Why Western Boundary Currents in Realistic Oceans are Inviscid: A Link between Form Stress and Bottom Pressure Torques

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(Manuscript received 27 September 2000, in final form 30 January 2001)

ABSTRACT

It is shown that wind stress curl is balanced by bottom pressure torque in a zonal integral over any strip wide enough to smooth out the effect of nonlinear terms (typically about 3° of latitude). The derivation is completely general as long as the zonal wind stress is balanced by form stress at each latitude, as is known to be the case in the ocean. This implies that viscous torques are not important in western boundary currents, their place being taken by bottom pressure torques. The prediction is confirmed in the context of a global, eddy-permitting, numerical ocean model. This link between form stress and bottom pressure torques makes it easier to consider Southern Ocean dynamics and subtropical gyre dynamics in the same conceptual framework, with topographic interactions being important in both cases.

1. Introduction

The role of bottom topography in ocean circulation is a strange subject. On the one hand, since the work of Stommel (1948) and Munk (1950), it has been assumed to play a rather minor role in the subtropical gyre circulation. On the other hand, in the Southern Ocean topography has been known to be needed to obtain a dynamical balance since the work of Munk and Palmén (1951), and topography clearly has a strong influence on other high-latitude flows.

a. Gyre circulations

Belief in the unimportance of topography in the interior of subtropical gyres stems initially from the success of Stommel's (1948) model in describing the major features of these circulations. The main purpose of Stommel's paper was to demonstrate the source of east–west asymmetry in the gyre circulation. This was shown to result from the variation of the Coriolis parameter with latitude (the beta effect), by means of the simplest model possible: a wind-stress-forced, flat-bottomed, barotropic, rectangular ocean in which the flow is linear, and the only friction is a linear bottom friction. In a sense, of course, this model does contain topography in the form of vertical sidewalls. The analytical model was then solved on an f plane and on a β plane to show that the introduction of β leads to western intensification. There was no reason to choose a flat bottom except for simplicity, but the reasonable circulation that resulted made the assumption that topography is unimportant seem plausible.

The theoretical basis for this assumption was strengthened by Munk (1950), who showed that the same interior flow (in a depth-integrated sense) results from the assumption that the flow does not penetrate deep enough to reach the bottom. The interior flow is then (as in Stommel's model) given by Sverdrup balance in which the depth-integrated meridional flow is determined by the wind stress curl. With this assumption there can be no bottom friction, so Munk substituted lateral friction to permit closure, resulting in what is now commonly known as a Munk boundary layer structure for the western boundary current. Still, as there was no a priori reason to assume that the flow could not reach the ocean floor, the assumption continued to be justified only by the reasonable circulation patterns that resulted.

A stronger theoretical reason for negligible flow at depth eventually came from spinup calculations, in which the flow was started from rest and the wind stress turned on at time zero. In the flat-bottom, linear, quasigeostrophic case (e.g., Young 1981), it is possible to show analytically that the interior Sverdrup circulation

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is set up by Rossby waves propagating from the eastern boundary. The fastest waves are barotropic and quickly set up a barotropic interior flow. Next comes the first baroclinic mode, which does not alter the depth integrated flow, but confines it to a flow above the thermocline. Higher modes traverse the basin more slowly, and successively confine the circulation to a thinner layer near the surface, until the flow in that layer becomes fast enough for the assumption of linearity in the density advection equation to break down, all with no change to the Sverdrup relation in the depth integral. Numerical two-layer simulations extended this reasoning to the case with topography (Anderson and Gill 1975; Anderson and Killworth 1977) in which case, while the initial barotropic adjustment is to a state quite different from Sverdrup balance (it is close to what is usually referred to as ‘‘topographic Sverdrup balance,’’ for which f/H contours replace latitude lines as the characteristics), the baroclinic adjustment cuts off the flow in the bottom layer, leading again to the conventional Sverdrup balance for the depth integral. We therefore have reason to believe that the wind-driven circulation in ocean gyre interiors should not be influenced by bottom topography, unless the flow speed becomes comparable with the propagation speed of the first baroclinic Rossby wave. With the Rossby wave speed decreasing from 10 cm s^{-1} or more in the mid Tropics, to less than 1 cm s^{-1} at high latitudes (Killworth et al. 1997), it is to be expected that Sverdrup balance will hold in the Tropics but will break down due to topographic interactions at higher latitudes.

Even in the Tropics, this argument only holds for the Sverdrup interior of wind-driven gyres. The deep branch of the thermohaline circulation must certainly interact with topography, and the western boundary currents interact strongly with either the sidewalls (Munk boundary layer) or the ocean bottom (Stommel boundary layer). The real ocean, of course, has a continental slope and shelf instead of vertical sidewalls, so the sidewalls must also be considered to be bottom topography.

The situation is complicated by the role of nonlinearity in western boundary currents, which can be significant. However, nonlinearity cannot balance wind stress curl in an integral sense, so friction remains crucial in these boundary currents in order to obtain a dynamical balance when the only relevant topography is in vertical sidewalls.

Theoretically, then, we might expect topography to be important at high latitudes (where the stratification is weak and Rossby wave speeds are slow), for deep flows, and in western boundary currents. Also, the argument for ignoring bottom topography in the interior of subtropical gyres only holds if it makes sense to consider a near-surface wind-driven circulation independently of the deep thermohaline flow. In fact, evidence for the validity of Sverdrup balance is still wanting, even in the subtropical North Atlantic (Wunsch and Roemmich 1985).

Evidence of the importance of topography in western boundary currents was first put forward in this context by Holland (1973), who noted that a continental slope could induce a recirculation in the western boundary current, increasing the nearshore current while compensating for this increase with an offshore return flow. For deep western boundary currents, the role of topography in permitting currents broader than the inertial or viscous scale was noted earlier by Stommel and Arons (1972). More recently, idealized flows have been constructed in which there is no need for viscous effects to permit a meridional flow (Salmon 1992, 1994; Straub et al. 1993; Becker and Salmon 1997; Griffiths and Veronis 1997, 1998; Becker 1999; Ford 2000). In fact, the simplest such model (linear and barotropic) is easily derived from a thermal analogy due to Welander (1968), which has recently been extended to include small departures from a fixed mean stratification by Salmon (1998). Friction is still important in these models, to obtain a global balance, but it is no longer the term which permits a meridional mass flux in the western boundary current. Instead, the flow conserves potential vorticity (following f/H contours in the barotropic cases), and the meridional flow is balanced by a bottom pressure torque.

In this paper, then, when we talk about viscosity being unimportant in western boundary currents, we are referring specifically to its role in upsetting the Sverdrup balance and permitting meridional flows. There is widespread belief in the importance of bottom friction to the energetic balance of the flow and in the important role of friction in selecting western boundary currents in preference to eastern boundary currents within the ocean. A probably related issue is the need for viscosity in various idealized but complete solutions (e.g., Ford 2000) to satisfy a no-normal-flow boundary condition. The theory and diagnostics presented here do not address these issues.

b. *The Southern Ocean*

In the Southern Ocean, attention has centered on the angular momentum (or zonal momentum) balance rather than on Sverdrup balance, although both have been considered. Munk and Palmén (1951) showed that pressure forces on the ocean bottom topography are the most likely candidates to balance the angular momentum input by zonal wind stress. Other possibilities are bottom or lateral friction, or the nonlinear advection of angular momentum, which have been shown to be too small by scaling analysis (Oort 1985), model studies (Treguier and McWilliams 1990; Wolff et al. 1991; Ponte and Rosen 1994; Stevens and Ivchenko 1997; Gille 1997), in situ studies (Bryden and Heath 1985), and satellite remote sensing measurements (Morrow et al. 1994). In this zonal and depth integral balance, the Coriolis force drops out, if there is no steady northward mass flux across each latitude, and becomes small for realistic mass fluxes due to evaporation and precipitation. The only remaining term is the pressure force on topographic obstacles. Munk and Palmén (1951) estimated that the zonal wind stress could be balanced by bottom pressure in the Southern Ocean if the pressure was higher on the western side of ridges than on the eastern side by an equivalent 4 cm of water for a series of ridges with height totaling 10 km.

