Cabling due to Isopycnal Mixing in Isopycnic Coordinate Models

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ABSTRACT

The cabling that arises as a consequence of isopycnal mixing in a North Atlantic model based on MICOM (the Miami Isopycnic Coordinate Model) is quantified. Annually averaged over the model Atlantic, the diapycnal volume flux associated with cabling reaches 1.5 Sv, with an associated net density flux of $2 \times 10^6$ kg s$^{-1}$ (equivalent to an annual-basin mean cooling of 0.6 W m$^{-2}$). Over the range of densities that incorporate the major water masses of the model Atlantic, cabling effectively weakens the density flux due to parameterized diapycnal turbulent mixing by $\approx 25\%$. The strength of cabling varies in proportion to the isopycinal mixing of heat and salt, the local buoyancy frequency, and a "cabling parameter" (which is inversely proportional to temperature). As a consequence of these dependencies, cabling is highly localized and seasonal. In the model, strongest cabling occurs during summer at the subpolar front in the northwest Atlantic.

Model cabling arises both physically (due to the independent mixing of heat and salt in isopycnic layers) and, to a lesser extent, nonphysically (due to the advection of heat and salt). Fields of layer thickness changes due to model cabling compare reasonably well with changes predicted by "physical" cabling. Physical cabling is therefore predicted for a global model (QGIM) based on a more recent version of MICOM, which features salinity-only advection and mixing (and hence no cabling). In the circumpolar Southern Ocean of QGIM, intermediate water would be transformed (by cabling) to higher density at rates of up to 7 Sv, primarily due to end-of-winter freshwater forcing around Antarctica. This suggests that the cabling associated with isopycnal mixing, although neglected in later versions of MICOM, may play a significant role in water mass transformation around the Southern Ocean. However, the layer temperature, salinity, and thickness fields used to initialize MICOM lead to unrealistically strong cabling around the Mediterranean outflow during the early stages of spinup, a problem which further highlights the unsuitability of $\sigma_z$ as a layer variable for water masses below $\approx 1000$ m.

1. Introduction

Due to the nonlinear nature of the equation of state for seawater, a conservation equation for potential density includes a source term (McDougall 1991). In a turbulent environment, this amounts to a "densification on mixing" proportional to the dissipation of thermal variance. Oriented laterally along or vertically across "neutral" surfaces, contributions to the thermal variance from epineutral (lateral) and dianeutral (vertical) heat fluxes are associated with distinctly separate physical effects. Lateral heat fluxes lead to observable changes in potential temperature and contribute to dianeutral advection, while vertical heat fluxes do not lead to observable consequences (McDougall and Garrett 1992). The process of density gain through isopycnal mixing of water parcels with the same density, but different temperature and salinity, has become known as "cabling." The rather obscure origins of this terminology are reviewed and discussed by Foster (1972).

Garrett and Horne (1978) find that the isopycnal mixing of heat and salt at a typical front leads to significant cabling and an associated sinking rate estimated to be about 1 m day$^{-1}$. Cabling may thus drive vertical circulations that maintain fronts against diffusion. Based on densely sampled vertical profiles of temperature and salinity, Horne (1978) concludes that cabling is indeed the dominant mechanism that drives vertical circulation at the subsurface front between warm slope water and Labrador slope water off Nova Scotia. Based on more sparsely sampled sections, Gordon et al. (1977) note strong cross-frontal isopycnal exchange during summer in the Antarctic Polar Frontal Zone (PFZ) in the western Scotia Sea, and show that the fraction $\Delta T/\Delta S$ (where $\Delta T$ and $\Delta S$ are temperature and salinity anomalies, relative to a frontal mean $T$--$S$ profile) is strongly dependent on a cabling parameter (see sec. 2 below). Gordon et al. (1977) point out that cross-frontal mixing may thus influence the characteristics and
thickness of the subtantarctic, nearly isohaline layer that subducts as Antarctic Intermediate Water in the vicinity of the PFZ.

Cabbeling may also lead to significant water mass transformation in the deeper ocean, although large uncertainties remain. McDougall (1987a) used the Levitus (1982) climatological dataset to compute fields of diapycnal velocities due to cabbeling and thermobaricity over the subtropical and midlatitude North Atlantic, choosing a spatially constant lateral eddy diffusivity \( K \) of 1000 m\(^2\) s\(^{-1}\). McDougall thus found that cabbeling generally dominates thermobaricity, associated downwelling of typically 0.5 \( \times 10^{-3} \) m s\(^{-1}\). At a depth of 1000 dB, this is half of, and opposite to, the global-mean upwelling velocity of \( 10^{-7} \) m s\(^{-1}\) required to upwell \( \sim 30 \) Sv (Sv = \( 10^{6} \) m\(^3\) s\(^{-1}\)) of bottom water to the base of the thermocline (Munk and Wunsch 1998).

McDougall (1987a) further stressed that his calculations of the diapycnal velocity due to cabbeling are conservative estimates, for two reasons. First, the chosen value for \( K \) is “probably low by at least a factor of 2” and spatially variable (dependent on the local eddy field). Estimates of \( K \) vary from 500 m\(^2\) s\(^{-1}\) [from inverse analysis of hydrographic data in a quiescent region of the eastern subtropical gyre (Armi and Stommel 1983)], through values of 1700–2900 m\(^2\) s\(^{-1}\) [from transient tracer distributions on \( \sigma_{26} = 26.5 \) and 26.8, closer to the eastern boundary (Thiele et al. 1986), up to 6500 m\(^2\) s\(^{-1}\) from SOFAR float trajectories at 700 dB in a region of moderate eddy activity in the western Atlantic (McWilliams et al. 1983)]. Most recently, the North Atlantic Tracer Release Experiment (again in the quiescent eastern subtropical gyre) has yielded estimates for \( K \) that are close to the canonical value of 1000 m\(^2\) s\(^{-1}\) [from both spreading of a sulphur hexafluoride patch (Ledwell et al. 1998) and the statistics of coreleased floats (Sundermeyer and Price 1998)].

A second reason for conservatism in estimating the effects of cabbeling is the dependance of downwelling rate on the squared isopycnal temperature gradient (see sec. 2), a quantity which is highly smoothed (over 7° of latitude and longitude) in the Levitus (1982) dataset. McDougall estimates that the smoothed temperature field may lead to underestimated of the downwelling velocity again by a factor of 2. Combined with the low constant value for \( K \), McDougall’s (1987a) downwelling rates due to cabbeling may therefore be underestimated by a factor of 4. However, different water masses of similar potential density may lie in close proximity [such as Mediterranean Water (MW) and Labrador Sea Water (LSW) in the northeast Atlantic] but mix laterally on quite separate neutral surfaces. In this situation, an assumption of along-isopycnal mixing would lead to considerable overestimation of the downwelling rate due to cabbeling. McDougall (1987a) shows that, since the isopycnal potential temperature gradient near the Mediterranean outflow is up to four times that on local neutral surfaces, a (conservatively estimated) downwelling rate of \( 2 \times 10^{-7} \) m s\(^{-1}\) may be overestimated to be \( 32 \times 10^{-7} \) m s\(^{-1}\).

