections are used (for further details, see Weaver et al. 2001). Vertical mixing in the control experiment is achieved using a horizontally uniform diffusivity that increases with depth, taking a value of $0.6 \text{ cm}^2/\text{s}$ at the surface and increasing to $1.6 \text{ cm}^2/\text{s}$ at the bottom. The effect of subgrid scale eddies on horizontal tracer transport is modelled by globally uniform horizontal diffusion. The diffusion coefficient in both horizontal directions is $A_h = 2 \times 10^7 \text{ cm}^2/\text{s}$. The constant lateral mixing coefficient for momentum is $2 \times 10^9 \text{ cm}^2/\text{s}$, and the constant vertical mixing coefficient for momentum is $10 \text{ cm}^2/\text{s}$. For economy of computation, with many multi-millennial integrations, no parametrisation of along-isopycnal diffusion (Redi 1982) or eddy-induced advective tracer transport (Gent and McWilliams 1990) is used in the standard set of experiments. Rather, our focus is on model THC sensitivity to vertical diffusive mixing. It is noted, however, that we have re-evaluated several of the experiments under GM and confirm that our results are robust in both GM and non-GM experiments.
To examine the effect of reducing $K_v$ in different ocean basins, we have integrated the following configurations of the model to equilibrium: (1) a control experiment CNTRL, where no reduction in $K_v$ is applied, (2) an experiment denoted $\frac{2}{3}K_V\text{Atl}$, where $K_v$ takes a value reduced to $2/3$ of CNTRL inside the Atlantic basin, (3) an experiment denoted $\frac{1}{3}K_V\text{Atl}$ where $K_v$ is multiplied inside the Atlantic basin by a factor $1/3$ and (4) an experiment where $K_v$ is similarly reduced inside the Pacific and Indian basins, denoted $\frac{1}{3}K_V\text{IP}$. Figure 2 (a) shows the two areas where reduction of $K_v$ is applied in the respective experiments. The reduction is only applied between 35°S and 48°N in each ocean basin. The above reductions in $K_v$ are also only applied between the surface and 1257 m depth, with no change in $K_v$ at levels deeper than 1257 m (Fig. 1). A version of $\frac{1}{3}K_V\text{Atl}$ using GM has been run to verify the robustness of our results with respect to the choice of parametrization of subgrid-scale eddy mixing of tracer properties.

To further examine whether our model exhibits an increase in NADW stability upon increasing $K_v$ globally at a depth range containing the pycnocline, as found by MS99, we have integrated an experiment $MS99K_V\text{glob}$ similar to their so-called “LVD” experiment where $K_v$ takes an increased global value of $1.3 \text{ cm}^2/\text{s}$ down to 2500 m depth (see Fig. 1). Below 2500 m depth $K_v$ follows the profile of CNTRL. The MS99 “LVD” experiment employs constant $K_v = 1.3 \text{ cm}^2/\text{s}$, however such a profile would yield a decrease in $K_v$ at depth compared to CNTRL. Hence, the profile shown in Fig. 1 is adopted for direct comparison with the CNTRL experiment. To elucidate the role of vertical mixing inside the Atlantic in the context of the experiments of MS99, we have also integrated an experiment $MS99K_V\text{Atl}$ where $K_v$ is changed in a similar fashion to $MS99K_V\text{glob}$, but only inside the Atlantic. The above
experiments should yield similar results to experiments using the actual MS99 profile with $K_v = 1.3 \text{ cm}^2/s$ throughout the water column, as diapycnal mixing inside the pycnocline, rather than abyssal mixing, is the main factor affecting the strength of NADW formation. To confirm this, we have run experiments using a value of $1.3 \text{ cm}^2/s \ K_v$ throughout the water column, as in MS99, and found results that are very similar to those of $MS99KV_{glob}$ and $MS99KV_{Atl}$. Hence, only the one set of experiments with increased $K_v$ are reported here.