It is now generally accepted that the zonal wind stress is mainly balanced by this pressure force—bottom form stress—in the Southern Ocean. Munk and Palmén (1951) hinted that this might provide a relationship between wind stress and the transport of the Antarctic Circumpolar Current (ACC), implicitly assuming that the form stress could be interpreted as a form drag related to the strength of the zonal flow at the ocean floor. In fact, the interpretation is not so simple.

Although the Coriolis force drops out in the depth and zonal integral, this is not the case for the zonal integral at each depth. The zonal wind stress puts angular momentum into a near-surface Ekman layer in which a northward flow occurs, permitting Coriolis to balance the wind stress. Mass conservation, however, requires a return flow at depth in which an equal and opposite Coriolis force must be balanced by a pressure force (form stress). Seen this way it is clear that the balance between wind stress and form stress is closely related to the meridional overturning circulation, but how (and whether) this is related to the strength of the ACC is not so clear.

Several authors have examined the possibility that the transport of the ACC is determined by the wind stress curl (Stommel 1957; Baker 1982; Chelton et al. 1990), construing the ACC as an eastward flow with a small southward component in Sverdrup balance with wind stress curl, closed by a northward flow in a western boundary current. More recently, Warren et al. (1996) advanced this viewpoint, while condemning consideration of the balance between wind stress and bottom form stress as ''obscurantist'' [see comments by Hughes (1997), Olbers (1998), and replies by Warren et al. (1997, 1998)]. It is not our purpose in this paper to discuss the complicated subject of what controls the strength of the ACC, but one result will be to show that, far from being obscurantist, consideration of form stress is, in fact, essential to the understanding of Sverdrup balance in the Southern Ocean, and elsewhere.

c. Combining the two concepts

Although the balance between wind stress and bottom form stress has been given attention in the Southern Ocean, it must hold at all latitudes. The only difference at latitudes outside Drake Passage is that the bottom topography reaches the surface in the shape of continents. There can still be no (or very little) mass transport across each latitude circle, making the Coriolis force negligible in the integral over longitude and depth. The zonal integral of zonal friction is still small, even if alongstream friction is important in narrow western boundary currents and nonlinear terms are also small (the Southern Ocean is where they are largest). The zonal wind stress is then balanced by a pressure difference across topographic features, of which the continents are just an extreme example. Again, this balance can be interpreted as a meridional overturning with northward flow in the Ekman layer and a geostrophic return flow at depth, or vice-versa. This is an aspect of Sverdrup gyres that is often overlooked. Although the depth-integrated flow in a gyre interior is given by Sverdrup balance and the wind stress curl, the partitioning between flow in the Ekman layer and the geostrophic return flow at depth depends on the wind stress, and not only on the wind stress curl. The dependence of the geostrophic flow on wind stress, for a constant wind stress curl, was recently highlighted by Veronis (1996).

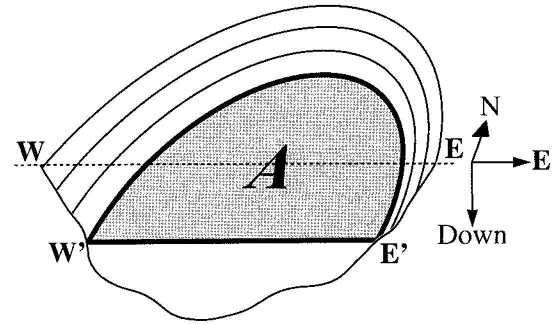

FIG. 1. Schematic to show a region of an ocean basin, cut off at a zonal section that includes surface points **W**, **E** and points **W**′, **E**′ on the constant depth surface ***A***.

When the ocean has sloping sidewalls, this has an interesting effect. Consider the section of an ocean basin shown in Fig. 1. Since we know that the zonal wind stress along the surface from **W** to **E** is balanced by form stress (i.e., pressure differences across topographic features), there must be some depth, say the depth of the surface ***A*** at which the pressure on the western boundary W′ is different from that at the eastern boundary **E**′. If we then consider the pressure along the isobath that joins **W**′ and **E**′, it is clear that there must be a place along this isobath where the pressure changes. As we will see in the next section, a change of pressure along an isobath leads to a bottom pressure torque, which must upset the Sverdrup balance. This suggests an important role for topography in gyre circulations and a relationship between form stress and Sverdrup balance that will also apply in the Southern Ocean.

This relationship also applies to the Stommel (1948) and Munk (1950) solutions in which the zonal momentum equation outside the Ekman layer was explicitly considered to be geostrophic, even in western boundary currents, so that zonal wind stress is balanced by an east–west pressure difference across the basin. However, in this case all of the variation of bottom pressure along depth contours occurs at the vertical sidewalls, resulting in a delta function of bottom pressure torque at the boundary. This cannot upset the Sverdrup balance and permit a meridional flow since such a flow cannot occur in a delta function at the boundary. Friction and nonlinear effects must therefore be the determining factors in permitting meridional flow in western boundary currents when topography only occurs in vertical sidewalls.

When the sidewalls are sloping though, the bottom pressure torque can balance a meridional flow. In the next section, it will be shown, as described briefly by Hughes (2000), that, in fact, the bottom pressure torque is of exactly the right size to balance the meridional flow required in a western boundary current. As long as that current flows above the sloping region (which it must do unless the sloping region is narrower than a viscous boundary layer), there is no need for viscous or nonlinear torques to balance the meridional flow in western boundary currents.

Section 3 then discusses diagnostics from a fully nonlinear, global, ¼° resolution numerical model: the Ocean Circulation and Climate Advanced Modelling project (OCCAM) model. These diagnostics confirm the predictions of section 2.

Finally, section 4 summarizes the main results and remaining open questions.

2. Form stress and Sverdrup balance

The two balances, Sverdrup balance and the balance between zonal wind stress and form stress, have one thing in common: they both concern the steady, depth-integrated momentum equation. In two-dimensional vector form, the steady, horizontal momentum equation may be written

$$\rho f \mathbf{k} \times \mathbf{u} = -\nabla p + \boldsymbol{\tau}_z + \mathbf{a} + \mathbf{b}, \qquad (1)$$

where f is the Coriolis parameter, $\mathbf{u}$ is the two-dimensional horizontal velocity, ρ is density, p is pressure, $\boldsymbol{\tau}$ is the viscous stress on a horizontal surface, $\mathbf{a}$ is the divergence of the remaining (lateral) viscous stress, and $\mathbf{b}$ represents terms nonlinear in $\mathbf{u}$. The individual components of $\mathbf{b}$ may be written as a zonal component $b^x = -\rho(\mathbf{u}\cdot\nabla u + wu_z + uv\tan\phi/r)$ and a meridional component $b^y = -\rho(\mathbf{u}\cdot\nabla v + wv_z + u^2\tan\phi/r)$, where r is the radius of the earth (this is the primitive equation form in which r is taken as a constant).

