Poleward Buoyancy Transport in the Ocean and Mesoscale Eddies

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ABSTRACT

There are many dynamic similarities between mesoscale eddies in the ocean, and cyclones and anticyclones in the earth's atmosphere. Observational data, however, are still not adequate to explore this analogy in detail. In the present study a new eddy-resolving ocean circulation model, which includes both wind-driving and buoyancy-driving, is used to determine whether mesoscale eddies play a role in poleward buoyancy transport in any way comparable to the role of synoptic scale motions in transporting heat in the atmosphere. Within an Eulerian reference frame, mesoscale eddies transport buoyancy poleward through two mechanisms. One involves the correlation of time-dependent fluctuations of horizontal velocity and buoyancy. The other transport mechanism involves wave-driven cells in the meridional plane. These cells are analogous to the Ferrel cell in the atmosphere, except that the geometry of the ocean basin allows them to be geostrophically balanced. In an eddy-resolving model of ocean circulation, the two mechanisms for buoyancy transport are almost perfectly compensating. Within a Lagrangian framework, the trajectories of the eddies are largely excursions on isopycnal surfaces. Heat transport may take place by eddies in the renal ocean without eddy buoyancy transport, since temperature gradients always exist on isopycnal surfaces and may be quite strong in polar regions. Mesoscale eddies and the thermohaline circulation in the model can be weakly coupled, because available potential energy created by the large-scale wind stirring provides a primary energy source for baroclinic instability. The model results indicate that the actual measurement of mesoscale eddy transports is extremely difficult, since it involves an accurate determination of the difference between transport by wave-driven, mean flows and by the correlation of the time-dependent fields.

1. Introduction

The exploration and description of the mesoscale ocean eddy field has been a major achievement of oceanography over the last two decades. Extensive measurements from ships of opportunity, moored platforms and drifting buoys provide a data base which will soon allow a first assessment of the role of mesoscale eddies in the general circulation of the ocean. An outstanding technical achievement has been the mapping of mesoscale eddies from space, using the radar altimeter, infrared measurements of sea surface temperature, and estimates of surface color. As a description of mesoscale eddies has emerged, it has been natural to compare meanders of the Gulf Stream and Kuroshio to atmospheric disturbances associated with the polar front. Synoptic disturbances in the troposphere appeared to be a natural analogue for mesoscale eddies in spite of the great differences in horizontal scale.

The fundamental role of synoptic-scale disturbances in both the heat and momentum balance of the atmosphere is well known. The poleward transport of momentum by atmospheric cyclones was first predicted by Jeffreys (1926) and later confirmed in observational studies by Starr (1948) and others. Several studies have attempted to confirm aspects of the analogue between mesoscale and eddies and synoptic disturbances in the atmosphere. Newton (1961) estimated the amount of heat transferred poleward by eddies in the Gulf Stream. He concluded from observations available to him at the time that rings breaking off from the stream would result in the net transfer of about 0.25 petawatts across the stream toward the cold northern wall. Bryden (1979) analyzed current meter measurements in the Drake Passage and found that the Circumpolar Current contained baroclinically unstable eddies transferring heat poleward. On the basis of his measurements, he suggested that a large fraction of the net surface heat loss of the Southern Ocean poleward of the Circumpolar Current could be balanced by the poleward flux of heat by mesoscale disturbances. DeSzoeke and Levine (1981) made a study of the poleward heat transport by geostrophic currents in the Southern Ocean. Combining geostrophically balanced and wind-induced Ekman currents in their poleward heat transport estimate, the authors compared their estimate with surface heat balance requirements for the sector poleward of the Circumpolar Current. They found a significant discrepancy, amounting to 0.5 plus or minus 0.3 PW, and concluded that poleward heat transport by mesoscale eddies was the most likely cause of the lack of balance in their heat budget calculations. The existence of an eddy heat flux signature in the Drake Passage has recently been confirmed from an extensive five-year data set by Nowlin et al. (1985).

