

NOTES AND CORRESPONDENCE

The Work Done by the Wind on the Oceanic General Circulation

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ABSTRACT

A new estimate is made using altimeter data of the rate at which the wind works on the oceanic general circulation. The value of about 1 TW is lower than previously estimated and is dominated by the work done by the mean zonal wind in the Southern Ocean. The meridional component of the mean wind contributes primarily in the eastern upwelling regions of the ocean. Fluctuating component contributions are small. A comparison with the results of a numerical model produces both the same total work as well as the same general geographical patterns but with detailed differences. Both observations and model show that the subtropical gyres are regions where the atmosphere is braking the ocean circulation. The input of wind energy is shown to be qualitatively consistent with estimates of the rates of decay of barotropic and baroclinic mesoscale variability. If most of the energy input into the Southern Ocean is dissipated there, this region could be a dominant factor in mixing the global ocean.

1. Introduction

The rate of working by the wind on the oceanic general circulation is one of the fundamental quantitative indicators of the coupling between the atmosphere and ocean. The wind, along with the tides (see Munk and Wunsch 1998, hereafter MW98), is apparently, and contrary to intuition, the fundamental determinant of the oceanic heat flux.

There seem to have been only a few efforts to estimate the rate at which the wind works on the ocean, including Faller (1968), Fofonoff (1981), and Oort et al. (1994). All three of these estimates make the assumption (Stern 1975, p. 114) that the work can be computed as

$$W = \langle \boldsymbol{\tau} \cdot \mathbf{v}_g \rangle, \quad (1)$$

where $\boldsymbol{\tau}$ is the wind stress, $\mathbf{v}_g = (u_g, v_g)$ is the *geostrophic* flow at the sea surface, and the bracket denotes a time average. The wind generates a multitude of small-scale oceanic motions: ripples, gravity waves, Langmuir cells, turbulent mixing motions, Ekman-like flows, etc. Lueck and Reid (1984) discuss the generation of these motions in the wider context of air–sea coupling. Use of Eq. (1) assumes that all such small-scale motions are dissipated within the surface mixed layer and do not directly produce any motions included in the general

circulation per se. Radiating surface and internal gravity waves, which give up their energy and momentum to the “mean” flow through critical layer and related phenomena, would create an error in this calculation that must eventually be examined. The equatorial band, in particular, must be treated separately. Equation (1) fails if there are extensive regions where the dynamics differ from simple geostrophy plus an Ekman layer. Note also that \mathbf{v}_g is determined globally, not locally, and that the regions of energy dissipation will often be remote from the regions of energy input.

The existing estimates of (1) are quite crude [e.g., Oort et al. (1994) used ship drift observations to estimate \mathbf{v}_g and a wind climatology for $\boldsymbol{\tau}$]. Here we will make a more quantitative estimate by exploiting the global surface elevation coverage available through the TOPEX/POSEIDON altimetry and the wind stress estimates available from the National Centers for Environmental Prediction (NCEP). Issues of reliability do, however, remain.

2. Altimetric estimates

The TOPEX/POSEIDON data are used in 10-day average form, based upon the spacecraft repeat cycle (for a review of TOPEX/POSEIDON data see Wunsch and Stammer 1998). The wind stresses from NCEP are averaged over this same 10-day period—and we thus cannot directly estimate the work done on the ocean by higher frequency wind fluctuations, but these are believed to be very small. The data were gridded at 2°

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TABLE 1. Wind work on the ocean surface from altimetry. All values are four-year averages; the mean surface geostrophic flow is relative to the EGM96 geoid. Averages are based upon an oceanic area of $3.2 \times 10^{14} \text{ m}^2$. S/C is the result from the Semtner–Chervin nominal $1/4^\circ$ global model calculation.

Component	Global integral (W)	Global Av. (W m ⁻²)	S/C global average (W m ⁻²)
$\langle \tau_x \rangle \langle u_g \rangle$	8×10^{11}	2.5×10^{-3}	2.4×10^{-3}
$\langle \tau'_x u'_g \rangle$	2.9×10^{10}	0.09×10^{-3}	0.1×10^{-3}
$\langle \tau_y \rangle \langle v_g \rangle$	4.1×10^{10}	0.13×10^{-3}	0.2×10^{-3}
$\langle \tau'_y v'_g \rangle$	1.0×10^{10}	0.03×10^{-3}	0.1×10^{-3}
Total	8.8×10^{11}	2.8×10^{-3}	2.8×10^{-3}

intervals (see Stammer and Wunsch 1994) and the two components of surface geostrophic velocity were computed conventionally.

Here W is divided into a mean and fluctuating part:

$$W = \langle \boldsymbol{\tau} \rangle \cdot \langle \mathbf{v}_g \rangle + \langle \boldsymbol{\tau}' \cdot \mathbf{v}'_g \rangle. \quad (2)$$

The primed fluctuations are taken relative to a four-year mean of altimetry and wind stresses. The resulting values are displayed in Table 1.