Performing a depth integral from the ocean floor ($z = -H$) to the free surface ($z = \eta$), we can write the depth-integrated mass transport as $\mathbf{U} = (U, V) = \int_{-H}^{\eta} \rho\mathbf{u}\, dz$. The integral of (1) then gives

$$f\mathbf{k} \times \mathbf{U} = -\int_{-H}^{\eta} \nabla p\, dz + \boldsymbol{\tau}_0 + \mathbf{A} + \mathbf{B}, \qquad (2)$$

where $\boldsymbol{\tau}_0 = \boldsymbol{\tau}(z = \eta) - \boldsymbol{\tau}(z = -H)$ is the wind stress minus the bottom frictional stress, and $\mathbf{A} = \int_{-H}^{\eta} \mathbf{a}\, dz$, $\mathbf{B} = \int_{-H}^{\eta} \mathbf{b}\, dz$. The gradient may then be taken outside the integral:

$$-\int_{-H}^{\eta} \nabla p\, dz = -\nabla \int_{-H}^{\eta} p\, dz + p_b\nabla H + p_b\nabla\eta, \qquad (3)$$

where p_b is the bottom pressure and p_a is atmospheric pressure at the sea surface. The atmospheric pressure term could be retained for completeness, but its contribution is small and it is convenient to discard it at this stage by assuming atmospheric pressure to be a constant and interpreting all pressures as the actual pressure minus atmospheric pressure, so $p_a = 0$. Writing $P = \int_{-H}^{\eta} p\, dz$, we then have

$$f\mathbf{k} \times \mathbf{U} = -\nabla P + p_b\nabla H + \boldsymbol{\tau}_0 + \mathbf{A} + \mathbf{B}. \qquad (4)$$

In a realistic ocean, with sloping sidewalls, the depth of the ocean goes to zero at its lateral boundaries ($-H = \eta$). At such boundaries, $p_b = p_a = 0$ and, being depth integrals of finite quantities, $\mathbf{U}$, P, $(-\nabla P + p_b\nabla H)$, $\mathbf{A}$, and $\mathbf{B}$ all go to zero. This is also true of $\boldsymbol{\tau}_0$, meaning that the bottom stress must equal the wind stress at the boundary. Thus, although it is probably reasonable to think of $\boldsymbol{\tau}_0$ as being the wind stress in the ocean interior, it must drop to zero at the boundary as the bottom stress increases to balance the wind stress there.

A zonal integral of (4) around either a closed latitude circle or from the western boundary to the eastern boundary of an ocean basin leads to the angular momentum, or zonal momentum balance:

$$f\int_W^E V\, dx = \int_W^E p_b H_x + \tau_0^x + A^x + B^x\, dx, \qquad (5)$$

where dx is shorthand for $r\cos\phi d\lambda$ (λ is longitude), and a superscript x represents the zonal component of a vector. If the basin is closed to the north or south or the integral is around a closed contour, then the left-hand side must be zero unless there is a source or sink of mass. In reality there may be sources or sinks of order 1 Sv (10^6 m^3 s^{-1}), compared with a typical Ekman transport of 10 Sv. The terms on the right-hand side represent form stress, zonal (wind − bottom) stress, lateral friction, and nonlinear terms. As discussed in the introduction, it is now firmly established that the main balance at each latitude is between form stress and wind stress, with the other terms playing a minor role.

Arguments involving Sverdrup balance stem from taking the curl of (4), which leads to what is usually termed the barotropic vorticity (BV) equation:

$$\nabla\cdot(f\mathbf{U}) = \nabla \times (p_b\nabla H) + \nabla \times \tau_0 + \nabla \times (\mathbf{A} + \mathbf{B}). \qquad (6)$$

See the appendix for a derivation of the same equation by taking the curl of the momentum equation (1) to produce the vorticity equation and then depth integrating. Although the main topographic influence appears here in the bottom pressure term, it is worth noting that

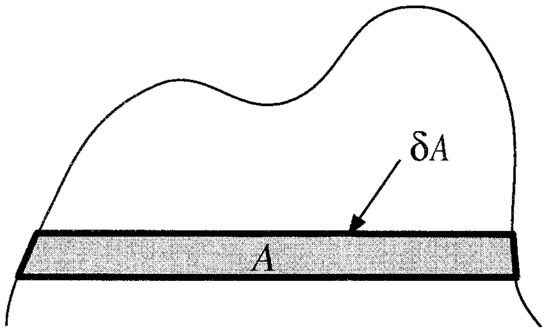

FIG. 2. Schematic to show an area $\boldsymbol{A}$, and its bounding curve $\delta\boldsymbol{A}$, which consists of two latitude lines and two stretches of coastline.

the nonlinear term may indirectly reflect topographic interactions, for example via its effect on oscillatory motions that may lead to a nonzero time average of $\mathbf{B}$. If there are no mass sources, $\nabla\cdot\mathbf{U} = 0$, so $\nabla\cdot(f\mathbf{U}) = \beta V$ where $\beta = f_y$. Sverdrup balance then results if the nonlinear, lateral viscous and bottom stress and bottom pressure torque terms are negligible, giving Sverdrup balance:

$$\beta V = \nabla \times \boldsymbol{\tau}_w, \qquad (7)$$

where $\boldsymbol{\tau}_w$ is the surface wind stress. The bottom pressure torque in (6) is $\nabla \times (p_b \nabla H) = \nabla p_b \times \nabla H$, so it is nonzero wherever there is a gradient in pressure along a depth contour. As noted in the introduction, this implies a relationship between bottom pressure torque and form stress and, hence, between bottom pressure torque and zonal wind stress.

That argument can be localized by considering the geometry of Fig. 2, a strip $\boldsymbol{A}$ of ocean basin bounded by a line $\delta\boldsymbol{A}$ that consists of two latitude lines and two short stretches of coastline (the coastline being defined as where the water depth goes to zero, $-H = \eta$). Consider the integral of the bottom pressure torque over this area. We can use Stokes' theorem to write

$$\int_A \nabla \times (p_b \nabla H)\, dS = \oint_{\delta A} p_b \nabla H \cdot \mathbf{ds}. \qquad (8)$$

Since $p_b = 0$ at the coast, there is no contribution to the line integral from the two short coastline stretches, leaving only the zonal integrals:

$$\int_A \nabla \times (p_b \nabla H)\, dS = \int_{\phi_1} p_b H_x\, dx - \int_{\phi_2} p_b H_x\, dx, \qquad (9)$$

where ϕ_1 and ϕ_2 represent the southern and northern bounding latitudes. So the bottom pressure torque, integrated over area A, is simply the difference between the zonal integrals of form stress at the two bounding latitudes.

The same argument can be made for each of the terms in the BV equation (6), using Stokes' theorem to relate the area integral to the corresponding zonal integral in the angular momentum balance (5). So, for the (wind − bottom) stress we have

$$\int_A \nabla \times \boldsymbol{\tau}_0\, dS = \int_{\phi_1} \tau_0^x\, dx - \int_{\phi_2} \tau_0^x\, dx. \qquad (10)$$

It therefore becomes clear from (9) and (10) that, if the zonal (wind − bottom) stress is balanced by form stress in zonal integrals at each latitude, then the area integral of (wind − bottom) stress curl is balanced by the area integral of bottom pressure torque for all areas bounded by latitude lines. If we consider an infinitesimal area bounded by two latitude lines separated by a distance δy, the area integral reduces to δy times the zonal line integral of terms in the BV equation (6), showing that we can write the line integral of each of these terms as d/dy of the corresponding term in the angular momentum balance (5). Since we know that the primary balance in (5) is between wind stress and form stress, we must conclude that wind stress curl is balanced by bottom pressure torque in the area integral of (6).

In the vertical sidewall case, that bottom pressure torque occurs as a delta function at the sidewalls, so viscosity and/or nonlinear terms are required at each latitude to balance the meridional flow, which cannot occur in a delta function. This situation would persist as long as any sloping sidewall region is narrower than a viscous or nonlinear boundary layer. If the sloping sidewall region is broader than a viscous boundary layer, then the western boundary current would flow in a region where bottom pressure torques occur and are sufficient to balance the boundary current βV. In this case there is no need for viscous torques to balance wind stress curl at each latitude.

