Effects of different closures for thickness diffusivity

Carsten Eden¹, Markus Jochum² and Gokhan Danabasoglu²

IFM-GEOMAR, Kiel, Germany
NCAR, Boulder, USA

Manuscript submitted to Ocean Modelling, 2008.

Abstract

The effect of lateral and vertical structure in the thickness diffusivity (K) appropriate to the parameterisation of Gent and McWilliams (1990) is assessed in a coarse resolution global ocean general circulation model. Simulations using three different closures yielding lateral and/or vertical variations in K are compared with a simulation using a constant value. Although the effect of changing K is in general small and all simulations remain biased compared to observations, we find systematic local sensitivities of the simulated circulation on K.

The depth of the equatorial thermocline lifts by increasing K locally in the near surface tropical ocean. The strength of the Antarctic Circumpolar Current decreases while the subpolar and subtropical gyre transports in the North Atlantic increase by increasing K locally. We also found that the lateral and vertical structure of K given by a recently proposed closure reduces the negative temperature biases in the western North Atlantic by adjusting the pathways of the Gulf Stream and the North Atlantic Current to a more realistic position.
1 Introduction

In many state-of-the-art, non-eddy-resolving ocean models the so-called thickness diffusivity $K$ appropriate to the Gent and McWilliams (1990) (GM) parameterisation is used. This lateral diffusivity is meant to account for the advective effect of the turbulent lateral mixing by mesoscale eddies. In the GM parameterisation, the value of $K$ has to be specified and was chosen in the past as a constant value of $o(1000 \, m^2/s)$. Modern ocean models, however, are beginning to incorporate spatially varying thickness diffusivities (Griffies et al., 2005; Danabasoglu and Marshall, 2007). It is the aim of this study to compare a representative selection of such closures for $K$ in a global ocean model and to document their impact on the simulated circulation and watermass characteristics. We compare four different choices for the thickness diffusivity:

- $K = 800 \, m^2/s$, i.e. a constant value (experiment CONST, hereafter)
- $K(x, y, t)$ as suggested by Visbeck et al. (1997), i.e. $K = \alpha L^2 \bar{\sigma}$ where $\bar{\sigma}$ is the Eady growth rate averaged over the main thermocline, $L$ an eddy length scale and $\alpha$ a tuning parameter (experiment VMHS)
- $K(x, y, z, t)$ as suggested by Danabasoglu and Marshall (2007) dependent on the local stability frequency ($N$), i.e. $K = K_0 N^2 N_{ref}^{-2}$ with the parameters $K_0$ and $N_{ref}$ as specified below (experiment NSQR)
- $K(x, y, z, t)$ dependent on an eddy length scale and time scale as suggested by Eden and Greatbatch (2008), i.e. $K = c L^2 \sigma$ where $\sigma$ is the local Eady growth rate and $L$ the minimum of Rossby radius and Rhines scale. (experiment EG)

In CONST, the thickness diffusivity $K$ shows no spatial or temporal variation (except for a tapered procedure) and can be considered as the zeroth order choice. This choice was often used in the past due to the lack of knowledge about a physically meaningful spatial dependency of $K$. In a first step beyond a constant value, Visbeck et al. (1997) suggested to use a mixing length approach, first proposed for geophysical application by Green (1970) and Stone (1972) and in accordance to the scaling given in Larichev and Held (1995) and Held and Larichev (1996). Visbeck et al. (1997) choose as the time scale the inverse Eady growth rate ($\sigma$) and as the length scale the width of the baroclinic zone (of the region of interest). They also assumed a vertically constant $K$.

On the other hand, it is often proposed that $K$ should also have vertical structure. Danabasoglu and McWilliams (1995) and Jochum (1997) proposed prescribed exponentially varying
vertical profiles for $K$. Ferreira et al. (2005) estimated the optimal value of $K$ in a non-eddy-
resolving model simulation using an adjoint technique while Eden et al. (2007) and Eden
(2006) diagnosed $K$ directly from results of eddy-resolving models. In these studies signifi-
cant horizontal and also vertical variations were found, with magnitudes ranging from zero to
more than 5000 m$^2$/s, consistent with previous estimates of eddy diffusivity from observations
and models (Rix and Willebrand, 1996; Ledwell et al., 1998; Stammer, 1998; Bryan et al.,
1999; Treguier, 1999; Nakamura and Chao, 2000; Roberts and Marshall, 2000; Drijfhout and
Hazeleger, 2001; Solovev et al., 2002; Zhurbas and Oh, 2004; Marshall et al., 2006). In general
all studies, and in particular Ferreira et al. (2005) and Eden et al. (2007), also agree in the
finding of large $K$ in the upper thermocline and small values below, which motivated Ferreira
and Marshall (2006) and later Danabasoglu and Marshall (2007) to investigate the impact of
using $K$ proportional to the square of the stability frequency ($N^2$) in a model. The choice
was motivated by the fact that $N^2$ shows a similar vertical structure as the diagnosed $K$ in
Ferreira et al. (2005), Eden et al. (2007) and Eden (2006). Our experiment NSQR is identical
to experiment ITN2 discussed in Danabasoglu and Marshall (2007).