Recently, long term records from current meter data in the North Atlantic have become available which permit estimates of eddy heat transport by transient motion. Bryden (1982) has analyzed data from the Local Dynamics Experiment (31°N, 69°W), the recirculation region of the Gulf Stream, and found an eddy transport by transient motions towards the equator. Very few measurements exist that shed light on the contribution of mesoscale eddies to the momentum balance of major currents. Webster (1961) analyzed momentum transfer by eddies in the Gulf Stream off Cape Hatteras and in the Florida Straits. He concluded that the net effect of the eddies was to enhance the mean flow by transferring downstream momentum into the stream. Nishida and White (1982) have used the TRANSPAC dataset to study the momentum balance of the Kuroshio Extension. Consistent with Webster's results off Hatteras, they found that transient eddies tended to increase the energy of the mean flow in the regions both to the east and west of the Shasky Rise. However, they also found that this tendency was somewhat compensated by standing eddies associated with bottom topography. We now have available numerical experiments which include both buoyancy generation processes and mesoscale eddies (Semtner and Mintz, 1977; Cox, 1985). These numerical experiments do not consider salinity effects, so heat and buoyancy are equivalent. The two experiments show a similar behavior in the poleward heat (actually buoyancy) transport integrals. Cox (1985) compares the heat transport of an eddy-resolving model to a similar model of lower resolution without eddies. The additional heat transport by transient disturbances is almost exactly compensated by a corresponding change in the heat transport due to the mean flow. The same compensation is evident in an analysis by Mintz (1979) of the Semtner and Mintz model. On the other hand, observations do not show a similar compensation between eddy heat transport by transient motion and mean flow in the troposphere. There is no convincing evidence that the compensation obtained by Semtner and Mintz (1977) and Cox (1985) is a purely model-dependent result, since the numerical models of the ocean are formulated in much the same way as atmospheric models that successfully simulate the poleward heat transport in the troposphere. Why are baroclinic disturbances in the ocean models and atmospheric models not transporting heat and buoyancy in the same way? Gill (1983) examined the results of the Semtner and Mintz (1977) model and concluded that the question of the role of mesoscale eddies in poleward heat transport was still open. This paper attempts to look at the question in more detail using the more complete calculations of Cox (1985).

A classical debate in oceanography involves the relative contribution of wind-forcing and buoyancy-forcing to the maintenance of the general circulation. In the troposphere, air parcels lose their original air mass properties rather quickly, so that the whole concept of a distinct air mass has gradually been abandoned. In the oceans, properties are much better conserved along trajectories, allowing the concept of water masses to become the practical basis for descriptive oceanography. Generally, nonadiabatic processes in the ocean appear to have a much longer time scale than mesoscale eddies, with the possible exception of wintertime convection in warm core eddies. Another factor which clearly differentiates the dynamics of the atmosphere and the oceans is the existence of coastal boundaries and midocean ridges. These boundaries are regions in which a large percentage of the mesoscale energy is localized. Certainly the existence of these barriers must have an important effect on the global energy balance of the oceans. The fact that the thermal driving of the ocean is weak compared to the atmosphere does not imply that the dissipation of kinetic energy is low relative to other processes. In fact, the kinetic energy of large-scale motions in the ocean appears to be recycled orders of magnitude faster than available potential energy. Clearly it is the weak thermal driving in the ocean which most clearly differentiates geostrophic turbulence in the ocean and the atmosphere. Work done by wind on the ocean creates a source of available potential energy that is capable of sustaining baroclinic instability without any direct link to the thermohaline circulation. The objective of this study is to determine what insight can be gained from numerical experiments, since observational datasets in the ocean are still too scanty to provide reliable second-order statistics that would allow a diagnostic comparison of mesoscale eddies in the ocean and synoptic disturbances in the atmosphere.