There are two complications to this argument that must be considered: First, that $\boldsymbol{\tau}_0$ is (wind–bottom) stress and the bottom stress must balance the wind stress at the coast. This means that we cannot simply interpret $\nabla \times \boldsymbol{\tau}_0$ as the wind stress curl in these integrals. However, the only case where this is a problem is when there is an alongshore wind stress at the coast. This can be seen by separating the integral of $\nabla \times \boldsymbol{\tau}_0$ into components due to wind stress ($\boldsymbol{\tau}_w$) and bottom stress ($\boldsymbol{\tau}_b$):

$$\int_A \nabla \times \boldsymbol{\tau}_0\, dS = \int_A \nabla \times \boldsymbol{\tau}_w\, dS - \int_A \nabla \times \boldsymbol{\tau}_b\, dS. \qquad (11)$$

Using Stokes' theorem, we have

$$\int_A \nabla \times \boldsymbol{\tau}_b\, dS = \oint_{\delta A} \boldsymbol{\tau}_b \cdot \mathbf{dS}. \qquad (12)$$

The line integral consists of two zonal integrals that we know to be negligible, since the predominant balance is between zonal wind stress and form stress, and two integrals along the coast, where we know that $\boldsymbol{\tau}_b = \boldsymbol{\tau}_w$. Hence, it is only the alongshore wind stress that significantly upsets the interpretation of $\nabla \times \boldsymbol{\tau}_0$ as $\nabla \times \boldsymbol{\tau}_w$. Most analytical and idealized modeling studies deliberately choose to have no alongshore wind stress in order to avoid this problem. In reality there will be an alongshore wind stress that will result in a viscous boundary layer (as $\boldsymbol{\tau}_b$ increases from zero away from the coast to $\boldsymbol{\tau}_w$ at the coast, $\nabla \times \boldsymbol{\tau}_b$ must become significant in the boundary layer). This is clearly a separate issue from the question of whether western boundary currents are viscous since it occurs at all boundaries, east or west (or others), and the size of the effect is completely determined by the strength of the alongshore wind stress. It is an effect that has been known for a long time and is described in detail for the vertical sidewall case by Pedlosky (1968).

The second complication comes from the different meridional length scales of terms in the angular momentum balance (5). Although the primary balance is between wind stress and form stress at each latitude, the area integral of the BV equation is related to the difference between terms in (5) evaluated at the two bounding latitudes. It is possible for wind stress and form stress to be the dominant terms at each of the bounding latitudes, but for the difference to be dominated by other terms. This is particularly true if the area A is very narrow or, in the extreme case, as the width tends to zero. In that case it is not the integral of wind stress, etc., that matters, but d/dy of those integrals. Although the nonlinear term is much smaller than the wind stress, it varies much more rapidly with latitude, so we should expect the nonlinear term to be large in a zonal integral of the BV equation. This is what Wells and de Cuevas (1995) found in diagnostics from the Fine Resolution Antarctic Model. Far from being a Sverdrup balance, the primary balance in the BV equation integrated zonally is between nonlinear torque and bottom pressure torque, with the wind stress curl a small residual. It is only when integrating over an area wide enough for nonlinear torques to average out that the wind stress curl should become significant. If the area is wide enough for the zonally integrated wind stress to change by a significant fraction of its total magnitude from north to south, then, clearly, smaller terms in the angular momentum balance cannot dominate the area integrated BV balance.

So, to summarize, the area integral of each of the terms in the BV equation (6), over a zonal strip as shown in Fig. 2, is equal to the difference between evaluations at the two bounding latitudes of the corresponding term in the angular momentum equation (5). Since the zonal wind stress is predominantly balanced by form stress in (5), this means that wind stress curl is balanced by bottom pressure torques in an area integral of (6), with two caveats: if the area is very narrow, terms that appear small in (5) but vary on short meridional length scales can dominate; any alongshore wind stress at the coast results in a viscous boundary layer near the coast in which the curl of viscous bottom stress balances some (easily calculable) part of the area-integrated wind stress curl.

These are integral balances that say nothing about where (at which longitude) the bottom pressure torques occur. Since any torque is balanced by βV, it seems natural to assume that the pressure torque is occurring in the western boundary current where βV is largest (except in the special case of vertical sidewalls when βV cannot occur in the same place as the pressure torque), but this cannot be confirmed from zonal integral arguments, so the variation in longitude must be resolved explicitly with a complete ocean model to clarify this.

3. Model diagnostics

a. The model

The model we will use is the Ocean Circulation and Climate Advanced Modelling project model, described in Killworth (1995). This is a global, primitive equation model in spherical polar coordinates on two grids. One grid has poles coinciding with the earth's rotational poles, and covers most of the World Ocean. The other has poles on the equator and an ''equator'' along the 38°W meridian and covers the North Atlantic and Arctic Oceans. Both grids are at 0.25° horizontal resolution and are joined seamlessly at the Atlantic equator and via a channel model at Bering Strait. There are 36 vertical levels with thickness increasing from 20 m at the surface to about 255 m at the maximum depth of 5500 m. Topography in this model (as in most other level models) can only occur at these 36 specified depths (or zero), resulting in a rather blocky topographic representation in some regions and particularly in difficulties in representing smooth slopes at depth (where the levels are thicker and the slopes more gradual).

The main differences from similar level models are that OCCAM has a free surface and uses the ''Split-QUICK'' advection scheme (Webb et al. 1998). The model was spun up over 4 yr by relaxation of model temperature and salinity fields toward the Levitus 94 annual mean climatological values (Levitus and Boyer 1994; Levitus et al. 1994) with a relaxation timescale of 30 days, with climatological wind stress forcing with monthly average values matching those of Siefridt and Barnier (1993). After 4 years, the relaxation to Levitus 94 was switched off in all but the top level where it was retained as a proxy for surface heat and freshwater fluxes (relaxation of salinity is performed by a flux of freshwater, so the relaxation also implies sources and sinks of mass). The model was run for a further 4 years to permit the eddy field to develop before reaching the ''analysis'' phase from which the results in this section are taken

