NOTES AND CORRESPONDENCE

The Creation of Reversed Baroclinicity and Subsurface Jets in Oceanic Eddies

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ABSTRACT

Reversals of baroclinicity and associated subsurface jets have often been observed in oceanic eddies, but without adequate explanation. In this study, a two-dimensional adiabatic frontogenesis model is applied and several features of this phenomenon are elucidated.

Observations show that reversals of baroclinicity and associated maxima of jet velocities are found primarily between the seasonal and main thermoclines at depths of about 100 m. This is accounted for in the model's initial conditions by superimposing a shallow seasonal thermocline with small isopycnal slopes over a more strongly baroclinic main thermocline. The initial model domain comprises a 1600-km wide strip which is then compressed by an external deformation field to simulate the formation of fronts by gyre-scale confluence.

After integrating over 30 model days, a front has formed and the sign of isopycnal slope has changed in the seasonal thermocline, leading to a maximum velocity in the frontal jet near 100-m depth. The reversal of slope is caused by a difference in direction of vertical motion between the cyclonic and anticyclonic sides of the jet due to stretching and compression of vortex tubes. The subsurface jet maximum is due simply to a sign change in the thermal wind.

These model results show that for realistic conditions, the reversals of baroclinicity and associated subsurface jet maxima in oceanic eddies can already have been produced during frontal formation and retained by the eddies after they become detached by hydrodynamic instabilities.

1. Introduction

It is a common feature of oceanic eddies that the slopes of isopycnals change sign with depth, i.e., there are reversals of baroclinicity. In many cases such reversals are accompanied by subsurface maxima in the azimuthal component of eddy velocity. This phenomenon has been observed in both cyclonic and anticyclonic eddies.

Olson (1980) described the structure of Gulf Stream cyclonic ring BOB, which had a diameter of about 160 km. The typical doming of isopycnals occurred only at depths greater than about 100 m, whereas in the surface layers they exhibited reversed slopes. This reversal of baroclinicity, according to gradient wind calculations, leads to a maximum in azimuthal velocity below the sea surface. Another cyclonic eddy, S2 of POLYMODE LDE, was observed to have a similar structure, though it was only 50 km across (Lindstrom et al. 1986). Detailed investigations of anticyclonic eddies have also documented the reversal of baroclinicity near 100-m depth. According to Joyce (1984) and Kennelly et al. (1985), Gulf Stream warm-core rings 81D and 82B, respectively, each with a diameter of about 160 km, exhibited a bowl-shaped structure of isopycnals below about 100 m, and a slight doming in the surface layers. Calculations of the azimuthal ring velocity based on the gradient wind method also revealed a subsurface maximum in ring 81D.

The four papers mentioned above are only a few examples extracted from a vast amount of literature describing the hydrography of eddies, many of which displayed reversals of baroclinicity. There are, however, no satisfactory explanations for the generation of this phenomenon. In this paper I propose a mechanism by which the slopes of isopycnals in oceanic eddies can change sign in the surface layers.

2. Basic concepts

Typically, oceanic eddies are products of potential vorticity fronts forced by either gyre-scale or synoptic-scale deformation fields (MacVean and Woods 1980; Bleck et al. 1988, referred to as BOW). General features of these fronts include strong baroclinicity and intense jets. These jets then become unstable due to hydrodynamic instabilities, form meanders, and ultimately shed cyclonic or anticyclonic eddies. Because frontal formation leads to the generation of eddies, it is reasonable to assume that some structures characteristic of eddies have already been created during the frontogenesis process before meanders have been formed.

The BOW frontogenesis model created a front that

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qualitatively exhibited several features typical of eddies, either cyclonic or anticyclonic. To conceptualize this, we may consider the BOW front as representing a radial section across any eddy (Fig. 1). If we assume the eddy center to be located at $x = -L/2$, then the density and velocity structures are similar to those found in cyclonic eddies—a doming of isopycnals in the eddy center, high static stability in the thermocline at the cyclone periphery (the expression "thermocline" will be used throughout in this paper instead of "pycnocline"; no conflicts arise, because salinity is held constant for simplicity), lower static stability close to the center, and a thermostad above the thermocline in the outer regions. The opposite is true if we imagine the eddy center to be located at $x = L/2$. Features characteristic of anticyclones are then identified: high static stability in the thermocline close to the eddy center, lower static stability at the periphery, and a large thermostad above the thermocline in the center of the anticyclone.