We further subject the obtained equilibria for CNTRL, $\frac{2}{3}KV_{Atl}$, $\frac{1}{3}KV_{Atl}$, $\frac{1}{3}KV_{IP}$, $MS99KV_{glob}$ and $MS99KV_{Atl}$ to FW pulses applied to the NA for the duration of 300 years. We vary the maximum magnitude of this pulse. In short, we vary both $K_v$ and the magnitude of the FW pulse in this study, and all other model parameters remain fixed. The FW perturbations are applied in the NA between the black lines shown in Figure 2 (a). Figure 2 (b) shows the FW perturbation against time. A linear increase from 0 to a maximum value $M$ occurs over the first 150 years, followed by a linear decline back to 0 over the following 150 years. After year 300, no further FW perturbation is applied. Note that we apply this perturbation for several values of $M$. This procedure, similar to that of Sijp and England (2005), is designed to examine the existence of multiple equilibria in the respective experiments.

The ocean exhibits substantial regional variations in $K_v$ (e.g. Ledwell et al. 2000), thought to be a function of tidal currents, subsurface bathymetry and local stratification. Several studies have examined the effects of locally enhanced vertical diffusivity over rough bathymetry in ocean models (e.g. Hasumi and Suginohara 1999; Saenko and Merryfield 2005). Unlike these previous studies, we make no attempt to examine the effects of contrasts in vertical
mixing that may exist between basins in the real ocean. Instead our experiments are specifically aimed to examine the role of the location of deepwater removal in determining the global THC and its stability.

3. Results

a. Equilibrated states

In this section we first analyse the steady states of the experiments. After discussing these equilibria we will present a series of perturbations to these steady states where the maximum value of the perturbation varies (section 3 b). Figure 3 shows the MOC in the Atlantic basin for the NADW “on” states and NADW “off” states of experiments CNTRL and $\frac{2}{3}K_{V_{Atl}}$. Also shown are the steady NADW “on” state of $\frac{1}{3}K_{V_{Atl}}$ and the only steady state found for $\frac{1}{3}K_{V_{IP}}$. A summary of the MOC transport rates for all experiments is given in Table 1. The NADW “on” state in CNTRL exhibits 21.5 Sv of NADW formation ($1\text{ Sv} = 10^6 \text{ m}^3 \text{ sec}^{-1}$; throughout this paper all transport values are quoted to the nearest 0.1 Sv). The CNTRL NADW “off” state was obtained from the NADW “on” state by applying a perturbation of maximum value $M=0.11\text{ Sv}$ (0.38 m/yr). This NADW “off” state exhibits an AAIW reverse cell of 9.3 Sv. This cell overlies an Antarctic Bottom Water (AABW) cell recirculating below 2000m depth with an AABW inflow into the Atlantic sector of 4.5 Sv.

The experiment $\frac{2}{3}K_{V_{Atl}}$ admits a NADW “on” state, as shown in Fig. 3 c, although with a
reduction of NADW recirculation within the Atlantic and an increase in NADW outflow. We obtain the NADW “off” state in $\frac{2}{3}K_{Atl}$ from the NADW “on” state in a similar fashion to the CNTRL experiment. In the NADW “off” state of experiment $\frac{2}{3}K_{Atl}$, the AAIW reverse cell is reduced by 3 Sv, taking a value of 6.3 Sv, and is restricted to shallower depths (Fig. 3d). This reduction in the strength and depth of the AAIW reverse cell results from a decrease in AAIW upwelling into the Atlantic thermocline due to lower vertical mixing rates there. This result illustrates the dependence of the AAIW reverse cell on vertical mixing inside the Atlantic basin. In contrast, the AABW recirculation in $\frac{2}{3}K_{Atl}$ is similar (4.3 Sv) to that of the NADW “off” state in CNTRL (4.5 Sv). This could be because AABW inflow into the Atlantic returns south at depth, and is not affected by changes in vertical mixing applied in the upper 1257 m, and changes in NADW formation are only small. The NADW “on” state in $\frac{1}{3}K_{Atl}$ exhibits a reduction of NADW formation from 21.5 Sv in CNTRL to 19.8 Sv. It is interesting to note that despite a reduction in NADW formation, NADW outflow \(^1\) is increased from 12.2 Sv in CNTRL to 14.8 Sv in $\frac{1}{3}K_{Atl}$. This implies a reduction of NADW recirculation inside the Atlantic \(^2\) from 9.3 Sv in the control experiment to 5.0 Sv when vertical mixing is reduced in $\frac{1}{3}K_{Atl}$. Thus, as expected, the rate of NADW recirculation inside the Atlantic depends on the rate of localised vertical mixing. A reduction of vertical mixing in the Atlantic therefore reduces this recirculation, which is compensated in part by reduced NADW formation, and in part by an increase in NADW outflow.