2. Eulerian mean and generalized Lagrangian mean flow

There are important differences between an ocean circulation confined to a basin bounded by walls to the east and west and a stratospheric model in an open channel. However, in both cases a zonal average is a useful tool if we wish to investigate poleward heat transport. Let angle brackets denote a zonal average and θ the potential temperature. If Q represents nonadiabatic effects, a steady state balance is given by

$$\langle \mathbf{v} \cdot \nabla \theta \rangle = \langle Q \rangle. \tag{2.1}$$

If we let $\overline{(\ )}$ represent the time-average and $(\ )'$ be the deviation from the time-average, (2.1) can be rewritten in the following form:

$$\langle \bar{\mathbf{v}} \cdot \nabla \bar{\theta} + \overline{\mathbf{v}' \cdot \nabla \theta'} \rangle = \langle \bar{Q} \rangle. \tag{2.2}$$

A Lagrangian average is taken along the trajectory of a moving particle. In addition, we can define a generalized Lagrangian average for an ensemble of paths sampled over a span of longitude and time (Dunkerton, 1980). In this case (2.2) can be written as

$$\bar{\mathbf{v}}^{\mathrm{L}} \cdot \nabla \bar{\theta}^{\mathrm{L}} = \langle \bar{Q} \rangle \tag{2.3}$$

where $\bar{\mathbf{v}}^{\mathrm{L}}$ is the Lagrangian mean velocity, and $\bar{\theta}^{\mathrm{L}}$ is the Lagrangian mean potential temperature. Note that no cancellation can take place in this case, so that a weak forcing implied by small Q requires that the component $\bar{\mathbf{v}}^{\mathrm{L}}$ normal to isothermal surfaces must also be small. By definition,

$$\bar{\mathbf{v}}^{\mathrm{L}} = \langle \bar{\mathbf{v}} \rangle + \bar{\mathbf{v}}^{\mathrm{s}}. \tag{2.4}$$

Here $\langle \bar{\mathbf{v}} \rangle$ is the Eulerian mean with respect to longitude and time, and $\bar{\mathbf{v}}^{\mathrm{s}}$ is the Stokes drift. While the component of $\bar{\mathbf{v}}^{\mathrm{L}}$ normal to isothermal surfaces must be small for the weak forcing case, the normal component of the Eulerian mean may not be. Thus we can expect that in an ocean circulation largely driven by wind-forcing and only weakly driven by heating, transport associated with the two terms on the right-hand side of (2.4) will largely cancel. The same type of balance does not necessarily apply to potential vorticity, however. As we noted in the introduction, mechanical dissipation can be relatively important in the ocean. Thus the weak forcing for the thermal field does not imply that the potential vorticity field is also weakly forced, particularly in the vicinity of boundaries.

3. Results from numerical models

A pioneering investigation of an eddy-resolving model that included thermodynamic processes was carried out by Semtner and Mintz (1977). The geometry of their model is shown in Fig. 1. To limit the scope of the calculation, their basin was chosen to be

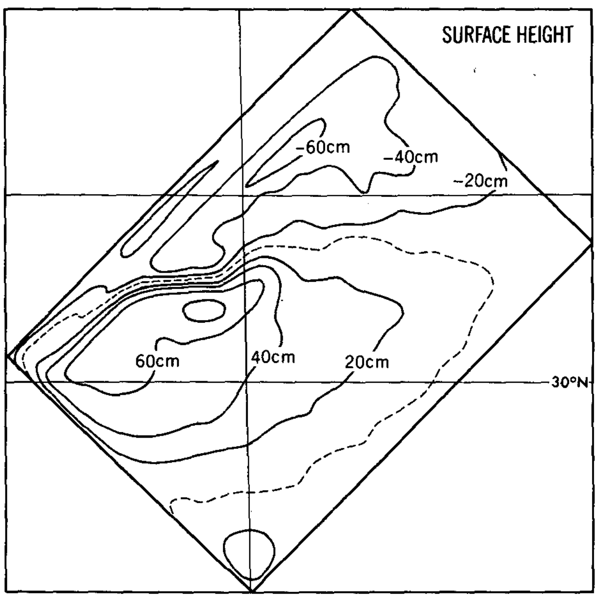

FIG. 1. Surface topography averaged with respect to time of the Semtner and Mintz (1977) model. The model extends from 15° to 50° of latitude. (Figure redrawn from Mintz, 1979.)

a relatively small, rectangular area. It is rotated 45° with respect to a global meridian in order to simulate a western boundary somewhat like the east coast of North America. The wind stress has only a zonal component and is independent of longitude and time. The zonal wind stress varies with latitude in such a way that the boundary between regions of cyclonic and anticyclonic wind torques lies at 35° of latitude. The wind torques are greatly amplified relative to observations to make the total Sverdrup transport equivalent to that of the North Atlantic in spite of the much smaller size of the basin. Surface heating is proportional to the difference between the local temperature of the surface layer and a reference temperature, which is a linearly decreasing function of latitude. This formulation is essentially the boundary condition introduced by Haney (1971). The constant of proportionality is 28 W m^{-2}/°C. The reference temperature is 30°C at the equatorward apex shown in Fig. 1 and 0°C at the poleward apex. All effects of salinity are neglected so that buoyancy transport and heat transport become equivalent. This must be kept in mind in interpreting the results for the case of the real ocean.