There is one major difference between the structures of observed eddies and those constructed from the BOW front: The BOW front has only one thermocline, whereas all observed eddies have two—the seasonal thermocline in the upper 100 m and the deeper main thermocline somewhere above 1000-m depth. Once again with respect to those observed eddies exhibiting reversals of baroclinicity, this reversal occurs between the seasonal and main thermoclines: In cyclones the isopycnals of the main thermocline are domed, whereas those of the seasonal thermocline reflect higher pressure in the eddy center. The opposite is true for anticyclones. This feature was not produced by the BOW model simply because it was designed to simulate frontogenesis in the seasonal thermocline. It was therefore not necessary to define a main thermocline explicitly in the initial conditions. However, in order to model the reversal of baroclinicity, it is required that two thermoclines be specified in the initial conditions.

The slopes of isopycnals in the seasonal thermocline are generally between one and two orders of magnitude less than those in the main thermocline, which can be seen in climatological atlases such as the one by Bauer and Woods (1984). In the region of the Gulf Stream extension east of the Grand Banks, their atlas reveals vertical displacements of isopycnals (such as $\sigma_t = 26.0$) in the summer (August) thermocline of only some tens of meters over horizontal distances of 1000 km, i.e., isopycnal slopes are on the order of only 10^{-5}. In contrast, contemporaneous vertical displacements in the main thermocline ($\sigma_t = 27.0$) are about 500 m over 500 km horizontal distance, yielding slopes of order 10^{-3}. This is an important feature that must be accounted for if reversals in eddy baroclinicity are to be explained in terms of frontogenetic dynamics, which will be accomplished in the following. To do so, I use the BOW model with altered initial conditions: Two thermoclines are specified which represent the seasonal and main thermoclines. Also, the BOW model was originally intended to simulate frontogenesis in a synoptic-scale deformation field, but here the model simulates the generation of fronts due to gyre-scale confluence from which synoptic scale eddies become detached. This is accomplished by compressing the north–south extent of the model domain during simulation.

3. The numerical experiment

The model in its original form was fully described by BOW. Therefore, only alterations made here to that model are outlined.

As compared with synoptic-scale confluence, gyre-scale confluence extends over much larger horizontal scales. To accomodate this in the model, the BOW initial domain is doubled in size from 800 to 1600 km. But the number of grid points in the horizontal is left unchanged from 256, thus the horizontal resolution is now 6.25 km. Also, gyre-scale deformation occurs at a slower rate than the synoptic-scale. This is taken into account by reducing the deformation rate to 10^{-6} s^{-1},

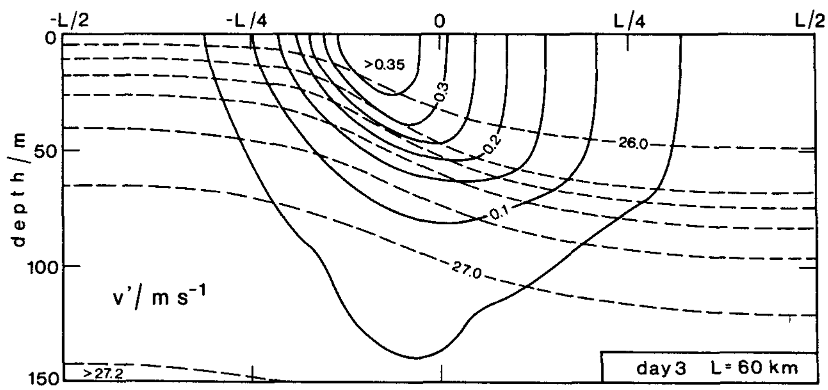

FIG. 1. The final mass and velocity fields of the model of Bleck et al. (1988, their Fig. 5b) in the upper 150 m of the model domain. Broken lines are σ_t-surfaces in increments of 0.2 kg m^{-3}. Thick lines are isotachs of v'-velocity in increments of 0.05 m s^{-1}.

which results in a time scale for frontogenesis of one month (as opposed to 3 days in BOW). Shown in Fig. 2 is a meridional section of the initial flow field of the model. It is geostrophically balanced in seven isopycnic layers and is intended to represent average summer conditions at the northern rim of the North Atlantic subtropical gyre: The main thermocline ($\sigma_t > 26.4$) exhibits strong isopycnal slopes, above which is a seasonal thermocline that is nearly horizontal. In order to provide a realistic lower boundary condition for the main thermocline, the bottom of the domain is set at 4000-m depth.

