Oceanic Response to Surface Loading Effects Neglected in Volume-Conserving Models

RUI M. PONTE
Atmospheric and Environmental Research, Inc., Lexington, Massachusetts

(Manuscript received 24 February 2005, in final form 8 July 2005)

ABSTRACT

Forcing by freshwater fluxes implies variable surface loads that are not treated in volume-conserving ocean models. A similar problem exists with the representation of volume changes implied by surface heat fluxes. Under the assumption of an equilibrium response, such surface loads merely lead to spatially uniform sea level fluctuations, which carry no dynamical significance. A barotropic model forced by realistic freshwater fluxes is used to test the validity of the equilibrium assumption on seasonal to daily time scales. The simulated nonequilibrium signals have amplitudes much weaker than those of the forcing, with standard deviations well below 1 mm over most of the deep ocean. Larger values (up to ~1 cm) can be found in shallow and semienclosed coastal areas, where the equilibrium assumption can lead to substantial errors even at monthly and longer time scales. Forcing by mean seasonal river runoff yields similar results, and heat flux effects lead to weaker nonequilibrium signals. In contrast, nonequilibrium signals driven by atmospheric pressure loading are at least an order of magnitude larger than those forced by freshwater fluxes. The exceptions occur for some shallow, coastal regions in the Tropics and at the longest time scales, in general, where forcing by freshwater flux is much stronger than by pressure.

1. Introduction

Surface freshwater fluxes $F$ and heat fluxes $H$ are an important driving mechanism for the ocean circulation. Through complex diabatic, turbulent processes, $F$ and $H$ lead to changes in the near-surface ocean density field. The importance of these thermohaline signals is well recognized, and most general circulation models attempt to represent them through treatment of the relevant thermodynamics. Forcing by $F$ and $H$ also imply changes in oceanic volume. However, as discussed by Huang (1993), Greatbatch (1994), and Mellor and Ezer (1995), ocean models that use a continuity equation of the form $\nabla \cdot \mathbf{U} = 0$ are necessarily volume conserving. Thus, sea level and circulation changes associated with volume changes implied by $F$ and $H$ are not represented in their solutions. Recent model formulations that use the natural freshwater flux boundary condition proposed by Huang (1993), instead of a virtual salt flux condition, do allow for freshwater source terms in the continuity equation and for corresponding changes in volume (e.g., Campin et al. 2004). These models can represent loading effects resulting from $F$ forcing, although they still neglect volume effects associated with $H$.

Some insight into the nature of the oceanic adjustment to these implicit volume changes comes from analytical studies in the case of steady $F$ forcing and the so-called Goldsbrough circulation (Stommel 1984). In this context, vortex stretching implicit in $F$ induces a meridional flow through the Sverdrup balance. However, under realistic forcing, this simple vorticity balance predicts a very weak (order $10^6$ m$^3$ s$^{-1}$) Goldsbrough circulation (Huang and Schmitt 1993), and its neglect in volume-conserving models is not of major concern. Effects of $H$ are expected to be even weaker (Greatbatch 1994).

Motivated by the need to understand steric sea level variability in volume-conserving models, Greatbatch (1994) considered the more general case of oceanic adjustment to time-varying $F$ and $H$. If one assumes neg-
eligible nonlinearities and coupling between barotropic and baroclinic pressure gradients, then signals excluded by volume-conserving models should evolve according to the vertically integrated equations

\[ \begin{align*}
U_t - fV &= -g \xi_x, \\
V_t + fU &= -g \xi_x, \quad \text{and} \\
\zeta + (HU)_x + (HV)_y &= \zeta^e, 
\end{align*} \]

where \( \xi \) is sea level, \( H \) is ocean depth, \( f \) is the Coriolis parameter, \( U \) and \( V \) are vertically integrated zonal and meridional velocities, respectively, and \( \zeta^e \) is the forcing term related to volume changes resulting from expansion or contraction of the water column or from surface mass fluxes. Rewriting (1)–(3) in terms of \( \xi' = \xi - \xi^e \) leads to

\[ \begin{align*}
U_t - fV &= -g(\xi' + \xi^e)_x, \\
V_t + fU &= -g(\xi' + \xi^e)_x, \quad \text{and} \\
\zeta' + (HU)_x + (HV)_y &= 0. 
\end{align*} \]

