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Effects of Bottom Friction on Nonlinear Equilibration of an Oceanic Baroclinic Jet

PASCAL RIVIÈRE

LEMAR, IUEM, Technopôle Brest-Iroise, Plouzané, France

ANNE MARIE TREGUIER AND PATRICE KLEIN

LPO, IFREMER Centre de Brest, Plouzané, France

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ABSTRACT

Bottom friction is an important sink of energy in the ocean. Indeed, high-resolution ocean models need bottom friction to achieve a satisfactory kinetic energy level at equilibrium. However, bottom friction has also subtle and discriminating effects on the different energy transfers and therefore on the 3D structure of the flow, some of which have to be clarified. In this study, those effects on an unstable baroclinic jet are reexamined using a primitive equation model. As in previous studies using quasigeostrophic models, it was found that bottom friction strongly affects the barotropic mode whereas the baroclinic modes are weakly changed. The new result is that bottom friction yields a significant space-scale selection. Analysis of the dynamics reveals strong agreement with previous quasigeostrophic studies at the mesoscale in the interior but differences in the eddy field at small scales close to the surface. A rationalization of these results is proposed by a comparison with preceding atmospheric studies. It is shown that the "barotropic governor" of James and Gray is not active in ocean simulations and that the scale selection induced by bottom friction is primarily induced by nonlinear interactions in the three-dimensional structure of the eddy field.

1. Introduction

Midlatitude ocean circulations are characterized by a strong variability in space and time, as revealed by satellite observations (Stammer 1998), that is due to ocean eddies. The mesoscale eddy dynamics are driven by strongly nonlinear mechanisms involving instabilities and interactions with the mean zonal circulation. Ocean eddies play an important role in the exchange of properties in and between the ocean basins (transport of heat and salt but also biogeochemical tracers). One of the main challenges is to better represent these dynamics in regional or global ocean models that become more and more realistic.

Numerous studies have investigated the sensitivity of mesoscale dynamics to parameters in linear and nonlinear approaches. They have revealed the existence of key parameters such as the β effect or the vertical and horizontal structure of the mean flow (Pedlosky 1983; James and Gray 1986). A less straightforward mechanism acting on mesoscale dynamics is the dissipative effect of friction over the ocean bottom. The interior of the ocean can be thought of as quasi inviscid, the viscosity terms being negligible when compared with Cor-

iolis and advection terms. Nevertheless, close to the bottom, viscosity becomes more important in the Ekman layer to cancel the velocity at the bottom. This dissipative effect acts on long time scales (several hundred days) as compared with the characteristic time scale of the baroclinic instability process (one to several weeks), but it probably represents the main energy sink for the mesoscale circulation. Indeed, estimates of bottom frictional effect on the Gulf Stream meanders and rings reveal that it may represent between 50% and 100% of the energy dissipation of the Gulf Stream system (Weatherly 1984). Moreover, the importance of bottom friction is also revealed by numerical simulations with high-resolution ocean models that need a significant bottom friction parameter value to achieve a satisfactory kinetic energy level at equilibrium.

Bottom friction does not only represent an energy sink. It has also some subtle effects on the linear and nonlinear interactions that affect the evolution and the 3D structure of the flow. Several theoretical and numerical studies have focused on the friction effect in the oceanic context but also in the atmospheric context in which the friction drag at the earth's surface plays the same role as at the ocean bottom. Figure 1 presents the regions of parameter space explored in those studies. This nondimensional parameter space is defined by choosing a length scale and velocity scale equal re-

Corresponding author address: Dr. Pascal Rivière, LEMAR, IUEM, Technopôle Brest-Iroise, 29280 Plouzané, France. E-mail: Pascal.Riviere@univ-brest.fr

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FIG. 1. Position of preceding studies in parameter space. See text for details. Three kinds of studies are indicated: "process" studies (Holopainen 1961; Weng and Barcilon 1991; Pedlosky 1983; Rivière and Klein 1997), "atmosphere" studies at midlatitude (Hua and Haidvogel 1986; James 1987; Cai and Mak 1990; Vautard and Legras 1988; Panetta 1993), and "ocean" studies at midlatitude (Treguier and Panetta 1994).

spectively to the internal deformation radius L_D and the vertical shear ΔU . The associated advective time scale is then $L_D/\Delta U$. This choice fixes all the nondimensional parameters except two: $\hat{\beta} = \beta (L_D^2 / \Delta U)$, where β is the latitudinal variation of the Coriolis frequency, and $\hat{R} =$ $R(L_D/\Delta U)$, where R is a bottom friction decay time [see Rivière (1995) for more details]. In this two-dimensional space $(\hat{R}, \hat{\beta}), \hat{\beta}$ can be thought of as a forcing parameter and \hat{R} as a dissipative parameter, as indicated in Fig. 1. In particular, linear calculations with the Phillips model (not shown) reveal a significant increase of instability when $\hat{\beta}$ increases. The instability is weakly sensitive to variations in \hat{R} . In Fig. 1 we can distinguish three kinds of studies: process studies, realistic atmosphere studies, and realistic ocean studies. These studies can be classified in terms of their degree of instability linked to the forcing parameter value $\hat{\beta}$: process studies are characterized by low forcing; realistic atmosphere studies and realistic ocean studies are characterized by stronger forcing (more unstable systems). In this parameter space we see that ocean dynamics is situated in a region more unstable and less dissipative than the atmosphere.

In the linear baroclinic instability context, Holopainen (1961), Weng and Barcilon (1991), and Pedlosky (1983) have shown that the vertical structure of the friction (including bottom and/or top friction) can induce an additional frictional instability. Nevertheless, the effects are very small and concern very weakly unstable baroclinic waves. The most unstable waves for baroclinic

instability are not significantly affected by friction. By opposition the nonlinear studies (Pedlosky 1983; Rivière and Klein 1997; Arbic and Flierl 2002, manuscript submitted to J. Phys. Oceanogr.) reveal a strong effect of bottom friction variations on the nonlinear dynamics of midocean eddies. In particular, for low to moderate unstable systems (Pedlosky 1983; Chou and Loesh 1986; Rivière and Klein 1997) bottom friction modifies the wave-zonal flow interactions and induces a decrease of the perturbations at equilibrium. James and Gray (1986), using a primitive equation model in a realistic context, have shown that a variation of bottom friction can affect crucially the storm-track dynamics in the midlatitude atmosphere. A reduction in bottom friction modifies the vertical and horizontal structure of the mean flow at the equilibrium. This particular structure, characterized by a strong barotropic part and horizontal zonal velocity gradients, inhibits the baroclinic instability of the system at equilibrium and then tends to favor the zonal winds against synoptic perturbations. James and Gray named this mechanism the "barotropic governor." This result has been rationalized in James (1987) and Dong and James (1997) and is referenced in numerous atmospheric studies but has not been confirmed in the oceanic context.