Figure 3 shows the mass and v'-fields (v' is the alongfront velocity component from which the deformation velocity has been previously subtracted) after integrating over 30 model days with 5-minute time steps. A comparison with Fig. 2 (please note the different domain widths in Fig. 2 and 3!) reveals that pressure has increased on all isopycnals on the cyclonically sheared side of the jet and has decreased on the anticyclonically sheared side (cf. BOW, Fig. 3). In the main thermocline this has only led to a reduction of isopycnic pressure difference between the sidewalls, but in the seasonal thermocline it has caused a reversal of baroclinicity, i.e., the sign of the slope has changed during frontogenesis. In the beginning, the slopes of all isopycnals, as measured against radial distance, had the same sign, but subsequent to frontogenesis comes a reversal of baroclinicity with depth. The second striking feature of Fig. 3, in contrast to Fig. 4b of BOW, is that the frontal jet now has its maximum speed not at the surface, but at a depth of about 100 m.

4. Interpretation of model results

In light of the BOW model results, it is rather easy to identify the mechanism that has generated the modeled structures. The external deformation field enhances frontal baroclinicity, which accelerates the frontal jet, in turn leading to an increase in the cyclonic and anticyclonic shear vorticities on the respective flanks of the jet. Conservation or potential vorticity then requires stretching and compression of vortex tubes and a compensating cross-frontal mass flux directed from the anticyclonically sheared side of the jet to the cyclonic. The divergence of mass flux then drives vertical motion—downwelling on the cyclonic side and upwelling on the anticyclonic. In the main thermocline this vertical motion causes only a reduction of the isopycnals' slopes, but in the seasonal thermocline, because of the very much weaker baroclinicity, only small changes in angle are required to level the isopycnals. If, as in the model, the external deformation continues beyond that time, then anticlockwise rotation of isopycnals about the position of no vertical displacement continues (cf. BOW, Fig. 3), resulting in a change of sign of isopycnal slopes. This reversal of baroclinicity occurs initially at the lower boundary of the seasonal thermocline because the vertical motion increases with depth (cf. BOW, Fig. 7), which can also be seen in Fig. 3: The signs of slope of the uppermost four isopycnals have not yet changed, but a change has already occured in the lower part of the seasonal thermocline ($25.6 < \sigma_t < 26.4$). In principle, these reversals could also have been produced by the BOW model, but first of all between 300 and 500 m, because of the lack of a strong potential vorticity gradient separating the seasonal and the main thermocline, which is rather unrealistic. Because the frontal jet is in approximate geostrophic balance (the Rossby number on day 30 is about 0.01), the mass field can be used to explain the structure of the jet velocity. At the depths where isopycnals slope downward to the right (in the main thermocline and at the top of the seasonal thermocline),

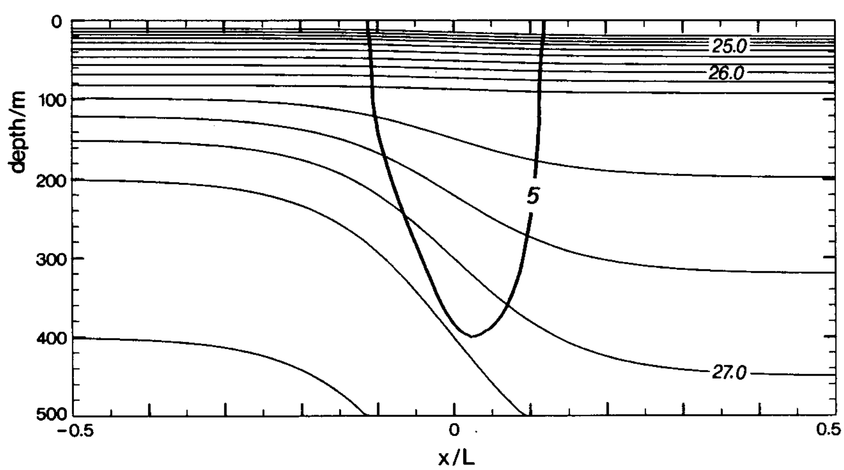

FIG. 2. The initial mass and velocity fields in the upper 500 m of the model domain. Thin lines are σ_t-surfaces in increments of 0.1 kg m^{-3}. The thick line is the 5 cm s^{-1} isotach of the geostrophic jet. The initial width of the model domain is L = 1600 km.