The nonlinear equation of state also implies “extra” vertical motion on diapycnal (or dianeutral) mixing (McDougall 1989; McDougall and You 1990). While the downwelling associated with cabbeling is due to a physically distinct process, this “extra” vertical motion is actually an artifact of the oversimplified notion of density as a conservative property on diapycnal mixing. However, assuming a constant diapycnal diffusivity of \( 10^{-4} \) m\(^2\) s\(^{-1}\), McDougall (1989) shows that, across the neutral surface \( \gamma = 27.25 \) (mean pressure 650 db), this manifestation of nonlinearity effectively “weakens” the dianeutral velocity (due to diapycnal mixing) to a substantial degree (by up to \( 10^{-7} \) m s\(^{-1}\)) over large areas of the subtropical Atlantic and Indian Oceans. Averaging over different subtropical gyres of the World Ocean, McDougall (1989) further shows that, in vertical profile, this additional “downwelling” due to nonlinearity is confined to the upper ocean (above about 1500 m), reaching maxima (in different gyres) of \( 1–5 \times 10^{-7} \) m s\(^{-1}\) in the main thermocline. In contrast, the nonlinear effect is equivalent to weak “upwelling” in the deeper ocean, effectively doubling diapycnal diffusivity in the depth range 1500–4000 m (McDougall 1988).

The ratio of the sinking on isopycnal mixing (through cabbeling) to this extra vertical motion on diapycnal mixing scales as \( K \nabla \theta \cdot \nabla \theta/(D D') \), where \( K \) and \( D \) are isopycnal and diapycnal diffusivities respectively, \( \theta \) is potential temperature, and the subscript \( i \) denotes an isopycnal surface (T. McDougall 1998, personal communication). Taking values for \( K \) and \( D \) of \( 10^3 \) m\(^2\) s\(^{-1}\) and \( 10^{-4} \) m\(^2\) s\(^{-1}\), this ratio is therefore unity for an angle between neutral surfaces and \( \theta \) surfaces (\(- \nabla \theta \nabla \theta\) ) of \( \sim 3 \times 10^{-4} \) (a typical value in the upper ocean).

Nonlinearity of the equation of state may also lead to densification on mixed layer entrainment. Using a one-dimensional model to simulate the seasonal cycle of mixed layer heating and cooling, Zahariev and Garrett (1997) find that, only with a nonlinear equation of state, an annual-mean apparent buoyancy gain is compensated by a corresponding buoyancy loss (which they call cabbeling) through entrainment during winter and spring. Based on upper ocean data and surface forcing observed in the Mediterranean, the extra buoyancy gains/losses thus attributed to nonlinearity in the equation of state are equivalent to heating/cooling at a rate of \( -6 \) W m\(^{-2}\). Neglecting “entrainment cabbeling” (which compensates an apparent surface buoyancy flux), Fig. 1 schematically illustrates three specific mixing circumstances in which nonlinearity in the equation of state may contribute significantly to diapycnal/dianeutral advection (i) through isopycnal mixing across strong near-surface subpolar fronts, (ii) through diapycnal mixing in the main thermocline of subtropical gyres, and (iii) through epineutral mixing of different water masses (such as MW and LSW) in the deep ocean. The distinction between near-surface isopycnal mixing and deep epineu-
Cabbeling may be globally quantified by diagnosing dianeutral transports in GCMs. Hirst et al. (1996) find that, in a 1.6° latitude by 2.8° longitude Bryan–Cox GCM, the dianeutral transport resulting from isopycnal diffusion is mostly small (<0.05 Sv per degree latitude, zonally integrated along neutral surfaces) compared to that associated with vertical and horizontal diffusion. More recently, Hirst and McDougall (1998) show that larger effects of cabbeling (due to isoneutral tracer diffusion) are confined to the ventilated surface layer where, at least on zonal average, cabbeling competes with the very strong diapycnal mixing that can occur in the mixed layer and by direct air–sea interaction. The dianeutral transport due to cabbeling and thermobaricity reaches 0.4 Sv per degree latitude in the unventilated Southern Ocean at $\gamma = 27.4$. In the sense that these dianeutral fluxes are unidirectional (downward), latitudinal integration yields significant transports. Considering again the neutral surface $\gamma = 27.4$, total dianeutral transports of $\sim 1.2$ Sv and $\sim 3.5$ Sv are obtained in the zones 25°–50°N and 35°–50°S, respectively (T. Hirst, personal communication). Hirst and McDougall (1998) note that isoneutral tracer diffusion makes the “next most substantial contribution” to dianeutral transport (after vertical diffusion).

While implicitly represented in $z$-coordinate GCMs, cabbeling is optionally included as an explicit process in the more recent generation of layered, isopycnic-coordinate GCMs that employ density as a vertical coordinate and layer thickness as a prognostic variable (Bleck et al. 1989; Oberhuber 1993). The option of excluding cabbeling in isopycnic-coordinate GCMs raises the question: how serious is such exclusion? In the present study the cabbeling in such models is diagnosed, specifically the diapycnal velocities and volume fluxes associated with cabbeling in the $\sim 1°$ resolution Atlantic Isopycnic Model (AIM: New et al. 1995). AIM is based on the Miami Isopycnic-Coordinate Ocean Model (MICOM, as described by Bleck et al. 1992), in which potential density referenced to the surface ($\sigma_0$) is the vertical coordinate.

AIM was originally based on an early version (v1.3) of MICOM in which both potential temperature ($\theta$) and salinity ($S$) are advected and diffused in isopycnic layers, giving rise to “physical” cabbeling due to isopycnal mixing of heat and salt and “nonphysical” cabbeling due to divergences of advective heat and salt fluxes. In order to maintain prescribed layer densities, a mass flux across layer interfaces is invoked: the advective consequence of cabbeling. The cabbeling mass fluxes across layer interfaces of the spunup AIM are integrated “online.” These integrated fluxes are compared with mass fluxes diagnosed from the diapycnal velocities associated with physical cabbeling (see McDougall 1984) to
C�®¢rels and thermobaric parameters [de®ned in (3) below], pressure for neutral surface, \( h \), ef®cients, and \( a \) and \( b \) are the thermal expansion and haline contraction co®®bents, respectively. The present study addresses the weak salinity and pressure dependence of the vertical density ®ux divergence of \( w \), that takes fully into account the nonlinear dependence of \( \rho \) on potential temperature \( \theta \), salinity \( S \) and pressure \( p \), and the tendency for water parcels to move/mix along “neutral” surfaces [which do not exactly coincide with isopycnal surfaces (McDougall 1987b)]. McDougall (1991) thus relates a diapycnal velocity, \( w \), to parameterized diapycnal and isopycral mixing:

\[
\begin{align*}
\omega &= \left[ \frac{1}{\rho_0} \frac{\partial \rho}{\partial z} \right] = \left[ \frac{1}{\rho_0} \frac{\partial \rho}{\partial z} \right] + D \frac{\partial \tilde{a}}{\partial \theta} \tilde{\theta}^2 + \tilde{a}(c - 1) \frac{1}{h} \nabla_n \cdot (h K \nabla_n \theta) + \frac{\tilde{\beta}}{\beta} K \left\{ C_n \nabla_n \theta \cdot \nabla_n \theta + T_n \nabla_n \theta \cdot \nabla_n p \right\},
\end{align*}
\]

(1)

where \( D \) and \( K \) are diapycnal and isopycral eddy di®®susivities, \( \nabla_n \) denotes the lateral gradient operator in a neutral surface, \( h \) is the thickness of a layer bounded by “local” and reference neutral surfaces (at in situ and reference pressures, \( p \) and \( p_r \)), \( C_n \) and \( T_n \) are cabbeling and thermobaric parameters [de®ned in (3) below], \( \tilde{a} \) and \( \tilde{\beta} \) are the thermal expansion and haline contraction co®®bents, and \( c = [\alpha/\beta] [\tilde{a}/\tilde{\beta}]^{-1} \).