The diagnostics presented here are all averages of

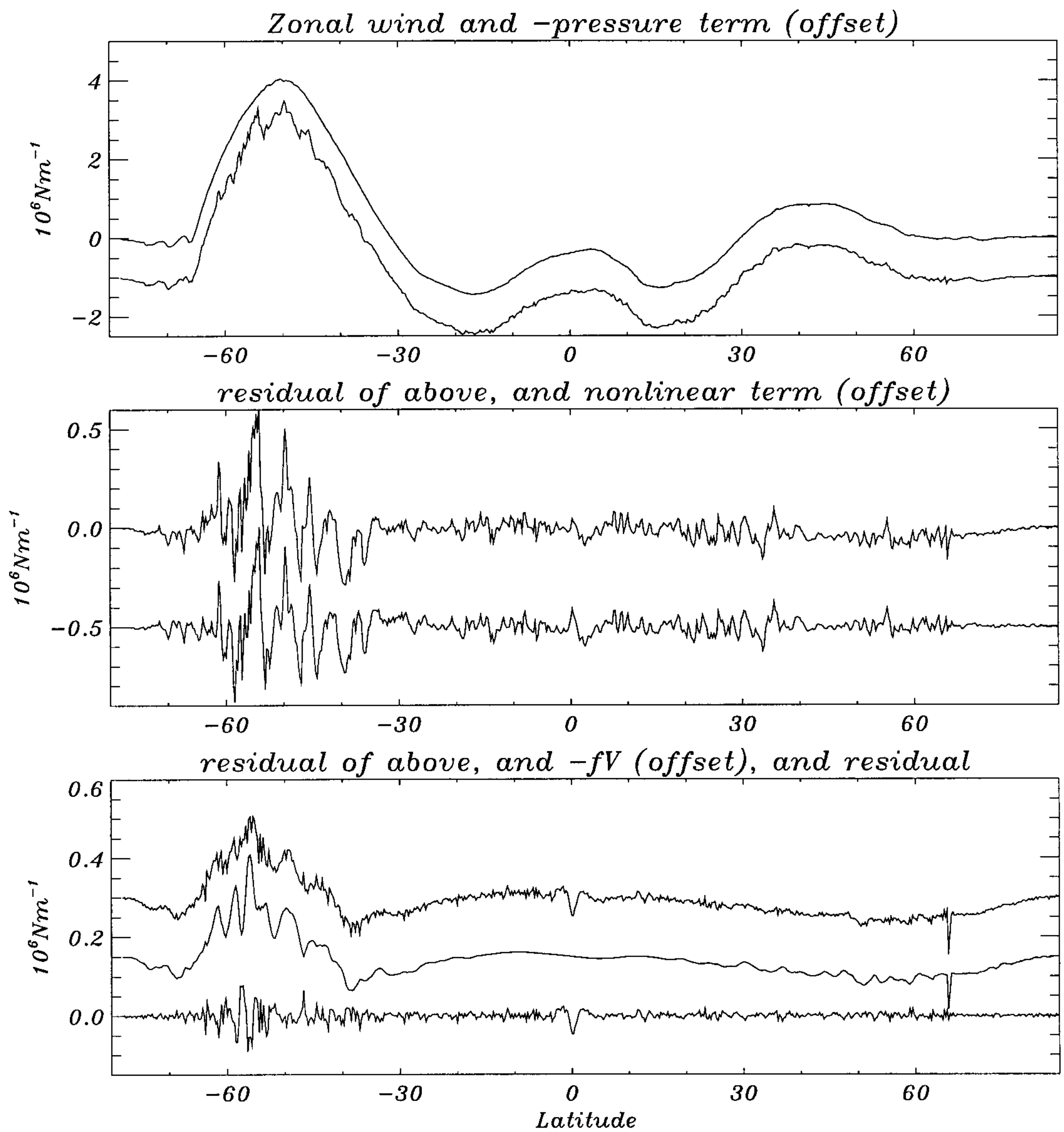

FIG. 3. Terms in the angular momentum balance (5) from OCCAM. The curves plotted are zonal integrals of, from top to bottom, τ^x (wind − bottom stress), $-p_bH_x$ (−form stress), $-(\tau^x + p_bH_x)$, b^x (nonlinear), $-(\tau^x + p_bH_x + b^x)$, $-fV$ (Coriolis), $-(\tau^x + p_bH_x + b^x + fV)$. See text for full definitions. Divide by $f\rho$ to get equivalent volume transports.

diagnostics from 98 model dumps, taken every 15 days, covering 4 model years (years 8.0 to 12.0, days 2925 to 4383).

b. The angular momentum balance

The terms in the angular momentum balance (5) were evaluated from the model in a manner consistent with the model dynamics. The zonal integration is straightforward on the main model grid. On the rotated grid, grid points do not lie along lines of latitude, so the integral was performed along a zigzag line along sides of grid boxes, approximating each latitude of the nonrotated grid. Tests with small changes to the path or method of integration showed that the results are not sensitive to this choice.

The results for a 4-yr average are shown in Fig. 3. The primary balance (top panel) is between (wind–bottom) stress and form stress. The residual of this balance, $-\oint \tau_0^x + p_bH_x\,dx$, shown in the middle panel (with vertical range reduced by a factor of 5) is mostly balanced by the nonlinear term. The bottom panel (with a range reduced by a further factor of 2) shows the residual of this balance, $-\oint \tau_0^x + p_bH_x + b^x\,dx$, to be mainly balanced by Coriolis ($-f\oint V\,dx$). The final residual is dominated by lateral friction (the time dependence term is another order of magnitude smaller for the 4-yr average).

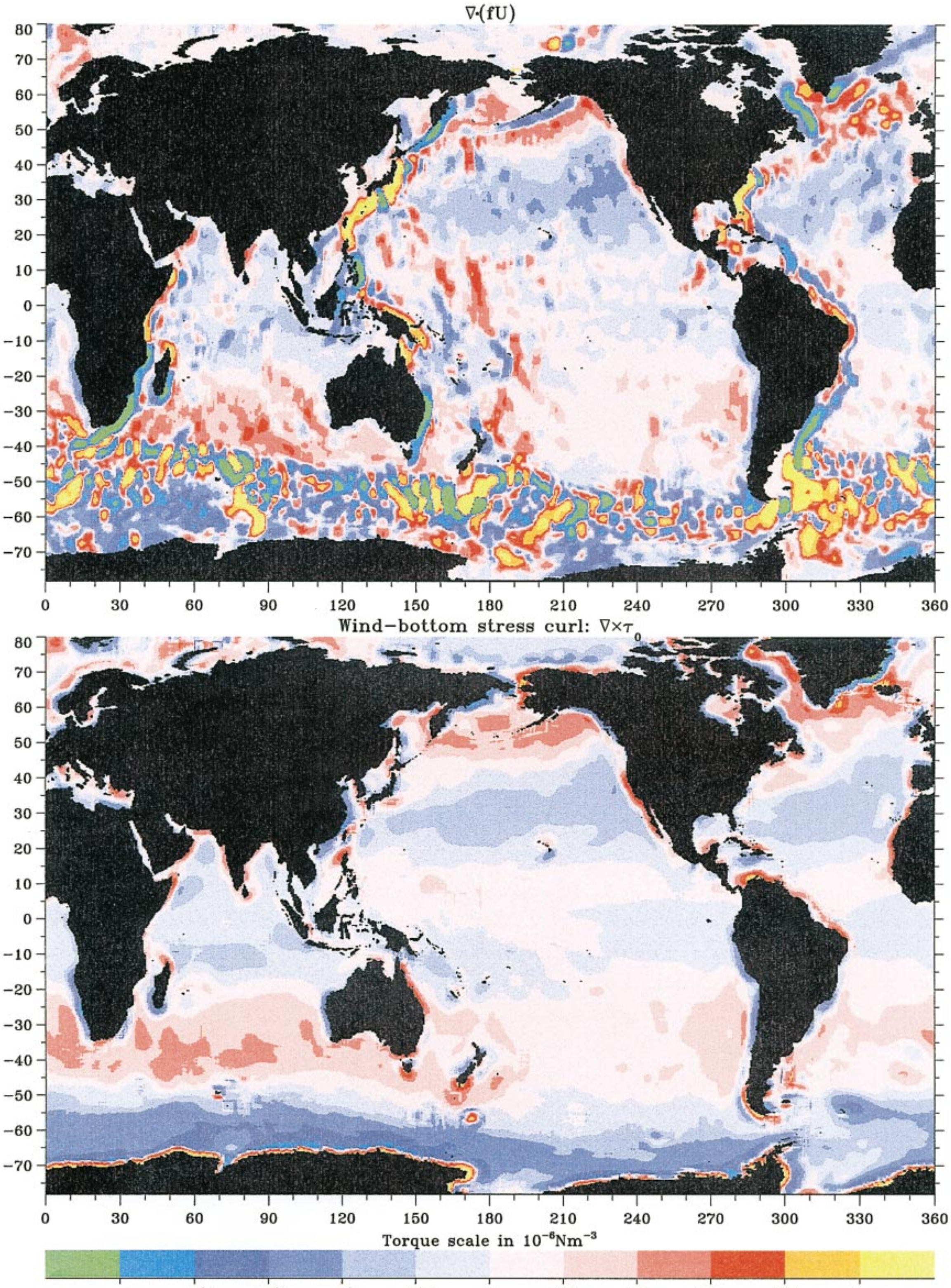

FIG. 4. Terms from the barotropic vorticity equation (6), diagnosed from OCCAM and averaged over 4 years and an area of 1.75° degrees of latitude by 4° degrees of longitude. The final panel, showing the lateral friction term, has an approximate streamfunction for the depth-integrated flow superimposed, to show the main features of the flow. The contour interval changes from 5×10^6 m^3 s^{-1} north of 40°S to 15×10^6 m^3 s^{-1} south of that latitude, in order to represent both gyre circulations and the ACC on the same plot.