These equations are equivalent to those forced by atmospheric pressure \( P_a \) (Gill 1982; Greatbatch 1994), with \( \xi' \) here taking the place of \( \xi \) in the latter case. A corresponding equilibrium solution, equivalent to an inverted barometer response to \( P_a \) (Ponte 1993), is simply \( \xi^e = \bar{\zeta} - \zeta^e \) or \( \xi = \bar{\zeta} \), where \( \bar{\zeta} \) is the area integral of the forcing over the global ocean. Based on studies of the response to \( P_a \) (Ponte 1993), Greatbatch (1994) suggested that the part of the time-dependent response neglected in volume-conserving models should approach equilibrium at seasonal and longer time scales. In this case, the ocean shifts mass around rapidly enough so as to maintain negligible horizontal pressure gradients. The only effect is a spatially uniform and, thus, dynamically irrelevant change in sea level equal to \( \bar{\zeta} \), which can be calculated a posteriori and used to correct sea level estimates in the volume-conserving models (Greatbatch 1994). Such corrections are important, for example, when examining the seasonal cycle in bottom pressure (Ponte 1999).

The work of Greatbatch (1994) partly motivated several ocean modeling efforts that included the full effects of volume changes. Whereas the results of Mellor and Ezer (1995) support the equilibrium, linearized arguments of Greatbatch (1994), larger deviations from that theory are obtained by Huang and Jin (2002) and Losch et al. (2004). Such deviations may result, on the one hand, from the complex role of nonlinearities and barotropic–baroclinic coupling in places like the Southern Ocean (Losch et al. 2004). On the other hand, these deviations could still occur for the simple reason that the equilibrium assumption breaks down, as discussed by Huang and Jin (2002).

The equilibrium hypothesis of Greatbatch (1994) is, of course, not expected to hold true always. The response to any surface load is scale dependent and is expected to be nonequilibrium at sufficiently short time scales (Ponte 1993). Other factors can also lead to deviations from equilibrium. Dynamics can play a more essential role, for example, in regions of closed \( f/H \) contours where resonance can occur, over shallow bathymetry where waves are slowed down, resulting in longer adjustment time scales, or in semienclosed seas where Helmholtz-type resonances can occur (Wright et al. 1987). The potential for a nonequilibrium response under realistic \( \zeta^e \) forcing remains to be assessed over the full range of time scales, from seasonal to daily.

In this work, we use a simple shallow-water model, with good representation of fast gravity waves, inclusion of shallow coastal regions, and forcing by realistic 6-hourly flux fields, to test the equilibrium hypothesis of Greatbatch (1994). In this way, we separate nonequilibrium issues from the more complex effects of nonlinearities and thermodynamics. Effects of \( F \) and \( H \) were both considered, but the analysis focuses on the former because of their relatively larger amplitudes. In what follows, we describe the modeling approach (section 2) and the relevant forcing fields (section 3), examine the numerical results focusing on sea level variability (section 4), and discuss relevant findings and conclusions (section 5).