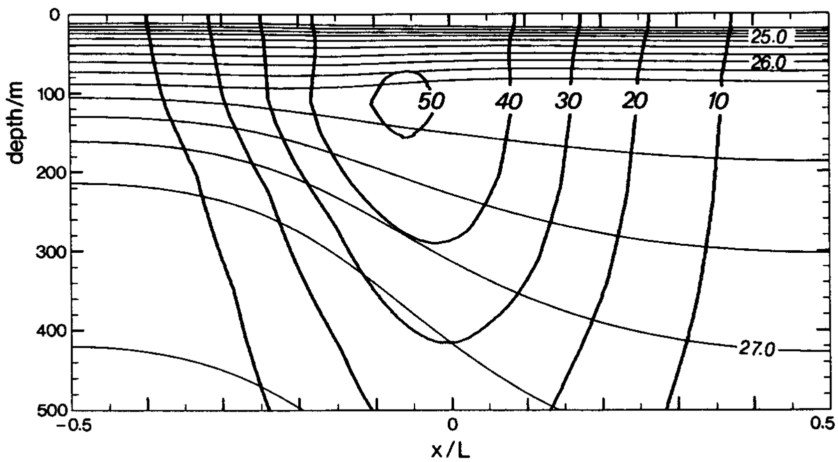

FIG. 3. The mass and velocity fields after 30 days of integration. Thin lines are σ_t-surfaces in increments of 0.1 kg m^{-3}. Thick lines are isotachs of v'-velocity in increments of 10 cm s^{-1}. The domain width is L = 120 km.

the vertical shear is positive; where the isopycnals slope downward to the left (in the lower part of the seasonal thermocline), the shear is negative. These sign changes of the vertical shear explain the subsurface maximum of the frontal jet and a slight increase of the jet speed very close to the sea surface.

5. Comparison with observations and discussion

In order to demonstrate that the front in the seasonal thermocline, as created by two-dimensional frontogenesis, already possesses features which can be found in detached oceanic eddies, it is useful to create mirror images of Fig. 3 at $x = L/2$ and $x = -L/2$. Figure 4a shows the result of the first manipulation, and may be considered to represent a radial section through an anticyclonic eddy. A similar section through a cyclonic eddy results from the second manipulation and is displayed in Fig. 4b. In each figure a horizontal kilometer scale symmetric to the eddy center has been added, and the signs of the v'-velocity have been adjusted to the mass fields.

The modeled mass and velocity fields in the seasonal thermocline are now very similar to their observed anticyclonic counterparts. In observed anticyclonic eddies the baroclinicity reverses between the seasonal and main thermoclines at about 100-m depth; near the surface the isopycnals are domed, whereas in the lower layers they are deepest in the eddy centers (cf. Joyce 1984, Fig. 4c; Kennelly et al. 1985, Fig. 7a). This is not exactly true for the modeled anticyclone for $|x| < 50$ km, but this is only a consequence of the symmetry axes being at $x = L/2$. Also, the magnitudes of isopycnic slopes in the modeled and observed anticyclone are comparable; vertical displacements in the seasonal thermocline, both in the model and in observations, are on the order of 10 m over horizontal distances of 100 km, yielding slopes of order 10^{-4}, whereas at greater depths the slopes are about one order of magnitude larger and have the opposite sign.

With respect to cyclonic eddies, the structure of the modeled mass field (Fig. 4b) is comparable to those observed by Olson (1980, Fig. 1) and Lindstrom et al. (1986, Fig. 3c). In both the model and observations the reversal of baroclinicity occurs near the ring periphery and at depths of about 100 m. The modeled field in Fig. 4 even shows the deepening of isopycnals in the seasonal thermocline which can be found in Olson's ring at 40 km from the ring center. But this may be an overinterpretation of the model results. Olson's potential vorticity analysis (his Fig. 7a) reveals that the isopycnic depth anomalies are correlated with anomalies in the potential vorticity field. Thus, it is likely that the isopycnic depth anomalies are not a product of frontogenesis, but have been created later as a consequence of advection of different potential vorticity "streamers" by the three-dimensional velocity field of the eddy. The modeled v'-velocity field shown in Fig. 4 exhibits nearly the same features as do sections of azimuthal velocity displayed in Joyce (1984, Fig. 9) and Olson (1980, Fig. 3). Except for the right half of Olson's eddy, all observed velocity sections have subsurface jet maxima between the seasonal and main thermoclines, as in the model.