In this study, the effects of bottom friction on the nonlinear equilibration of an unstable vertically sheared mean flow are reexamined in an oceanic context using a primitive equation (PE) model. The region of the parameter space we explore is in the shaded area referenced as "ocean" in Fig. 1.

The model setup and bottom friction parameterization are presented in section 2. In section 3 we present the results concerning the effects of friction on the horizontal and vertical scale selection, and the role of linear processes is investigated. In section 4, the role of nonlinear dynamics is described in terms of surface dynamics and zonal momentum balance. Then energetics at equilibrium are analyzed in section 5 and a discussion on the role of the barotropic structure of the flow is developed. In this section we perform a comparison with quasigeostrophic simulations to replace our results within the context of preceding (atmospheric) studies.

2. Model setup and bottom friction parameterization

a. Initial conditions and forcing

As mentioned in the introduction, we consider a baroclinically unstable midlatitude ocean current. The two important parameters are the degree of baroclinic instability of the jet (linked to the forcing) and the bottom friction (linked to the dissipation). A particular choice in the region of the parameter space corresponding to oceanic zonal currents (Fig. 1) will not significantly affect the results. We have taken parameters relevant for the Antarctic Circumpolar Current (ACC) in the In-

dian sector of the austral ocean (in terms of stratification and mean current). This current has the particularity to present a strong barotropic transport (Nowlin and Klinck 1986; Inoue 1985) as compared with other oceanic midlatitude currents. This choice was motivated by previous atmospheric studies concerning the barotropic governor of James and Gray. Our results will show that it does not induce any loss of generality concerning the effects of the bottom friction. Concerning the bottom friction parameter, there is a strong uncertainty on its value in the ocean, as mentioned in the introduction, and particularly in the ACC. We thus investigate the effects of varying bottom friction. The different values are presented in the next section. The nonlinear equilibrium of the oceanic jet results from interactions between mesoscale eddies arising from baroclinic instability, largescale meanders, and the mean zonal flow. A forcing mechanism is needed to achieve such an equilibrium. In the real ocean, wind forcing, differential heating, and bottom topography play an important part. In order to isolate the effect of bottom friction, we restrict ourselves to the flat-bottom case. In this context, the simplest forcing that allows the maintenance of a realistic baroclinic jet is relaxation to a mean density profile. We use a relaxation time of 150 days. This time scale is large when compared with time scales associated with baroclinic instability, as will be confirmed by linear stability analysis. Preliminary tests with other values have not revealed qualitative differences with respect to the effect of bottom friction.

The geometry of the problem is a zonal channel, zonally periodic with walls at the north and south, on a β plane with $f_0 = -1 \times 10^{-4} \text{ s}^{-1}$ and $\beta = 1.6 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$. Zonal and meridional extension are $L_x = 1000 \text{ km}$ and $L_y = 2000 \text{ km}$, respectively. Depth is H = 4000 m.

The initial density field is also the forcing field. It is calculated using simplified analytic vertical and meridional profiles of potential density and zonal velocity without zonal variations. Figure 2a shows the Brunt-Väisälä frequency vertical profile at the center of the frontal zone. The main pycnocline is at 1000-m depth, and a seasonal pycnocline is also present at 300 m. The associated vertical modes are presented in Fig. 2b. They will be used in the next sections to discuss the vertical structure of the nonlinear equilibrium. The first two Rossby deformation radii are 21 and 13 km. Figure 2c shows a section of the initial velocity and temperature (deduced from the density by a linear equation of state). Velocity decreases from 0.45 m s⁻¹ at the surface to zero at the bottom. The meridional extension of the frontal zone is about 300 km.

To initiate the instability, a small perturbation in wavenumbers 1-10 (relative to the channel length) is added to this temperature profile. The initial currents are geostrophic. Experiments are performed for a duration of 5 yr with different values of the bottom friction.

b. Models

Two kinds of models are used in this study. The main results have been obtained with a primitive equation model, but we have used a quasigeostrophic (QG) model for linear instability analysis and comparison with preceding studies on the barotropic governor.

The PE model is the one developed at LODYC (Madec et al. 1991, 1999). Salinity is held constant for simplicity, and a linear equation of state is used. Horizontal resolution is 9 km \times 9 km, and vertical resolution varies with depth: 26 levels spaced from 10 m at the surface to 200 m at the bottom. Horizontal mixing of momentum and density is biharmonic, with coefficient 4 \times 10⁹ m⁴ s⁻¹. The model is eddy resolving so that the horizontal mixing coefficient is fixed to a value as small as possible. For the vertical mixing, we have used the classical second-order closure of the model (Blanke and Delecluse 1993) with an enhanced vertical mixing coefficient in the case of static instability. Vertical mixing is negligible in our experiments, and so the details of the parameterization do not matter.

The QG model is the one of Treguier and McWilliams (1990). The setup is as close as possible to the PE model except for the vertical grid: the horizontal grid is the same as well as the physical parameters (biharmonic, Coriolis, . . .). We use three layers in the vertical direction. This kind of vertical discretization has been largely used in process studies with QG models. It does not capture the high vertical modes present in the PE model, but, as we will see in the last section, it captures well the effects of bottom friction on mesoscale dynamics. The depth of each interface is estimated by the second baroclinic mode zero-crossing points (Fig. 2b) following Flierl (1978). Reduced gravity at each interface is deduced from Brunt-Väisälä frequency values (Fig. 2a), and a vertically mean velocity is calculated in each layer. The first two Rossby deformation radii are thus similar in the two models.

c. Bottom friction parameterization

The interior of the ocean can be considered as a quasiinviscid fluid (friction terms are negligible when compared with Coriolis or advection terms). Nevertheless, close to the bottom, a viscous layer is needed to satisfy boundary conditions. The dynamics of a viscous layer in rotating fluids is well described by the Ekman theory: in a thin layer, momentum diffusion has to be strong enough to shut down the geostrophic equilibrium. However, in numerical ocean models the parameterization of the Ekman layer depends on the vertical resolution.