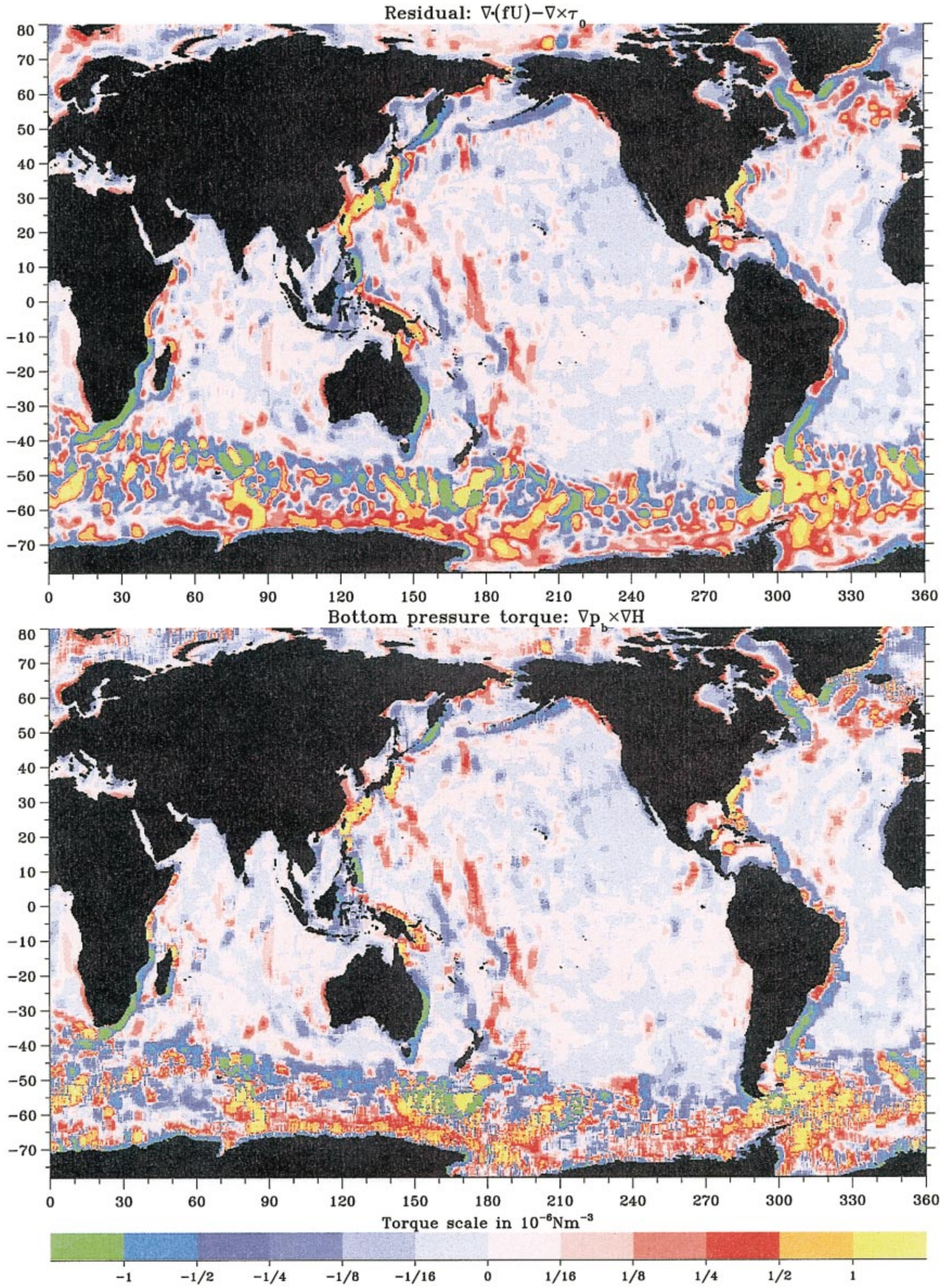

FIG. 4. (*Continued*)

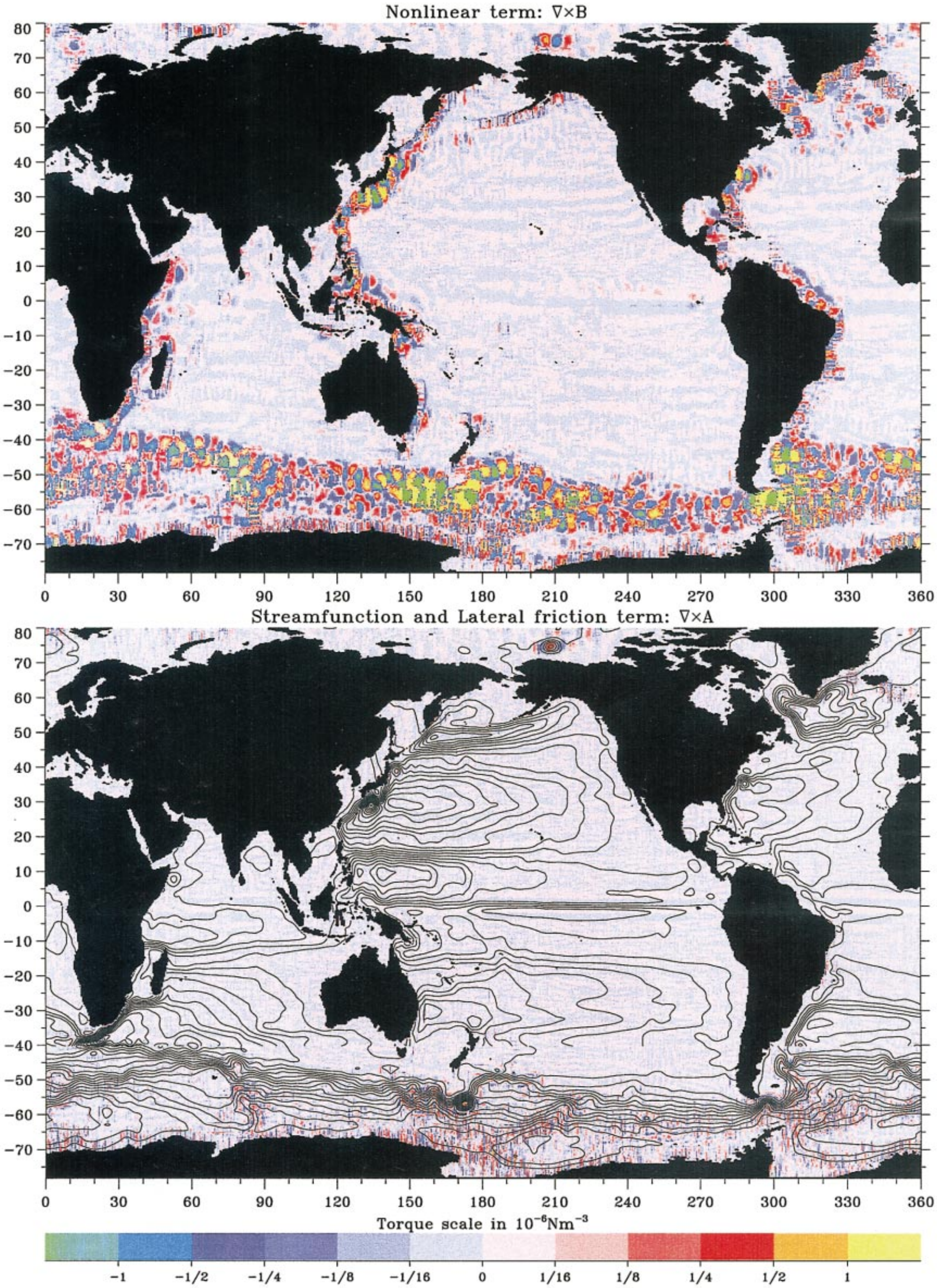

FIG. 4. (*Continued*)

The Coriolis term is larger than in the real ocean as a result of the surface freshwater flux that acts to restore the surface salinity values to climatology. For comparison, at midlatitudes (taking $f = 10^{-4}$ s^{-1}), a northward transport of 1 Sv (10^6 m^3 s^{-1}) represents a Coriolis force of 10^5 N m^{-1}. Also, the spike at about 66°N is due to the channel model at Bering Strait, which has not been included in the diagnostic calculation: some flow occurs in the channel model and is therefore absent from the main model, as are the dynamical terms balancing that flow.

Although there is a balance between wind stress and form stress at each latitude to within about 10%, the relatively short length scale of variability in the nonlinear and form stress terms means that the gradients do not balance at each latitude, so the bottom pressure torque does not balance wind stress curl in a zonal integral at a single latitude. Typically, the latitude difference for which differences in zonal wind stress and form stress approximately balance is about three degrees of latitude, although a larger separation is necessary in parts of the Southern Ocean. Despite the shorter length scale of the lateral friction term, the main contribution to the imbalance of gradient comes from the nonlinear term, so the balance of gradients at each latitude is dominated by pressure and nonlinear terms, as in Wells and de Cuevas (1995). Apart from its role in balancing the Coriolis force due to the artificially large sources and sinks of freshwater in the Southern Ocean, the only place where lateral friction has a significant effect is at the equator.

c. The barotropic vorticity balance

The barotropic vorticity balance for OCCAM western boundary currents in the Pacific was evaluated by Saunders et al. (1999), demonstrating, as the theory suggests, that bottom pressure torques are the dominant term balancing βV in these regions, with nonlinear torques also important in some cases. Here, we extend this analysis to include all regions of the World Ocean.

