Annual Cycle in Southern Tropical Indian Ocean Bottom Pressure

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ABSTRACT

The seasonal monsoon drives a dynamic response in the southern tropical Indian Ocean, previously observed in baroclinic Rossby wave signatures in annual sea level and thermocline depth anomalies. In this paper, monthly mass grids based on Release-05 Gravity Recovery and Climate Experiment (GRACE) data are used to study the annual cycle in southern tropical Indian Ocean bottom pressure. To interpret the satellite data, a linear model of the ocean’s response to wind forcing—based on the theory of vertical normal modes and comprising baroclinic and barotropic components—is considered. The model is evaluated using stratification from an ocean atlas and winds from an atmospheric reanalysis. Good correspondence between model and data is found over the southern tropical Indian Ocean: the model explains 81% of the annual variance in the data on average between 10° and 25°S. Model solutions suggest that, while the annual baroclinic Rossby wave has a seafloor signature, the annual cycle in the deep sea generally involves important barotropic dynamics, in contrast to the response in the upper ocean, which is largely baroclinic.

1. Introduction

The tropical Indian Ocean sector is characterized by a strong seasonal monsoon, driven by reversals in the temperature gradient between land and sea, and associated with changes in atmospheric circulation and precipitation (Trenberth et al. 2000). The corresponding dynamic response of the ocean to the monsoon is partly facilitated by a pronounced first baroclinic mode Rossby wave in the southern tropical Indian Ocean (STIO), first noted in model simulations (Woodberry et al. 1989) and later in observations of sea level from altimeters (Périgaud and Delecluse 1992; Fu and Smith 1996) as well as thermocline depth and subsurface temperature from expendable bathythermographs (Masumoto and Meyers 1998; Wijffels and Meyers 2004). More recently, Johnson (2011) observed annual Rossby wave signatures in density and velocity signals from floats in the deep (1600–1900 dbar) ocean; he proposed that the annual response might reach even farther down into the abyss.

How deep does this annual Rossby wave response penetrate? For atmospheric forcing at frequencies and wavenumbers appropriate for the excitation of vertical modes, the response of an ocean with surface-intensified stratification should in theory reach down to the seafloor, such that first baroclinic mode oscillations in sea level and bottom pressure $p_b$ will be in antiphase, with $p_b$ amplitudes generally much smaller than those at the surface (e.g., Gill 1982). While most previous studies have interpreted large-scale $p_b$ observations using barotropic frameworks (Boening et al. 2011; Cheng et al. 2013), Piecuch (2013) recently interpreted $p_b$ anomalies in the northwestern tropical Pacific Ocean from the Gravity Recovery and Climate Experiment (GRACE) mainly in terms of baroclinic Rossby waves, observationally demonstrating that baroclinic adjustment can be manifested at the seafloor. Johnson and Chambers (2013) used GRACE to study the seasonal cycle in $p_b$ over the ocean, revealing significant annual $p_b$ variability in the STIO, but they did not explicitly investigate the nature of the associated physics; following earlier works (Gill and Niiler 1973; Ponte 1999), the breakdown between barotropic and baroclinic contributions remains unclear.

We seek a deeper understanding of the dynamics underlying the annual cycle in $p_b$ in the STIO, inquiring whether it reflects the seafloor signature of the annual baroclinic Rossby wave observed near the surface or deep-sea physics distinct from the upper ocean. We consider

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fields derived from Release-05 GRACE data over the STIO, interpreting them using vertical normal mode theory. Our results suggest that, while the annual baroclinic Rossby wave has a seafloor signature, the annual cycle in \( p_b \) involves important barotropic dynamics. Our findings highlight the improved quality of the latest GRACE data over the ocean and their usefulness for regional ocean circulation studies.

2. Observations

The annual cycle in \( p_b \) in the STIO has been touched on in previous analyses over the global ocean based on models (Ponte 1999; Vinogradov et al. 2008) and observations (Ponte et al. 2007; Johnson and Chambers 2013). We revisit this topic using monthly \( p_b \) grids over 2003–12 based on Release-05 GRACE time-variable gravity coefficients. Data are processed at the University of Texas Center for Space Research (Bettadpur 2012) and post-processed by Don P. Chambers (University of South Florida) using methods discussed in detail by Chambers and Bonin (2012) that include smoothing the data with a 500-km Gaussian filter, which reduces short-wavelength errors in the satellite recoveries, but also, by removing small-scale oceanic signatures, potentially attenuates the true magnitudes of the \( p_b \) signals. Fields are not considered close to land to reduce the likelihood of using data contaminated by the leakage of terrestrial water storage or other geophysical noise (Chambers and Bonin 2012). Global-mean \( p_b \) values are subtracted from the grids and \( p_b \) values are quoted in units of equivalent water thickness. We extract amplitudes and phases of the annual cycle from detrended GRACE \( p_b \) time series using harmonic analysis.