The five-level model is integrated from a state of rest over a period of 15 years with a coarse 75 km × 75 km resolution. The horizontal resolution is then reduced by a factor of 0.5 and integration continued for 6 years. During the final 2.5 years of the integration, a biharmonic form of lateral viscosity was introduced in the model. It is clear that the calculation does not span the very long time-scales of the deep ocean circulation, and the deeper layers of the model must be far from thermal equilibrium. The heat balance of the model was analyzed in a later paper by Mintz (1979). The poleward heat transport curve shown in Fig. 2 shows the relative contributions of the time-averaged circulation and deviations from time-average to the poleward heat transport. The area of the basin is too small at low latitudes to allow a large uptake of heat by the ocean. This peculiar geometry shown in Fig. 1 explains why the total heat transport in curve C rises rather slowly in low latitudes. If the model was in a state of thermal equilibrium, curve C in Fig. 2 would indicate that the ocean is generally receiving heat equatorward of 32° of latitude and giving off heat poleward of that latitude. Since the model is not in a steady state, we cannot be completely sure of this interpretation. The important point relates to the transport of heat by transient motions. In the region between 20° and 30° latitude the transient motions transport heat equatorward. This behavior seems strange, but it must be remembered that the thermocline slopes upward toward the equator. The average temperature gradient in the water column is therefore opposite the surface gradient in low latitudes. An equatorward heat transport by transient motions is consistent with the idea of downgradient mixing in the subtropical gyre. Note that the sign of the transport by transient eddies reverses

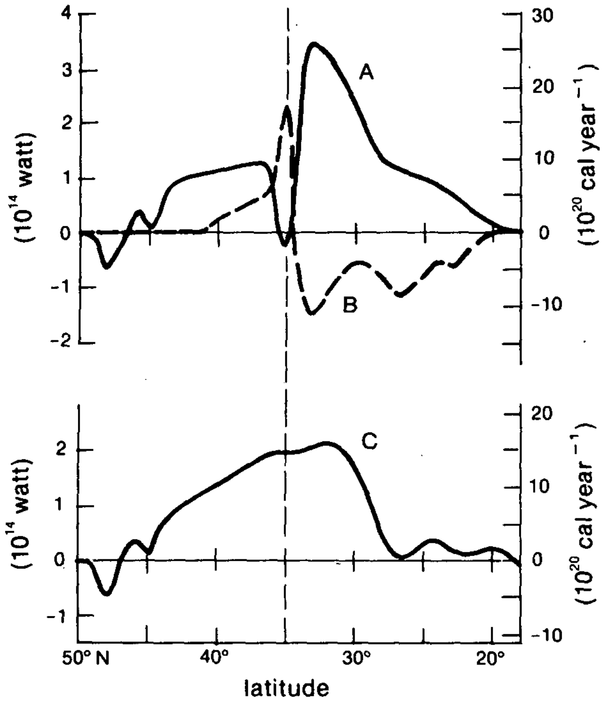

FIG. 2. Poleward heat transport in the Semtner and Mintz (1977) model. Curve A is the transport due to the time-averaged flow, and curve B is the contribution to fluctuations. Curve C is the total. The thermohaline circulation may not be in adjustment with the thermal boundary conditions after only a few decades of numerical integration. (Figure reproduced from Mintz, 1979.)

along the boundary between the subarctic and subtropical gyres. The total heat transport varies quite smoothly between 30° and 40°N, while curves A and B go through rapid variations with respect to latitude. Mintz (1979) attributed this apparent compensation to thermally indirect meridional circulations analogous to the Ferrel cell in the atmosphere. This important idea has not received the attention it deserves.