The left-hand term of (1) represents a material rate of change of \( \rho_0 \). On the right-hand side, term (A) is the ®®usive ®lux divergence of \( \rho_0 \) (vertical mixing of \( \rho_0 \), as if it was a conservative property), and term (B) is a “correction” that accounts for the effect of nonlinearities in the equation of state (approximated by ignoring the weak salinity and pressure dependence of \( \tilde{a} \)). Terms (A) and (B) together represent the “true” diapycnal ®lux due to turbulent mixing of nonconservative \( \rho_0 \) (see also McDougall 1989). Term (C) “corrects” the
of density change due to isopycnal mixing that is really happening on neutral surfaces. Terms (D1) and (D2) represent the cabbeling and thermobaricity that arise from purely isopycnal mixing. While the subject of this paper is term (D1) [briefly compared to (A)], the importance of “missing” term (C) is highlighted and discussed (see sec. 5c).

McDougall (1984) earlier used the conservation equations for potential temperature and salinity on an isopycnal surface to derive a generalized diapycnal velocity equation (now correctly written in terms of a vertical coordinate $z$, rather than the diapycnal orientation $d$—T. McDougall 1998, personal communication, and as summarized in appendix A of McDougall 1995):

$$(e - D_i)g^{-1}N^2 = K(a\nabla^2\theta - \beta\nabla^2S) + D(\alpha\nabla\theta - \beta\nabla S),$$

where the subscripts $i$ and $z$ denote differentiation with respect to isopycnal and vertical coordinates respectively, $e$ is a net diapycnal velocity (positive upwards), $g$ is gravitational acceleration, $N$ is the buoyancy frequency, and $F^S$ and $F^\theta$ are the double-diffusive fluxes of salt and potential temperature. Taking the isopycnal divergence of the identity $\beta \nabla S = \alpha \nabla \theta$ (which defines a neutral surface), McDougall (1984) expanded the first term on the right-hand side of (2) (representing isopycnal mixing of $\theta$ and $S$) as follows:

$${K(a\nabla^2\theta - \beta\nabla^2S) = -K[\nabla,\theta] \left[ \frac{\partial\alpha}{\partial\theta} + 2\frac{\partial\alpha}{\partial \beta} \frac{\partial S}{\partial \beta} - \frac{\alpha \beta}{\beta^2} \frac{\partial S}{\partial \beta} \right].}$$

The first and second terms on the right-hand side of (3) represent the effects of cabbeling and compressibility respectively. A diapycnal velocity due to cabbeling can therefore be expressed as

$$e_{\text{cab}} = -gN^{-1}K[\nabla,\theta] \left[ \frac{\partial\alpha}{\partial\theta} + 2\frac{\partial\alpha}{\partial \beta} \frac{\partial S}{\partial \beta} - \frac{\alpha \beta}{\beta^2} \frac{\partial S}{\partial \beta} \right].$$

The term $[\partial\alpha/\partial\theta + \cdots]$ is the “cabbeling parameter,” $C(\theta, S, p)$. The pressure dependence of $C(\theta, S, p)$ is rather weak (see the appendix of McDougall 1987a). A cabbeling parameter referenced to the surface (applicable to $\sigma_0$ versions of MICOM), $C(\theta, S, 0)$ [henceforth $C(\theta, S)$], can be fitted by the polynomial of McDougall (1987a):

$$C(\theta, S) = 0.136108 \times 10^{-4} - 0.325332 \times 10^{-6} \theta + 0.663168 \times 10^{-8} \theta^2 - 0.732603 \times 10^{-10} \theta^3 + (S - 35.0) \times \{-0.106105 \times 10^{-6} + 0.205719 \times 10^{-8} \theta\}.$$  

Figure 2 shows $C(\theta, S)$ contoured in $\theta$–$S$ space. Note that $C(\theta, S)$ is primarily a function of $\theta$ (through $\partial C/\partial \theta$) and that it varies by almost a factor of 2 over the range $-2^\circ C < \theta < 30^\circ C$. This implies that cabbeling becomes a more “efficient” process as temperature decreases.

3. The models

a. Model configuration and surface forcing

The basic configuration of the Atlantic Isopycnic Model, AIM, is as described in New et al. (1995). Only brief details are outlined here. The model grid is based on a rotated Mercator projection, ranging approximately from 15$^\circ$S to 82$^\circ$N, with a horizontal resolution of $1\degree$. The northern and southern boundaries are solid walls so that there is no advection or diffusion of mass or tracers in or out of the model domain. Based on AIM, the Quasi Global Isopycnic Model (QGIM: see Marsh et al. 2000) is implemented on a 1.25$\degree$ × 1.25$\degree$ grid, extending from Antarctica to 74$^\circ$N (thus avoiding numerical problems associated with divergence of meridians towards the North Pole). The northern boundary at 74$^\circ$N is represented by a solid vertical wall (as in AIM). The land mask has resolution 2.5$\degree$ latitude by 3.75$\degree$ longitude (the resolution of an atmospheric GCM to which the OGCM is to be coupled).

Beneath the mixed layer (index 1) 19 isopycnic layers (index 2–20) represent the ocean interior. In AIM, layer potential densities range from 25.65 (layer 2) to 27.75 (layer 16) in intervals of 0.15, and from 25.75 to 28.15 (layer 20) in intervals of 0.10. In QGIM, layer densities
are chosen in the range $24.70 < \sigma_0 < 28.13$, to represent the global range of water masses. Both model layer densities are listed in Table 1. All density values used in the present paper will be values of $\sigma_0$ (henceforth referred to as $\sigma$) and as such will be dimensionless.