A further closure which we consider here (experiment EG) is a simplified version of the
parameterisation of Eden and Greatbatch (2008). It is based on a mixing length approach
by Green (1970) and Stone (1972) and a prognostic budget for eddy kinetic energy with
parameterised sources (due to baroclinic and barotropic instability) transports and sinks
(dissipation). A simplified and localised form of the closure (as used here in experiment EG)
was shown to agree in mid-latitudes with the scaling laws of Larichev and Held (1995) and
Held and Larichev (1996) and to be similar to the parameterisation of Visbeck et al. (1997),
although three-dimensional and with a different eddy length scale. We consider this closure
here because it combines vertical variations (NSQR) with horizontal (VMHS) ones.

It was found in Eden and Greatbatch (2008) that the application of the new closure did not
change much a North Atlantic model simulation of $o(1^\circ)$ resolution compared to a simulation
using a constant value of $K$. This result is qualified here since in agreement with Danabasoglu
and Marshall (2007) we find indeed certain improvements of a global ocean model simulation
of $o(3^\circ)$ resolution and it is our aim to document these improvements. We also report some
systematic model dependencies of the circulation on the value of the thickness diffusivity,
such as the strength of the subpolar gyre with impacts on convection activity and ventilation
rates of the North Atlantic and the depth of the equatorial thermocline with potential effects
on the simulation in coupled climate models.
In all experiments, a state-of-the-art coarse-resolution ocean model based on the Parallel Ocean Program of the Los Alamos National Laboratory (Smith and Gent, 2004) is used, which is also the ocean component of the Community Climate System Model (CCSM). It is a global, level-coordinate model with the grid North Pole displaced into Greenland with nominal 3° horizontal resolution (Yeager et al., 2006). There are 25 vertical levels, monotonically increasing from 8 m near the surface to about 500 m in the abyssal ocean. The surface forcing is given by the normal year (CORE) forcing data set provided by Large and Yeager (2004) at T62 resolution. There is no active sea ice model. The vertical mixing coefficients are determined using the K-Profile Parameterization of Large et al. (1994), as modified by Danabasoglu et al. (2006). In the ocean interior, the background internal wave mixing diffusivity varies in the vertical from $0.1 \times 10^{-4} \text{m}^2\text{s}^{-1}$ near the surface to $1.0 \times 10^{-4} \text{m}^2\text{s}^{-1}$ in the abyssal ocean. The vertical viscosity has the same shape as the vertical diffusivity, but 10 times larger. More details about the model setup can be found in e.g. Danabasoglu et al. (2008), in the following we only discuss issues relevant for the thickness diffusivity.

In each case, a preliminary version of the near-surface eddy flux parameterization of Ferrari et al. (2008), as implemented by Danabasoglu et al. (2008), is used in the model tracer equations. In this new approach, a transition layer separates the quasi-adiabatic interior where eddy fluxes are oriented along isopycnals from the diabatic, near-surface regions (e.g., the boundary layer) where diapycnal mesoscale fluxes are directed along the ocean surface. In the interior, eddy fluxes are still represented using the isopycnal diffusion tensor (Redi, 1982; Cox, 1987) and the Gent and McWilliams (1990) parameterization for eddy induced advection (with, eventually, horizontally and/or vertically varying identical diffusivities for both). As the surface is approached, the mesoscale eddy fluxes become parallel to the ocean surface, crossing outcropping density surfaces. This behavior is parameterized using a down-gradient horizontal diffusion with a mixing coefficient of $K_H$, which has the same value as the interior isopycnal/thickness diffusivity, i.e., $K$, indicating that the interior and near-surface mixing rates remain the same. A linear combination of horizontal and isopycnally-oriented mixing occurs within the transition layer. The eddy-induced velocity is parallel to the ocean surface and has no vertical shear within the boundary layer. It must then develop vertical shear within the transition layer to match the interior values. We use the mixed layer depth as defined in Large et al. (1997) to represent the diabatic layer depth in the present simulations. We note

1 Zonal resolution is uniform at 3.6°, but the meridional resolution varies from 0.4° and 0.6° in the subpolar northeast Atlantic and at the equator, respectively, to about 3° in Southern Hemisphere mid-latitudes and northwest Pacific.
that this new scheme eliminates the need for any ad-hoc, near-surface taper functions usually
used with the Gent and McWilliams (1990) scheme. Further details of the near-surface eddy
parameterization are given in Danabasoglu et al. (2008).

In the quasi-adiabatic interior, the four different choices for the thickness diffusivities are
used. Note that we use identical values for thickness diffusivity and isopycnal diffusivity.
We expect only minor effects by the choice of isopycnal diffusivity on buoyancy and the
circulation, but compare Danabasoglu and Marshall (2007) for numerical experiments with
the present model concerning the choice of isopycnal diffusivity. For passive tracers with large
gradients on isopycnals the choice of isopycnal diffusivity becomes however important and are
briefly discussed based on an ideal age tracer.

The main experiments (CONST, VMHS, NSQR and EG) have been integrated for 500
years. We note that this integration length might be too short for a complete diffusive equilib-
rium of the ocean model, for which the time scale would be several 1000 years. Consequently,
in this study we focus on the dynamical effects of the thickness diffusivity, for which the
integration length should be sufficient. Note that the diffusive equilibrium will also depend to
a large extent on parameterisation of diapycnal diffusivity and surface forcing, issues which
are left aside in the present study. We have integrated additionally a number of sensitivity
experiments for 100 years. A large number of further model details have to be clarified in the
remainder of this section. They are discussed separately for each experiment.