In QG layered models, bottom friction is introduced as a vertical velocity at the base of the deepest layer (e.g., Eckman pumping at the top of the bottom Ekman layer). The friction depends on the velocity in that layer and is, in fact, equivalent to a body force. The deepest layer thickness is generally large (over 2000 m in our



FIG. 2. Initial conditions for the PE model (a) N^2 vertical profile at the center of the frontal zone (s⁻²). (b) Normalized vertical modes: barotropic mode and two first baroclinic modes are indicated by heavy lines, the following three baroclinic modes by thin lines. (c) Temperature (°C) (gray) and zonal velocity (m s⁻¹) (isocontours) in meridional and vertical section (vertical coordinates are in meters; meridional coordinates are in kilometers).

model), and so this choice induces a strong viscous constraint over a large fraction of the ocean depth. The Rayleigh damping in the bottom layer vorticity equation can be related to a dissipative time scale τ_b .

In *z*-coordinates PE models, the vertical resolution needs to be higher than in quasigeostrophic models (the deepest layer thickness is generally 100–200 m). The bottom friction effect is parameterized as a momentum flux at the bottom, which serves as a boundary condition for the vertical mixing of momentum:

$$\mathbf{D}^{U} = \left[\frac{\partial}{\partial z} \left(\nu_{v} \frac{\partial u}{\partial z} \right), \frac{\partial}{\partial z} \left(\nu_{v} \frac{\partial v}{\partial z} \right) \right]$$

For linear friction we impose at the last level $\nu_v(\partial u/\partial z) = ru_b$ and $\nu_v(\partial v/\partial z) = rv_b$, where (u_b, v_b) is the bottom velocity and *r* is the friction parameter (m s⁻¹). When the interior mixing of momentum ν_v is small (we take a background value of $\nu_v = 10^{-5} \text{ m}^2 \text{ s}^{-1}$), the theoretical thickness of the bottom Ekman layer is very small in comparison with the deepest model layer thickness. In that case, the vertical diffusion operator is again equivalent to a body force acting in the deepest model layer:

$$\mathbf{D}^{U} = \left(-\frac{r}{\Delta z}u_{b}, -\frac{r}{\Delta z}v_{b}\right).$$

The bottom friction can also be nonlinear. In that case the bottom momentum flux may be expressed as

$$\nu_{v}\frac{\partial u}{\partial z} = C_{d}\sqrt{(u_{b}^{2} + v_{b}^{2} + e_{b})}u_{b} \quad \text{and}$$
$$\nu_{v}\frac{\partial v}{\partial z} = C_{d}\sqrt{(u_{b}^{2} + v_{b}^{2} + e_{b})}v_{b},$$

where C_d is a drag coefficient. Modelers often add a background bottom turbulent kinetic energy e_b to represent the effect of unresolved high-frequency currents like tides (Willebrand et al. 2001).

Because of these different representations, the calibration of bottom friction parameters for a comparison of QG and PE models is not straightforward. Here we choose the friction parameter in each model by setting a typical decay time τ over the whole ocean depth H. Then τ is related to the QG decay time τ_b by $\tau = \tau_b (H_b/T)$ H), H_b being the lower layer thickness, and to the friction parameter r in PE by $\tau = H/r$. An analysis of the momentum balance demonstrates the consistency of this choice. Typical τ values used in OGCMs are close to 150 days. Our simulations correspond to $\tau = 100$ days (high bottom friction) and $\tau = 800$ days (low bottom friction). A case without bottom friction (r = 0) is also analyzed. Intermediate τ values (200 and 400 days) have also been tested and confirm our findings. We have also compared linear and nonlinear friction in the PE model. No significant differences have been noticed, and so the results with nonlinear friction are not presented in this paper.

3. Horizontal and vertical scale selection

a. Spatial scales at equilibrium

The equilibrium achieved by the PE model with high and low bottom friction is characterized by strong instabilities typical of realistic ocean currents in midlatitudes. Figure 3 shows, for high and low bottom friction, two snapshots of temperature close to the thermocline at 1000 m. With high bottom friction (Fig. 3a) we observe an important mesoscale activity with meanders and eddy detachments: large-scale dynamics are dominated by 500-km-wavelength meanders. Smaller-scale meanders (250-km wavelength) are also observed and correspond to the scale of detached eddies. Eddy radii are close to 100 km north of the frontal zone and a little smaller to the south. We observe in particular one cold core ring in the north (Fig. 3a) that has been detached 20 days before and that reinteracts with frontal dynamics 20 days after. These meander and eddy characteristics illustrate the main dynamical features of an unstable zonal current: a statistical equilibrium in the energy transfers among mesoscale, large-scale, and mean zonal flow dynamics. A comparison with Fig. 3b for the lowbottom-friction case reveals important and striking differences: with low bottom friction the flow is now dominated by a very energetic large-scale meander of 1000km wavelength (the longest permitted by the channel geometry). Maximum velocities in the frontal zone are almost doubled (1 m s⁻¹) when compared with the highfriction case (0.6 m s⁻¹). Eddy detachments are observed, but their signature appears to be much weaker than the large-scale dynamics.

Figure 4 shows zonal wavenumber spectra for barotropic and baroclinic kinetic energy at equilibrium. Bottom friction has a significant effect on the threedimensional structure of the kinetic energy. When bottom friction decreases, the barotropic kinetic energy strongly increases. This effect is particularly important for wavenumbers $1 \le k_x \le 2$. The effect on baroclinic kinetic energy is the opposite: the baroclinic kinetic energy decreases for all wavenumbers except at $k_x =$ 1. Thus bottom friction strongly affects the vertical structure of the flow. Furthermore, for both barotropic and baroclinic energy, low friction favors the lowest horizontal scales.

Two different mechanisms may lead to the observed space-scale selection: the instability of the mean zonal flow (either the basic state or the mean zonal flow at equilibrium) or the wave-wave interactions (which drive the inverse energy cascade). Friction may affect one of these mechanisms or both (Rivière and Klein 1997).

b. Role of the linear instability mechanisms

The linear instability of the initial velocity profile (Fig. 5a) is first estimated using the three-layer QG model (see Beckmann 1988 for details). Results (thin lines



FIG. 3. Snapshots of the temperature (°C) at 1000 m for (a) high bottom friction (dissipative time scale 100 days) and (b) low bottom friction (dissipative time scale 800 days) during the fifth year.

in Fig. 5b) reveal one peak at $k_x = 4$ and no significant effect of bottom friction. Then we estimate the linear instability of the same initial velocity profile, but now using the vertical discretization used in the PE model. Results (heavy line on Fig. 5b) now reveal three peaks, respectively at $k_x = 4$, $k_x = 10$, and $k_x = 15$. The peak at $k_x = 4$ displays a similar growth rate for

The peak at $k_x = 4$ displays a similar growth rate for the three-layer QG model and the 26-level PE model (Fig. 5b). The corresponding most unstable wave (250km wavelength) is related to the first deformation radius and is therefore captured by the first baroclinic mode. Its growth rate corresponds to a time scale of 7.6 days.