Both models are initialized from the September values of the Levitus (1982) dataset, and, in a 30-yr spinup, are forced with monthly mean climatological fluxes [Hellerman and Rosenstein (1983) wind stress, Ebsen- sen and Kushnir (1981) heat fluxes and evaporation rate, Jaeger (1976) precipitation rate] plus relaxation fluxes to ensure that sea surface temperature and salinity remain close to monthly mean climatological values (Levitus 1982). The relaxation method used here is similar to that of Haney (1971), although a fixed relaxation coefficient, $\lambda$, is used. The surface heat and freshwater fluxes applied to the model, $Q$ and $E - P$, are formulated as

$$Q = Q_{obs} + \lambda (T_{obs} - T)$$

(6a)

$$E - P = (E - P)_{obs} + \frac{\lambda}{C_p} \left( \frac{S_{obs}}{S} - 1 \right),$$

(6b)

where, in the model, $Q$ is the net heat flux (the sum of radiative, latent and sensible fluxes of heat, taken to be positive into the ocean), $E - P$ is the net freshwater flux (evaporation minus precipitation), and $T$ and $S$ are mixed layer (surface) temperature and salinity; $Q_{obs}$, $(E - P)_{obs}$, $S_{obs}$ and $T_{obs}$ are corresponding climatological values; $\lambda$ is constant at $35$ W m$^{-2}$ K$^{-1}$, a value consistent with climatological estimates (Barnier et al. 1995). An equivalent relaxation coefficient for the freshwater flux is $\lambda(C_p \rho_0) \approx 8.5 \times 10^{-6}$ m s$^{-1}$ (given a specific heat of seawater $C_p = 4000$ J kg$^{-1}$ K$^{-1}$, a mean density for seawater, $\rho_0 = 1025$ kg m$^{-3}$). For a mixed layer water column of 50 m, these coefficients imply relaxation timescales of approximately 2 months.

In version 2.4 of MICOM (used in this study), the advective consequence of diapycnal mixing is implemented as a diapycnal velocity, $e_{i+1/2}$ (defined positive downwards), across the interface between layers $k$ and $k + 1$, defined by Hu (1996) as

$$e_{k+1/2} = \frac{K_k \frac{\delta \rho}{\delta z}_k - K_{k+1} \frac{\delta \rho}{\delta z}_{k+1}}{2(\rho_k - \rho_{k+1})},$$

(7)

where $(\delta \rho/\delta z)_k$ is the “local” stratification (at the mid-point of layer $k$) and $K_k$ is a diapycnal eddy diffusion coefficient, stratification dependent and parameterized as $K_k = c/\sigma$. The value of $c$ is a constant, the choice of which determines the strength of diapycnal mixing. For the version of AIM with diapycnal mixing, $c = 10^{-3}$ cm$^2$ s$^{-1}$, but this was decreased to $0.575 \times 10^{-3}$ cm$^2$ s$^{-1}$ for QGIM, values which, in the deep ocean, yield diapycnal diffusivities of $O(10^{-4})$ cm$^2$ s$^{-1}$ and diapycnal velocities of $O(10^{-7})$ m s$^{-1}$. Other model parameterizations are as outlined in New et al. (1995) and Marsh et al. (2000).

b. Density perturbation due to the independent advection/diffusion of $\theta$ and $S$ in isopycnic layers of MICOM

Consider $\theta_{i,j,k}^t$ and $S_{i,j,k}^t$, the time $t$ temperature and salinity of layer $k$ (potential density $\sigma_j$) at mass point $(i,j)$ on the horizontal grid, which satisfy $\sigma(\theta_{i,j,k}^t, S_{i,j,k}^t) = \sigma_j$. Over an advective/diffusive timestep $1D$, heat, and salt is advected and diffused between mass points $(i,j)$ and up to four adjacent (nonzero thickness) mass points: $(i-1,j)$, $(i,j-1)$, $(i+1,j)$ and $(i,j+1)$. The updated $(i,j,k)$ temperature and salinity $(\theta_{i,j,k}^{t+1}, S_{i,j,k}^{t+1})$ may result in the coordinate violation, $\sigma(\theta_{i,j,k}^{t+1}, S_{i,j,k}^{t+1}) \neq \sigma_j$. As a consequence of heat and salt diffusion, $\sigma(\theta_{i,j,k}^{t+1}, S_{i,j,k}^{t+1}) > \sigma_j$ (layer $k$ becomes too dense). However, the separate advection of heat and salt can also result in either $\sigma(\theta_{i,j,k}^{t+1}, S_{i,j,k}^{t+1}) < \sigma_j$ or $\sigma(\theta_{i,j,k}^{t+1}, S_{i,j,k}^{t+1}) < \sigma_j$ (layer $k$ becomes too light). To restore layer $k$ density to the prescribed value $(\sigma_j)$, a fraction of the mass at $(i,j,k)$ is redistributed vertically. Details of this “isopycnic restoration” are outlined in Appendix A of Bleck et al. (1992). A downward mass flux due to the density gain on isopycnic mixing (diffusion) is henceforth referred to as “physical” cabling, while any vertical transfer due to advection is noted as “unphysical” cabling (although this is, by definition, not cabling). The vertical mass flux due to combined effects of advection and diffusion is termed “model” cabling. It is shown below (sec. 5a) that (desirably) the majority of model cabling is “physical” (i.e., compared to diffusion, advection plays a minor role in perturbing layer density).
c. Model experiments

The four 30-yr MICOM experiments used in this study are summarized in Table 2. In version 1.3 of MICOM, temperature and salinity are advected and diffused separately (i.e., allowing cabbeling). In version 2.4 of MICOM, salinity is advected and diffused in isopycnic layers and temperature is determined from an inversion of the equation of state (i.e., cabbeling is disabled), while the Hu (1996) parameterization of diapycnal advection across interior layer interfaces is optional. The two 30-yr spinups AIMv1 and AIMv2 are each followed by a further year (year 31) of detailed “online” diagnosis. The volume fluxes due to cabbeling (in AIMv1) and diapycnal mixing (in AIMv2) are explicitly derived from annually-integrated vertical fluxes of mass across layer interfaces at each grid point.

The models have all been used in separate related studies: AIMv1 is described in New et al. (1995); AIMv2 was used to diagnose water mass formation rates from air–sea fluxes and mixing (Nurser et al. 1999), a study which motivated the present paper; AIMv2ND was used along with AIMv2 in an intercomparison exercise (Roberts et al. 1996; Marsh et al. 2000); QGIM was used to quantify water mass transformation in the Southern Ocean (Marsh et al. 2000).

It is worth remarking here on problems now recognized as inherent in the Hu (1996) diapycnal advection scheme. While the diapycnal advection parameterized by that scheme appears to play a reasonable and realistic role in water mass transformation (Nurser et al. 1999), grave concerns regarding the nonconservation of heat (and the advection-only—that is, nondiffusive—character of the scheme) motivated development of the more recent McDougall and Dewar (1998) mixing scheme. Recognizing this problem, and with emphasis here on the vertical motion (cabbeling) due to isopycnic mixing (in AIMv1), the vertical velocities and mass fluxes obtained with Hu’s scheme (in AIMv2) are only used here in a comparative exercise (see sec. 5a).