Annual amplitudes in GRACE STIO \( p_b \) are around 14 mm on average, but local amplitude extrema of 19 and 17 mm are apparent east of the Ninetyeast Ridge at 17°S, 92°E and south of the Mascarene Plateau at 25°S, 62°E, respectively (Fig. 1a); the former point is close to the location of maximum annual sea level variance in altimeter data (12°S, 90°E) that has been ascribed to baroclinic Rossby waves (Périgaud and Delecluse 1992; Fu and Smith 1996), but the observed \( p_b \) amplitude is much smaller than the measured sea level amplitude. Phases in \( p_b \) are such that maximum values are generally reached during austral winter, but spatial phase gradients are evident (Fig. 1b), possibly indicating wave propagation; while it is in mid-August in the center of the domain, the peak value in the east of the basin occurs in late June/early July, which is roughly out of phase with the altimetric sea level annual cycle in that region, which is in late November/early December (Périgaud and Delecluse 1992). These differences in amplitude and phase between \( p_b \) and sea level suggest a first baroclinic mode response in the east of the basin, consistent with previous work arguing that the annual cycle in sea level, which is a measure of the mass changes observed by GRACE along with any density fluctuations, is mostly steric in nature in this region (Vinogradov et al. 2008).

What is the physical nature of this \( p_b \) behavior? Signals generated remotely over the equatorial Pacific propagating through the Indonesian Throughflow can be important near the coast of Australia (Birol and Morrow 2001; Potemra 2001), but because the annual cycle in the STIO more generally has been ascribed to baroclinic adjustment driven by winds over the tropical Indian Basin (Masumoto and Meyers 1998; Yang et al. 1998; Johnson 2011), we now consider solutions to a linear model of the ocean’s response to interior wind forcing.

3. Interpretation

a. Framework

The modal bottom pressure response \( p_{b,n} \) to wind forcing can be computed by integrating a linear vorticity equation along characteristics from the eastern boundary \( \lambda_e \) (Piecuch 2013):

\[
p_{b,n}(\lambda, \phi, t) = \frac{a \cos \phi}{c_{r,n}} \alpha_n(-H) \int_{\lambda_e}^{\lambda} w_E(\lambda', \phi, t - l') \, d\lambda'. \tag{1}
\]

This expression assumes a response mediated by large-scale, linear waves in a continuously stratified, flat-bottomed ocean. Here \( n \) is a nonnegative integer denoting mode number, \( \lambda \) is the longitude, \( \phi \) is the latitude, \( t \) is time, and \( a \) is the planetary radius;

\[
w_E = \frac{1}{\rho_0 a \cos \phi} \left[ \frac{\partial}{\partial \lambda} \left( \frac{\tau^\phi}{f} \right) - \frac{\partial}{\partial \phi} \left( \frac{\tau^\lambda}{f} \cos \phi \right) \right] \tag{2}
\]

is the Ekman pumping velocity, where \( \rho_0 \) is a reference density, \( f \) is the Coriolis parameter, and \( \tau^\phi \) and \( \tau^\lambda \) are the zonal and meridional wind stresses, respectively;

\[
c_{r,n} = \frac{c_{\phi,n}^2}{f^2} \left( \frac{1}{a} \frac{df}{d\phi} \right) \tag{3}
\]

is the mode \( n \) long Rossby wave phase speed;

\[
\alpha_n(-H) = \frac{\psi_n(-H)}{\psi_n(0)} \frac{H_n^e}{H_n^l} \tag{4}
\]

is a nondimensional scale factor, such that
is the modal equivalent depth, with \( g \) being gravitational acceleration, and

\[
H'_n = \frac{1}{\psi_n(0)} \int_0^H \psi_n^2(z') \, dz'
\]  