The other peaks, at $k_x = 10$ and $k_x = 15$ are only present when the high vertical discretization is taken into account. They represent the contribution of the higher baroclinic modes present in the realistic profiles (see Fig. 2b). These vertical modes affect the linear baroclinic and barotropic instability and induce a destabilization of horizontal scales much smaller than the first deformation radius. This result is coherent with Samelson (1999), who showed that, when stratification is characterized by both thermocline and seasonal thermocline, the fastest growing waves may have scales much smaller than the Rossby radius. They are induced by the shear structure of mean flow associated with the seasonal thermocline. These unstable small scales are observed in our nonlinear PE simulations, but only during the first days of the simulation. They have no significant signature at equilibrium: they are thought to be damped by energy transfers into larger scales close to the first deformation radius and then may play a role in the energy balance of the system (Smith and Vallis 2001; Fu and Flierl 1980).

The initial instability is thus characterized by three unstable modes corresponding to $k_x = 4$, 10, and 15 without significant effect of the bottom friction. Let us now consider the stability of the zonal flow (averaged over 2 yr) at the nonlinear equilibrium. The linear stability analysis is presented in Fig. 6 for the three-layer QG model. Time-mean zonal flow is strongly affected by the bottom friction (Fig. 6a): meridional and vertical structures are modified. With high and low friction a thin eastward current is observed at the center of the domain with westward currents on each side. Low friction induces a larger jet width and a strong barotropic component with high meridional gradient of zonal velocity north and south of the jet. The baroclinic component is weakly affected by friction. Nevertheless the corresponding growth rates are identical for high or low friction (Fig. 6b). We can explain this result by consid-



FIG. 4. Zonal wavenumber kinetic energy spectra (temporal average over 3 yr) of the (a) barotropic mode and (b) first baroclinic mode in the PE model for high friction (continuous line) and low friction (dashed line).

ering the vertical shears corresponding to each friction case characterized by the baroclinic signature in velocity: in the center of the jet they are quasi identical so that baroclinic instability is quasi similar in terms of supercriticality. The meridional structure linked to the barotropic countercurrents appears to have no effect on the instability: it is an indication of the baroclinic instability dominance over barotropic instability. The only effect of bottom friction is a weak stabilization when it increases. This result seems to be in contradiction with the barotropic governor mechanism of James and Gray (1986), mentioned in the introduction, and will be discussed in the last section.

In conclusion, the linear study does not reveal any significant effect of bottom friction at nonlinear equilibrium, and so the strong differences (depicted in the preceding part on scale selection) are induced by nonlinear processes, and more precisely by a change in the



FIG. 5. Linear stability results for initial conditions: (a) initial barotropic (heavy line) and first baroclinic (thin line) components of the zonal velocity. (b) Linear growth rates (s^{-1}) as a function of zonal wavenumbers from initial state linear stability analysis: thin lines for the QG model (continuous line for high friction, dashed line for low friction) and heavy line for the PE model.

barotropic inverse cascade as shown by the energy spectra in Fig. 4.

4. Role of the nonlinear dynamics

a. Surface dynamics

One important difference between PE and QG models lies in their surface dynamics. The PE model with its enhanced vertical resolution and its ability to generate strong vertical velocities is expected to produce stronger dynamics close to the surface in the frontal zone. The Fig. 7 diplays snapshots of the temperature in the surface layer: the PE surface dynamics reveals not only a mesoscale activity but also thinner and elongated structures in the temperature field. This is particularly clear when friction is low. Surface temperature spectra are shown in Fig. 8. The spectra are presented at several depths from 10 m to the thermocline. The slope appears to be sensitive to depth in the PE model with lower slope at the surface $(k^{-2} \text{ to } k^{-3})$ as compared with slopes in the interior (k^{-5}) . This k^{-5} slope is the one commonly observed with QG models (Held et al. 1995). This result



FIG. 6. Linear stability results at equilibrium: (a) barotropic (heavy line) and first baroclinic (thin line) components of the zonal velocity at the equilibrium (temporal mean on 3 yr) for low friction (continuous line) and high friction (dashed line). (b) Linear growth rates (s^{-1}) as a function of zonal wavenumbers from the equilibrium state linear stability analysis for the QG model: continuous line for high friction, dashed line for low friction.

confirms that the QG model captures well the mesoscale dynamics close to the thermocline but fails to reproduce strong and small-scale dynamics and its effects on tracers close to the surface.

Two factors can be invoked to explain the differences, observed with the PE model, between the temperature spectra near the surface and at 300 m. The first one is the contribution of the baroclinic modes higher than mode one. From the Fig. 2b, these modes are mostly trapped within the first 150 m below the surface. Analysis of their kinetic energy spectra (not shown) reveals that the sum of the amplitudes of these modes near the surface is not negligible when compared with mode one. Since the energetic scales of these modes are usually smaller than those of the first baroclinic mode (see Hua and Haidvogel 1986), their contribution should lessen the temperature spectrum slope. Another factor is the intensification of the ageostrophic circulation near the surface that is permitted by the primitive equations and not by the quasigeostrophic equations. This intensification (usually associated with frontogenesis and frontolysis processes) accelerates the production of strong gradients and small scales of any passive or active tracer such as temperature. These two factors are absent in the QG model because of the vertical resolution usually used and the equations considered.

Friction has little impact on the surface-layer temperature in the PE model because of the relative importance of small scales. Temperature spectra show that friction affects only the largest scales as shown by energy spectra in the preceding section.

b. Zonal momentum balance

The role of bottom friction has often been investigated in QG models using the momentum balance (Mc Williams and Chow 1981; Panetta 1993). When the flow is forced by relaxation to a mean shear, there is no net momentum input, and the various terms in the balance redistribute momentum horizontally and vertically (Panetta 1993; Treguier and Panetta 1994). Panetta shows that the flow tends to organize itself into multiple jets. This is not the case in our experiment because we impose a meridional scale for the front and our channel is not very wide.