\(^1\)NADW outflow is measured as the local maximum at intermediate depth of the meridional transport streamfunction at 30 °S.

\(^2\)We calculate this recirculation by subtracting the NADW outflow rate from the NADW formation rate.
NADW formation is absent in the steady state MOC in \( \frac{1}{3}KV_{IP} \), despite being set up identical to CNTRL apart from a reduction of \( K_v \) in the tropical Indian and Pacific Oceans. This suggests that the maintenance of a NADW overturning circulation depends in part on vertical mixing rates in the ocean basins outside the Atlantic. For \( \frac{1}{3}KV_{IP} \), the Atlantic MOC is similar to that of the NADW “off” state of CNTRL (compare Fig. 3b and Fig. 3f), with an AAIW reverse cell of 10.2 Sv overlying an AABW cell recirculating below 2000m depth. This suggests that the Atlantic circulation of the NADW “off” state in CNTRL is not significantly influenced by vertical mixing outside the Atlantic basin. The similarity between the CNTRL NADW “off” state and the unperturbed \( \frac{1}{3}KV_{IP} \) experiment arises because the AAIW reverse cell relies almost exclusively on vertical mixing inside the Atlantic for its upwelling branch.

In contrast to experiments CNTRL and \( \frac{2}{3}KV_{Atl} \), we were unable to obtain stable NADW “off” states in \( \frac{1}{3}KV_{Atl} \). This was attempted with additions of FW perturbations of up to \( M= 0.58 \) Sv. This shows a significant increase in NADW robustness with respect to FW perturbations in the NA when \( K_v \) is reduced in the Atlantic sector. This is because a strong reduction of \( K_v \) in the Atlantic weakens the AAIW reverse cell that is otherwise required to prevent the eventual re-establishment of the NADW cell after the forcing is removed. As we will see later, the reverse cell appears temporarily during our FW perturbation, but with reduced vigour in \( \frac{1}{3}KV_{Atl} \). Unlike \( \frac{1}{3}KV_{Atl} \), the existence of a stable NADW “off” state in the \( \frac{2}{3}KV_{Atl} \) experiment allows us to examine the effect of vertical mixing in the Atlantic on the AAIW reverse cell.

To test the relationship between the AAIW reverse cell and the value of \( K_v \) inside the Atl-
Atlantic, we have run several experiments similar to $\frac{1}{3}KV_{Atl}$, only using multiplicative factors of 0.75, 0.85 and 1.25. In all cases we obtain NADW “off” states using a FW perturbation. Figure 4 shows the strength of the AAIW reverse cell in the NADW “off” states against the multiplicative factor applied to $K_v$ for these experiments, along with $\frac{2}{3}KV_{Atl}$ and CNTRL. The solid line is a linear “best fit” using the method of least squares. The close fit of this line to our experimental results suggests that the strength of the AAIW reverse cell has a linear dependence on the rate of vertical mixing inside the Atlantic, as would arise if a simple advection-diffusion balance were in place. From this analysis we can assume that vertical mixing inside the Atlantic constitutes a limiting factor on the strength of the AAIW reverse cell.