The equatorial region (within $\pm 3^\circ$ of the equator) is omitted from these calculations and the error so introduced needs to be estimated. It is a region of high dissipation (e.g., Gregg 1987) and probably should be treated as a top-to-bottom lateral boundary layer. Such a discussion is beyond our present scope.

a. Mean fields

The mean altimetric flow fields were computed relative to a geoid height field given by the EGM96

solution (Lemoine et al. 1997) and geoid errors will dominate the uncertainty of the resulting mean flow field—which is shown in Fig. 1. This and all other geoid estimates contain errors, particularly conspicuous in the Southern Ocean, but there is no known evidence of any systematic sign error in the geoid slope there. By far the largest contribution to the total work comes from the $\langle \tau_x \rangle \langle u_g \rangle$ term, with meridional wind work contributing only about 4% of the zonal term. Figure 2 displays the global distribution of the two terms $\langle \tau_x u_g \rangle$, $\langle \tau_y v_g \rangle$ and Fig. 3 that from $\langle \tau'_x u'_g \rangle$, $\langle \tau'_y v'_g \rangle$. The result is dominated by the positive input of energy by the zonal wind in the Southern Ocean and in the Kuroshio and Gulf Stream/North Atlantic Current regions. Figure 4 shows the percentage contribution to the global total of $\langle \tau_x \rangle \langle u_g \rangle$ in each 2° latitude band and the cumulative total (to 100%) obtained by integrating from 63°S to 63°N (the limits of TOPEX/POSEIDON data), and the same quantities from all four terms contributing to W (not shown) are visually almost indistinguishable. About 70+ % of the estimated energy input from both the dominant $\langle \tau_x \rangle \langle u_g \rangle$ and the total W occurs south of 40°S . This result indicates the enormous importance of the Southern Ocean in the general circulation energy budget, and probably in oceanic mixing (see MW98).

Figure 2a shows that the Tropics are broadly regions of negative work by the mean wind. The negative values over much of the North Atlantic subtropical gyre are a consequence of the weak mean eastward component of velocity visible in Fig. 1. This feature is contrary to the conventional picture of the subtropical gyre and could be due to error in EGM96. On the other hand, the same

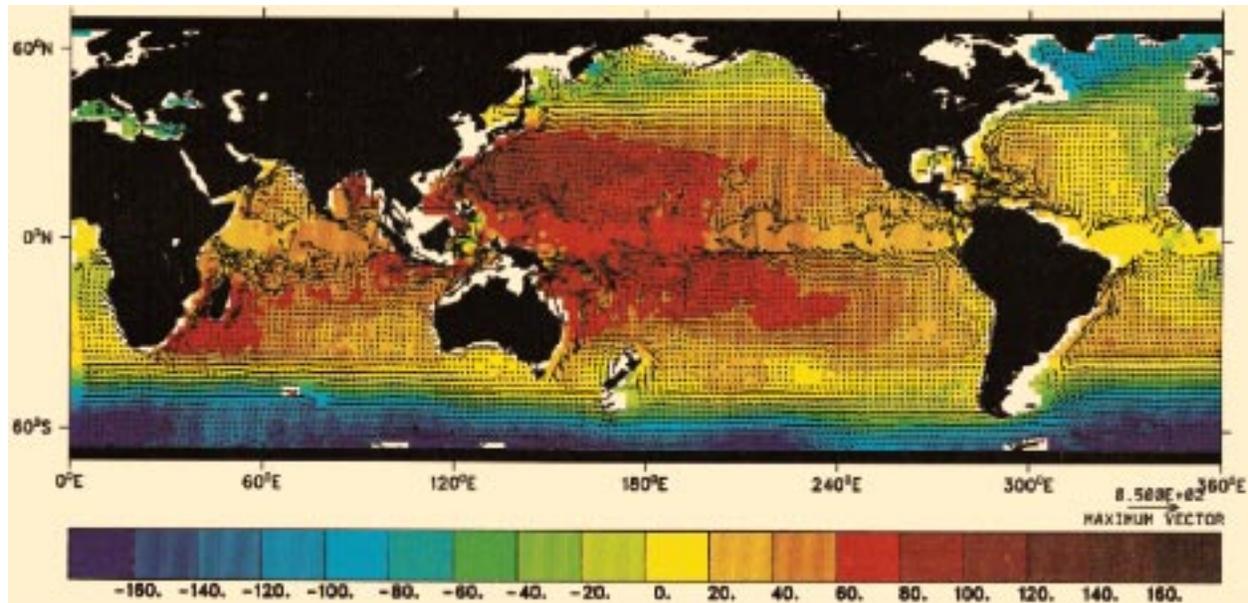


FIG. 1. Four-year average (Dec 1992–Dec 1996) mean surface topography from TOPEX/POSEIDON altimetry relative to the EGM96 geoid height estimate. Any geoid errors will appear here as errors in flow magnitude and direction. Units are centimeters and centimeters per second.