If we represent the time average by an overbar and zonal average by brackets, we can define the perturbation by $u'(x, y, z, t) = u(x, y, z, t) - \overline{u}(x, y, z)$ and $v'(x, y, z, t) = v(x, y, z, t) - \overline{v}(x, y, z)$. The zonal momentum balance at equilibrium can then be written at the vertical level *k*:

$$\frac{\partial \langle \overline{u_k} \rangle}{\partial t} = 0 = -\frac{\partial \langle u'_k v'_k \rangle}{\partial y} - \frac{\partial \langle \overline{u_k} \overline{v_k} \rangle}{\partial y} + f \langle \overline{v_k} \rangle - \delta_k \nu_b \langle \overline{u_k} \rangle,$$

where $\langle \rangle = L_x^{-1} \int_0^{L_x} dx$.

We have verified that horizontal and vertical interior diffusion are negligible in this equation. In the upper layers, there is a balance between the Reynolds stress divergence and the Coriolis term. Here the Reynolds stress divergence is dominated by the turbulent contribution because there is no bottom topography. In the bottom layer, the balance is between bottom friction and the Coriolis term (the familiar Ekman layer balance). How is this balance affected when friction varies?

When the friction coefficient is reduced by a factor of 8, the bottom friction term is reduced by only a factor of 2. Indeed, the reduction of bottom friction induces an increase of the barotropic jet and the bottom velocity by a factor of 4. The resulting bottom meridional velocity is then decreased by a factor of 2 to maintain the balance between Coriolis and friction term.

In the surface layers, the amplitude of the Reynolds stress is almost not affected by the bottom friction (Fig. 9a). However the bottom friction affects the momentum balance by changing the vertical structure of the Reynolds stress as shown in Fig. 9b. In this figure the dominant terms of the momentum balance integrated over the bottom layer are shown as bars at the bottom; the cumulative integral of the Reynolds stress (dotted lines) and Coriolis term (black lines), starting just outside the



FIG. 7. Snapshots of the surface temperature (°C) in the PE model in the two friction cases: (a) high bottom friction (dissipative time scale 100 days) and (b) low bottom friction (dissipative time scale 800 days) during the fifth year.

bottom layer, is shown as a curve above. At the surface, the cumulative integral of the interior terms balances the bottom contribution for both the low-friction case (thin lines) and high-friction case (thick lines). The different momentum sink at the bottom is compensated by a different vertical scale of the Reynolds stresses: With high bottom friction the Reynolds stress term is important over the whole water column above the Ekman



FIG. 8. Zonal wavenumber temperature spectra at different depths in the PE model for (a) high bottom friction and (b) low bottom friction. Depths concerning the PE model are 10 (continuous line), 150 (dashed line), 200 (dashed–dotted line), and 1200 m (dotted line).



FIG. 9. Zonal momentum balance in the PE model for low and high bottom friction: black continuous line for the Coriolis term, black dashed line for the turbulent Reynolds stress term, and gray continuous line for the bottom friction term. Heavy lines are for low bottom friction, and thin lines are for high bottom friction. (a) Values of the Coriolis and turbulent Reynolds stress terms in the top layer for low and high bottom friction. (b) Bars at the bottom represent integrated values over the bottom layer, and lines above represent cumulative integrated values over the water column above the bottom layer (see text for precisions).

layer. With low bottom friction the Reynolds stress values are significant only close to the surface.

Thus we can point out two effects of the bottom friction parameter variation. First, at the bottom the strong effect on the barotropic structure of the flow modifies the amplitude of the friction and Coriolis terms. Then the vertical structure of the Reynolds stress in the interior is modified to compensate this effect, inducing no significant change of the balance observed at the surface. However, we do not observe a strong effect of the Reynolds stress variations on the mean frontal structure because of the meridional scale for the front imposed in our simulations.

Another useful indicator of the effects of eddy-mean flow interactions is shown in Fig. 10 by the Eliassen-Palm flux and its divergence. Eddy momentum and heat fluxes are also indicated in this figure. Eliassen-Palm fluxes are calculated as in Edmon et al. (1980):

$$(F_{y}, F_{z}) = \left(-\overline{u'v'}, \frac{f_{0}\overline{v'\theta'}}{\partial \Theta/\partial z}\right)$$

The divergence of this vector indicates where and how the zonally averaged time-mean flow experiences the local net effect of eddies. The figure shows a maximum of eddy heat fluxes near the bottom as predicted by baroclinic instability. The maximum of eddy momentum fluxes is observed near the surface as shown before concerning the momentum budget. The Eliassen-Palm flux divergence is positive close to the bottom and negative near 500 m. The Eliassen-Palm vectors show an upward propagation of eddy activity and a meridional propagation out of the baroclinic region close to the surface. This general scheme is also observed in atmospheric studies (Edmon et al. 1980; James and Gray 1986). However, we observe two main differences with atmospheric results. First, there is a strong maximum of divergence near the surface where the QG assumption fails because of high vertical modes signature and strong ageostrophic velocities. This is a major characteristic of ocean surface dynamics. Second, the E-P fluxes are very symmetric and confined into the baroclinic region in our simulation. This may be induced by the channel geometry of the model.



-500 -1000 -1500 -2500 -2500 -3500 -500 1000 1500 Y

v'ť

FIG. 10. Meridional cross section at the equilibrium of (a) eddy momentum fluxes (contour interval: $2.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-2}$), (b) eddy heat fluxes (contour interval: $1.6 \ 10^{-4} \text{ }^{\circ}\text{C} \text{ m s}^{-1}$), and (c) Eliassen–Palm flux and its divergence (contour interval: $5.4 \times 10^{-7} \text{ m s}^{-2}$). Dashed lines indicate negative values.

5. Energetics of the nonlinear equilibration

a. Energy and enstrophy equilibration

Figure 11 shows the temporal evolution of the potential energy (deviation from the initial potential energy), kinetic energy, and enstrophy during five years for high bottom friction, low bottom friction, and no bottom friction. Bottom friction has a strong effect on kinetic energy and potential energy at equilibrium (Figs. 11a,b). When friction is multiplied by a factor of 8, kinetic energy is divided by a factor of 3 and potential energy deviation is multiplied by a factor of 1.5 (meaning a decrease in total potential energy). The higher bottom friction is, the sooner kinetic and potential energy saturation occurs (after 4 months for a 100-day dissipation time scale and 6 months for an 800-day dissipation time scale). This result is also observed for intermediate dissipation values (not shown). Concerning the enstrophy evolution (Fig. 11c) no significant effect of bottom friction is observed over the entire simulations: same saturation time scale and values, and same enstrophy level at equilibrium.