is the modal equivalent forcing depth, where \( H \) is ocean depth; and

\[
t' = \frac{a \cos \phi}{c_{\tau,n}} (\lambda' - \lambda)
\]

is defined so that \( t - t' \) is the time it takes a signal to transit from \( \lambda' \) to \( \lambda \). Terms \( c_n \) and \( \psi_n \) are eigenvalues and eigenfunctions of the Sturm–Liouville system (Kundu and Cohen 2004):

\[
\frac{d}{dz} \left( \frac{1}{N^2} \frac{d\psi_n}{dz} \right) + \frac{1}{c_n^2} \psi_n = 0,
\]

\[
\left. \frac{d\psi_n}{dz} \right|_{z=0} + \frac{N^2}{g} \psi_n \bigg|_{z=0} = 0, \quad \text{and}
\]

\[
\left. \frac{d\psi_n}{dz} \right|_{z=-H} = 0,
\]

where \( N \) is the stability frequency, and \( z \) is the vertical coordinate.

A summand that is omitted from the right-hand side of Eq. (1) is the eastern boundary condition \( p_{n,0}(\lambda, \phi, t) \). This term is not considered on account of the potential contamination of the GRACE data over the near-coastal ocean (which would provide the boundary condition) by leakage of terrestrial water storage, which can overwhelm the oceanic signals, even after attempting to correct for this leakage (Chambers and Bonin 2012). For example, magnitudes of month-to-month terrestrial water storage anomalies on Australia can be >10 cm (Syed et al. 2008; Landerer and Swenson 2012; Fasullo et al. 2013), which is an order of magnitude larger than the oceanic signals of interest (Fig. 1). Based on a set of reduced gravity model experiments, Potemra (2001) reasoned that, while baroclinic signals emanating from the eastern boundary (generated over the equatorial Pacific) are responsible for \( \approx70\% \) of the energy in regions east of 120°E, their relative influence decays rapidly to the west, accounting for \( \lesssim30\% \) of the energy in the tropical Indian Ocean areas west of 100°E. Thus, to the extent that \( p_n \) signals at the eastern boundary are baroclinic in nature, we anticipate that disagreements between model and data affected by the omission of the eastern boundary condition in Eq. (1) will be largely confined to near the Indonesian Throughflow and the coast of Australia.

b. Evaluation

Results discussed in the preceding section motivate us to evaluate Eq. (1) for the first baroclinic \((n = 1)\) mode,
but should other modes [e.g., the barotropic \((n = 0)\) mode] also be considered? Gill and Niiler (1973) provided the first investigation of the seasonal tropical \(p_b\) response. They considered the linearized, depth-averaged momentum and mass conservation equations, hypothesizing that ocean density variations arise from vertical isopycnal heaving due to Ekman pumping; assuming idealized vertical structures for stratification and vertical velocity, these authors then determined that the ratio of the baroclinic to the barotropic \(p_b\) response is given by the quotient (their section 8)

\[
\frac{0.037}{\sin \phi \tan \phi}. \tag{9}
\]

This ratio equals 1.2 and 0.29 at \(\phi = 10^\circ\) and \(\phi = 20^\circ\), respectively, hinting that baroclinic and barotropic effects might both contribute to annual motions in the deep tropical ocean; hence, we solve Eq. (1) for both barotropic and first baroclinic modes.

The evaluation of Eq. (1) requires knowledge of the time-variable surface wind stress. We compute maps of \(w_E\) using monthly means of instantaneous turbulent surface stresses from the Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim) product (Dee et al. 2011) according to Eq. (2) using second-order finite differences.

Also generally required is knowledge of the time-mean ocean stratification. Parameters \(c_{r, \alpha}\) and \(\alpha_{\alpha}\) derive from the \(c_n\) and \(\psi_n\) terms, so we seek solutions to the Sturm–Liouville system [Eq. (8)]. Barotropic solutions are well known (Kundu and Cohen 2004); \(c_0 = \sqrt{gH}\) is the gravity wave phase speed \((\approx 200 \text{ m s}^{-1} \text{ assuming } H = 4000 \text{ m})\), and \(\psi_0\) is depth independent. Hence, \(\alpha_0(-H) = 1\) is a constant, and \(c_{r, \alpha}\) is sufficiently large that \(\ell^\prime\) effectively vanishes for all realizable \((\lambda - \ell)^\prime\) values, reducing Eq. (1) to a barotropic Sverdrup balance in this case.