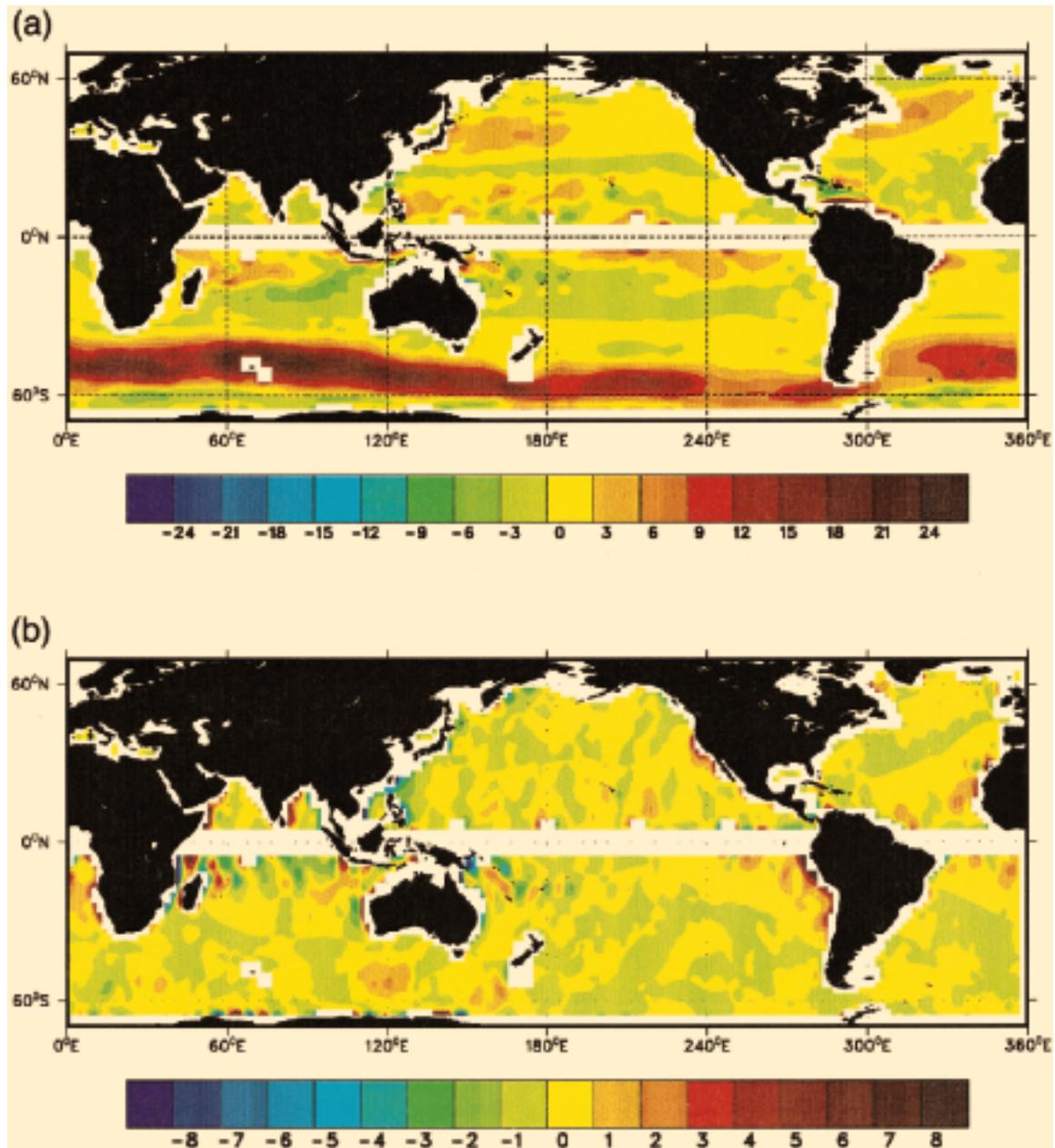


FIG. 2. (a) Four-year mean $\langle \tau_x \rangle \langle u_x \rangle$ and (b) $\langle \tau_y \rangle \langle v_y \rangle$ in units of 10^{-3} W m^{-2} . Some of the small-scale, large values at the edges, e.g., near the Caribbean island arc in (a), are unlikely to be reliable as these are regions of strong geoid gradients. There is a similar issue in (b), e.g. in the region of rapid geoid gradients just west of the Andean mountain chain

feature is present in both the $\frac{1}{4}^\circ$ Semtner and Chervin (1992) general circulation model, as well as the assimilation result of Stammer et al. (1997). The meridional mean contribution to the net work is much less than the zonal one, with the major regions of comparatively strong wind work being in the upwelling regions on the eastern boundaries of the ocean.

The overall picture of wind work is qualitatively similar to that of Oort et al. (1994), but their values are about twice as large in the Southern Ocean and three

times larger in the Gulf Stream and other western boundary currents. Possibly their use of ship drifts, with unaccounted for direct windage effects and inclusion of Ekman velocities, led to an overestimate of the term, although based upon values computed from the FGGE drifters, they conclude that they had actually *underestimated* it. The drifters would, however, suffer many of the same biases as would ships. (Because the ship drift and climatological wind data are even more heavily spatially averaged than the altimeter and NCEP analyses,