One striking result concerns the simulation without bottom friction: in this case we do not observe kinetic energy saturation (Fig. 11b). The model does not equilibrate and kinetic energy grows quasi-linearly during the simulation whereas potential energy saturates and equilibrates at a value close to the low-bottom-friction case (Fig. 11a). This nonviscous solution is mentioned in several studies (as, e.g., in Panetta 1993) and will be discussed in the next section with the energy transfers analysis. Without bottom friction, enstrophy equilibrates at a higher value than in the two dissipative cases.

A more detailed energy analysis has been performed to examine the effects of friction on the zonal flow and perturbation energy. Table 1 summarizes the kinetic energy level achieved at the equilibrium for high and low bottom friction in the two models (averaged over the whole domain and the last 2 yr). We distinguish the total kinetic energy (ke), the kinetic energy of the zonal mean flow (zke), and the perturbation kinetic energy (pke = ke - zke). The vertical structure of those quantities is further analyzed in terms of barotropic (bt) and baroclinic (bc) kinetic energy. We first discuss the results concerning the PE model and then the QG model.

For the PE model, the first line in Table 1 confirms the tendency observed in Fig. 12: the total ke increases when the dissipation time scale goes from 100 to 800 days. The kebc and kebt values indicate that this increase in total kinetic energy results from two mechanisms: on the one hand, a strong increase in the barotropic kinetic energy (kebt) by a factor of 3.6 and, on the other hand, a decrease in the baroclinic energy (kebc) by a factor of 0.5. The net effect is to increase by a factor of 5 the barotropic-to-baroclinic energy ratio. We also see that zke is more affected by friction than pke, but a reduction of friction produces the same effect: both zke and pke increase. The bottom friction has no



FIG. 11. Temporal evolution of (top) potential energy (difference from initial potential energy; J m⁻³), (middle) kinetic energy (m² s⁻²), and (bottom) enstrophy (s⁻²) with a high bottom friction (continuous line), low bottom friction (dashed line), and no bottom friction (dotted line). Values are meant over the whole domain.

significant effect on zonal baroclinic energy (zkebc), whereas pertubation baroclinic energy (pkebc) decreases by a factor of 0.7 when dissipation time scale goes from 100 to 800 days. Thus the decrease in total kebc, mentioned before, results principally from the effect of bottom friction on the perturbation. Inversely, the bottom friction effect on barotropic kinetic energy is more important on the zonal mean flow (zkebt) than the perturbation (pkebt) but induces globally an increase in these two quantities. In conclusion, a bottom friction decrease induces a strong barotropization of the flow (increase in zkebt and pkebt). Moreover, it increases significantly zonal and perturbation kinetic energy but favors the zonal kinetic energy dominance.

Comparison with the three-layer QG model in the last columns of Table 1 shows that these effects are well captured by QG dynamics. For high friction, ke levels in the two models are similar. However, in the case of low friction, the flow barotropization is significantly increased by the QG dynamics, particularly for the zonal flow for which the ratio zkebt/bc is 7.3 instead of 4 in the PE model. The consequence is an increase in total kinetic energy when friction is low in the QG model when compared with the PE model.

b. Energy transfers

In this section we use the barotropic kinetic energy equation to understand how bottom friction acts on the energy transfers.

The barotropic energy equation is obtained classically by projecting the horizontal momentum equations on the barotropic mode, multiplying by barotropic velocities and then integrating over the entire domain:

$$\frac{\partial \mathrm{KE}_{T}}{\partial t} = C_{\mathrm{KE}} + C_{\mathrm{fric}}$$

with

$$\begin{split} C_{\rm KE} &= -\left\langle \frac{1}{2} u_T [(\overline{u_c^2})_x + (\overline{v_c^2})_y] \right\rangle - \left\langle u_T \overline{v_c \partial_y u_c} \right\rangle \\ &- \left\langle v_T \overline{u_c \partial_x v_c} \right\rangle - \left\langle u_T \overline{w \partial_z u_c} \right\rangle - \left\langle v_T \overline{w \partial_z v_c} \right\rangle \\ C_{\rm fric} &= -\left\langle u_T \overline{\Pi v_{\rm bot} u_{\rm bot}} \right\rangle - \left\langle v_T \overline{\Pi v_{\rm bot} v_{\rm bot}} \right\rangle, \end{split}$$

where $\langle \rangle = 1/L_x L_y H \iiint dx dy dz$, and $\overline{} = F_0 \int_{-H}^0 F_0 dz$ with $F_0(z) = 1/\sqrt{H}$ the barotropic normalized mode; C_{KE} represents the baroclinic interactions through the advection terms, and C_{fric} represents the dissipative term induced by the bottom friction (the other dissipative terms in the interior are neglected for simplicity). Here $\Pi = 1$ in the bottom layer and 0 elsewhere, $u_T = \overline{u}$ and $v_T = \overline{v}$ are the zonal and meridional barotropic velocities, and $u_C = u - u_T$ and $v_C = v - v_T$ are the zonal and meridional baroclinic velocities.

Let us consider first the barotropic equation in the absence of bottom friction. In this particular case one question is: Why does the kinetic energy grow linearly in time and not reach equilibrium? One important result in that case (not shown) is that the baroclinic energy is well equilibrated in time and thus the total kinetic energy growth is induced only by the barotropic energy evolution. Figure 12a shows the time evolution of the total nonlinear term (C_{KE}) and the bottom friction term (C_{fric}) of the barotropic energy equation during the fifth year. The time evolution of the barotropic energy is driven by the nonlinear terms that are always positive and induce a quasi-constant energy transfer from baroclinic to barotropic energy. This constant barotropic energy source thus implies a constant barotropic energy increase in time. To maintain an equilibrium, the barotropic flow needs in that case a dissipative sink. Here, in the absence of bottom friction, baroclinic energy is equilibrated but nothing can equilibrate the barotropic energy.

Figures 12b and 12c show the same temporal evo-

TABLE 1. Kinetic energy values ($m^2 s^{-2}$) in the PE and QG models for low and high friction as ke: total kinetic energy, pke: perturbation kinetic energy, zke: zonal flow kinetic energy (see text for details), bt: barotropic part, bc: baroclinic part, and bt/bc: the ratio of the two preceding values. Values are averaged over 3 yr and the whole domain; "mult fact" gives the multiplicative factor from high to low friction: greater than 1 if values increase when friction is decreased.