Recently, Cox (1985) has completed calculations with a model which allows us to examine the heat transport by transient ocean disturbances in more detail. It is based on the same rigid-lid, primitive equation system considered by Semtner and Mintz (1977). Again, salinity is not included in the model so that heat and buoyancy transport are equivalent. However, modern computers have allowed a more ambitious calculation. The basin of Cox's model extends from the equator to a latitude of 65°, and in the east–west direction it covers a 60° span of longitude, making it more nearly comparable to a basin the size of the North Atlantic. The east–west component of the wind stress is a function of latitude only, as in the Semtner and Mintz (1977) model, and heating at the surface is specified using the same method, but with a different linear dependence of reference temperature with latitude. The model has 18 vertical levels distributed to provide the best resolution in the upper thermocline. With a coarse horizontal resolution of 1° × 1.2° in latitude and longitude, the model was initially brought in nearly exact equilibrium with the imposed upper boundary conditions. For this resolution, transient mesoscale eddies are suppressed, allowing the "distorted physics" approach (Bryan, 1984) to be applied to reduce the number of iterations required for convergence. The next step was to integrate the eddy-resolving form of the model without "distorted physics" and a resolution of 0.33° × 0.40° in latitude and longitude over a period of 16 years. A check on the distribution of kinetic energy in the upper thermocline indicates that the model reaches a new equilibrium at the end of approximately one decade, although some transients of much longer time-scale exist at deeper levels. Another check was made on the heat balance which indicated that the convergence of poleward heat transport almost exactly matched the time-averaged surface heating at each latitude. This meant that long term transients were not causing a redistribution of heat with respect to latitude. This important requirement is often overlooked in analyzing the heat balance of ocean–climate models.

Cox (1985) has carefully compared the solutions for the eddy-resolving and corresponding non-eddy-resolving models. One of the most striking differences in the solutions is caused by eddy mixing of potential vorticity and tracers along density surfaces. In fact, the eddy mixing is so strong along some isopycnal surfaces that the shear dispersion process envisioned by Rhines and Young (1982) plays only a small role. The eddies rapidly erase gradients in potential vorticity caused by ventilation of low vorticity water in the eastern part of the subtropical gyre. As Cox (1985) points out, the process seems to be self-regulating, in that gradients of potential vorticity created by the ventilation process cause baroclinic instability, which in turn diffuses the gradients. This cycle confirms the importance of diffusion of potential vorticity by mesoscale eddies outlined by Rhines and Holland (1979). It must be remembered that Cox's model does not contain salinity effects. If salinity were included there would be regions where temperature gradients would exist along isopycnal surfaces. In those places we would expect to see a diffusive transport of heat by mesoscale eddies in the same way tracers and potential vorticity are being diffused in limited regions in the present model.

The poleward buoyancy transport in Cox's model is shown in Fig. 3a. Note that the total heat transport in the eddy- and non-eddy-resolving models was nearly the same in spite of the very vigorous eddy activity in the high resolution case. Figure 3b shows the $\langle \overline{V'T'} \rangle$ component of heat transport. Note that the dominant effect of $\langle \overline{V'T'} \rangle$ is to transport heat equatorward in the subtropical gyre. This behavior is very similar to that found by Mintz (1979), although there is no $\langle \overline{V'T'} \rangle$ component at the latitude at which the wind curl changes sign. Some confirmation that equatorward heat