**Experiment CONST**

The constant value of $K = 800 \text{ m}^2/\text{s}$ in experiment CONST, was found to produce simulation
in good agreement with observations in Danabasoglu (2004). We use therefore this value as
well here and consider it as the optimal (constant) value for this particular ocean model.

**Experiment VMHS**

In experiment VMHS the thickness diffusivity is given by $K = \alpha L^2 \bar{\sigma}$. The mean Eady growth
rate $\bar{\sigma} = f \bar{Ri}^{-1/2}$ is calculated from the Richardson number $\bar{Ri} = \bar{N}^2 \left| \frac{\partial}{\partial z} u_h \right|^{-2}$ averaged
between 100 m and 2000 m depth as being representative of the Eady growth rate within
the main thermocline of the ocean. The eddy length scale $L$ was defined in Visbeck et al.
(1997) somewhat unspecific as the width of the baroclinic zone, which is difficult to estimate
in a realistic numerical model. On the other hand, it is often assumed that the local Rossby
radius determines the relevant eddy length scale (Krauss et al., 1990; Stammer, 1997). For
\( N = \text{const} \) the Rossby radius is given by \( Nh(f\pi)^{-1} \) and as before for the Eady growth rate we use a vertically averaged \( N \) as being representative for the main thermocline, i.e. \( \bar{L}_r = \bar{N} hf^{-1} \) with \( \bar{N} = (\bar{N}^2)^{1/2} \) where the vertical mean is taken over same depth range \( h \) as for \( \bar{Ri} \). However, we noticed that using \( L = \bar{L}_r \) in the model yields too low values of \( K = \alpha L^2 \sigma \) at higher latitudes, or, in turn, too high values in the subtropics (by adjusting the parameter \( \alpha \)). We decided to use \( L = \max(\bar{L}_r, \Delta x) \) where \( \Delta x \) denotes the local horizontal grid spacing. Although the optimal value for the parameter \( \alpha \) was found by Visbeck et al. (1997) as \( \alpha = 0.015 \), a larger value \( \alpha = 0.13 \) is used here. The resulting diffusivity \( K \) is furthermore bounded by \( 300 \, m^2/s \leq K \leq 4000 \, m^2/s \).

Note that the length scale \( L \) is important for the parameterisation of Visbeck et al. (1997) and certainly related to the choice of the parameter \( \alpha \). Other choices for \( L \) are possible, e.g. a constant length scale as in Griffies et al. (2005) or more complicated choices such as the width of the baroclinic zone defined by some kind of algorithm (Visbeck et al., 1997; Pacanowski and Griffies, 1999). We consider experiment VMHS as an example for many other possibilities, which, however, have a vertically constant \( K \) in common.

In VMHS, the local Rossby radius is replaced from its mid-latitude form to the equatorial Rossby radius near the equator. However, the Eady growth rate, \( \sigma = f|\partial_z u_h|/N \), and thus \( K \) will go to zero near the equator. This is because the closure is based on heuristic considerations relevant for mid-latitude baroclinic instability theory, which become inappropriate at the equator. To prevent \( K \) to become zero near the equator, \( f \) is replaced with \( \sqrt{2 \beta e} \) as the relevant equatorial time scale in VMHS near the equator.

**Experiment NSQR**

For experiment NSQR, Danabasoglu and Marshall (2007) proposed as optimal choice \( K_0 = 4000 \, m^2/s \). The reference value for the stability frequency \( N_{ref} \) is chosen at each horizontal position as the uppermost value of \( N \) in the quasi-adiabatic interior layer provided that \( N^2 > 0 \) there, otherwise, \( N_{ref} \) is given by the first \( N^2 > 0 \) further below. The resulting thickness diffusivity is bounded by \( K_0/10 \leq K \leq K_0 \).

**Experiment EG**

In experiment EG, the thickness diffusivity is given by \( K = cL^2 \sigma \). The eddy length scale \( L \) is given as the minimum between Rossby radius \( L_r \) and Rhines scale \( L_{Rh} = \sqrt{\bar{e}^{1/2} \beta^{-1}} \), where \( \bar{e} \) denotes EKE. This choice for \( L \) was found to be consistent with independent estimates of
eddy length scales from satellite observations and eddy-resolving model results (Eden, 2007) and with recent theoretical considerations (Theiss, 2004). \( L_{Rhi} \) is estimated from variables of the coarse model as \( L_{Rhi} = \sigma/\beta \) (Eden and Greatbatch, 2008) while \( L_r \) is given by \( L_r = \min \left[ \frac{c_r}{f}, \sqrt{\frac{c_r}{2\pi}} \right] \) where \( c_r \) denotes the first baroclinic Rossby wave speed, which is calculated approximately following Chelton et al. (1998) by \( c_r \approx f|_h N/\pi \) dz, where \( h \) denotes the local water depth. The inverse eddy time scale \( \sigma \) was found by Eden and Greatbatch (2008) as given by \( \sigma = |\nabla \bar{b}|/N \), which is, using the thermal wind relation, for mid-latitudes identical to the Eady growth rate \( \sigma = fRi^{-1/2} \). Here, \( Ri = N^2|\frac{\partial}{\partial z} u_h|^2 \) denotes the local Richardson number in contrast to a vertically averaged \( Ri \) as in VMHS.