		PE model			QG model		
		$\tau = 100 \text{ days}$	τ = 800 days	Mult fact	$\tau = 100 \text{ days}$	τ = 800 days	Mult fact
ke	ke	0.014	0.045	3.2	0.015	0.063	4.2
	bt	0.012	0.043	3.6	0.013	0.062	4.8
	bc	0.0022	0.0016	0.7	0.0019	0.0015	0.8
	bt/bc	5.4	26.9	5	6.6	41.7	6.3
pke	pke	0.0087	0.0252	2.9	0.0111	0.0413	3.7
•	bt	0.0078	0.026	3.3	0.010	0.040	4.2
	bc	0.0018	0.0012	0.7	0.0016	0.0011	0.7
	bt/bc	4.4	21.3	4.8	6.1	34.9	5.7
zke	zke	0.0048	0.0183	3.8	0.0034	0.0221	6.5
	bt	0.0044	0.0179	4.1	0.0031	0.0218	7
	bc	0.000 42	0.000 43	1	0.000 31	0.000 30	1
	bt/bc	10.5	41.6	4	9.9	72.0	7.3
zke/pke		0.5	0.7	1.4	0.3	0.5	1.7



FIG. 12. Temporal evolution of the different terms in the barotropic kinetic energy balance in the PE model for (a) no bottom friction, (b) low bottom friction, and (c) high bottom friction: $\partial KE_T/\partial t$ (continuous line), $C_{\rm bebc}^1 + C_{\rm bebc}^2 + C_{\rm bebc}^3$ (dashed line), $C_{\rm fric}$ (dashed–dotted line) (see text for details).

lution but, respectively, for low and high bottom friction. The introduction of a very low bottom friction has almost no effect on the baroclinic energy level, but it acts on the barotropic energy balance: the energy transfers are always directed toward the barotropic reservoir, with a time mean strength comparable to the no-bottomfriction case, but it is now equilibrated by the bottom friction sink. When a high bottom friction is introduced (Fig. 12c), we observed an increase in the energy transfer from baroclinic into barotropic energy.

In the three experiments, wave–wave interactions always represent a positive barotropic energy source. However, the corresponding energy fluxes have a strong amplitude with high friction and a small amplitude with no friction. This is in agreement with the smaller Reynolds stresses found for low friction in section 4b. So a high friction induces larger energy exchanges from baroclinic to barotropic components, leading to a factor of 2 between high and low friction values. This result is in agreement with Hua and Haidvogel (1986), who show that the bottom friction term introduces an additional coupling between barotropic and baroclinic modes and can act as a source term in the baroclinic energy equation. This coupling effect induces stronger energy transfers when friction is increased.

c. The barotropic governor

Several studies have mentioned the barotropization of the flow when friction decreases. One important consequence of this barotropization in the atmospheric context has been highlighted in James and Gray (1986) and James (1987). These papers have shown that, depending on friction strength, the horizontal structure of the barotropic flow is able to moderate significantly the baroclinic instability of the jets induced by vertical shear. This mechanism, which they called barotropic governor, can be preponderant in certain cases corresponding to life cycles of nonlinear baroclinic waves in the atmosphere (Simmons and Hoskins 1978).

In the present study we have reexamined the question of the bottom friction effect with a PE model in an oceanic current configuration. One striking result as compared with James and Gray (1986) concerns the zonal current instability that appears to be very little affected by the barotropic structure of the flow in our simulations, although this latter can be very strong when friction is low (see Fig. 6a). More precisely the energy balance in the preceding section shows that, when friction decreases, both zke and pke increase. This is in direct contradiction with James and Gray (1986), who observe an increase in zke but a decrease in pke, which is explained by the barotropic governor mechanism (see, e.g., their Fig. 7a). This stabilizing mechanism appears to be inactive in our configuration as it is confirmed by our linear stability analysis at equilibrium.

To ascertain and explain these discrepancies we have characterized the oceanic and atmospheric baroclinic jets in term of instability in parameter space and then tested the efficiency of the barotropic governor mechanism in this space with our QG model. First of all we notice that with our calibration of the bottom friction coefficient (as a decay time for the barotropic mode), velocities averaged over the jet width in the bottom layer of the QG and PE models are similar. Our QG and PE results at the thermocline are very similar in terms of horizontal scale selection and energy levels. The only significant difference observed between the two models concerns the surface dynamics: the QG model fails to reproduce the strong and small-scale dynamics close to the surface.

In the nondimensional parameter space described in the introduction (Fig. 1), the present study is characterized by $\hat{\beta} = 0.0235$ and $\hat{R} = 8.12 \times 10^{-3}$ to 10^{-3} . James (1987) studied the barotropic governor mechanism in a simple QG model of the atmosphere. The corresponding values for this study are $\hat{\beta} = 0.2$ and \hat{R} = 0.029 to 1.5×10^{-3} . Our oceanic configuration is thus characterized by a stronger forcing, and a similar to lower friction as compared with the atmospheric configuration of James (1987). As mentioned in the introduction, the ocean dynamics appear to be more unstable than atmosphere dynamics in midlatitudes.

Results of new simulations with our three-layer QG model using parameters \hat{R} and $\hat{\beta}$ close to the James (1987) configuration are presented in Table 2 in terms of zke and pke. These results can be directly compared with Table 1 (with QG model) because the dissipative time scales are the same. The energy levels in Table 2 are lower because of lower instability in the new simulations. We observe an increase in zke when friction decreases; however, pke decreases as friction decreases, contrary to our oceanic case in Table 1. We have also checked that the stability at equilibrium in these new simulations is modified by the bottom friction strength:

TABLE 2. As in Table 1 but for experiments with the QG model in a configuration close to James (1987).

		τ = 100 days	τ = 800 days	Mult fact
pke	pke	0.0008	0.0005	0.6
-	bt	0.0005	0.0004	0.8
	bc	0.0003	0.0001	0.3
zke	zke	0.0009	0.0014	1.5
	bt	0.0004	0.0008	2.0
	bc	0.0005	0.0006	1.2

when friction decreases, the growth rates of unstable waves decrease very close to zero in contrast to the previous oceanic simulations.

It appears that these new simulations with lower instability give coherent results with James (1987) on the barotropic governor mechanism: the barotropic structure of the mean zonal flow constrained by the reduction of bottom friction at equilibrium inhibits the baroclinic instability and favors zonal structures instead of mesoscale structures. Oceanic simulations presented before do not reveal this kind of sensitivity: although the barotropic component of the mean zonal flow is strong and very sensitive to bottom friction, it does not modify the linear instability at the equilibrium. Thus the stronger baroclinic instability process in ocean currents appears to inhibit the barotropic governor mechanism of James and Gray (1986). This is the most significant difference from the atmospheric counterpart.