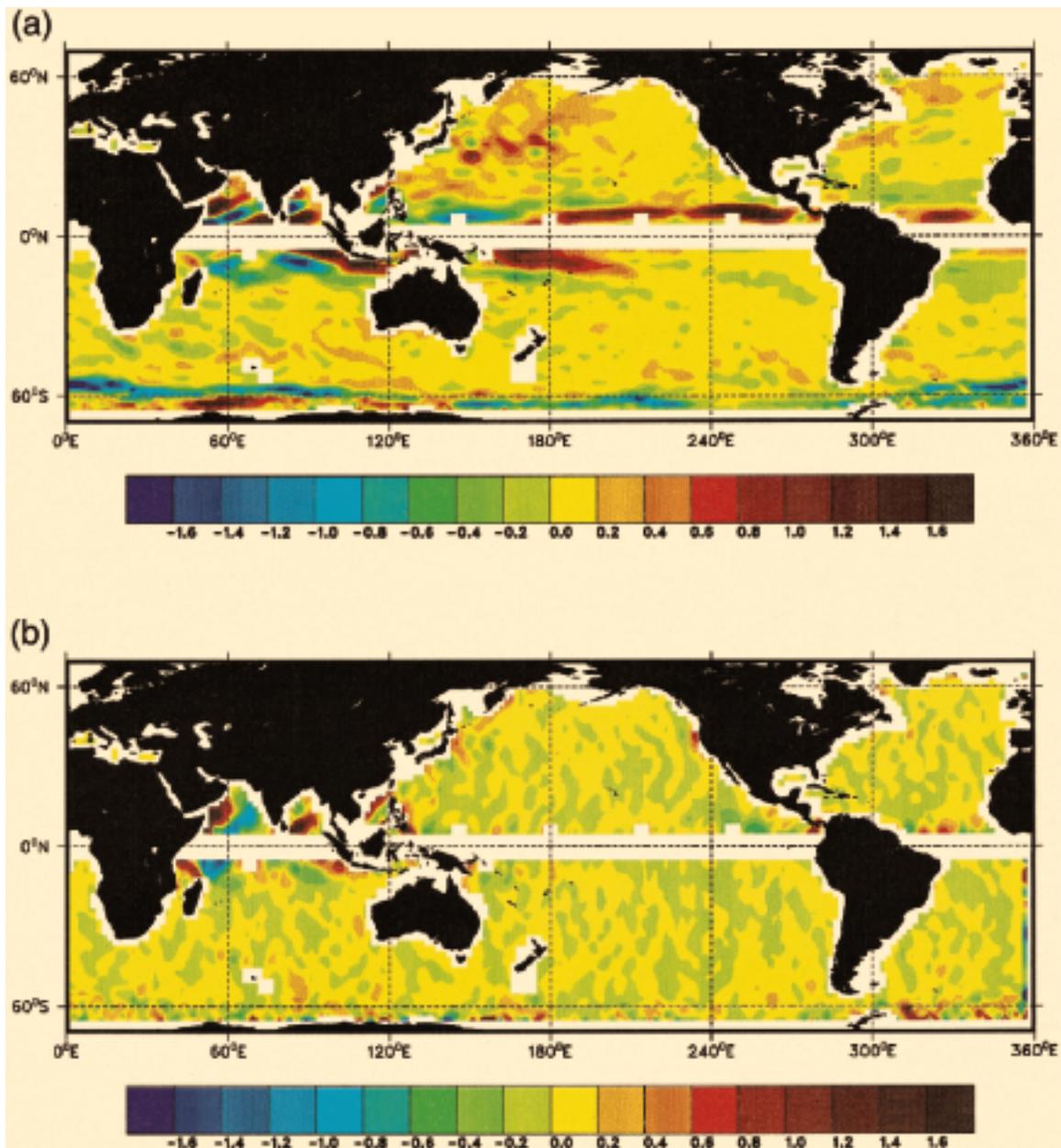


FIG. 3. (a) Four-year mean $\langle \tau'_x u'_g \rangle$ and (b) $\langle \tau'_y v'_g \rangle$ in units of 10^{-3} W m^{-2} .

one cannot attribute their higher values to the use of data with higher spatial resolution.)

The Southern Ocean is clearly important in this calculation, and an estimate needs to be made of the bias error incurred because the coarse resolution of the altimetric estimates produces a reduced velocity in the Antarctic Circumpolar Current. (It is compensated at least in part by the resultant broadening of the current; wind field scales are much larger than the width of the high speed core of the oceanic flow.) Bryden (1979) estimated the wind work in the Drake Passage to be about $3.4 \times 10^{-3} \text{ W m}^{-2}$, a value higher than the global

mean of $\langle \tau'_x u'_g \rangle$ but significantly less than the values near $12 \times 10^{-3} \text{ W m}^{-2}$ estimated in Fig. 2a near Drake Passage.

A proper error estimate for these results would be very desirable, as it is difficult otherwise to assess their reliability. The main problem is that the flow field is computed from a geoid height estimate with strongly spatially correlated errors (Lemoine et al. 1997), and the errors in the NCEP stress field probably also have strongly correlated errors. Even if the geoid slope errors and the wind stress errors are supposed to be uncorrelated (a doubtful assumption because there is reason to

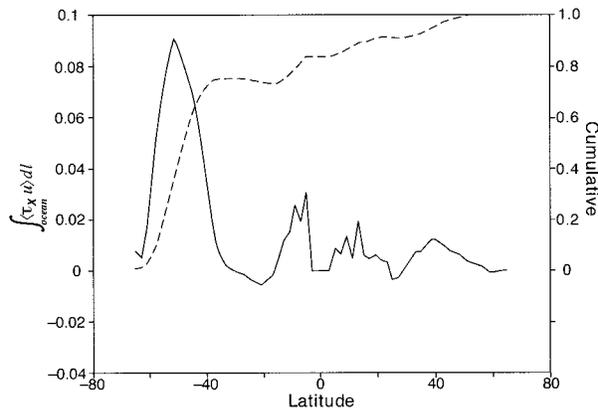


FIG. 4. Percentage contribution to $\langle \tau_x \rangle \langle u_g \rangle$ from each 2° latitude band, integrated zonally around the globe (solid line) and the running sum starting the Southern Ocean. The curves for the total W are visually nearly identical to these. Results from the numerical model are generally similar, but noisier.