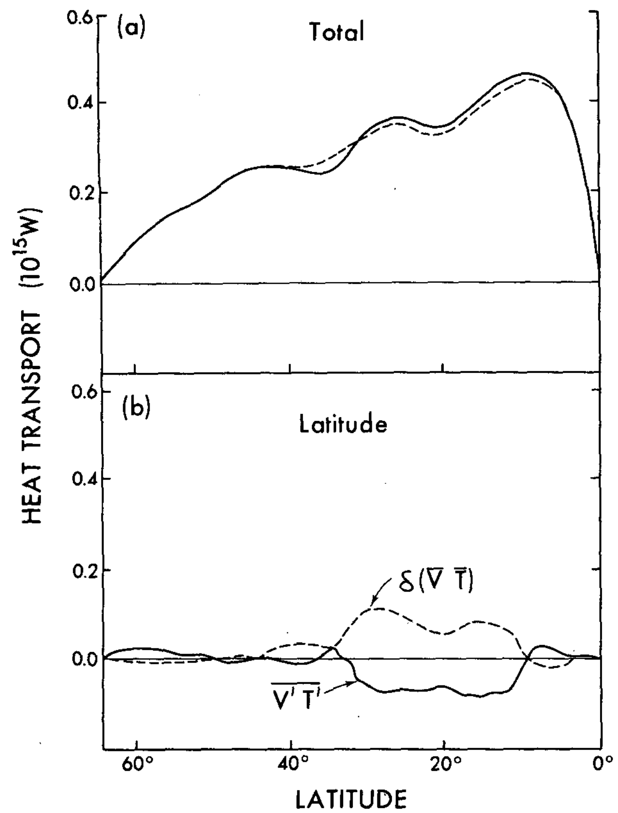

FIG. 3. (a) Total time-averaged buoyancy transport from Cox's (1985) model. The eddy resolving model (solid) and the coarse resolution, non-eddy resolving model (dashed). (b) Heat flux due to transient fluctuations (solid) and the enhanced transport by time-averaged motions in the eddy resolving model minus the transport by time-averaged motions in the coarse resolution model.

flux by mesoscale eddies does exist in the recirculation of the Gulf Stream has been obtained by Bryden (1982) based on analysis of data from the Local Dynamics Experiment. The increase of heat transport by the time-averaged flow is represented as $\delta(\bar{v}\bar{T})$. Note that it almost exactly compensates the effect of the time-dependent motions. The small differences in total transport shown in Fig. 3a are due to the existence of diffusive transport in the low resolution case and small differences between the two curves shown in Fig. 3b. The effects of diffusion on the total poleward heat transport in the high resolution case are entirely negligible.

To understand how changes in the time-averaged circulation can compensate for the heat transport due to transient motions, we have to examine the total heat transport in the meridional plane shown in Fig. 4a. The pattern is similar to that shown in previous studies by Bryan and Cox (1968) and Holland (1971). A shallow, wind-driven cell exists near the equator. It consists of an upward branch at the equator and surface outflow associated with Ekman transport in the trade wind zone. Another shallow cell of opposite sign in middle latitudes is driven by eastward wind stress. At depth, the overturning is dominated by the thermohaline circulation with a downward branch at the poleward boundary and broad upward motion which supports the thermocline. The maximum transport of the meridional overturning is 12 Mt s^{-1} (1 megaton $= 10^{12}$ g).

The difference between the meridional circulation shown in Fig. 4a and the corresponding pattern for the non-eddy-resolving case is shown in Fig. 4b. Note that the sinking near the poleward wall is somewhat less intense, but the most significant feature from the heat transport standpoint is increased poleward flow near the surface in the subtropical gyre, compensated by equatorward flow at lower levels. There is a significant temperature contrast between the upper and lower branches so that this change in circulation has a large effect on the heat transport. In deep water, on the other hand, the temperature differences are so small that the large over-turning in the opposite direction has little impact on the poleward heat flux.