2. Model

Ponte (1993) used a barotropic numerical model to study the global ocean response under \( P_a \) loading. The same approach is used here to examine the ocean response under loading by realistic \( F \) patterns, given the close analogy between the problem at hand and that of \( P_a \) forcing (Gill 1982). The model solves the nonlinear shallow-water equations on a sphere, for an ocean with realistic boundaries and bottom topography. Linear bottom friction and horizontal viscosity terms are included in the momentum equations. No-normal-flow and no-slip conditions are applied at the solid boundaries. More details on the numerical model are provided in Ponte (1993) and Hirose et al. (2001), and references therein. The model has been applied successfully to the study of pressure- and wind-driven high-frequency variability in sea level and bottom pressure (Hirose et al. 2001; Ponte and Hirose 2004) and was found to capture well the deviations from the inverted barometer (i.e., equilibrium) response to pressure observed in altimeter data (Ponte and Gaspar 1999).
The present configuration of the model uses the “fine” topography of Hirose et al. (2001), with minimum depth set to 100 m for stability. Domain and bathymetry shown in Fig. 1 include many shallow and semienclosed seas (e.g., Hudson Bay, North Sea, Mediterranean Sea, and Patagonian and northern Australian shelves). Bottom friction and horizontal viscosity coefficients are set to 2 cm s\(^{-1}\) and 10\(^{8}\) cm\(^2\) s\(^{-1}\), respectively. These settings yielded the best variance reduction in altimeter data when corrected by simulated sea level fields, as well as good correlation with bottom pressure data (Hirose et al. 2001). A version of the model similar to the one used here is currently being run operationally for purposes of dealiasing data from the Gravity Recovery and Climate Experiment (GRACE) mission.

Following the discussion of (4)–(6), we avoid unnecessary modifications of the model code by solving for \(\dot{\zeta} = \dot{\zeta}^F\). With this substitution, the forcing appears as \(\dot{\zeta}^F\) and \(\dot{\zeta}^f\) in the zonal and meridional momentum equations, just as for the case of \(P_a\) loading. The procedure implicitly assumes that nonlinearities in the continuity equation arising from the terms \((\dot{\zeta} u)\) and \((\dot{\zeta} v)\), are negligible, which is a good approximation. At the initial time step, \(\dot{\zeta}^F\) is set to zero and integration starts from a resting state, without need for spinup.

3. Forcing

To focus on the effects of \(F\), the model was for run 1 yr (2001), forced by fields of precipitation minus evaporation \((P - E)\) from the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis (Simmons and Gibson 2000). For comparison, \(P_a\) and heat flux fields were also obtained. The year chosen is the last complete year in the dataset. Fields were downloaded from the ECMWF data server at the available resolution (2.5° × 2.5° grids every 6 h) and were linearly interpolated to the model time step (60 s) and grid (1.125° × 1.125°). Biases in \(P\) fields have recently been discussed by Troccoli and Källberg (2004). In more general terms, fields of \(P - E\) are likely to be noisy, especially at high frequencies. These errors are not a major concern for our purposes of determining the characteristics of the sea level response to \(F\), assuming that the statistics (variance, correlation scales, etc.) of the forcing fields approximate those of the true fields.

To help to understand the ocean response to \(F\), discussed in the following section, it is useful to examine the variability of the forcing \(\int (P - E)\) in comparison with \(P_a\). The standard deviation of \(\dot{\zeta}^F\) shows typical values of 10–20 cm over most of the ocean, but with significantly larger values in the Tropics, particularly along the intertropical convergence zones and in the western Pacific Ocean (Fig. 2a). Peak values can reach more than 100 cm. Taking 1 cm of water as equivalent to 1 hPa of pressure, the variability in ocean loading associated with \(\dot{\zeta}^F\) can be directly compared with that of \(P_a\), which ranges from less than 2 hPa in the Tropics to approximately 15 hPa at high latitudes (Ponte 1993). Forcing by \(P - E\) thus can be much stron-
A substantial part of the variability in Fig. 2a results from a strong trend that is associated with the mean values of $P - E$. The standard deviation of detrended $\xi^p$ series (Fig. 2b) is significantly smaller (<10 cm over extratropical regions and mostly <50 cm in the Tropics). Still, forcing by $\xi^p$ at low latitudes can be one order of magnitude or more larger than that by $P_o$.