It is clear that \( \sigma \) has a singularity for \( N \to 0 \). To prevent this singularity we use \( \sigma = f(Ri + \gamma)^{-1/2} \) with \( \gamma > 0 \) which acts effectively as an upper limit for \( \sigma \) and consequently for \( K \). Note that the latter setting is motivated by the fact that effective growth rates are replaced by \( f(Ri + 1)^{-1/2} \) for baroclinic instability under weak stratification (Stone, 1971, 1972). The effect of \( \gamma \) and the tuning parameter \( c \) will be explored below, they were set initially to \( \gamma = 200 \) and \( c = 2 \) in experiment EG. However, note that \( Ri \gg 200 \) over the main thermocline and below except near the surface diabatic layer (and in abyssal almost unstratified regions), such that the choice of \( \gamma \) is only relevant approaching the mixed layer where the tapering scheme becomes active as well.

Note that using \( \sigma = f(Ri + \gamma)^{-1/2} \) in EG, the time scale and thus \( K \) goes to zero at the equator, as before in experiment VMHS. To prevent this, \( f \) is replaced with \( \sqrt{2\beta c_r} \) also in EG as before for VMHS. Note also that it is in principle also possible to include in addition to baroclinic instability horizontal shear (barotropic) instability in the new closure (Eden and Greatbatch, 2008), which is however not considered in the present study.

3 Results

Thickness diffusivity

The horizontal distribution of the thickness diffusivity \( K \) in the main thermocline in the four basic model simulations are shown in Fig. 1, while the vertical distribution in the Pacific Ocean is shown in Fig. 2. In experiment CONST the effect of the tapering scheme of Ferrari et al. (2008) can be seen, which becomes active in the polar regions where deep mixed layers are met, near lateral boundaries and in a relatively thin near surface layer. Note that this tapering scheme is identical for all experiments.
In VMHS, $K$ is large, up to the upper bound of 4000 m$^2$/s near the equator, in the Southern Ocean and in western boundary currents but vertically constant below the diabatic layer near the surface. Since the length scale $L$ in VMHS is given by the local grid spacing $\Delta x$, except for the tropics where the Rossby radius $\bar{L}_r$ becomes larger than $\Delta x$, the horizontal structure in higher latitudes depends to a large degree on the mean Eady growth rate $\bar{\sigma}$ (the effect of the reduced grid spacing in $K$ towards the poles is minor). The maximum in $K$ near the equator can be explained by the large $\bar{L}_r$ while all remaining maxima of $K$ in VMHS, i.e. in the Southern Ocean and in western boundary currents, coincide with maxima in $\bar{\sigma}$. Therefore, maxima in off-equatorial values of $K$ in VMHS are related to maxima in horizontal buoyancy gradients and vertical shears associated with the Antarctic Circumpolar Current (ACC) and the western boundary currents.

In experiment NSQR, values of $K$ in the main thermocline are in general smaller than in VMHS, but note that they reach $K = 4000$ m$^2$/s at each water column at the base of the surface diabatic layer by construction. As expected $K$ is large near the surface and decreasing with depth. In regions of strong near surface stratification, such as in the tropical Pacific with its sharp thermocline, $K$ shows a faster decay with depth compared to regions where the stratification varies less with depth, as e.g. in the Southern Ocean. The result are large values of $K$ in the near surface Pacific, comparable to VMHS, and moderate values in the Southern Ocean.

In experiment EG, $K$ is large where large horizontal density gradients and vertical shears can be found. This is similar to VMHS, with however smaller values of $K$ in the Southern Ocean and the tropics due to the smaller length scale $L$. Note that $L$ becomes the Rhines scale $L_{Rhi}$ equatorwards of about 30° latitude (and in regions of very weak eddy energy), otherwise the local Rossby radius as discussed in Eden and Greatbatch (2008) and Eden (2007). The vertical structure of $K$ in EG is similar to NSQR, but note that the horizontal maxima of $K$ in NSQR differ from EG. In NSQR, maxima can be found in the centre of the subtropical gyres, which is in particular evident for the North Atlantic, while in EG the maxima are near the western boundary currents and in regions of strong isopycnal slopes.

**Watermass characteristics**

The changes in thickness diffusivities in the individual experiments show up in turn in differences in the simulated watermasses and circulation. We compare in this section the temperature distribution after 500 years integrations of the individual experiments with observational estimates of Levitus and Boyer (1994). We first note that in all experiments large temperature
biases up to 8° C in regions of strong lateral or vertical temperature gradients can be found. Biases of similar relative magnitude can be seen in the salinity distribution of the simulations. These biases change and sometimes decrease having adjusted the thickness diffusivity, which we discuss here for two exemplaric regions, namely the North Atlantic Ocean and the equatorial ocean. Note, however, that in none of the simulations the temperature and salinity biases are reduced to a value well below the observational error margin (which we can only guess) such that all simulation are still biased.