MS99 found an increased stability of NADW when globally increasing $K_v$ to 1.3 $cm^2/s$ in their so-called LVD experiment. This result appears to be in contradiction to our finding that decreasing $K_v$ inside the Atlantic results in an increase in NADW stability. Therefore, we now examine the results of our experiment $MS99KV_{glob}$ where we set $K_v$ to 1.3 $cm^2/s$ in the upper 2500 m globally, similar in design to the LVD experiment of MS99. A second experiment $MS99KV_{Atl}$ is evaluated where we apply this increase of $K_v$ only inside the Atlantic. Note that our CNTRL experiment employs a value of 0.6 $cm^2/s$ at the top layer. Experiments $MS99KV_{glob}$ and $MS99KV_{Atl}$ therefore employ larger values within the global and Atlantic pycnocline, respectively, than our CNTRL experiment. Figure 5 shows the MOC in the Atlantic basin for (a) the NADW “on” state of $MS99KV_{Atl}$, (b) the NADW “off” state of $MS99KV_{Atl}$ and (c) the NADW “on” state of $MS99KV_{glob}$. Increasing $K_v$ only inside the Atlantic ($MS99KV_{Atl}$) results in a 2.2 Sv increase in NADW formation to
23.7 Sv and yet a significant reduction in NADW outflow (from 12.2 Sv down to 7.8 Sv) for the NADW “on” state (Table 1). Enhancing vertical mixing inside the Atlantic therefore results in a significant increase in NADW recirculation within the Atlantic at the expense of NADW outflow. AABW inflow remains at similar values to CNTRL. As expected, the NADW “off” state for $MS99KV_{Atl}$ (Figure 5 b) shows a significantly increased AAIW reverse cell, attaining a strength of 15.1 Sv. In agreement with the results shown in Fig. 4, this indicates that the strength of the AAIW reverse cell is set by vertical mixing inside the Atlantic. Enhancing $K_v$ only inside the Atlantic results in a significant increase in strength of the AAIW reverse cell due to the advective-diffusive balance that the upwelling branch of this cell relies on. Strong overturning (28.0 Sv) occurs in $MS99KV_{glob}$ (Figure 5 c), and the streamfunction in this experiment is very similar to that shown in Figure 9a of MS99. Finally, it is noted that in agreement with MS99, we have been unable to obtain a stable NADW “off” state for $MS99KV_{glob}$.

b. *Freshwater perturbations*

As stated in Section 2 we have subjected the equilibria obtained for all experiments to 300 year FW pulses of different magnitude. We also apply FW pulses to experiments with $K_v$ multiplied by factors of 1/4 and 1/2 in the Atlantic (denoted $\frac{1}{4}KV_{Atl}$ and $\frac{1}{2}KV_{Atl}$ respectively). Figures 6-9 summarize the results. Figure 6 shows the timeseries of NADW formation and the AAIW reverse cell for experiments $\frac{1}{4}KV_{Atl}$, $\frac{1}{3}KV_{Atl}$, $\frac{1}{2}KV_{Atl}$ and $\frac{2}{3}KV_{Atl}$ under a FW perturbation attaining a maximum value M of 0.29 Sv (1.02 m/yr). In other words, in
this set of experiments, the maximum value $M$ of the FW pulse is fixed, and $K_v$ varies. The FW pulse is sufficient to shut down NADW formation in $\frac{2}{3}K_{V_{Atl}}$, whereas in $\frac{1}{4}K_{V_{Atl}}$ and $\frac{1}{4}K_{V_{Atl}}$ a recovery of NADW formation occurs after the pulse has ceased. In $\frac{1}{2}K_{V_{Atl}}$ we see a slower recovery of NADW formation and an associated decline of the AAIW reverse cell after the FW pulse has ceased. This set of experiments shows that decreasing $K_v$ in the Atlantic results in an increased robustness of NADW formation with respect to a given FW pulse.

Figure 7 shows timeseries of NADW formation and the AAIW reverse cell in $\frac{1}{3}K_{V_{Atl}}$ in response to the application of a series of 300 year FW pulses of varying magnitude $M$. Thus, in this analysis, we have kept the reduction of $K_v$ fixed while varying the magnitude $M$ of the FW flux. For comparison, we have included the response of CNTRL to a smaller FW pulse. A perturbation of maximum value $M = 0.11$ Sv (0.38 m/yr) yields a stable NADW “off” state in the control experiment. In contrast, when FW pulses attaining higher maximum values of $M$ up to 0.43 Sv (1.53 m/yr) are applied to $\frac{1}{3}K_{V_{Atl}}$, NADW formation eventually recovers (Figure 7a). The NADW shutdown in CNTRL coincides with the emergence of the AAIW reverse cell (Fig. 7b), which eventually attains a value of 9.3 Sv. When $K_v$ is reduced in the Atlantic, however, the role of the AAIW reverse cell appears to be diminished. Despite the initial suppression of the NADW cell in response to the FW perturbations applied to $\frac{1}{3}K_{V_{Atl}}$, and the temporary establishment of a higher maximum strength AAIW reverse cell in one of the experiments, there is no permanent sustenance of this cell. After the perturbation has terminated NADW formation recovers over a period of 1500-2500 years. This indicates that in the absence of an external FW flux, circulation changes alone cannot maintain the light
surface conditions at the NADW formation regions required to prevent sinking when $K_v$ is reduced by a factor of 1/3 in the Atlantic. Indeed, Gregory et al. (2003) suggest that the AAIW reverse cell is responsible for importing FW into the Atlantic, maintaining buoyant NA surface conditions that prevent a transition to a NADW “on” state. In contrast to the AAIW reverse cell, NADW finds its upwelling areas outside the Atlantic basin, and is thus not subject to the same impediment as the AAIW reverse cell when $K_v$ is reduced in $\frac{1}{3}K_{V_{At}}$. This indicates that a reduction of $K_v$ inside the Atlantic basin favours stability of the NADW cell by inhibiting the AAIW reverse cell.