believe that both have their largest errors in the Southern Ocean), one must double sum globally products of the form

$$R_{\text{geoid}}(i, j) R_{\text{stress}}(i, j), \quad (3)$$

where $R_{\text{geoid}}(i, j)$ is the geoid slope error covariance in the zonal or meridional direction at lateral grid point i , j and $R_{\text{stress}}(i, j)$ is the equivalent for the wind stress field components. Information is available for computing $R_{\text{geoid}}(i, j)$, although it is a formidable calculation (see, e.g., Ganachaud et al. 1997); no such information is available for determining $R_{\text{stress}}(i, j)$.

b. Fluctuating fields

The zonal fluctuation term dominates the contribution from wind and ocean current variability. The frequency spectrum of oceanic variability in general declines rapidly at periods shorter than the 20-day Nyquist period of TOPEX/POSEIDON and it seems unlikely that omission of the higher frequency wind and current fields represents any significant amount of missing energy. From Fig. 3, the dominant contributions to $\langle \tau' \cdot \mathbf{v}'_g \rangle$ are the positive values in the Tropics (suggesting a possible negative bias through the omission of the near-equatorial flow field) and the negative values in the Southern Ocean. The most interesting feature of the weak input from the variable wind component is the negative contribution from $\langle \tau'_x u'_g \rangle$ in the region of mean easterlies south of the Antarctic Circumpolar Current. Here $\langle \tau'_y v'_g \rangle$ shows a significant contribution of both signs in the western monsoon region of the Indian Ocean.

The small-scale spatial variability of the atmosphere and ocean, which has been removed by spatial averaging, contributes an error that needs to be estimated, but which is probably small owing to the mismatch of atmospheric and oceanic scales.

Rounding the estimated total value upward, we obtain

about 1 TW for the wind work on the oceanic general circulation—a remarkably small value.¹ This value is about one-half that estimated by Fofonoff (1981) and Oort et al. (1994), but we note again its potentially large uncertainty.

3. Energy of oceanic variability

To put these results in context, it is useful to compare this rate of working to estimates of the rate at which the ocean dissipates energy. We focus on the dissipation rates of the variability, as little appears known of the rates for the mean circulation. Much of the latter is probably dominated by the transfer of mean kinetic and potential energy into eddy kinetic and potential energy, which is then dissipated. The discussion is best divided into barotropic and baroclinic motions because there is a difference in the knowledge of how they dissipate. Here “barotropic” and “baroclinic” have the specific meaning of the linear modes of a flat-bottom resting ocean—an approximation (Wunsch 1997) that provides a moderately efficient representation of variability in the real ocean.

As a reference point, the mean-square slope of oceanic variability relative to the 4-yr TOPEX/POSEIDON average is 2.3×10^{-12} . An approximate global rms surface kinetic energy can be obtained from this value by multiplying by $[g/f(30^\circ)]^2$, employing a fixed mid-latitude value of f , which produces $v_{g-rms} = 20 \text{ cm s}^{-1}$ for the surface velocity. The mean surface kinetic energy of the North Atlantic obtained from the current meter moorings used by Wunsch (1997) is an rms velocity of 16 cm s^{-1} ; the North Pacific gives 29 cm s^{-1} —a reasonable bracketing of the altimetric value [if the modes are assumed uncorrelated in time (i.e., not phase locked), the North Atlantic current meter value increases slightly, but the North Pacific value is reduced to 23 cm s^{-1}]. Because of the spatial averaging in the altimetry and the tendency to place moorings in “interesting” high energy regions, current meters are expected to produce higher values on average than does the altimetry.

Using current meter moorings, Wunsch (1997) estimated the vertical average kinetic energy $\overline{\text{KE}}$ in the variability as well as the proportions in the barotropic and baroclinic modes. Accounting for the uncertainty of the averages and assuming a 500-km decorrelation scale, the spatial-average North Atlantic value of $\overline{\text{KE}}$ is $\langle \overline{\text{KE}} \rangle = 61 \pm 30 \text{ g cm}^{-3} (\text{cm s}^{-1})^2 = 6.1 \pm 3.0 \text{ J m}^{-3}$. (The errors stated are simple variances; kinetic energy is probably close to being distributed in χ^2 and the variances should not be interpreted as for a Gaussian variable.) The North Pacific mean was $\langle \overline{\text{KE}} \rangle = 27 \pm 27 \text{ g cm}^{-3} (\text{cm s}^{-1})^2 = (2.7 \pm 2.7) \text{ J m}^{-3}$, which is only

¹ The value is very small compared to the net heat exchange at the seasurface (about 2 PW) and is only one-third the tidal dissipation (about 3 TW); see MW98.