Fig. 3 shows a prominent example of such a model bias, related to a missing Northwest Corner of the North Atlantic Current (NAC). This well-known artifact, coming along with a displaced Gulf Stream, leaves its imprint in a large negative temperature bias in the subpolar North Atlantic in experiment CONS. The model bias is still present in experiment NSQR but decreases much in experiment EG and VMHS. In fact, the negative temperature bias east of Newfoundland of up to 5° C in experiment CONS and NSQR almost vanish in EG and VMHS and a small region with an even positive bias near the coast shows up. The average rms temperature bias between 40° and 60°N at 200 m depth in the North Atlantic is 2°, 2.2°, 1.8° and 1.7° C for CONS, NSQR, VMHS and EG respectively. It is clear that the vanishing negative temperature bias in EG and VMHS is related to a different pathway of the Gulf Stream and the NAC around Newfoundland and the interior subpolar gyre. Note that this pathway determines to a large extent the subpolar front in the North Atlantic and is apparently in better agreement with the observed paths in experiment EG and VMHS compared to the other model solutions.

Fig. 4 shows a meridional section of the temperature biases and zonal velocity within the upper thermocline of the North Atlantic. The figure shows that the core of the NAC is indeed displaced by several hundred km in CONS and NSQR compared to the other experiments. Negative temperature bias due to the southward displacement of the NAC in CONS and NSQR extents down to almost 1000 m depth. The average rms temperature bias for the region shown in Fig. 4 is 1.5°, 1.8°, 2.4° and 1.2° C for CONS, NSQR, VMHS and EG respectively. In VMHS, the NAC is much weaker compared to EG, and a rather larger negative temperature bias shows up below the main thermocline in the subtropical gyre. These artifacts in VMHS can be attributed to the differences in the vertical structure of K, i.e. the missing decay of K with depth in VMHS. Note, however, that the horizontal near surface structure of K is similar in VMHS and EG in the North Atlantic (Fig. 1), leading to the improvement with respect to the path of the NAC in both experiments.

There are also large temperature biases in the tropical Atlantic. They are dominantly
positive in CONST, NSQR and EG and slightly smaller in VMHS (at 200 m depth). This comes along with a sharper and shallower equatorial thermocline in VMHS. Fig. 5 shows a zonal section of the temperature biases in all 4 experiments along the equator. Within the upper 200 m the experiments differ only slightly, largest differences can be found below 200 m. A too deep equatorial thermocline can be seen in the Indian and Atlantic Ocean in CONST, NSQR and EG, while in VMHS this bias is reduced. In the Pacific Ocean, the thermocline appears to be even too shallow in VMHS. The larger (and vertically constant) thickness diffusivity at the equator in VMHS compared to the other experiments apparently reduces the temperature bias in equatorial ocean by lifting the thermocline.

Circulation

The relative changes in the simulated circulation are in general small. Exceptions are changes in the western boundary currents of the North Atlantic in experiment EG related to the temperature bias in that region as discussed above and changes in Southern Ocean in VMHS. The maximal strength of the Eulerian mean meridional overturning circulation (MOC) between 0°N and 70°N in the Atlantic Ocean shown in Fig. 6 are 14.8, 15.2, 9.9 and 11.1 Sv for experiments CONST, NSQR, VMHS and EG respectively. Note that the vertical structure of the deep overturning cell in the North Atlantic is different in VMHS compared to the other experiments. The maximal transports of the subtropical (subpolar) gyre in the North Atlantic (not shown) are 31.1, 32.0, 31.9 and 26.3 Sv (31.3, 32.1, 23.6 and 13.7 Sv) for experiments CONST, NSQR, VMHS and EG respectively. It is clear that the horizontal gyre strengths, in particular the strength of the subpolar gyre, are significantly smaller in experiment EG compared to the other experiments, and also compared to observational estimates of around 40 Sv (Pickart et al., 2002), an issue explored below in more detail.

Despite the reduced MOC strength in experiment EG, the reduced strength of the subpolar gyre in the North Atlantic in EG comes along with larger ventilation rates. The latter was estimated using a standard ideal age tracer \( \tau \) integrated simultaneously during the model simulations. The source function for \( \tau \) is given by one unit per model year. Fig. 7 shows \( \tau \) in the lower limb of the MOC in the North Atlantic after 500 years of model integration. It is clear that the ventilation of the deep North Atlantic appears to be strongest in EG, and that the ideal age of the watermasses within the deep western boundary current is significantly younger and more concentrated at the western boundary in experiment EG compared to the other experiments. The ideal age is smaller in EG compared to the other experiments since the ideal age is already up to 20 years younger in the subpolar North Atlantic where the deep
watermasses are formed, which appears to be in turn related to the depth reached by the
deep convection in the Labrador Sea: Maximal mixed layer depths in CONST, NSQR and
VMHS are not deeper than 1500 m, while in EG the winter time convection reaches depths of
up to 2500 m in the central Labrador Sea. In VMHS, we find the weakest ventilation of the
subpolar North Atlantic coming along with the shallowest mixed layer depths. Note that we
find strongest gradients in ventilation age in the North Atlantic for experiment EG, weaker
gradients in CONST and NSQR and even weaker ones in VMHS as a consequence of using
different isopycnal diffusivities in the experiments. Note also that the deep boundary flow
appears to to be detached from the coastline of the subtropical North Atlantic in VMHS.