4. Diagnostics

a. The diapycnal velocity associated with “physical” cabbeling

Given the vertical discretization of MICOM, (4) is adapted to compute \( \varepsilon_{cab}^{k+1/2} \), a diapycnal velocity due to the “physical” cabbeling, across an interface \( k + \frac{1}{2} \) (the lower interface of layer \( k \)). This vertical velocity is now defined as positive downward, consistent with the sign convention for vertical mass fluxes in MICOM (see Bleck et al. 1992), and follows as

\[
\varepsilon_{cab}^{k+1/2} = gN^2 \frac{1}{2} \int \nabla \theta_k \cdot C(\theta_k, S_k),
\]

where the buoyancy frequency \( N = \sqrt{-(\rho/\rho_0)(\rho_{k+1} - \rho_k)/(H_{k+1} - H_k)} \) (\( \rho_0 \) is the mean density of seawater, \( \rho_k \) and \( H_k \) are the density and midpoint depth of layer \( k \)), \( K = u_L L \), where \( L \) is the length of a gridbox cell wall across which heat and salt are diffused, and \( u_s \) is a diffusion velocity constant [in the present MICOM implementations, \( u_s = 1 \) cm s\(^{-1}\), yielding typical values for \( K \) of \( 10^4 \) m\(^2\) s\(^{-1}\), the value chosen by McDougall (1987a)], and \( C(\theta_k, S_k) \) is computed using (5).

b. Water mass formation rates and diapycnal volume fluxes

Following the pioneering approach of Wallin (1982), and as outlined in Nurser et al. (1999), \( M(\sigma)\Delta \sigma \) is defined as the formation rate (Sv) of water with potential density between \( \sigma \) and \( \sigma + \Delta \sigma \), and \( G(\sigma) \) is a diapycnal volume flux (Sv) across \( \sigma \). A formation rate \( M(\sigma) \), with units of Sv (kg m\(^{-3}\))\(^{-1}\), is therefore the diapycnal divergence of \( G \):

\[
M(\sigma) = \frac{\partial G}{\partial \sigma}
\]

A positive value of \( G \) represents a diapycnal volume flux directed to higher \( \sigma \), and a positive value of \( M \) represents formation of a water mass with density \( \sigma \). For an incompressible ocean, \( G \) comprises terms representing three different diapycnal processes:

\[
G(\sigma) = G_{surf}(\sigma) + G_{diff}(\sigma) + G_{cab}(\sigma),
\]

where \( G_{surf} \) is due to surface density forcing, \( G_{diff} \) is due to turbulent diapycnal mixing (in MICOM this includes entrainment across the mixed layer base, lateral mixing in the mixed layer, and, if included, diapycnal mixing across isopycnic layer interfaces) and \( G_{cab} \) is due to cabbeling.

In this paper, the diapycnal volume flux due to cabbeling is obtained as follows. An annual-basin-mean volume flux across the lower interface of layer \( k \), \( \langle G_{cab}(\sigma_{k+1/2}) \rangle \), is obtained by area integrating \( \langle \varepsilon_{cab}^{k+1/2} \rangle \), the annual-mean interfacial diapycnal velocity associated with isopycnic restoration of layer \( k \) at each time step:

\[
\langle G_{cab}(\sigma_{k+1/2}) \rangle = \sum_{i,j} \langle \varepsilon_{cab}^{i,j+k+1/2} \rangle A_{i,j},
\]

where \( A_{i,j} \) is the surface area of each gridbox. With \( \varepsilon_{cab} \) defined as positive downward, \( G_{cab} > 0 \) (i.e., a volume flux towards higher density). If layer \( k \) spans a density range \( \Delta \sigma \), the layer \( k \) formation rate due to cabbeling

\[
M(\sigma)_{floor} = \int_{\sigma_{floor}}^{\sigma_{top}} M(\sigma) \, d\sigma
\]
is \( \langle M_{\text{cab}}(\sigma_x)\Delta \sigma_x \rangle \), and can be calculated from layer-to-layer divergences of \( \langle e^{\text{cab}}_{i,k-1/2} \rangle \):

\[
\langle M_{\text{cab}}(\sigma_x)\Delta \sigma_x \rangle = \sum_{i,j} \{ \langle e^{\text{cab}}_{i,k-1/2} \rangle - \langle e^{\text{cab}}_{i,k+1/2} \rangle \} A_{i,j}.
\]

A positive layer formation rate arises when there is a convergence of vertical velocities across the upper and lower interfaces of the layer. Annual diapycnal volume fluxes and formation rates due to interior diapycnal mixing can be likewise calculated by substituting \( e \) for \( e^{\text{cab}} \) in (11) and (12).

c. The basin-mean mass source and heat flux due to cabbeling

Over a given period, the basin-integrated result of cabbeling is a net gain in density. This can be equated to \( \mathcal{M} \), a basin-mean mass source (in kg s\(^{-1}\)):

\[
\mathcal{M} = \sum_{i,j} \{ \langle M_{\text{cab}}(\sigma_x)\Delta \sigma_x \rangle \Delta \sigma_x \},
\]

which, when multiplied by the specific heat of seawater and divided by a basin-mean thermal expansion coefficient (\( \overline{\alpha} \)) and basin area, is equivalent to \( Q^{\text{cab}} \), a basin-mean heat flux (W m\(^{-2}\)):

\[
Q^{\text{cab}} = \frac{C_p \mathcal{M}}{\overline{\alpha} \sum_{i,j} A_{i,j}}.
\]

where \( \overline{\alpha} \) is computed as

\[
\overline{\alpha} = \frac{\sum_{i,j} \langle \alpha_{i,j} \rangle(h_{i,j})A_{i,j}}{\sum_{i,j} (h_{i,j})A_{i,j}}.
\]

Of course, if density is assumed to be a conservative property (which it is not!), diapycnal mixing conserves density and heat.

5. Results

a. Cabbeling in AIM

As an example of how diapycnal mixing compares geographically with cabbeling, Fig. 3 shows year 31 fields of \( \langle e^{\text{cab}} \rangle \) and \( \langle e \rangle \) (computed from integrated mass fluxes in AIMv1 and AIMv2 respectively) across the lower interface of layer 8 (for which \( \sigma = 26.55 \), typical of the subtropical thermocline). Note that, in general, \( e < 0 \) (upwelling), while \( e^{\text{cab}} > 0 \) (downwelling). The limited occurrence of unrealistic \( e > 0 \) is a consequence of using a stratification-dependent diapycnal mixing coefficient, where layer 8 is thin and lies directly beneath a thicker mixed layer. Diapycnal mixing is far more widespread than cabbeling, with associated upwelling rates of typically 2–10 m yr\(^{-1}\). However, strong cabbeling does arise in a small region of the northwest Atlantic, where associated downwelling rates exceed 50 m yr\(^{-1}\).

Differences between \( \langle e^{\text{cab}} \rangle \) at the upper and lower interfaces of a layer result in an annual change of that layer thickness due to cabbeling. Figure 4 shows fields of this change in thickness (in AIMv1), for layers 7–10 in the northwest Atlantic. The effect of cabbeling is to thicken layers 9–10 and to thin layers 7–8, by as much as \( \pm 100 \) m. Area-integrated over the basin, the fields of layer thickness change yield the layer 2–20 formation rates shown in Fig. 5. For layers 7–10, cabbeling drives annual-basin-mean layer formation rates of around \( \pm 0.5 \) Sv (Sv = \( 10^6 \) m\(^3\) s\(^{-1}\)), likewise for layers 15–19 (strong cabbeling in these denser layers is found farther north, in the subpolar gyre).