Figure 3 shows an estimate of the average power density in $\xi^p$. The spectrum is calculated by detrending $\xi^p$ series at each model grid point and averaging all of the periodograms. For comparison, a similar spectral estimate for $P_o$ is also shown. The spectrum of $\xi^p$ is red with a dependence on frequency close to $\omega^{-2}$ over most of the range, which is in contrast with the whiter (red-
der) $P_a$ spectrum at low (high) frequencies. The variance-preserving plot in Fig. 3 clearly shows that most of the variance in $/H9256_F$ occurs at monthly and longer periods, but at submonthly periods for $/H9256_P_a$.

The amplitude of equilibrium signals under $F$ and $P_a$ forcing is also very different. Under an equilibrium response to $F$, sea level should just equal $/H9256_F$. The time series of $/H9256_F$, shown in Fig. 4, corresponds to a small residual of the local forcing fields and thus exhibits much weaker variability than $/H9256_F$. The standard deviation of the series in Fig. 4 is approximately 2 mm, with most variability contained at monthly and longer scales. Equilibrium signals under $P_a$ loading are at least an order of magnitude higher, besides being spatially variable (Ponte 1993).

4. Simulated ocean response to $F$

The departures from equilibrium defined in terms of dynamic sea level $/H9256_d = /H9256 - /H11005_F$ can be readily calculated from the model output. The standard deviation of $/H9256_d$, shown in Fig. 5, is very small. Typical values over most of the deep oceans are well below 1 mm. Larger variability occurs in several shallow and semienclosed regions (e.g., Hudson Bay, North Sea, and Japan Sea). Maximum nonequilibrium response (~1 cm rms) is found in the shallow and constricted Gulf of Carpentaria, north of Australia. Note that, although small, departures from equilibrium of a few millimeters are comparable to predicted equilibrium signals (Fig. 4). Thus, the simulations indicate that over shallow regions the equilibrium assumption is not applicable.

Enhanced $/H9256_d$ signals in Fig. 5 do not coincide, for the most part, with regions of strongest forcing (Fig. 2), and they are more plausibly explained by the stronger dynamical constraints imposed by shallow depths and semienclosed regions on the mass adjustments necessary to achieve an equilibrium solution. Time series of $/H9256_d$ and respective forcing $/H9256_F$ are shown in Fig. 6 for three areas with enhanced variability in Fig. 5 (Gulf of Carpentaria, Japan Sea, and Hudson Bay). For all series shown, while slow variability dominates the forcing, the largest changes in $/H9256_d$ are associated with relatively rapid signals. This result is consistent with the expected tendency for larger nonequilibrium signals at high frequencies. There is, nevertheless, noticeable variability at relatively long time scales. A few “monthly” events with amplitudes on the order of 1 cm can be seen, for example, in the Gulf of Carpentaria and Hudson Bay, and seasonal fluctuations of a few millimeters are present in all series.

Many of the $/H9256_d$ signals can be traced to local forcing events, particularly to the slope of $/H9256$, which is equivalent to $P - E$ values, also shown in Fig. 6. This is clear for rapid variability in $/H9256_d$ as well as for lower-frequency signals, particularly in the Gulf of Carpentaria and Japan Sea. The noted behavior indicates that the dynamic response is partly locally forced and likely results from an incomplete adjustment to local freshwater fluxes.
Other variability suggests the importance of nonlocal effects, however. This appears to be the case, for example, with the negative $\zeta_d$ signals in Hudson Bay for the first few months, while the forcing is positive and weak.

At a maximum of 1 cm rms (Fig. 5), variability in $\zeta_d$ is typically less than 1% that of $\zeta^F$ (Fig. 2). The relative weakness of the nonequilibrium response holds across all frequencies, as can be seen by comparing the globally averaged $\zeta_d$ spectrum in Fig. 7 with that of $\zeta^F$ in Fig. 3. The red $\zeta_d$ spectrum may seem surprising at first in light of the expected larger nonequilibrium signals at high frequencies. Note, however, that with an approximately $\omega^{-4/3}$ dependence, the spectrum of $\zeta_d$ does not decay with frequency as fast as the $\zeta^F$ spectrum (cf. Fig. 3). Thus, departures from equilibrium at short periods are indeed relatively larger.