The enhanced stability of the NADW cell in $\frac{1}{3}K_{V_{At}}$ is not the result of changes in surface density or isotherm depth. Firstly, the surface conditions in $\frac{1}{3}K_{V_{At}}$ indicate a cooling and freshening of the NADW formation regions relative to CNTRL (figure not shown), with the net effect of this change a reduction of surface density in the North Atlantic. This proves that changes in NA surface density in $\frac{1}{3}K_{V_{At}}$ are not responsible for the increased stability of NADW. Second, as expected, we observe a shoaling of the isotherms in the upper 2000 m of the Atlantic in $\frac{1}{3}K_{V_{At}}$ relative to CNTRL (figure not shown). A deepening of Atlantic isotherms is known to enhance NADW formation (Gnanadesikan 1999) and perhaps its stability, yet we find shoaling of the Atlantic isotherms in the upper 2000 m in $\frac{1}{3}K_{V_{At}}$. Moreover, the slightly reduced rate of NADW formation in $\frac{1}{3}K_{V_{At}}$ (19.8 Sv) might simplistically suggest reduced NADW stability, the opposite to what we have found. We have shown in contrast that a reduction in vertical mixing inside the Atlantic leads to an increased stability of NADW formation. This results from the weakening of the AAIW reverse cell caused by the reduced vertical mixing in the Atlantic sector.
A very different picture emerges when $K_v$ is reduced in the Indian and Pacific Oceans. We have tried to excite a transition to a NADW “on” state in $\frac{1}{3}K_VIP$ by applying FW extractions from the NA without success. For example, Figure 8 shows the timeseries for experiment $\frac{1}{3}K_VIP$ under a FW extraction attaining a maximum of 1.7 m/yr after 150 years. Time-series are shown for the NADW production rate and the AAIW reverse cell, clearly showing the inability of the $\frac{1}{3}K_VIP$ case to maintain stable NADW production despite a strong FW extraction. Apparently, in our model, the Indian and Pacific Oceans comprise an area of significant deepwater removal by vertical mixing. This sensitivity of NADW to vertical mixing inside the Indian and Pacific Oceans indicates that in our model, alternative removal mechanisms, such as mechanically driven upwelling due to wind stress in the Southern Ocean, are not sufficient alone to maintain the NADW cell in the absence of strong Indo-Pacific vertical mixing. A dye tracer experiment tagging NADW recirculation (figure not shown) shows upwelling of NADW in the lower latitudes of the Indian and Pacific Oceans, as well as at some locations in the high latitudes of the Southern Ocean. This confirms the importance of the Indian/ Pacific and Southern Oceans in the removal of NADW from the deep oceans in this model. In this case, a reduction of vertical mixing inside the Indian and Pacific Oceans shifts the competitive advantage of the AAIW/ NADW cells in favour of the AAIW cell. We have also run an experiment reducing $K_v$ globally by the factor 1/3, and again find weak overturning in the Atlantic basin. In this case, both the AAIW reverse cell and the NADW cell are suppressed.

Figure 9 shows a timeseries of the CNTRL experiment under a FW pulse attaining a maximum value $M$ of 0.1 Sv (0.34 m/yr). A recovery of NADW formation occurs under this
perturbation. As in the experiments of Gregory et al. (2003), the recovery of the NADW cell is simultaneous with the disappearance of the AAIW reverse cell. This is in contrast to the experiments $\frac{1}{4}KV_{Atl}$ and $\frac{1}{3}KV_{Atl}$ shown in Fig. 6, where NADW formation recovers some time after the AAIW reverse cell has disappeared. This suggests that reducing $K_v$ inside the Atlantic also removes the synchronicity between the re-establishment of the NADW cell and the disappearance of the AAIW reverse cell in ocean models.

c. Importance of location of change in $K_v$.