Panetta (1993) has studied the QG dynamics of zonal jets in a wide parameter space region with, in particular, a varying bottom friction parameter. He mentioned that the barotropic governor mechanism of James and Gray may explain most of his results concerning sensitivity of poleward heat fluxes to friction, except in the region of the parameter space corresponding to the most unstable baroclinic currents. Our kinetic energy diagnostics agree qualitatively with Panetta (1993) in this region.

A more complete comparison of Tables 1 and 2 highlights the difference between the ocean and atmosphere dynamics response to friction variations. The values of Tables 1 and 2 concerning the barotropic and baroclinic pke are summarized in Fig. 13. We observe that the barotropic pke varies inversely in the two cases: in the atmospheric case (Fig. 13b) it decreases when friction decreases as a consequence of the barotropic governor of James and Gray, but it increases in the ocean case. On the other hand, the baroclinic pke decreases in the two cases. There is a strong analogy between the sensitivity of zonal flow (zke) and eddies (pke) to friction for the atmosphere and the sensitivity of barotropic (pkebt) and baroclinic (pkebc) eddies in the ocean: barotropic eddies may have a stabilizing effect on baroclinic eddies in the ocean, instead of the zonal barotropic flow in the atmosphere, but we have not yet rationalized this question.



FIG. 13. Illustration of the variations of barotropic and baroclinic perterbation kinetic energy in the QG model as a function of dissipative time scale in (a) the ocean configuration (Table 1) and (b) the atmospheric configuration from James (1987) (Table 2).

6. Conclusions

Many numerical studies have addressed the role of bottom friction in baroclinic unstable currents in the ocean with QG models and weakly to moderately unstable systems. In this paper we have reexamined effects of bottom friction on the nonlinear equilibration of a fully unstable oceanic current with a frontal structure using two different models: a 25-level PE model and a three-layer QG model. Nonlinear simulations have been analyzed with two extreme bottom friction values: a low bottom friction corresponding to an 800-day decay time and a high bottom friction corresponding to a 100-day decay time. This latter value is close to the OGCM values.

We have shown that bottom friction has a subtle effect on the energy transfers and then on the three-dimensional structure of the flow. Horizontal large scales are favored by a low bottom friction and mesoscales by high bottom friction. Linear stability arguments cannot explain this scale selection: the growth rates of unstable waves at equilibrium are the same in high and low friction simulations, although the mean flows differ by their barotropic structure. Kinetic energy spectra projected on vertical modes show that low bottom friction induces a strong barotropic kinetic energy cascade favoring large-scale dominance. A low bottom friction also increases the large-scale baroclinic kinetic energy but reduces the baroclinic energy at mesoscale and small scales as compared with a high friction.

Comparison between QG and PE simulations show that this scale selection is present in the two models. The QG model captures well the sensitivity of the 3D structure to bottom friction, but it amplifies the barotropization of the flow when friction is low and makes the baroclinic structure less sensitive to the bottom friction value. When friction is large, the barotropic and baroclinic kinetic energy spectra are close in the two models.

Although the QG model captures well this scale selection concerning barotropic and first baroclinic modes, it fails to capture surface dynamics reproduced by the PE model. Surface-layer temperature spectra in the QG model reveals a k^{-6} slope instead of the k^{-3} slope in the PE model close to the surface. This result does not depend on the bottom friction strength. Thus the QG model appears to be robust concerning the mesoscale and large-scale features but fails to reproduce smaller scales induced by surface dynamics close to the frontal zone.

The classical momentum balance at nonlinear equilibrium has been reexamined in the PE model simulations, and its sensitivity to bottom friction has been analyzed. The surface values of Coriolis and turbulent Reynolds stress terms are very weakly sensitive to bottom friction variations: bottom friction increase is compensated by a change of the vertical structure of the turbulent Reynolds stress over the whole water column.

Energy balances show that bottom friction acts effectively on the potential and kinetic energy level at equilibrium without significant change in the enstrophy. Nonlinear equilibrium cannot be achieved without bottom friction: baroclinic energy equilibrates between sources due to heat fluxes arising from the mean flow instability and sinks due to nonlinear energy transfers, but these nonlinear fluxes always convert baroclinic into barotropic energy so that barotropic energy does not equilibrate.

The role of the barotropic structure has been investigated in comparison with atmospheric studies of James and Gray (1986) and Panetta (1993). In the oceanic simulations of the present paper the barotropic structure of the mean zonal flow is very sensitive to bottom friction variations but has no significant effect on the stability of baroclinic waves at equilibrium. This is in contradiction with the barotropic governor hypothesis of James and Gray (1986) for atmospheric dynamics. In the atmosphere, when friction decreases, zonal kinetic energy increases and induces a stabilization of eddies and thus a decrease of perturbation kinetic energy. In ocean simulations (characterized by a stronger forcing term as shown in Fig. 1) zonal kinetic energy increases but also perturbation kinetic energy. We verified that this is not a consequence of the special forcing chosen in our model and conclude that the stronger instability of ocean dynamics at midlatitude, as compared with the atmosphere, implies that baroclinic instability at equilibrium inhibits the barotropic governor mechanism of James and Gray.

This result does not mean that the barotropic structure of the flow has no effect but only that instability of the barotropized mean zonal flow is not significantly changed. The changes induced by bottom friction are more subtle and involve the 3D structure of the eddy field through nonlinear interactions. In particular, we observe a strong analogy between the variations of barotropic and baroclinic energy in the eddy field for ocean simulations and zonal and eddy kinetic energy in the atmosphere. This may indicate that large-scale barotropic eddies (and not barotropic zonal flow) may have a nonlinear stabilizing effect on mesoscale baroclinic eddies in the ocean. This question is left to future investigations.

This study was focused on the bottom friction effects so that we chose a very simple model setup without wind forcing and topography. These two types of forcing play an important role in the real ocean, especially in the ACC. In particular, they organize the vertical structure of the mean flow. Topography can act as a momentum sink and reduce significantly the zonal transport of the ACC (McWilliams et al. 1978) and also generate standing eddies with a strong barotropic component. Wind stress is one of the main factors driving the ACC, and its effect may be an additional damping of the eddies close to the surface modifying the vertical structure of the dissipation (McWilliams and Chow 1981). The interactions between these mechanisms and the bottom friction effects shown in the study have not been investigated and are left to future work.

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