The seasonality of model cabbeling (again in AIMv1) is shown by the monthly basin-integrated layer formation rates in Fig. 6. Note that the strongest formation (up to 1.2 Sv for layer 9 water of \( \sigma = 26.70 \)) and destruction (up to \( -1.6 \) Sv for layer 7 water of \( \sigma = 26.40 \)) of intermediate density layers takes place in June, when freshly ventilated layers (hence strong \( \theta-S \) con-
Fig. 4. Year 31 northwest Atlantic change in layer thickness (m, contoured at 2, 5, 10, 20, 50 m) in AIMv1 due to model cabbeling, for layers 7–10, (a)–(d) respectively.

The strong cabbeling-related formation of layer 9 in June is associated with large lateral gradients of $\theta$ and $S$ in the overlying layer 8 (Figs. 7a,b). Monthly AIMv1 fields of $\theta$ and $S$ (such as in Figs. 7a,b) are used with (8) to compute annual-mean diapycnal velocities at layer interfaces. Downwelling rates at the base of layer 8 in June (Fig. 7c) reach 100 m yr$^{-1}$. This “physical” cabbeling (driven by isopycnal mixing) leads to annual changes in layer thickness, obtained as layer-to-layer divergences of downwelling fields (such as Fig. 7c), and shown in Fig. 8 (for the same layers as Fig. 4). Discrepancies between Fig. 4 and Fig. 8 (aside from undersampling errors in the latter) can be attributed to the
independent advection of temperature and salinity (see sec. 3b). The supplementary cabling which arises from independent advective fluxes of heat and salt between gridboxes can be considered as “unphysical.” In the northwest Atlantic, the model North Atlantic Current (NAC) strongly advects heat and salt to the east and unphysical cabling is strong. As a consequence, model cabling (Fig. 4) is underpredicted by physical cabling (Fig. 8), and there is a tendency for strong cabling to extend eastward (along the NAC) in the model (e.g., compare Fig. 4c and Fig. 8c).

To quantify the relative roles of cabling (model and physical) and diapycnal mixing, respective diapycnal volume fluxes are obtained [see (11)]. Figure 9 shows the annual-basin mean diapycnal volume fluxes attributed to model cabling and diagnosed physical cabling (in AIMv1) and to diapycnal mixing (in AIMv2). Note in Fig. 9 the peaks of about 1.3 Sv in model cabling fluxes at subtropical and subpolar mode water densities (the curve returns to zero at high and low density limits, reflecting a conservation of total volume).

**Fig. 7.** June year 31 AIMv1 layer 8 fields: (a) $\theta$, (b) $S$, and (c) diagnosed “physical” $\varepsilon^{\Phi}$ (m yr$^{-1}$).

**Fig. 8.** As Fig. 4 but due to diagnosed physical cabling in AIMv1.
The volume fluxes based on model and physical cabbeling agree quite well at low and intermediate density [at high density, physical cabbeling “overpredicts” the model cabbeling by a factor of ~3, probably due to some ambiguity in the definition of $N_{i+1/2}^2$—see (8)]. The vertical grey bar represents the lightest (isopycnic) layer density (layer 2, for which $\sigma = 25.65$). There is no isopycnal mixing, and hence no cabbeling, at $\sigma < 25.65$. The effect of diapycnal mixing is to transform water towards lighter density at $\sigma > 25.65$, and vice versa at $\sigma < 25.65$, essentially mixing interior water with mixed layer water. The volume flux associated with diapycnal mixing reaches $-8$ Sv at $\sigma = 26.5$. At intermediate and high densities, cabbeling (as implemented in AIMv1) would therefore weaken the effectiveness of diapycnal mixing (transforming such waters towards lower densities in AIMv2) by up to ~20%. While diapycnal mixing (of “conservative” density) does not involve a net mass source, the net effect of cabbeling (both model and physical) in the model Atlantic is an apparent mass gain, $M \approx 2 \times 10^6$ kg s$^{-1}$ [see (13)], equivalent to a cooling rate, $Q_{\text{cab}} \approx 0.6$ W m$^{-2}$ [see (14)]. Compared to the climatology used here to force AIM, this cooling rate (due to cabbeling) is about 10% of the annual-mean net cooling due to air–sea fluxes over the North Atlantic.

These results suggest that cabbeling may play a significant role in the thermohaline circulation, a conclusion not reached by Hirst and McDougall (1998). Consider, therefore, a comparison of the meridional overturning as a function of potential density, $\psi(\Theta, \sigma)$, in year 30 of (a) AIMv1 and (b) AIMv2ND and (c) the “AIMv1 minus AIMv2ND” difference [solid/dashed lines indicate year-round maximum mixed layer density at a given latitude; black masking applies at densities lower/higher than minimum/maximum values at each latitude].

Note also the increase in maximum surface (mixed layer) density in the equatorial region from $\sigma = 24$ in AIMv2ND to $\sigma = 25$ in AIMv1, associated with enhanced upwelling (although this difference is not obviously related to cabbeling). The cabbeling associated with the subpolar front is apparent as a small 1.0 Sv cell centred on $\sigma = 26.6$ and spanning 25$^\circ$–45$^\circ$N. Finally, Fig. 11 shows the year 0–30 basin-mean layer thickness trends in AIMv1 [Fig. 11a, computed as in New et al. 1995], and the differences from AIMv2ND [Fig. 11b, AIMv1 minus AIMv2ND]. It is now clear that the greatest large-scale impact of cabbeling is at high density (layers 16–19). Big differences between AIMv1 and AIMv2ND (increases/decreases in basin-
mean layer 18/17 thickness of ±25 m) have developed after only 5 years. While the models subsequently equilibrate to a large extent, differences (notably in layer 19 thickness) are still growing at the end of the 30-yr spinup.

b. The physical cabbeling implied by $\theta$ and $S$ fields in the Southern Ocean of QGIM

There is evidence that Antarctic Intermediate Water (AAIW) is strongly modified by isopycnal mixing across the Polar Front (Gordon et al. 1977). The melting of sea ice around Antarctica fluxes freshwater into the surface layer, leading to low surface salinities where the winter ice has retreated. In QGIM this seasonal freshwater influx is not explicitly represented, but rather is mimicked by the relaxation of surface salinity towards the monthly climatological values of Levitus (1982). Figures 12a,b show fields of layer 9 ($\sigma = 27.22$, typical of AAIW) temperature and salinity in March, when the sea ice around Antarctica is at a maximum retreat and surface salinities reach seasonally minimum values. Note in particular the coincidence of low temperatures and salinity values south and west of Drake Passage. Locally strong meridional gradients of $\theta$ and $S$ crudely represent a model Polar Front, surface-expressed by the $2^\circ$C isotherm in Drake Passage (Park et al. 1993). Figure 12c shows the annual-mean vertical velocity due to cabbeling across the lower interface of layer 9 implied by monthly $\theta$ and $S$ fields. Figure 12d shows the resulting annual change in layer thickness. Corresponding with the strongest lateral $\theta$ and $S$ gradients (Figs. 12a,b), cabbeling is strongest in and around Drake Passage. Cabbeling-related vertical velocities of $O(10$ m yr$^{-1}$)
Fig. 12. March year 30 Southern Hemisphere QGIM layer 9 ($\sigma_0 = 27.22$) fields of: (a) $\theta$ (shaded where $\theta < 0^\circ$C), (b) $S$ (shaded where $S < 33.9$ psu), (c) implied annual-mean $\epsilon^{ab}$ (m yr$^{-1}$) across lower interface of layer 9, and (d) annual change in thickness due to the implied cabbeling (shaded where layer is thinning at rates greater than 2 m yr$^{-1}$).
Fig. 13. Year 30 annual-mean layer formation rates (Sv) due to implied cabbeling in the Southern Ocean of QGIM.