It is known that the ACC transport depends to a large degree on sub-grid-scale param-
terisations and, in particular, on the thickness diffusivity (Danabasoglu and McWilliams,
1995). Consistently, the Drake Passage transport is lowest in experiment VMHS, only 71.7
Sv, since here \( K \) was largest in all experiments, and higher, i.e. 133.1, 143.8 and 140.6 Sv in
the experiments CONST, NSQR and EG respectively, thus more consistent with the observa-
tional estimate of 134 ± 13 Sv (Whitworth and Peterson, 1985). This model bias in VMHS
comes along with a drastic change in the MOC of the Southern Ocean. Fig. 8 shows the total
MOC, i.e. Eulerian mean plus transient eddy contribution (parameterised by the thickness
diffusivity) in the individual experiments. While in experiments CONST, NSQR and EG a
meridional circulation of similar strength and familiar structure shows up, the circulation in
the upper 2000 m in experiment VMHS is rather different from the other experiments coming
along with the larger thickness diffusivity in the Southern Ocean. In fact, the near surface
polewards transport between 65° and 45° S has almost vanished in VMHS.

Dependency on parameters

The large values of the thickness diffusivity \( K = \alpha L^2 \bar{\sigma} \) in the Southern Ocean in the simu-
lation VMHS are related to a reduced ACC transport. We have reduced \( K \) by changing the
parameter \( \alpha \) from \( \alpha = 0.13 \) in VMHS to \( \alpha = 0.04 \) in the experiment VMHS_LOW (which
we have integrated for 100 years only). The resulting \( K \) in the main thermocline is shown in
Fig. 9 a). The change in thickness diffusivity yields indeed a higher Drake passage transport
of 103.6 Sv, in better agreement to the observational estimates, although still too low, such
that one might tend to reduce \( \alpha \) even further. On the other hand, we notice that the thick-
ess diffusivities in the North Atlantic are now also much lower in VMHS_LOW compared to
VMHS and all previous experiments. This comes along with an increased cold temperature
bias related to a missing Northwest corner and a southward displacement of the NAC (Fig. 9
b). Since $K$ in VMHS\_LOW is also smaller in the tropical Atlantic compared to VMHS, the bias of too deep equatorial thermocline is also enhanced again. We conclude that by tuning the single parameter $\alpha$ it appears impossible to improve the parameterisation of $K$ and the model simulation at all locations at the same time. In order to reduce all biases one would have to use a different length scale $L$ or a different time scale $\sigma$ in VMHS (or a spatially dependent parameter $\alpha$ which would be equivalent).

In experiment EG, the thickness diffusivity is given by $K = cL^2\sigma$. The effect of a smaller or larger value of the parameter $c$ is consistent with the previous discussed effects of the value of $K$ on the circulation. Changing $c$ from $c = 2$ in EG to $c = 1$ in the experiment EG\_LOW (to $c = 3$ in the experiment EG\_HIGH), the Drake passage transport after 100 years of integration is increased (decreased) from 151.2 to 170.3 $Sv$ (139.6 $Sv$). The strength of the subtropical gyre in the North Atlantic decreases after 100 years of integration from 28.4 $Sv$ in EG\_HIGH to 25.3 $Sv$ in EG and 25.6 $Sv$ in EG\_LOW, and the subpolar gyre strength decreases from 16.1 $Sv$ in EG\_HIGH to 14.0 $Sv$ in EG and 12.6 $Sv$ in EG\_LOW (the maximal mixed layer depth in the Labrador Sea and the ideal age in EG, EG\_HIGH and EG\_LOW vary however only little).

On the other hand, the improvement of the cold bias in the North Atlantic in EG is very similar after 100 years of integration in EG\_LOW and EG\_HIGH (not shown), i.e. the pathway of the NAC appears rather insensitive to the magnitude of $K$. In contrast, the equatorial thermocline depth shows a significant sensitivity on the parameter $c$ (Fig. 10). For higher (lower) $K$ in the tropical ocean, the equatorial thermocline gets higher (deeper). In fact, the temperature in EG\_HIGH is now in good agreement below ca. 200 m in the tropical Pacific and Indian Ocean, although the tropical Atlantic still shows a warm bias below the thermocline. Consistently, the equatorial temperature biases are getting larger in EG\_LOW.

Note that the dependency of the subpolar gyre strength on $K$ is also consistent with NSQR and VMHS: Here, $K$ is larger in both experiments NSQR and VMHS since lower bounds for $K$ have been prescribed, which we have not used in EG, EG\_LOW and EG\_HIGH. Thus, larger subpolar gyre strengths correspond to higher values of $K$. To test the influence of a lower bound of $K$ on the circulation in experiment EG\_BOUND we have employed the lower bound $K \geq 300 \ m^2/s$ on the thickness diffusivity. In experiment EG\_BOUND, the Drake passage transport stays almost the same as in EG, while the subpolar (subtropical) gyre strength in the North Atlantic increases after 100 years of integration from 14.0 (25.4) $Sv$ in EG to 17.2 (26.0) $Sv$ in EG\_BOUND. We also notice that maximal convection depth in the Labrador Sea and the ventilation rates as simulated by the ideal age tracer are reduced
in EG_BOUND compared to EG. On the other hand, the temperature distribution in the upper North Atlantic was almost unchanged in EG_BOUND compared to EG. Finally we have assessed the sensitivity to the parameter $\gamma$ by changing it from $\gamma = 200$ in EG_HIGH to $\gamma = 50$ in EG_GAMMA. The differences are as expected only very small. However, reducing the parameter further to $\gamma = 1$, maximal values of $K$ are getting too large in regions where the stratification becomes almost zero and a stable integration becomes impossible. This behaviour could of course be omitted by simply by employing an upper bound for $K$ (as in the other experiments).