The other noticeable influence of ocean dynamics on the $\zeta_d$ spectrum in Fig. 7 is the peaks near diurnal and semidiurnal periods, similar to the findings of Ponte (1993) in the context of $P_a$ forcing. There is only a hint of corresponding peaks in the forcing spectrum (Fig. 3). The peaks in $\zeta_d$ are instead related to the unusual large-scale coherence of the daily cycle in $F$, with a predominantly zonal wavenumber-2 propagating pattern much like the atmospheric pressure tides (not shown), which causes a tidelike, quasi-resonant response in the ocean, although of much weaker amplitude than that forced by the actual pressure tides.

A strict measure of the validity of the equilibrium assumption as a function of time scale involves comparing the spectrum of $\zeta_d$ with that of $\zeta^F$. A particularly relevant issue is whether the correction suggested by Greatbatch (1994) is a good approximation at seasonal time scales. Albeit very small relative to $\zeta^F$, variability in $\zeta_d$ is more comparable to that in $\zeta^F$, because the latter is only a small residual of the large local forcing (Fig. 4). From the spectrum of $\zeta^F$, also shown in Fig. 7, the equilibrium assumption seems mostly valid for monthly and longer time scales, with the possible exception of the annual period. These results are sensitive, however, to how well balanced the $P - E$ fields are over the ocean domain, which determines the variability in $\zeta^F$.

For comparison, we also performed model experiments...
using \( F \) fields from the National Centers for Environmental Prediction–National Center for Atmospheric Research reanalysis. In this case, \( \xi^F \) showed stronger variability (standard deviation \( \sim 4 \) mm), including a substantially larger annual cycle, while the \( \xi^d \) spectrum was very similar to that shown in Fig. 7 for ECMWF forcing. Our tentative conclusion is that at seasonal time scales the equilibrium assumption is a good one, except in shallow and semienclosed coastal areas discussed in Fig. 5.

5. Discussion and summary

To assess the relative importance of nonequilibrium signals associated with \( F \), we compare them with similar signals associated with \( P_a \) loading, which were obtained by running the model with \( P_a \) forcing from ECMWF for the same year. The average spectrum of the dynamic signals driven by \( P_a \), included in Fig. 7, contains much more power than that forced by \( F \). The dominance of \( P_a \)-forced signals holds at all frequencies, except at the seasonal time scale. Results are consistent with the stronger forcing by \( P_a \) relative to \( \xi^F \) at high frequencies (cf. Fig. 3), for which the largest dynamic effects are expected.

On a regional scale, the only places where dynamic signals driven by \( F \) and \( P_a \) effects are comparable are at low latitudes. Such is the case for the regions between Australia and New Guinea, to which we have already alluded, and between Indonesia and Malaysia, based on the estimated ratio of \( \xi^d \) standard deviations (not shown). In these regions, total variance of \( F \) forcing is much stronger than that in \( P_a \) (Fig. 2), which can partly compensate for the redder spectral character of \( F \) forcing (Fig. 3). Such a characteristic of \( F \) forcing also gives rise to \( \xi^d \) signals with a stronger low-frequency content than those forced by \( P_a \) (Figs. 6 and 7).

The \( F \) fields used in this study do not account for variability in runoff, which is a difficult quantity to determine even at seasonal time scales. We have also tested for these effects using a recently created runoff dataset (Röske 2006) that provides a representation of the mean seasonal cycle. Deviations from equilibrium were found to be small relative to those seen in Fig. 5, except as expected in a few semienclosed coastal regions with substantial river input, such as the Persian Gulf. A major future challenge in modeling runoff effects in these regions will be to obtain adequate forcing fields.