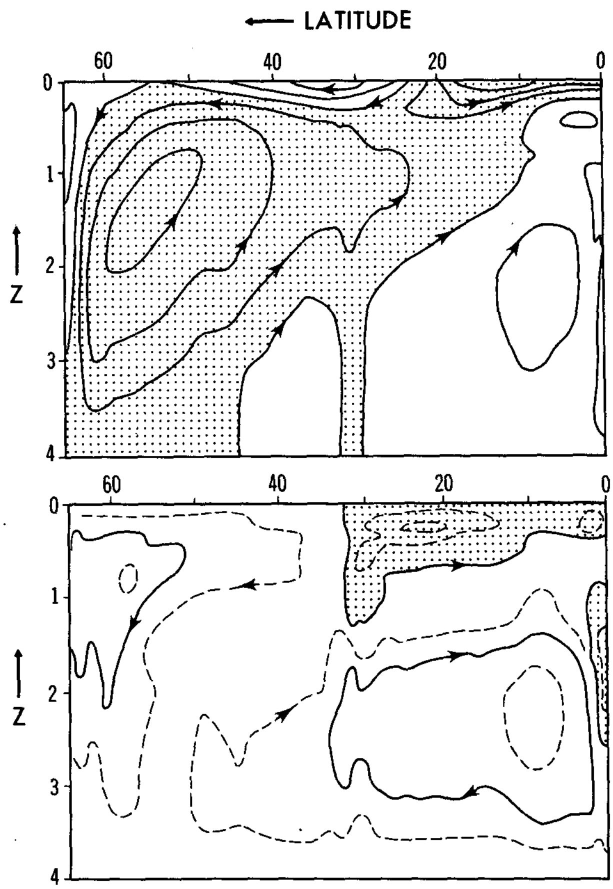

FIG. 4. (a) The total time-averaged mass transport in the meridional-depth plane in Cox's (1985) high resolution model. The contour interval in 4×10^9 kg s^{-1}. (b) The enhanced meridional circulation in the eddy resolving model relative to the coarse resolution model. Solid contours represent an interval of 2×10^9 kg s^{-1} and the dashed contours 1×10^9 kg s^{-1}.

The magnitude of the upper cell in the subtropics associated with transients is about 2 Mt s^{-1} (10^9 kg s^{-1}). To estimate its effect on heat transport we note that the heat capacity of water is approximately 4000 J $kg^{-1}/°C$ or 4 MJ $t^{-1}/°C$. If the upper and lower branches of the cell differ by as much as 20°C,

$$2 \text{ Mt s}^{-1} \times 4 \text{ MJ t}^{-1}/°\text{C} \times 20°\text{C} = 0.16 \times 10^{15} \text{ W}.$$

This calculation illustrates how a strong vertical temperature contrast allows even a very weak vertical overturning to have a significant effect on the poleward heat transport.

4. Discussion

Cox's (1985) model of the ocean circulation in a basin of planetary dimensions is the first eddy-resolving model to include both the effects of wind and a fully developed thermohaline circulation that is in near equilibrium with surface forcing. It provides an opportunity to examine the coupling between mesoscale eddies and the thermohaline circulation, a coupling which is extremely difficult to diagnose from the scant observational data base presently available. The eddies in the model are highly energetic and account for a large share of the total kinetic energy. Yet the eddies appear to be almost entirely decoupled from the thermohaline circulation. When the eddy-resolving model is compared to a similar model of lower resolution it is found that the meridional circulation is enhanced slightly in the subtropical gyre region. This enhanced Eulerian mean meridional circulation carries light surface water poleward and denser water at depth equatorward, almost exactly balancing the eddy transfer of buoyancy in the subtropical gyre region of the model. The concept of the generalized Lagrangian mean with respect to time and the zonal direction is useful. Actual particle trajectories associated with the mesoscale eddies appear to be largely excursions parallel to isopycnal surfaces. As a result, there is little contribution to the component of the generalized Lagrangian mean which is normal to isopycnal surfaces. It is this component that is important for the convergence of buoyancy flux and determines the thermohaline circulation.

To provide another way of looking at the heat transport compensation found in the model, consider a train of waves in the x-direction. Let buoyancy gradients in the z-direction and y-direction be denoted by $\partial_z Bo$ and $\partial_y Bo$.

In this case a small buoyancy perturbation can be expressed as

$$B' = -Z'\partial_z Bo - Y'\partial_y Bo. \tag{4.1}$$

Here Z' and Y' are small displacements in the z- and y-directions, respectively. The primes denote departures from a time average. Let us consider a positive $\overline{V'B'}$ correlation. In this case, positive V' should be associated with a negative displacement, Z', and negative V' with a positive displacement in the vertical. The trajectories of particles in the wave would then execute loops in the meridional plane that would be clockwise looking in the positive x-direction. In Cox's model the main buoyancy transport due to transient motion is equatorward so the looping motion is in the opposite sense. This is reflected in the difference of the Eulerian mean meridional circulations shown in Fig. 4b. The counterclockwise shallow overturning in the subtropical gyre corresponds to the latitude of maximum eddy heat transport. The physical arguments given above are purely kinematic. What is the dynamic balance of the enhanced meridional circulation associated with the mesoscale waves in the subtropical gyre?