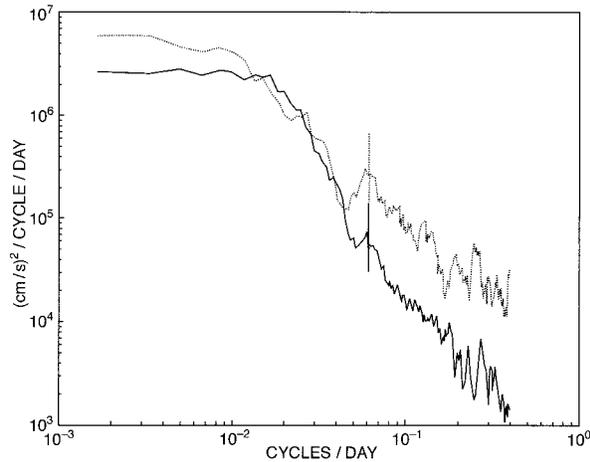


FIG. 5. Frequency spectral density of the barotropic mode (solid) and the first baroclinic mode (dashed) from the MODE center. These spectra are the Thompson-multitaper estimates as described in Percival and Walden (1993). An approximate average 95% confidence interval is shown.

about 40% of the North Atlantic value. Take 3 J m^{-3} as a rough global average. The area of the oceans is about $3.2 \times 10^{14} \text{ m}^2$ (Dietrich et al. 1980) with a mean depth of 4000 m. We then have a global total variability kinetic energy of about

$$\text{KE}_{\text{global}} = 3.8 \times 10^{18} \text{ J} = 3.8 \text{ ExaJ}. \quad (4)$$

a. Barotropic mode energy

The current meter moorings suggest that on average about 50% of the total kinetic energy of the North Atlantic is in the barotropic mode (in the Pacific it is closer to 30%), with about 30% of the North Atlantic total being in the first baroclinic mode (about 40% in the North Pacific). The North Atlantic coverage is much better than the North Pacific so we tentatively adopt a value of 50% as being barotropic, or

$$\text{KE}_{\text{global}/0} = 2 \times 10^{18} \text{ J} = 2 \text{ ExaJ}. \quad (5)$$

The question is how rapidly is this energy dissipated? There are a few indicators. Define

$$Q = 2\pi \frac{E}{\dot{E}T} \quad (6)$$

(Jackson 1975), where E is the energy of the motion, \dot{E} is the mean rate of dissipation, and T is the timescale (strictly speaking, the period of the motion). For the apparent resonant 5-day barotropic mode of the Pacific Ocean Luther (1982) suggested $Q \approx 4$. Wunsch et al. (1998) estimated $Q \approx 5$ for both the fortnightly (Mf) and monthly (Mm) tides. These values indicate that such barotropic motions decay on a timescale somewhat less than their period. We need an estimate of the timescale of barotropic motions.

Figure 5 shows the frequency power density spectrum

of the barotropic mode from a mooring near 28°N , 70°W (the MODE site) in the North Atlantic. The spectrum becomes nearly white for periods longer than about 100 days (this behavior is typical of the spectra from the few long mooring records available). Because of the logarithmic frequency scale, most of the energy lies not far from 100 days. Assuming therefore that all this energy is dissipated in about 100 days = $8.6 \times 10^6 \text{ s}$, the global dissipation rate would be

$$\begin{aligned} 2 \times 10^{18} \text{ J} / 8.6 \times 10^6 \text{ s} &= 2.3 \times 10^{11} \text{ J s}^{-1} \\ &= 2.3 \times 10^{11} \text{ W} = 0.2 \text{ TW}, \end{aligned} \quad (7)$$

equivalent to $6.2 \times 10^{-4} \text{ W m}^{-1}$. [For eddies near 31°N , 70°W , in the energetic, highly barotropic, Gulf Stream recirculation experiment, Bryden (1982) estimated a total eddy energy loss of about $3 \times 10^{-6} \text{ J m}^{-3}/\text{s}$, which would be about 10^{-2} W m^{-1} in 4000 m of water. If this energetic region were typical of the ocean as a whole, the global rate would be about 3 TW; a much smaller global mean value is reasonable.]

The ratio of kinetic to potential energy in a barotropic mode is approximately (e.g., Wunsch et al. 1998)

$$\mathcal{R} = \frac{\text{KE}}{\text{PE}} = (2\pi)^2 \left(\frac{R_0}{\lambda} \right)^2, \quad (8)$$

where R_0 is the barotropic Rossby radius (several thousand kilometers) and λ is the wavelength of the disturbance. The spatial scale of barotropic motions is poorly known. If, for want of a better estimate, we take $\lambda = 2\pi R_0$ (probably somewhat large), the potential energy is equal to the kinetic energy and we should double the dissipation rate above to about 0.4 TW globally—approximately half the estimated wind work.

b. First baroclinic mode energy

For baroclinic mode n , Eq. (8) generalizes simply as

$$\mathcal{R} = \frac{k^2 + l^2}{f^2 \gamma_n^2}, \quad (9)$$

where γ_n is a separation constant. By definition, $\gamma_n^2 = 1/(gh_n)$ where h_n is the equivalent depth. Hence,

$$\mathcal{R} = \frac{(k^2 + l^2)gh_n}{f^2} = (2\pi)^2 \left(\frac{R_n}{\lambda} \right)^2. \quad (10)$$

If the first baroclinic deformation radius $R_1 = \sqrt{gh_1}/f$ is taken as 35 km and the dominant wavelength of the eddies is 350 km., then $\mathcal{R} \approx (2\pi)^2 \times 10^{-4} = 4 \times 10^{-3}$; that is, the energy is primarily potential—about 250 times larger than the kinetic energy.