Baroclinic solutions are evaluated using time-mean, three-dimensional maps of \(N\) from the Ocean Comprehensible Atlas, version 2 (Forget 2010). Computed profiles of \(\psi_1\) over the STIO sector generally have inflection points roughly between 100 and 150 m, which is comparable to the mean depth of the 20°C isotherm at these latitudes (Masumoto and Meyers 1998) and cross zero around 1000–1500 m (not shown). Values of \(c_{r, 1}\) based on these solutions vary by no more than a factor of 2 over the STIO (Fig. 2a). Such variations arise from differences in the stratification over the basin. For simplicity, we use a constant value of \(c_{r, 1}\) (Fig. 2b). While showing some variation with longitude at fixed latitude, reflecting gradients in the deformation radius (Chelton et al. 1998), values of \(c_{\alpha, 1}\) are mostly controlled by the change of \(f\) with latitude (Fig. 2b); hence, we use zonally averaged \(c_{\alpha, 1}\) values to evaluate first baroclinic mode solutions.

Using these values for the model parameters and surface forcing, model solutions are evaluated numerically on a regular 1° grid spanning the STIO from Africa to Australia and Indonesia (as far east as 130°E) between approximately 5° and 30°S. Solutions are not computed equatorward of the equatorial deformation radius, because equatorial physics is not described by the model. For comparison against GRACE data, we spatially smooth the solutions with a 500-km Gaussian filter, extract the annual cycle using harmonic analysis, mask the points near land, and consider the solutions between 10° and 25°S (cf. Fig. 1).
FIG. 3. (a) Amplitude (mm) and (b) phase (month) of $p_b$ annual cycle from barotropic model solution over STIO. (c),(d) As in (a),(b), but for baroclinic solution. (e),(f) As in (a),(b), but for linear superposition of solutions.
c. Comparing model and data

Annual amplitudes and phases in $p_b$ from the barotropic and baroclinic model solutions are quite different in character (Fig. 3). Barotropic amplitudes evidence a gyre-like structure (Fig. 3a), increasing monotonically from east to west, and barotropic phases are nearly uniform across the basin (Fig. 3b), hinting at the approximately coherent nature of annual fluctuations in the wind stress curl over most of the basin at these latitudes (Johnson 2011).

In the northeast and west of the domain, where baroclinic Rossby wave signatures have been noted in altimetric sea level data (Périgaud and Delecluse 1992; Fu and Smith 1996), baroclinic amplitudes are elevated (Fig. 3c). Because of the westward propagation of the Rossby waves and the decrease in phase speed with increasing latitude, baroclinic phases apparently show south-southwest propagation (Fig. 3d), progressing from early June to late July in the northeast of the basin. Because our model omits dissipation and topographic effects, variable baroclinic amplitudes in the northeast of the domain result wholly from the forcing scales and the wave speeds [and not, e.g., from flow–topography interactions around the Ninetyeast Ridge (cf. Fig. 1)], consistent with past work (Birol and Morrow 2001; Wang et al. 2001).

The barotropic and baroclinic solutions are largely in phase or quadrature (Figs. 3b,d), so they generally add constructively. Thus, the linear superposition of the two model solutions has phases mainly in austral winter (Fig. 3f) and amplitudes (11 mm on average; Fig. 3e) that are mostly larger than the individual amplitudes of the barotropic and baroclinic solutions (10 and 4 mm on average, respectively; Figs. 3a,c).

In general, the barotropic model solution explains a majority of the variance in the GRACE data (75% on average; Fig. 4a), partly reflecting gross similarities between data and model phase (cf. Figs. 1b and 3b), where the fractional variance $\nu \in (-\infty, 1]$ in some time series $x$ explained by another time series $y$ has been defined as

$$\nu(x, y) = 1 - \frac{\sigma^2(x - y)}{\sigma^2(x)},$$

where $\sigma^2$ connotes temporal variance. While the baroclinic solution does not explain considerable data variance on average (Fig. 4b), a region of elevated data variance explained (up to 37%) appears toward the
northeast of the domain (Fig. 4b), overlapping with the area of enhanced annual sea level variance (Pérgaud and Delecluse 1992; Fu and Smith 1996). The linear superposition of model solutions generally explains more GRACE variance (81% on average; Fig. 4c) than either of the individual model solutions taken in isolation, most notably where baroclinic amplitudes are enhanced (Fig. 3c), emphasizing that both barotropic and baroclinic processes contribute importantly to the observed variability.