Kuo (1956) was able to show that in the atmosphere the Ferrel cell was primarily balanced by the large-scale Reynolds stress due to midlatitude disturbances. In the ocean, it is possible for a meridional circulation to be entirely geostrophic due to the existence of lateral walls, which exist at all latitudes except in the vicinity of the Drake Passage in the Southern Hemisphere. A very small difference in the net pressure difference between the eastern and western boundary would be enough to balance the enhanced meridional circulation of Cox's eddy-resolving model. An analysis of the time-averaged solutions does show that, indeed, the additional east–west pressure drop of the eddy-resolving, as compared to the non-eddy-resolving, case is enough to account for the change in meridional circulation seen in Fig. 4b.

It is not clear what type of wave-induced mean flows would occur in the ocean in the absence of meridional barriers, which would be the case in the Drake Passage. Eddy-generated meridional Reynolds stresses would have to be very strong to sustain the equivalent of the atmospheric Ferrel cell.

The nonbuoyancy transport property of mesoscale waves illustrated in Cox's model study could break down in the real ocean. Diabatic changes in water mass properties can be very intense locally. Warm core rings are a possible example. Since the salinity of warm core rings north of the Gulf Stream is much higher than ambient waters, winter cooling can cause a sudden loss of buoyancy and lead to rapid overturning. It should also be noted that in the model the parameterization of vertical mixing is completely decoupled from the mesoscale eddies. As more information about vertical mixing in the ocean becomes available it may not turn out to be a very accurate assumption and there may be a link between the Lagrangian mean flow across isopycnal surfaces and mesoscale eddies that is completely missing from Cox's simplified model. Clearly, as more detailed data on mesoscale eddies and surface heat exchange become available many other examples of breakdown of nonbuoyancy transport behavior will be found. We therefore only expect the idea of nonbuoyancy transport by eddies to be useful as a general rule, not a universal principle.

Returning to the question posed in the introduction concerning the difference between eddy heat transport in the atmosphere and eddy buoyancy transport in the ocean, we note diabatic processes are strong in the troposphere but relatively weak in the ocean. At synoptic scales in the atmosphere, radiative damping and heating due to moisture condensation have the same time scales as the motion itself. Thus particles lose their properties along synoptic displacements in the atmosphere, so that eddy transport and net transport of heat are nearly the same. On the other hand, the conservation of buoyancy along synoptic displacements in the ocean model means that transport of buoyancy due to transient motions can take place with no net transport. How do we interpret the measurements of eddy heat flux in the Gulf Stream (Newton, 1961) and in the Antarctic Current (Bryden, 1979; Nowlin et al., 1985)? Our model results show that equating eddy heat flux with total heat flux in the absence of other evidence may be misleading. In the case of the Antarctic Circumpolar Current, a good case may be made for a positive net poleward heat flux, since there is a temperature gradient within isopycnal surfaces (Reid, 1965). Mesoscale eddies would tend to diffuse heat along the isopycnal surfaces. A quantitative evaluation of the net poleward heat flux, however, would require estimates of the wave-induced overturning in addition to the complexities of measuring heat transport due to transient fluctuations.

In the next decade, new observational programs should provide a greatly expanded data base on the properties of mesoscale eddies. Our model results on buoyancy transport should not be interpreted to mean that mesoscale eddies do not play an important role in the ocean circulation. More detailed measurements will probably confirm the prediction made by Holland and Lin (1975) that eddies are extremely important in transferring vorticity downward in the water column. In the same way, Rhines and Young (1982) have emphasized the important role of mesoscale eddies in mixing potential vorticity laterally along isopycnal surfaces. The important conclusion to be drawn from the present study is that the horizontal eddy flux of any quantity which is nearly uniform along isopycnal surfaces will be extremely difficult to measure experimentally. Measuring eddy correlations is not enough. It is also necessary to measure eddy-driven, time-averaged flows with great accuracy. In many cases, inverse modeling based on measurements of large scale patterns may turn out to be the most practical way to evaluate the transport role of mesoscale eddies.

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