As stated above, the current meters suggest about 35% of the water column kinetic energy is in the first baroclinic mode. Globally, this would be about $1.3 \times 10^{18} \text{ J}$. Multiplying by 250 produces an estimate of

$$\text{KE}_{\text{global}/1} \approx 3 \times 10^{20} \text{ J} = 300 \text{ ExaJ}. \quad (11)$$

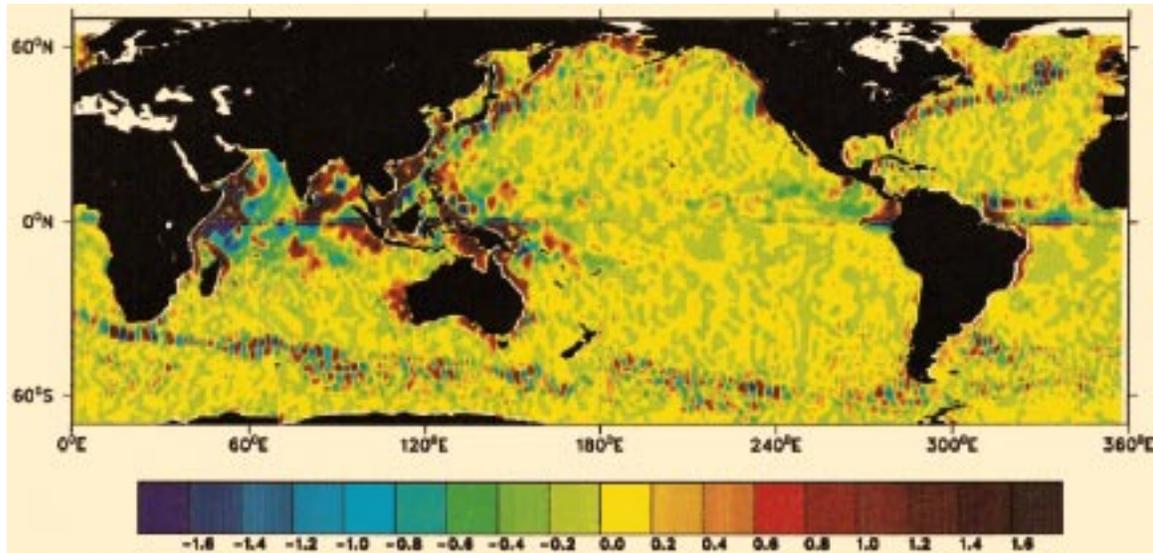


FIG. 6. Four-yr mean $\langle \tau'_x v'_g \rangle$ in units of 10^{-3} W m $^{-2}$ from the Semtner–Chervin model.

How rapidly does this energy dissipate? Little hard evidence exists. Some Gulf Stream rings have been observed to last with remarkably little change for over three years before reentering the Gulf Stream (e.g., Richardson 1983). Such reentry does not by itself represent dissipation—the energy presumably being reacquired by the general circulation. Lueck and Osborn (1986; see also Joyce and Wiebe 1992) found a warm-core ring spindown time of 2–3 yr. Even the very small Mediterranean eddies (meddies), which have higher kinetic to potential energy ratios, have been observed at great distances from their generation regions and appear to have decay times of a few years (e.g., Hebert et al. 1990). Both these types of mesoscale variability have dissipation mechanisms (e.g., water mass intrusions and double diffusive instabilities) not normally operating for more general eddy types and their decay rates may be unusually strong.

We can turn the calculation around and ask what the dissipation rate would be if the estimated baroclinic eddy energy were to be supplied by the wind alone. [If the buoyancy work contribution to the general circulation is assumed to vanish (e.g., Faller 1968), then only the wind is available to provide the energy necessary to sustain the baroclinic eddy field.] Oort et al. (1994) estimate the global average heat flux work as $(4 \pm 2.4) \times 10^{-3}$ W m $^{-2} = (1.2 \pm 0.7) \times 10^{12}$ W, but their error estimate is very optimistic—being equivalent to an air-sea heat flux error of only 10 W m $^{-2}$. There is a small additional contribution from evaporation and precipitation of even greater uncertainty.) In round numbers, using 1 TW of wind energy, one has

$$T \approx 3 \times 10^{20} \text{ J} / (1 \times 10^{12} \text{ W}) = 3 \times 10^8 \text{ s}, \quad (12)$$

that is, about 10 years to renew the baroclinic energy, a duration there is no reason to reject. The equivalent

Q (taking an eddy timescale of 100 days) is about 240. Thus, the estimated wind work is the right order of magnitude to sustain the general circulation, assuming that most of its dissipation is via the generation of eddies, which in turn decay at the rates estimated.

4. Generation of available potential energy and other interpretations

Theories often discuss the energetics of the general circulation in terms of the work done by Ekman pumping against the underlying oceanic stratification, thus generating available potential energy. For example, Gill et al. (1974) calculate the rate of working by the pumping of an Ekman layer against an underlying layer in Sverdrup balance. Simple manipulation of their expression (3.10) and the actual Sverdrup solution readily shows that it reduces to Eq. (1) above. Their estimate is 10^{-3} W m $^{-2}$ for the work done in the subtropical gyre—comparable with the spatial average over the subtropical gyre estimated here. In general terms, the total mechanical energy input is expressed by Eq. (1); this energy can then be removed in a variety of ways, including the direct generation of oceanic kinetic energy and through the generation of available potential energy, which then becomes an intermediate reservoir available for kinetic energy generation.