The terms in the BV equation (6) were evaluated from the model in the form

$$\nabla \times \left(\boldsymbol{\tau}_0 - \int_{-H}^{\eta} \nabla p \, dz - f\mathbf{k} \times \mathbf{U} + \mathbf{A} + \mathbf{B}\right) = 0. \tag{13}$$

For the continuous equations, these are exactly equivalent to the terms in (6). They were evaluated in this form in order to be completely consistent with the finite difference form of the momentum equation used by the model.

Consideration of the gradients in Fig. 3 suggests that a plot of this balance at each grid point would be dominated by large positive and negative values on small length scales, and it turns out just so. In fact, this problem is exacerbated by the way in which the model handles bottom topography: since only 36 ocean depths are permitted, there are many grid points at which the depth gradient is zero. The topography in deep regions takes the form of a series of terraces, so, if the effect of slopes is to be investigated, it is necessary to average over areas larger than the width of the terraces. Since the bottom pressure torque is nonzero only where there is a slope, the terraces introduce small scales and artificially high viscous and nonlinear forces into the dynamics. Some smoothing is therefore necessary if anything useful is to be seen.

Taking the length scales from Fig. 3 as a guide, the terms in (13) were smoothed by taking an area-weighted average of the values within 2 degrees of longitude and 1.375 degrees of latitude of each point (this represents an average over approximately 472 $\cos\phi \times 305$ km, or 17×11 grid points on the main grid). Other length scales were also investigated, giving a slow reduction in the nonlinear terms as scale increases and a rapid increase in nonlinear terms as the scale decreases.

These terms, together with $\nabla\cdot(f\mathbf{U}) - \nabla \times \boldsymbol{\tau}_0$, [i.e., the part of $\nabla\cdot(f\mathbf{U})$ that is not explained by Sverdrup balance] are plotted in Fig. 4. An approximate barotropic streamfunction (calculated as the integral of U/ρ_0 from south to north), smoothed in the same way, is superimposed on the lateral friction torque $\nabla \times \mathbf{A}$, with a contour interval of 5 Sv north of 40°S, and 15 Sv south of that latitude (such a change in contour interval is necessary to represent the Southern Ocean on the same plot as the rest of the world, so large are the transports in this region). It is worth noting that, since there are sources and sinks of mass in the model, $\nabla\cdot(f\mathbf{U})$ is not exactly equal to βV, although the difference is very small.

Clearly, the lateral friction term is very small and plays a negligible role in western boundary currents (in fact almost everywhere). This is somewhat surprising given the model's representation of topography, which artificially increases the predominance of near vertical slopes bounding the flat terraces. Indeed, without spatial smoothing, the friction term is large but with grid-scale oscillations that average to near zero on smoothing. The remnants of these large frictional terms can be seen in parts of the Southern Ocean where grid-scale noise remains in the friction term even after smoothing. On any scale small enough that the ''steppiness'' of the topography is apparent, the details of the model flow must therefore be considered suspect. This problem can only be solved in a level model by permitting partial grid cells at the ocean floor so that the topography is not limited to a fixed number of levels.

Given the smallness of the lateral friction term, a gyre in Sverdrup balance with a viscous western boundary current would show as a balance between βV and $\nabla \times \boldsymbol{\tau}_0$ (since $\nabla \times \boldsymbol{\tau}_0$ includes bottom stress). Boundary layers do show in $\nabla \times \boldsymbol{\tau}_0$, as expected, since they are needed for bottom friction to balance the alongshore wind stress. These are equally strong at eastern and western boundaries and are much smaller in amplitude than βV in western boundary currents. At some eastern boundaries, such as the west coasts of South America and southern Africa, these viscous torques are balanced by βV in boundary currents. In other places, such as the west coasts of North Africa and Australia, they are mainly balanced by the bottom pressure torque, and in some places there is a three-way balance. Away from the boundaries, however, the wind stress curl does balance βV on the large scale, over much of the tropical and subtropical ocean, demonstrating the relevance of Sverdrup balance to these regions.

In the western boundary currents, the Southern Ocean, and high northern latitudes, the main term balancing βV is the bottom pressure torque, as surmised from the integral balance. In a zonal integral, of course, the wind stress curl must be as important as the bottom pressure torque, since βV integrates to (almost) zero. However, most of the basin-scale structure in the Antarctic Circumpolar Current is clearly due to the bottom pressure torque, as is the northward deflection of the current in the vicinity of Drake Passage. The idea that much of the Southern Ocean is in Sverdrup balance is clearly inapplicable since much of the southward flow occurs in regions of very small wind stress curl, with somewhat weaker southward flow in the regions of strong negative wind stress curl farther south. This situation is possible because of the strong negative bottom pressure torque along the northern part of the ACC, balanced by large positive values farther south. The bottom pressure torque is also important in a number of small-scale structures embedded in the Sverdrup regions, particularly in the eastern Pacific and central Atlantic. Although these features are visible in the streamfunction, they are generally not visible in sea surface height fields, being mostly due to deep boundary currents flowing along topographic features.

The nonlinear term continues to be important locally, especially in the Southern Ocean and in western boundary currents, although it seems to acount for small-scale structure and meanders in the streamfunction rather than any basin-scale effects. The bottom pressure torque closely matches the Sverdrup imbalance (difference between $\nabla\cdot(f\mathbf{U})$ and $\nabla \times \boldsymbol{\tau}_0$), with the nonlinear term filling in details. In western boundary currents, the nonlinear term is strongest in regions of strong curvature, such as the meanders in the model Kuroshio, or the Gulf Stream separation point.

So, the model analysis confirms that the main term acting to balance βV in western boundary currents is the bottom pressure torque, which is also the term responsible for the northward deflection and other meridional excursions of the ACC and other high-latitude currents, and is balanced by the wind stress curl in a zonal integral averaged over a few degrees of latitude. The Southern Ocean can therefore be considered in the same conceptual framework as the rest of the World Ocean, a framework in which these integral balances are essentially inviscid. Within the model it is also the case that the basin-scale dynamics of subtropical gyres can be roughly described by Sverdrup balance, deep boundary currents excepted. This is not the case at high latitudes, either the Southern Ocean or North Atlantic (the North Pacific is arguable, with its awkward geometry), where bottom pressure torques are important at all longitudes and cannot be considered to be confined to western boundaries. As noted in the introduction, this is to be expected, as the baroclinic Rossby wave speed is slow at high latitudes (the stratification is weak), so the flow tends to be closer to barotropic and to follow f/H contours.

The model, of course, has its limitations. The resolution permits rather than resolves eddies, the lateral friction is set by requirements of numerical stability rather than physical arguments, and the representation of topography upsets the flow on small scales. Nonetheless, it produces a realistic circulation with essentially inviscid western boundary currents in which nonlinearity plays a significant, but secondary, role. Given that the real ocean is less viscous than any model and given the generality of the argument predicting inviscid western boundary currents, it seems reasonable to conclude that the real ocean also behaves this way. Caution would be advisable, though, in interpreting the details of the nonlinear term that depends on processes acting at scales near the model resolution. Higher resolution and better representation of topography could significantly alter things like meander scales (influenced by how narrow a jet is, for a given transport) and boundary current separation.