Amplitudes derived from the linear superposition of model solutions are 21% smaller than GRACE data amplitudes on average (Fig. 5a), which would be expected in the case of random data noise, for example, arising from satellite errors (uncertainties in the star cameras, etc.) or nontoceanic processes (leakage of terrestrial water storage, etc.) impacting the $p_b$ grids. But differences between model and data amplitudes can be more pronounced in some areas, rendering lower variance–explained values (Fig. 4c): in the southeast, model amplitudes are much smaller than data amplitudes, potentially implicating signals coming from the eastern boundary absent from the model but important near the coast of Australia and close to the Indonesian Throughflow (Birol and Morrow 2001; Potemra 2001); in the northwest, model amplitudes are considerably larger than data amplitudes, possibly implying missing model physics, for example, interactions with important features of the bottom topography, such as the shoaling Mascarene Plateau northeast of Madagascar (cf. Fig. 1).

The linear superposition of model solutions leads (lags) GRACE in the northeast (southwest) of the basin, but the absolute difference between model and data phase over the basin is only 2 weeks on average (Fig. 5b). This strong phase correspondence elucidates the generally good variance–explained values mentioned earlier: for large differences between model and data amplitudes ($\leq 40\%$; Fig. 5a), given a phase difference of two weeks ($\approx \pi/13$ radians; Fig. 5b), the model can still explain $\geq 80\%$ of the data variance (Fig. 4c).

4. Discussion

We used GRACE to study the annual cycle in $p_b$ over the STIO. To interpret the data (Fig. 1), we considered solutions to a linear model of the ocean’s response to wind forcing (Fig. 3), finding qualitative similarities between model and data (Fig. 4). Quantitative differences between the two (Fig. 5) reflect a mixture of data noise and missing model physics. Model solutions suggest that, while baroclinic processes contribute in some areas, the annual cycle in the deep ocean generally involves important barotropic dynamics, in contrast to the annual cycle in the upper ocean, which is largely baroclinic (Pérgaud and Delecluse 1992).

Our results corroborate general scaling arguments previously advanced by Gill and Niiler (1973). For example, averaging expression (9) between 10° and 25° latitude gives a value of 0.49, meaning that barotropic adjustment should be about twice as important as baroclinic response at these latitudes. This is roughly consistent with average values for baroclinic and barotropic amplitudes given in the preceding section.

Fluctuations in $p_b$ relate to deep circulation changes (Wahr et al. 2002). Model amplitudes (Fig. 3) crudely
resemble the mean dynamic topography over the basin (Schott et al. 2009; Rio et al. 2011), suggesting that the variations represent modifications to dominant regional circulation patterns. The annual cycle in $p_b$ related to barotropic response (Figs. 3a,b) implies maximum anticyclonic (cycloidal) anomalies in tropical gyre circulation during austral winter (summer); the maximum zonal difference in barotropic pressure across the basin is around 10 mm, which (assuming an ocean depth of 4000 m) corresponds to a meridional geostrophic transport of roughly 10 Sverdrups (Sv; 1 Sv = 10$^6$ m$^3$ s$^{-1}$); assuming a basin width of 5000 km, this transport is equivalent to mean meridional barotropic velocities and excursions around 0.5 mm s$^{-1}$ and 2.5 km, respectively. These patterns and values are comparable to those previously found, for example, by Lee and Marotzke (1998), who considered seasonal barotropic streamfunction changes over the Indian Ocean from a general circulation model.

The distinct natures of the deep-sea and upper-ocean responses, along with the recently improved quality of the GRACE data evidenced in this and other recent studies (Chambers and Bonin 2012; Johnson and Chambers 2013; Piecuch 2013; Piecuch et al. 2013), hint that GRACE ocean mass grids could provide valuable dynamical constraints to complement other datasets, mostly concentrated in the upper ocean (sea level from altimeters, subsurface hydrography from profiling floats, etc.), in state estimation efforts (Wunsch and Heimbach 2013). Because of their physical consistency, more accurate state estimates would facilitate better process-based understanding of circulation and climate in the data-sparse deep ocean.

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