Fig. 14. The diapycnal volume flux (Sv) due to implied physical cabbeling in the Southern Ocean of QGIM (dashed, consistent with Fig. 9).

compare to local annual-mean Ekman pumping rates of similar magnitude, and imply strongest annual thickness changes of \( \approx 50 \text{ m yr}^{-1} \), compared to total thickness (in Drake Passage) of up to 400 m (Marsh et al. 2000). By area integrating fields of cabbeling-implied thickness change south of \( 25^\circ \text{S} \), layer formation rates are obtained (Fig. 13). Compared to the North Atlantic (Fig. 5), stronger formation rates are found in the more expansive Southern Ocean, reaching \( \pm 2-3 \text{ Sv} \) for layers 9–16. The corresponding diapycnal volume flux (Fig. 14) reaches \( 7 \text{ Sv at } \sigma = 27.4 \), with a net mass flux of \( M \approx 5 \times 10^6 \text{ kg s}^{-1} \), equivalent to a cooling rate, \( Q_{\text{cabc}} \approx 2.3 \text{ W m}^{-2} \).

c. Cabbeling in MICOM and the problem of using \( \sigma_0 \) as a layer variable

Certain water masses share essentially the same potential density, but, displaying quite different \( \theta-S \) characteristics, otherwise occupy very distinct and separate neutral surfaces (McDougall 1987b). In AIM and QGIM, initial fields of layer thickness, temperature and salinity are obtained by vertically-interpolating Levitus (1982) fields of \( \theta \) and \( S \) (New et al. 1995). An unrealistic outcome of this interpolation is the close proximity of different water masses (characterized by \( \theta-S \) relations) in layers of the same prescribed potential density, with implications for the inclusion/exclusion of cabbeling in MICOM.

This problem notably arises in the representation of Mediterranean Water (MW) and Labrador Sea Water (LSW) in the initial state of AIM. In the northeast Atlantic, these two water masses share a similar potential density (\( \sigma_0 \approx 27.75 \)), and hence both occupy layer 16. Figure 15 shows the initial (September) potential temperature and salinity of layer 16 (Figs. 15a,b) and the diapycnal velocity at the base of layer 16 that would arise as a consequence of isopycnal mixing as parameterized in AIM (Fig. 15c). Cold (\( \theta < 4^\circ \text{C} \)) and fresh (\( S < 35.0 \text{ psu} \)) LSW spreads from the northwest, while warm (\( \theta > 10^\circ \text{C} \)) and salty (\( S > 36.0 \text{ psu} \)) MW spreads from the Strait of Gibraltar. The resulting strong gradients in \( \theta \) and \( S \) due west of Iberia, coincident with weak stratification, imply diapycnal velocities of over 250 m yr\(^{-1} \). This downwelling is \( O(100) \) times larger than an upwelling rate of \( 10^{-7} \text{ m s}^{-1} (\approx 3 \text{ m yr}^{-1} ) \) necessary to balance the global formation of bottom water at an estimated rate of 30 Sv (Munk and Wunsch 1998).

To illustrate the way in which cabbeling influences the equilibration of model masses during spinup, Fig. 16 shows the annual-mean layer formation rates due to physical cabbeling over the 30-yr integration (diagnosed from monthly fields of layer thickness, \( \theta \) and \( S \), as model cabbeling mass fluxes are only saved in year 31). As a consequence of the initial close proximity of different water masses in the same isopycnic layers, there is very strong cabbeling during years 1–5, especially in dense layers (\( \sigma > 27.5 \)), but by year 15 isopycnal mixing has mixed away the initially strong gradients in \( \theta \) and \( S \), and relatively steady cabbeling formation rates are established. With no Gibraltar inflow in AIM, some of the MW (present in the initial state) is thus transformed to higher density over the 30 yr spinup. Evidence for MW consumption through cabbeling can be seen in Fig. 17, which shows fields of \( \text{AIMv1 minus AIMv2ND} \) differences in layer 16 thickness after 1, 5, and 10 years of spinup. In the region of MW outflow, layer 16 thickness—typically 300 m in the initialization (see New et al. 1995)—decreases by over 100 m (west of Ireland) in the first 5 years in AIMv1 (due to cabbeling), compared to AIMv2ND (with no cabbeling). Negative layer 16 thickness anomalies appear to form on the northern flank of MW outflow in the very initial stages of spinup.
that AABW reaches higher in situ densities than NADW. Once in the same geographical vicinity, NADW and AABW mix preferentially along quite separate neutral surfaces. NADW and AABW thus preserve quite distinct \( \theta-S \) properties far from their respective formation regions.

While AABW does not form in QGIM (due to poor representation of the formation mechanism), a signature of AABW persists in layer 15 (\( \sigma = 27.88 \)) around Antarctica throughout the 30-yr spinup of QGIM—a “remnant” of the initialization procedure. Figure 18 shows the Atlantic sector fields of potential temperature and salinity for layer 15 as initialized in QGIM. North of about 30\(^\circ\)S, layer 15 is dominated by NADW, with \( \theta > 1.0^\circ\text{C} \) and \( S > 34.8 \) psu everywhere (except in the cold, fresh East Greenland Current), largely homogenized in the fairly narrow ranges 2.0 < \( \theta < 3.0^\circ\text{C} \) and 34.85 < \( S < 35.00 \) psu. The influence of Gibraltar outflow is evident where \( \theta \) and \( S \) exceed 3.0\(^\circ\text{C} \) and 35.0 psu. South of ~30\(^\circ\)S, \( \theta < 1.0^\circ\text{C} \) and \( S < 34.8 \) psu (gray-shaded in Fig. 18), and layer 15 is dominated by AABW properties, homogenized in the ranges -0.5 < \( \theta < 0.5^\circ\text{C} \) and 34.65 < \( S < 34.70 \) psu. In the narrow subtropical zone between these two regions, \( \theta \) and \( S \) “on” the isopycnal \( \sigma = 27.88 \) can increase, rather unrealistically, by 2\(^\circ\text{C} \) and 0.15 psu over just 5\(^\circ\) of latitude. The associated isopycnal mixing of both heat and salt across this zone (not allowed in QGIM) implies strong downwelling (not shown) and unrealistic inflation of denser layers (namely layer 16).