Figure 10 shows timeseries of NADW production rate and the AAIW reverse cell for experiment $MS99KV_{Atl}$ under FW perturbations attaining a maximum of 0.08 Sv and 0.11 Sv and for experiment $MS99KV_{glob}$ under FW perturbations attaining a maximum of 0.11 Sv and 0.76 Sv. In agreement with MS99 we find a significantly enhanced NADW stability in $MS99KV_{glob}$, where the 0.76 Sv perturbation is not sufficient to permanently shut down NADW formation. This should be compared to the much lower value of 0.11 Sv that is sufficient to excite a transition to a stable NADW “off” state in CNTRL (Figure 7). The NADW collapse in $MS99KV_{Atl}$ under a perturbation of only 0.08 Sv (Fig. 10), however, shows that increasing $K_v$ only inside the Atlantic results in a reduction of NADW stability compared to the enhanced stability found under a global increase in $K_v$. The reduction in NADW stability in $MS99KV_{Atl}$ arises from the significant increase in strength of the AAIW reverse cell in response to increased vertical mixing inside the Atlantic. This is consistent with our previous findings in experiments $\frac{1}{4}KV_{Atl}$ and $\frac{2}{3}KV_{Atl}$. In contrast, we could not obtain a stable “off”
state for $MS99 K_{glob}$. In summary, a markedly different response of the ocean’s MOC is obtained when altering $K_v$ globally as compared to $K_v$ changes limited to the Atlantic Ocean.

4. Summary and Conclusions

In this note we have demonstrated that regional variations in vertical mixing can fundamentally affect the global meridional overturning circulation and its stability to FW perturbations. An important reason for applying such changes in $K_v$ in our model is to demonstrate the role of the AAIW reverse cell in maintaining the stability of circulation states. Our results show that a reduction of $K_v$ inside the Atlantic basin can drastically increase NADW stability with respect to FW perturbations applied to the NA. Conversely, a reduction of $K_v$ inside the Indian and Pacific basins can inhibit NADW formation, enabling the establishment of an AAIW reverse cell driven by upwelling inside the Atlantic basin. Furthermore, the FW perturbations we apply are unable to excite transitions to a stable NADW “off” state for the cases where $K_v$ is multiplied by a factor of 1/2 or less in the Atlantic Ocean. Multiplication of $K_v$ in the Atlantic by a factor of 2/3, however, allows the model to admit a transition to a stable NADW “off” state in response to the FW perturbations. The reduction in strength and depth of the AAIW reverse cell in this NADW “off” state illustrates that the AAIW reverse cell is indeed inhibited by a reduction of $K_v$ in the Atlantic Ocean. It should be noted that we have rerun a version of $\frac{1}{2} K V_{Atl}$ using the computationally more intensive Gent and McWilliams (1990) eddy advection scheme combined with along-isopycnal mixing and no horizontal diffusion, and also found increased NADW stability in this experiment.
Because a reduction in vertical mixing in the Atlantic inhibits the AAIW reverse cell, it in turn enhances the stability of the NADW cell. Indeed, Saenko et al. (2003) and Gregory et al. (2003) suggest that the AAIW reverse cell is responsible for the stability of the NADW “off” state, which we have also demonstrated in this study. After an initial suppression of NADW due to freshening of the NADW formation regions by a FW pulse (Figure 6), no stable AAIW reverse cell develops when $K_v$ is reduced significantly in the Atlantic. The recovery of NADW formation in the absence of this cell in $\frac{1}{3}K V_{Atl}$ supports the idea that the AAIW reverse cell (associated with NADW “off” states in ocean models) is a contributing factor to the stability of collapsed NADW states. Furthermore, changes in surface conditions and pycnocline depth inside the Atlantic do not contribute to the enhanced NADW stability we find in $\frac{1}{3}K V_{Atl}$. Rather, it appears to be exclusively due to the suppression of the AAIW reverse cell in the Southern Hemisphere in this experiment.