4 Summary and discussion

We have described the effect of different closures for the thickness diffusivity $K$ appropriate for the Gent and McWilliams (1990) parameterisation in a coarse resolution general circulation model of the global ocean. Our choices for $K$ include a constant value (CONST), a variant of the Visbeck et al. (1997) parameterisation for $K$ (VMHS), a value dependent on the stability frequency $N$ (Danabasoglu and Marshall, 2007) (NSQR) and a new closure (Eden and Greatbatch, 2008) (EG). The thickness diffusivity $K$ in VMHS is horizontally varying and proportional to the Eady growth rate representative for the main thermocline but vertically constant. The vertical profile of $K$ in NSQR is proportional to the local $N^2$ profile while no other horizontal structure was prescribed. $K$ in EG is both horizontally and vertically varying, proportional to the local Eady growth rate and the square of an eddy length scale given by the minimum of Rossby radius and Rhines scale.

In general, the comparison and valuation of the simulated thickness diffusivity in the individual experiments is hampered by the lack of large-scale coverage of observational estimates for sub-surface eddy statistics. However, quantitative model based estimates of $K$ (Ferreira et al., 2005; Eden et al., 2007; Eden, 2006) agree in the finding that $K$ should be large within western boundary currents and on the northern flank of the ACC and smaller elsewhere. Furthermore, $K$ should be in general large, $o(1000 \text{ m}^2/\text{s})$, near the surface and rapidly decaying with depth, which is the case in experiments NSQR and EG but not in VMHS. On the other hand, the former requirement is qualitatively met by all experiments with spatially varying $K$, although there are relatively large variations in the relative magnitudes of $K$. In particular, the horizontal distribution of $K$ in NSQR show elevated values, in fact even maxima, within the centre of the subtropical gyres instead of the western margins, in both the Atlantic and Pacific Ocean and in both hemispheres (compare Fig. 1 b).

Using our choice of the eddy length scale in VMHS, effectively a constant in off-equatorial
regions but decreasing with latitude when the grid spacing does, the thickness diffusivity $K$ in VMHS is excessively high in the Southern Ocean or, when “tuning” the parameterisation by changing the parameter $\alpha$, too small in the mid-latitude western boundary regions. We have concluded that it is not possible to obtain realistic values of $K$ at all locations at the same time without changing the length scale (or time scale) in VMHS. Note that it is also difficult to obtain values of $K$ in agreement with our a priori knowledge about the lateral distribution of $K$ using the local Rossby radius as the eddy length scale in VMHS. Other choices for $L$ are often proposed, such as a constant length scale as in Griffies et al. (2005) or more complicated choices such as the width of the baroclinic zone defined by some kind of algorithm (Visbeck et al., 1997; Pacanowski and Griffies, 1999) which we have not tested in the study. In any case, we stress that a vertically constant $K$ as in VMHS is not in agreement with our a priori knowledge about $K$.

The changes in $K$ affect the simulated watermass characteristics in a relatively weak but systematic manner. Although none of the experiments are unbiased in temperature (or salinity), we noticed improvements in the near surface North Atlantic and in the equatorial thermocline in some of the experiments. In all sensitivity experiments with the new closure in EG, the pathway of the Gulf Stream and the NAC was more realistic and the negative temperature bias near the subpolar front due to the missing Northwest Corner of the NAC was reduced with potentially positive effects on a coupled model simulation as recently discussed by Weese and Bryan (2006) for the CCSM climate model.

Since this improvement was insensitive to the choice of parameters in EG, i.e. almost identical in EG_LOW, EG_HIGH and EG_BOUND, it might therefore be related to the detailed horizontal structure of $K$ in the North Atlantic, i.e. high within the Gulf Stream/NAC system and low outside that currents. In experiment EG and in VMHS the horizontal structure of the simulated thickness diffusivity in the North Atlantic is similar and agrees better with our a priori knowledge of $K$ compared to NSQR, which shows maxima of $K$ in the centre of the subtropical gyre. It should be pointed out, though, that in spite of their different origin, the thermocline structure of EG and NSQR are rather similar, since the vertical structure of the Eady growth rate is similar to the structure of $N^{-2}$.

A further systematic effect of the magnitude of the thickness diffusivity in the model was the depth of the equatorial thermocline. All experiments agree in the finding that increasing $K$ in the interior of the ocean lifts the equatorial thermocline. Danabasoglu and Marshall (2007) partly related that effect to increased equatorial upwelling. In addition, a direct effect of the thickness diffusivity can be seen. Fig. 2 illustrates how the Equatorial Undercurrent (EUC),
for which the zonal flow is in geostrophic balance at the equator, bends the isotherms at the
equator: Upwards above the core of the EUC and downwards below it. Compared to CONST,
all experiments have an increased thickness diffusivity in the equatorial thermocline, which
flattens the isotherms there and leads to an equatorial cooling. This lifting and flattening of
the thermocline might be welcomed since the model shows a bias of a too deep equatorial
thermocline in the reference experiment CONST. However, the improvement of the bias may
also come as a surprise, since the physical motivation of the flattening of the isopycnals is
based on baroclinic instability theory relevant for mid-latitudes (Gent et al., 1995), while at
the equator a different dynamical regime might apply. Therefore, we suggest that the impact
of thickness diffusivity on the simulated equatorial dynamics should be viewed with caution.