Huang and Jin (2002) suggest that the response to seasonal heat flux effects could significantly deviate from equilibrium given the large scale of the relevant surface loading. Using Greatbatch’s (1994) simple pa-

Fig. 6. Time series of \( \xi^d \) (thick solid line; cm) and \( \xi^F \) (thin solid line; m) from three different regions with enhanced nonequilibrium response in Fig. 5. Time series of \( P - E \) in centimeters of accumulated water over 6 h are also shown (dotted line).

Fig. 7. As in Fig. 3, but for nonequilibrium sea level signals \( \xi^d \) forced by freshwater fluxes (solid) and \( P_a \) (dotted).
rameterization of heat flux contributions to the forcing $\zeta^e$, we have also estimated the model’s dynamical response to loading by realistic $H$ effects based on the reanalysis fields. As with $F$ forcing, the largest nonequilibrium signals are confined to coastal areas, but their amplitudes are generally weaker than those shown in Fig. 5, consistent with the relative amplitudes of the two loading terms.

According to these results, the fact that seasonal heat flux effects on sea level can be easily traced in altimeter observations, in contrast with freshwater flux effects, as pointed out by Huang and Jin (2002), does not seem to be related to any stronger tendency for nonequilibrium. An alternative explanation has to do with the different thermodynamic effects of $H$ and $F$. Given the difference in thermal and haline expansion coefficients of seawater, $H$ has a much stronger effect on water density and respective baroclinic adjustment processes on the seasonal time scale. The surface signature of this baroclinic response is what gives rise to the observed sea level patterns and can in fact be reasonably well simulated by volume-conserving models (Greatbatch 1994).

To summarize the findings from the numerical experiments, the dynamic ocean response to surface loads not represented in volume-conserving models is expected to be order 1 cm rms or smaller. The effects of $F$, including those related to river runoff, are more important than those of $H$ but are still typically weak in comparison with those from $P_a$ loading, except in some low-latitude areas. Over most of the oceans, the equilibrium assumption used in the sea level correction proposed by Greatbatch (1994) should work well at seasonal time scales. For the purposes of modeling the large-scale ocean circulation, away from coastal areas, the neglect of nonequilibrium signals related to $\zeta^e$ does not represent a major source of error, as compared with other sources of uncertainty in current models (e.g., Losch et al. 2004). In fact, the more energetic $P_a$ effects continue to be neglected in most ocean modeling efforts. In this regard, the omission of $P_a$ loading effects may result in larger model errors than the neglect of $\zeta^e$ loading effects in volume-conserving models.

If one considers coastal regions, simulated $\zeta^e$ signals seem large enough to affect interpretation of in situ tide gauge observations of sea level, and some of the largest signals in low-latitude regions might be observable by current altimeter and gravity satellite missions (Chelton et al. 2001; Wahr et al. 2004). Our model results point to measurable nonequilibrium effects even at monthly and longer time scales, which is surprisingly in contrast with results from $P_a$ loading. Modeling of such signals should be useful for interpreting the tide gauge and satellite data at a variety of time scales, including seasonal. Under the assumption of linearity and uncoupled baroclinic and barotropic motions, the modeling approach used here provides a simple way of accounting for effects of $\zeta^e$. In a similar way, models that use natural freshwater flux boundary conditions and allow for freshwater source terms in the continuity equation (e.g., Campin et al. 2004) can account for the loading effects from $F$ and are thus preferable to the use of volume-conserving formulations. In any case, proper modeling of coastal regions will certainly require better resolution of the coastal domain and representation of the relevant dynamics as well as good estimates of the forcing, including runoff.

Acknowledgments. The original model mask and topography files were kindly provided by A. Ali. Helpful comments by two anonymous reviewers are also acknowledged. Support for this work was provided by NASA Jason-1 Project under Contract 1206432 with the Jet Propulsion Laboratory and by NASA SENH Grant NNG04GE98G.

REFERENCES