6. Discussion

The cabbeling that arises from isopycnal mixing may play a significant role in the global volumetric balance
of water masses. A realistically configured and forced isopycnic model of the North Atlantic (AIMv1), based on the Miami Isopycnic Coordinate Model (MICOM), has been used to quantify the diapycnal volume fluxes due to cabling. In the model layer density range $26.65 < \sigma_0 < 28.15$, cabbeling drives a positive (light to heavy) volume flux of typically 1 Sv. The annual-basin mean density gain which results from cabbeling is around 10% of that due to net effect of cooling and evaporation over the model North Atlantic. At intermediate to high densities ($\sigma_0 > 26.65$), the mass flux due to cabling is $\sim 20\%$ of, and opposite to, that due

Fig. 17. Fields of differences in layer 16 thickness between AIMv1 and AIMv2ND (AIMv1 minus AIMv2ND): (a) after 1 year, (b) after 5 years, and (c) after 10 years.

Fig. 18. Initial QGIM layer 15 ($\sigma_0 = 27.88$) fields in the Atlantic sector of (a) $\theta$ and (b) $S$ (shaded where $\theta < 1^\circ{C}$, $S < 34.8$ psu).
to diapycnal turbulent mixing (as parameterized in a later version of the model, AIMv2). Cabbeling tends to be highly localized and seasonal in the upper thermocline. In AIMv1, cabbeling is strongest at the northwest Atlantic subpolar front in summer, where isopycnal temperature gradients are strong. Fluxes associated with “model cabbeling” generally compare favorably with those due to the “physical cabbeling” predicted from model fields of layer thickness, $\theta$ and $S$.

Strong cabbeling in AIMv1 is also shown to transform initially present Mediterranean Water (MW) to higher density through the isopycnal mixing of cold, fresh LSW and warm, salty MW during the early stages of spinup. Recent Atlantic basin experiments with MICOM v2 (without this cabbeling) have included relaxation in various “buffer zones” (Chassignet et al., 1996; DYNAMO Group, 1997). In particular, the Mediterranean Sea is typically excluded, and $T$–$S$ properties are restored to Levitus (1982) values in a buffer zone just outside the Gibraltar Straits. If incorporated as in MICOM v1, cabbeling would presumably compete with, and to some extent compromise, such a relaxation scheme.

However, the cabbeling associated with isopycnal mixing in these relatively deep layers of AIM is unrealistic. While LSW and MW share the same value of $\sigma_0$, observations of conserved properties (such as salinity) suggest that these quite distinct water masses mix most effectively along neutral surfaces (McDougall, 1987b), with associated cabbeling velocities which are estimated 4–16 weaker than those associated with isopycnal mixing (McDougall, 1987a). Clearly then, the isopycnal mixing in $\sigma_0$ versions of MICOM drives too much cabbeling in the deep ocean, where large discrepancies arise between isopycnal and epineutral slopes.

This problem can now be alleviated by the choice of a MICOM layer coordinate which better describes stratification and along-layer mixing in the deep ocean. In particular, Sun (1997) has developed a $\sigma_2$ version of MICOM in which AABW is correctly represented in the South Atlantic as a northward-spreading layer which lies beneath a NADW layer. The recent replacement of potential density with “virtual potential density” allows for the incorporation of thermobaric effects in the horizontal pressure gradient (Sun et al., 1999). A reference level chosen to correspond with pressures on isopycnal surfaces representing LSW and MW in the region of MW outflow (1000–2000 db) would minimize the error in mixing at that level (although errors may then appear higher in the water column, with adverse consequences for near-surface cabbeling elsewhere, such as at the subpolar front). While the choice of a “better” coordinate surface in layered models is governed by the twin desires, (i) for along-layer mixing as neutral as possible and (ii) to evaluate horizontal pressure gradients in an accurate way, more realistic deep cabbeling in, say, a $\sigma_2$ version of MICOM (which allowed cabbeling) would be a desirable bonus.

By examining the cabbeling in AIMv1, and inferring the cabbeling that is “missing” in QGIM (a global model based on MICOM version 2.4), the importance of cabbeling in different regions of the World Ocean is quantified as an equivalent heat flux. In AIMv1, the density gain associated with cabbeling is equivalent to a basin-mean cooling (over the Atlantic) of 0.6 W m$^{-2}$, while the $\theta$, $S$, and thickness fields south of 25°S in QGIM imply regional-mean cooling of 2.3 W m$^{-2}$ (over the Southern Ocean). In particular, implied cabbeling is strong in the midlatitude Southern Ocean, driving a diapycnal volume flux (towards higher density) of up to 7 Sv at the density of lower intermediate water ($\sigma_0 = 27.4$). Marsh et al. (2000) showed that, in QGIM, water masses in the region south of 42.5°S are surface-transformed by a combination of cooling and freshening. Regionally averaged, the annual-mean cooling is 12.5 W m$^{-2}$, while the freshwater forcing is equivalent to warming at a rate of 24.3 W m$^{-2}$. Net buoyancy forcing over the region is thus equivalent to a fairly weak warming of 11.8 W m$^{-2}$. In the intermediate density range over which cabbeling-implied cooling is strongest, surface heat and freshwater fluxes are in an even closer balance (almost cancelling each other). Given this rather weak surface forcing, cabbeling (if included) might significantly contribute to the transformation of intermediate water masses in the Southern Ocean of QGIM.

Differences in meridional overturning between AIMv1 (cabbeling) and AIMv2ND (no cabbeling) suggest that cabbeling enhances the North Atlantic meridional circulation by up to $\sim$1 Sv at high density. The diapycnal volume flux implied by isopycnal mixing in the Southern Ocean of QGIM (up to 7 Sv) suggests larger contributions of cabbeling to the global overturning. These “transports” somewhat exceed the unventilated dianeutral transports associated with isoneutral tracer diffusion in the 1.6° latitude by 2.8° longitude GFDL-based model of Hirst and McDougall (1998). With isoneutral tracer diffusion in the GFDL model parameterized by $K = 10^3$ m$^2$ s$^{-1}$, more or less as in AIM and QGIM, larger diapycnal transports (on isopycnal mixing) in MICOM may reflect that stronger isopycnal gradients are better preserved in MICOM (which does not suffer spurious horizontal diffusion), although the GFDL cabbeling is diagnosed for isoneutral mixing (generally somewhat weaker than isopycnal mixing). Cabbeling in the GFDL model may also be offset by the generally upward dianeutral transport due to unphysical “slope clipping” [employed in the tracer diffusion scheme (Cox, 1987)], which sets an upper limit on the neutral surface of 1:100 (Hirst and McDougall, 1998). Steeper slopes (i.e., stronger epineutral mixing and more cabbeling) would otherwise arise in frontal zones and near the Mediterranean outflow of the GFDL model ocean (T. Hirst, personal communication).

In conclusion, cabbeling may yet prove to be a very necessary ingredient in MICOM. On long timescales the balance of intermediate and dense layers determines
the strength of the meridional overturning. Cabbeling plays a significant role in the transformation of the water masses represented by these layers. It is the author’s opinion that this justifies the computational cost of restoring to MICOM the independent isopycnal advection and diffusion of heat and salt.

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