Although detrimental to the AAIW reverse cell, reduced $K_v$ inside the Atlantic does not inhibit the NADW cell, as its deepwater removal branch is located elsewhere, in the Indian, Pacific and Southern Oceans. Our inability to obtain a stable NADW “on” state in experiment $\frac{1}{3}K V_{IP}$ shows that the Indian and Pacific basins are important locations for the removal of NADW in our model. Indeed, reduced vertical mixing inside the Indian and Pacific basins in $\frac{1}{3}K V_{IP}$ inhibits the eventual removal of NADW from the Atlantic basin and is therefore detrimental to NADW formation. In contrast, the similarity between the Atlantic MOC of the NADW “off” state in CNTRL and $\frac{1}{3}K V_{IP}$ shows that the AAIW reverse cell is not affected by the reduction of vertical mixing in the Indian and Pacific Oceans. This is not surprising as its upwelling branch occurs inside the Atlantic basin. Therefore, a reduction of
$K_v$ inside the Indian and Pacific Oceans shifts the competitive advantage from the NADW cell in favour of the AAIW reverse cell. While not explored here, this may have implications for ancient states of the global THC, as the distribution of seafloor “roughness” has evolved gradually over geological time-scales. In particular, changes in the Southern Ocean could play an important role.

A global increase of $K_v$ in our model results in significantly enhanced NADW stability. However, an identical enhancement of $K_v$ only inside the Atlantic ocean results in a reduction in NADW stability with respect to the control experiment CNTRL. Our results are thus not in contradiction with the enhanced NADW stability under *globally* increased $K_v$ found by Manabe and Stouffer (1999). However, we find a markedly different response when $K_v$ is increased only within the Atlantic Ocean, as in that case NADW stability is diminished. This suggests that the enhanced stability of the NADW “on” state found by MS99 under globally increased $K_v$ is due to the increase in vertical mixing outside the Atlantic Ocean, where NADW resurfaces.

The polarity of the global MOC is determined by interhemispheric competition between the AAIW reverse cell and the NADW cell (in today’s climate the AABW cell plays a minor role due to the Drake Passage effect, Sijp and England 2005). Unlike the idealised symmetric competition between a SH cell and a NH cell described by Bryan (1986), the NADW and AAIW cells depend on upwelling at different locations. In contrast to the AAIW reverse cell, the NADW cell depends on removal of its deepwater in areas outside the Atlantic, including the Southern and low-latitude Indian/Pacific Oceans. While vertical mixing plays an impor-
tant role in NADW removal in our results, wind-driven upwelling in the Southern Ocean is thought to be another important mechanism to remove NADW from the deep ocean (Toggweiler and Samuels 1995). NADW is sensitive to vertical mixing in the Indian and Pacific Oceans in our model as a substantial component of NADW upwells into those sectors in the simulations (as in Gordon 1986). If wind-driven upwelling was a more prominent NADW removal mechanism in our model, perhaps by employing vertical mixing schemes with lower background diffusivity values, we might expect a reduction in this sensitivity. For instance, Saenko and Merryfield (2005) employ a parametrization of tidally-driven deep mixing combining mixing near rough topography with a low background diffusivity of $10^{-1} \text{cm}^2/\text{s}$. As this tidally-driven mixing over rough topography mainly occurs below the pycnocline, vertical mixing inside the pycnocline is small. In this model set-up, Saenko and Merryfield (2005) find a dominance of Southern Ocean deepwater removal, over the classical notion of uniform abyssal upwelling (Munk 1966; Stommel and Arons 1960), and the conveyor pathway schematic of Gordon (1986). We would nonetheless expect the sensitivity of the AAIW reverse cell to vertical mixing in the Atlantic to be robust in experiments wherein wind-driven upwelling in the Southern Ocean dominates the removal of NADW.

We have shown that the relative strength of the NADW/AAIW cells, and therefore NADW stability, depends not only on surface density at the NADW and AAIW formation regions, but also on the relative strength of their removal mechanisms. This removal takes place in different geographic locations, so that the relative strengths of the NADW/AAIW cells can be altered by changing the vertical mixing coefficient in these regions. In the real ocean, NADW recirculation depends on vertical mixing across the thermocline and mechanical wind-driven
upwelling in the Southern Ocean. Of these fundamentally different processes, vertical mixing appears to be regulating the removal of deepwater when the AAIW reverse cell operates in a NADW “off” state. This has implications for the understanding of the global thermohaline circulation in past and future climates.

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<table>
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<th>AAIW reverse cell</th>
<th>NADW Stability</th>
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<td>7.8</td>
<td>15.1</td>
<td>unstable</td>
</tr>
</tbody>
</table>
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