4. Summary and discussion

Comparison of the angular momentum equation (zonal and depth integral of the zonal momentum equation) with the area integral of the BV equation over a zonal strip, as in Fig. 2, has shown that, whatever terms dominate the former balance, the corresponding terms dominate the latter, with the caveat that terms with short meridional length scales can have a disproportionate effect for very narrow strips. This means that, for strips a few degrees of latitude wide, the dominant balance in the area-integrated BV equation must be between wind stress curl and bottom pressure torque (with bottom stress curl also playing a role when there is alongshore wind stress at the coast). This removes the need, present in the vertical sidewall case, for a viscous western boundary current in which viscous torques balance the wind stress curl. Instead, if the ocean interior is in Sverdrup balance, the return flow will be in a topographically steered current in which βV is balanced by the bottom pressure torque.

Note that this does not mean viscosity is unimportant in obtaining a balance, only that it is unimportant in these particular zonal and/or depth integrals. In just the same way, large-scale currents in the ocean are close to geostrophic balance, even though ageostrophic effects control the position and strength of those currents. This is a diagnostic, not a prognostic, relationship.

In the introduction it was noted that several idealized models have been constructed recently in which the western boundary currents are inviscid. This new result suggests that inviscid western boundary currents should be expected in any ocean in which the zonal wind stress is balanced by bottom form stress, as is believed to be the case for the real ocean. Inviscid western boundary currents should be considered the rule, and the viscous boundary currents found in oceans with vertical sidewalls an exception that only occurs when the continental slope width is narrower than the Munk boundary layer.

This prediction was tested by calculating diagnostics from a less idealized model, OCCAM. The results confirm the prediction: in OCCAM western boundary currents, βV is predominantly balanced by the bottom pressure torque. Nonlinear torques also play an important local role in certain regions, but lateral viscous torques are negligible and bottom stress curl introduces weaker boundary layers at all coasts where there is alongshore wind stress.

For a purely barotropic flow, the dominance of bottom pressure torque over viscous and nonlinear terms would imply flow along f/H contours. With stratification, it seems reasonable to assume that the equivalent flow would be planetary geostrophic in nature, conserving linearized potential vorticity, but only a depth-integrated (and spatially smoothed) balance has been investigated here, so other possibilities cannot be ruled out.

These zonal integral relationships say nothing about where the bottom pressure torques occur or why we should expect western boundary currents rather than eastern boundary currents. In fact, we should not expect a clear identification of the position of the boundary current from general arguments including topography. This can be seen by considering the case of a barotropic ocean with bottom topography. In this case, Welander (1968) neatly demonstrated how a circulation forms with boundary currents situated at an ‘‘effective western boundary’’ determined by f/H contours rather than f contours. The position of this effective western boundary depends on the geometry of the ocean basin: for a basin with vertical sidewalls and f/H decreasing toward the pole it is, in fact, the eastern boundary.

In a stratified ocean, the corresponding characteristics can vary between f contours and f/H contours depending on the strength of stratification, which is determined internally by many complicated processes. Clearly a general argument cannot specify where a boundary current appears; that depends on the thermohaline forcing and internal dynamics. Salmon’s (1998) paper is quite enlightening and goes some way to explaining this but, being based on a linearization about a fixed stratification, it cannot account for how large variations in stratification are maintained or how they modify the flow.

So where does this leave us? If western boundary currents are essentially inviscid, it leaves us with a big question about the role of viscosity in the ocean: does it actually control the circulation? If it does, where does it act? It could still act in western boundary currents, but in a manner that does not upset the integral balances described here, or it could act in the eddy-rich regions downstream of western boundary currents after they separate from the continental slope. We are also left with a greater appreciation of the importance of bottom pressure as a parameter that relates to western boundary current dynamics. With the Gravity Recovery and Climate Experiment satellite (Hughes et al. 2000) due for launch in 2001 and promising to be able to measure changes in ocean bottom pressure with millimeter accuracy (averaged over large areas), this may be significant. Perhaps most importantly, we can see how the dynamics of the Southern Ocean connect with the dynamics of ocean basins. The link between the Munk and Palmén (1951) concept of balance between wind stress and form stress in the Southern Ocean, and the Stommel (1948) and Munk (1950) concept of a gyre in Sverdrup balance closed by a western boundary current, has been made much more transparent. In both cases, the wind stress is balanced by form stress, and in both cases this leads to a bottom pressure torque that permits meridional motions to close the circulation implied by the wind stress curl.

Of course the Southern Ocean remains significantly different from other regions of the World Ocean, as the diagnostics in Fig. 4 show quite clearly. What we have shown here is that the reason for this difference does not lie in either the question of how zonal wind stress is balanced or the question of how wind stress curl is balanced in a zonal integral, as the answers to those questions are the same for both regions. It must be other characteristics of the Southern Ocean that make it special, such as the possibility of closed linearized potential vorticity contours that are close to zonal or are close to being parallel to the wind stress.

Acknowledgments. I would like to thank the many people who have read and commented on versions of this paper over the years, and the two anonymous referees who offered constructive suggestions. This work was supported by the U.K. Natural Environment Research Council and the U.K. Defence Research Agency.

APPENDIX

Alternative Derivation of the Barotropic Vorticity Equation

The barotropic vorticity equation (6) can also be derived by integrating the vorticity equation over depth, as follows. Starting with the momentum equation (1), repeated here,

$$\rho f \mathbf{k} \times \mathbf{u} = -\nabla p + \boldsymbol{\tau}_z + \mathbf{a} + \mathbf{b}, \qquad \text{(A1)}$$

take the curl to obtain the vorticity equation:

$$\nabla\cdot(\rho f\mathbf{u}) = \nabla \times (\boldsymbol{\tau}_z + \mathbf{a} + \mathbf{b}) \qquad \text{(A2)}$$

(all vectors here are taken to be two-dimensional in the horizontal, except for $\mathbf{k}$ which is the vertical unit vector. Accordingly, the curl operator can be taken as shorthand for $\mathbf{k}\cdot\nabla\times$). An integral over depth (for simplicity, the integral is taken from the ocean floor at $z = -H$ to $z = 0$; the effect of a free surface can be taken into account in the same way that the variable bottom topography is treated here) gives

$$\int_{-H}^{0} \nabla \cdot (\rho f \mathbf{u})\, dz = \int_{-H}^{0} \nabla \times (\boldsymbol{\tau}_z + \mathbf{a} + \mathbf{b})\, dz. \qquad \text{(A3)}$$

The curl and divergence operators can be taken outside the vertical integration at the expense of introducing terms due to the horizontal variation of the limits of the integral (H in this case), giving

$$\begin{aligned}\nabla \cdot (f\mathbf{U})dz &= \nabla \times (\boldsymbol{\tau}_0 + \mathbf{A} + \mathbf{B}) \\ &\quad + ((\boldsymbol{\tau}_z)_b + \mathbf{a}_b + \mathbf{b}_b) \times \nabla H \\ &\quad + \rho_b f \mathbf{u}_b \cdot \nabla H, \qquad \text{(A4)}\end{aligned}$$

where the subscript b represents quantities evaluated at the bottom ($z = -H$). As expected, pressure has been eliminated by taking the curl. However, the bottom pressure can be reintroduced using (A1) $\times\ \nabla H$ evaluated at the bottom to give

$$\begin{aligned}&((\boldsymbol{\tau}_z)_b + \mathbf{a}_b + \mathbf{b}_b) \times \nabla H \\ &\quad = -\rho_b f \mathbf{u}_b \cdot \nabla H + \nabla p_b \times \nabla H. \qquad \text{(A5)}\end{aligned}$$

Combining (A4) and (A5) then gives the BV equation, (6):

$$\begin{aligned}&\nabla \cdot (f\mathbf{U})dz \\ &\quad = \nabla \times (\boldsymbol{\tau}_0 + \mathbf{A} + \mathbf{B}) + \nabla p_b \times \nabla H. \qquad \text{(A6)}\end{aligned}$$

From (A5) we see that the bottom pressure torque can be interpreted as the uphill flow at the bottom for a free-slip boundary condition and no friction at the bottom, in the linear case. With a no-slip bottom boundary condition, the bottom pressure torque can be rewritten in terms of viscous terms, although it is worth noting that these are the viscous stress divergence at the bottom, that is, the viscous force per unit volume on water at the bottom, not the viscous stress exerted on the ocean floor.

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