Yet another interpretation of the wind work—given by Fofonoff (1981, p. 132)—is that it is done by forcing the net Ekman flux upgradient across the lines of constant pressure. Other formulations are clearly possible too.

5. Comparison to a model result

Without formal error bars, the reliability of the results is unclear. Hence, to determine if they are robust, they

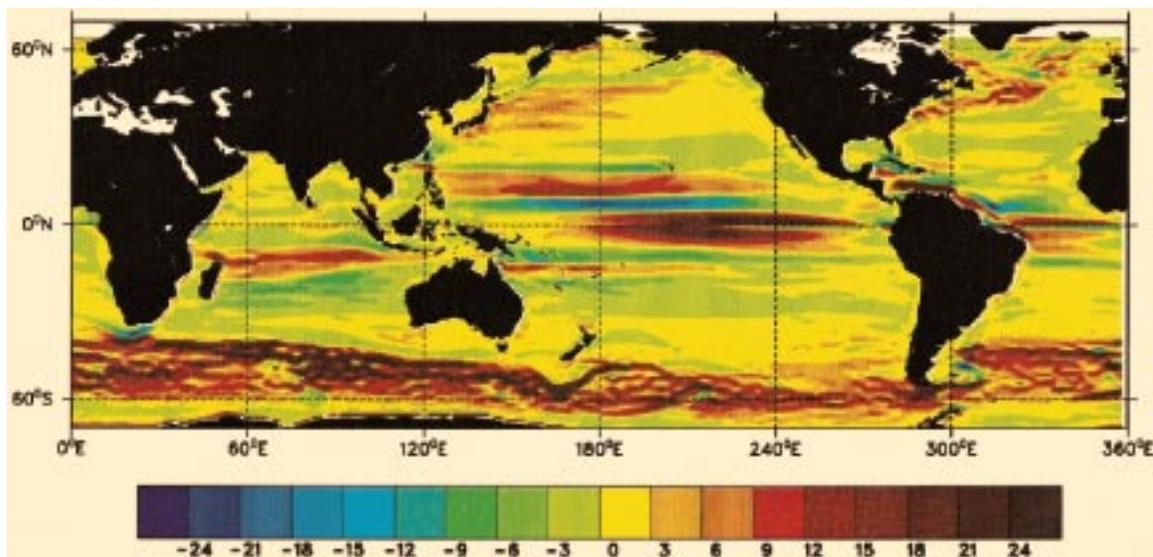


FIG. 7. Four-year mean $\langle \tau_x \rangle \langle u_g \rangle$ in units of 10^{-3} W m^{-2} from the model.

were compared to the equivalent calculation from the nominal $1/4^\circ$ horizontal resolution model of Semtner and Chervin (1992; see also Stammer et al. 1996). The same four years of wind data and model output were employed, and the model output was averaged over 10 days to mimic the TOPEX/POSEIDON data, but the model velocity field was averaged spatially to 1° , consistent with the wind field estimates. Wind stress errors are thus identical in both calculations.

Results of this computation are listed in Table 1; a region within $\pm 3^\circ$ of the equator was omitted in computing the values listed. Fortunately, the global total value is the same as from the altimetric calculation. The largest difference among the four terms is in the higher model contribution from $\langle \tau'_y v'_g \rangle$. Much of the spatial structure of this term (Fig. 6) is a sequence of high positive and negative features associated with the Antarctic Circumpolar Current, Agulhas Retroflexion, and Gulf Stream regions. The result is again dominated by $\langle \tau_x \rangle \langle u_g \rangle$. The spatial pattern from the model (Fig. 7) is the same as from the altimetry, albeit much “streakier,” but the maximum in the Southern Ocean is only about one-half that obtained from the altimeter. Even the negative values in the subtropical gyre are reproduced. The model extremes are generally larger than in the altimetric results. The equatorial band, if included, contributes a significant positive contribution to the work, but it is not clear that this work is available to the general circulation; it may be dissipated locally. Further exploration of model energy fluxes and balances would help elucidate the oceanic circulation energy budget.

6. Concluding remarks

The results here are only a piece of the full discussion of oceanic energetics (see Lueck and Reid 1984; Oort

et al. 1994). It remains to quantify the accuracy of the results—something which should emerge from ongoing efforts to better understand and use the geoid estimates, and also to obtain proper error estimates for numerical weather center analyses of surface stress.

The result that a majority of the work done by the wind on the ocean occurs in the Southern Ocean appears to be robust. What fraction of this energy is dissipated locally, how much is available for cross-isopycnal mixing and how much is exported is not very clear. Munk and Wunsch (1998) have suggested that the integrated oceanic meridional circulation and its corresponding flux of heat are controlled by the power available to mix fluid across deep isopycnal surfaces. They estimated that about 2 TW was required—of which about one-half can be of tidal origin—and primarily in the Northern Hemisphere oceans. Much of the remaining mixing would be wind driven and could be occurring primarily in the Southern Ocean with global consequences for the oceanic heat and other budgets.

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