The agreements seen here between model results and observations are notable, but the author does not claim that the proposed mechanism is the only one capable of producing reversals of baroclinicity and subsurface jets. These can also be achieved by divergences of the radial flow in eddies after they have become detached from fronts. Direct measurements of the velocity structures in rings (Joyce and Kennelly 1985; Kennelly et al. 1985; Joyce 1984) reveal radial velocities of order 10 cm s^{-1} in the seasonal thermocline directed inward

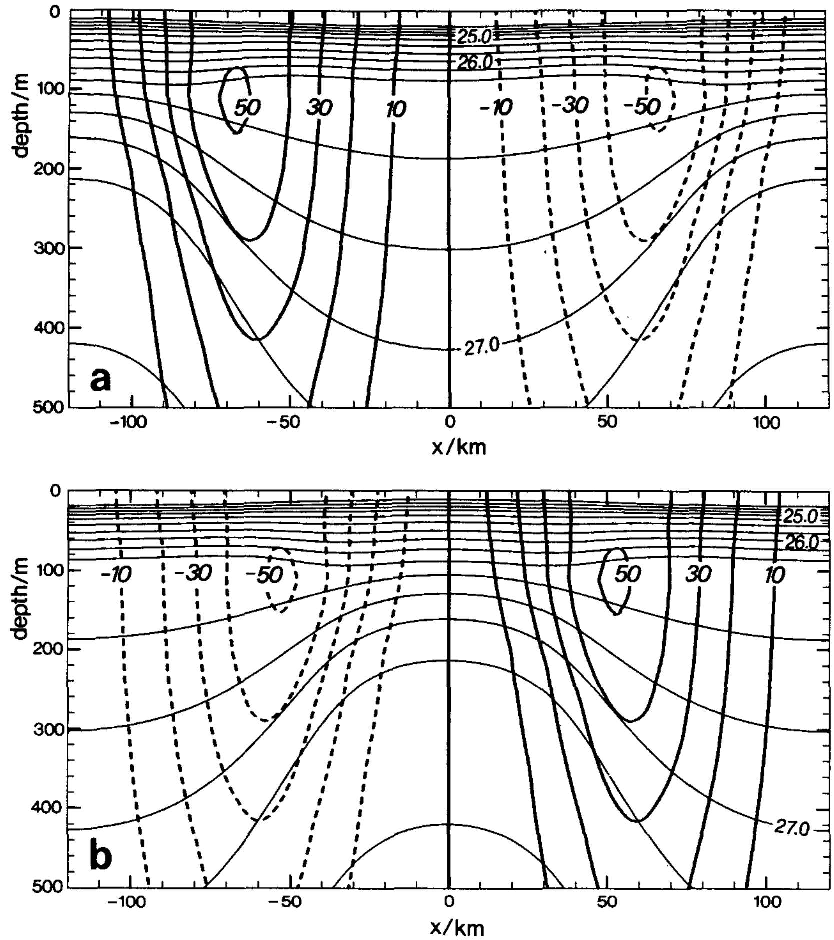

FIG. 4. Radial sections through (a) an anticyclone and (b) a cyclone. The sections have been reconstructed from Fig. 3 (for further explanation see text). The contour intervals are the same as in Fig. 3.

in cyclones and outward in anticyclones. These are probably forced by vortex stretching due to intensification of relative vorticity during ring spinup. The direction of radial flow then leads to convergence (downwelling) in the center of cyclones and divergence (upwelling) in anticyclones, which in turn enhances the reversal of baroclinicity in the qualitative same way as shown by the model. This conceptual picture of radial flow has been confirmed by Olson (1980). He analyzed the vertical displacements of isotherms in cold-core ring BOB over a period of 77 days and estimated downwelling velocities of 1.3 m d^{-1} in the ring center and upward motion of 0.2 m d^{-1} outside the ring, which are of the same order of magnitude as those in the model. Therefore, it is difficult to isolate the mechanism that caused the reversal of baroclinicity, a problem compounded by the time scale of gyre-scale deformation (order of month) being the same as the spinup time for synoptic-scale eddies. Another mechanism which might cause reversals of baroclinicity has been recently proposed by Williams (1988). He pointed out that diabatic mixing due to different surface heat fluxes between the centers of eddies and the ambient waters is capable of altering the density stratification to the point of producing reversals of baroclinicity. But the time scale over which this mechanism occurs is of order one year and is not consistent with observations of reversals in 'young' eddies.

6. Conclusions

A two-dimensional frontogenesis model has been used, together with conservation of potential vorticity arguments, to explain the reversal of baroclinicity and creation of subsurface jets that are often observed in oceanic eddies. Results from the model point to the possibility that these eddy features are carry overs from earlier processes of frontal formation and are retained by the eddies long after detachment from fronts due to baroclinic instability. Reversal of baroclinicity is a consequence of vortex stretching and occurs between the seasonal and main thermoclines. The reversed baroclinicity leads to a change in sign of the thermal wind, resulting in a jet maximum beneath the surface.

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