In agreement to previous studies (Danabasoglu and McWilliams, 1995) we found that the
ACC transport is related to the value of the thickness diffusivity in the model, i.e. larger
$K$ lowers the ACC transport, which offers also a clear physical interpretation of the effect
(Danabasoglu and McWilliams, 1995). Here we also identified a dependency of the subpolar
and subtropical gyre strength in the North Atlantic to the magnitude of $K$. A larger $K$ leads
to a larger transport both in the subpolar and the subtropical gyre, although the effect is
stronger and clearer for the subpolar gyre in the individual experiments. A weaker subpolar
gyre was also related to deeper convection depths and larger ventilation rates as indicated by
an ideal age tracer in the model simulation.

We speculate that the weaker parameterised eddy heat flux which results from smaller
$K$ leads to deeper convection in the Labrador Sea because of less mixing of buoyant water
from the boundary currents into the interior Labrador Sea and thus weaker restratification
after convection. The mean circulation, in particular the barotropic flow and consequently
the subpolar gyre strength was found to be weaker for deeper convection depth, which could
be due to the interaction of the barotropic circulation with the topography in the presence
of stratification (JEBAR, see e.g. (Olbers and Eden, 2003)). However, note that this in
contrast to what is known from the response of the subpolar gyre strength to atmospheric
forcing variability, in which deeper convection depths in years of enhance buoyancy loss over
the Labrador Sea are related to a spinup of the subpolar gyre, which was also interpreted as
a consequence of JEBAR (Eden and Willebrand, 2001). On the other hand, freshwater might
also be important. Treguier et al. (2005) demonstrated how the peculiar balance between
gyre strength, deep convection and freshwater balance of the subpolar North Atlantic can
lead to large biases even in high-resolution ocean models. The role of the thickness diffusivity
for this balance and to what extent the effects of $K$ on the subpolar gyre are specific to the
particular model which we have used, remain to be demonstrated.
Acknowledgements

This study was supported by the Deutsche Forschungsgemeinschaft within the SPP 1158 and by the NSF grant OCE-0336827 for the Climate Process Team on Eddy Mixed-Layer Interactions (CPT-EMILIE). The computational resources were provided by the Scientific Computing Division of the National Center for Atmospheric Research (NCAR). NCAR is sponsored by the National Science Foundation.

References


Figure 1: Annual mean thickness diffusivity ($K$) in $m^2/s$ at 300 $m$ depth in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration. Values of $K$ are shown for the interior region only, i.e. values of $K$ in the (seasonal maximum) diabatic surface and transition layer are not shown. Note the non-linear colour scale.
Figure 2: Annual mean thickness diffusivity ($K$) in $m^2/s$ in the Pacific Ocean in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration. Shown is the zonally averaged $K$ between $140^\circ E$ and $120^\circ W$ (shading) and contour lines of the zonally averaged temperature ($2 K$ contour interval). Values of $K$ in the (seasonal maximum) diabatic surface and transition layer are not shown.
Figure 3: Annual mean temperature difference between simulation and the climatology of Levitus and Boyer (1994) in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration at 200 m depth. Also shown are contour line of temperature (2 °C contour spacing).
Figure 4: Annual mean temperature difference between simulation and the climatology of Levitus and Boyer (1994) in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration as average between 40° and 30°W. Also shown are isotaches of annual mean zonal velocity (1 cm/s contour spacing).

Figure 5: Annual mean temperature difference between simulation and the climatology of Levitus and Boyer (1994) in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration at the equator. Also shown are contour line of temperature (2 °C contour spacing).
Figure 6: Annual mean meridional overturning streamfunction in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration in $Sv$ in the Atlantic Ocean. Shown is the streamfunction for the Eulerian mean advection velocity.
Figure 7: Ideal age tracer $\tau$ in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration in years (shading). Shown is the vertical average of the annual mean $\tau$ between 1500 and 2500 m depth. Also shown are isolines (red lines) of the maximal mixed layer depth in the individual experiments with 500 m contour spacing.
Figure 8: Annual mean meridional overturning streamfunction in experiment CONST (a), NSQR (b), VMHS (c) and EG (d) after 500 years integration in Sv in the Southern Ocean. Shown is the streamfunction for the total tracer advection velocity, i.e. the Eulerian mean velocity plus the transient eddy contribution given by the parameterised bolus velocities.

Figure 9: a) Annual mean thickness diffusivity ($K$) in $m^2/s$ at 300 m depth in experiment VMHS_LOW after 100 years integration. b) Annual mean temperature difference between VMHS_LOW and the climatology of Levitus and Boyer (1994) at 200 m depth. Also shown are contour line of temperature ($2 ^\circ C$ contour spacing).

Figure 10: Annual mean temperature difference between simulation and the climatology of Levitus and Boyer (1994) in experiment EG_LOW (a) and EG_HIGH (b) after 100 years integration at the equator. Also shown are contour line of temperature ($2 ^\circ C$ contour spacing).