Buoyancy-Driven Interannual Sea Level Changes in the Tropical South Atlantic

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ABSTRACT

Linear models of dynamical ocean adjustment to wind field changes, local atmospheric driving, and eastern boundary forcing are often invoked to explain observed patterns of interannual regional sea level variability. While skillful in some regions, these processes alone cannot explain low levels of interannual sea level variability observed in the tropical Atlantic. In this study, through a set of modeling approaches, interannual sea level changes in the tropical South Atlantic are attributed and the dynamical influence of buoyancy forcing is elucidated. Similar to recent findings in the southeast tropical Pacific, sea level patterns in the tropical South Atlantic (as estimated from a data-constrained ocean general circulation model) are found to result from the action of both surface wind and buoyancy forcing; in addition to static local effects, the buoyancy-driven changes comprise important nonlocal ocean dynamical processes. It is shown that the buoyancy-driven sea level changes can be understood within the framework of a linear first baroclinic mode Rossby wave model forced by atmospheric fields and variability along the eastern boundary. To lowest order, the linear model framework also reproduces qualitative patterns of basinwide compensation between wind- and buoyancy-driven sea level changes, which are mostly tied to the anticorrelation of both surface and boundary forcing. Results suggest that the ocean’s dynamical adjustment to buoyancy forcing exerts an important influence on interannual sea level changes across all tropical oceans.

1. Introduction

Ocean surface topography changes relate to aspects of the variable ocean circulation and climate, for example, surface geostrophic currents and subsurface heat storage. As such, historical records of local sea level from tide gauges and contemporary measurements of the global ocean surface topography field from satellite altimetry missions represent powerful observational means for studying oceanic variability. At low latitudes and on interannual time scales, regional sea level changes measured by altimeters are relatively high along the Indian and Pacific whereas observed variability in the Atlantic is strikingly low (Fig. 1a), possibly a reflection of the fundamentally distinct natures of El Niño–Southern Oscillation, which strongly influences ocean climate in the Indian and Pacific, and tropical Atlantic variability (e.g., Chang et al. 2006). Using linear models of eastern boundary generation, local ocean response to atmospheric forcing, or dynamical ocean adjustment to wind driving, past studies have sought to attribute the dynamics underlying these patterns in particular basins [e.g., South Pacific (Qiu and Chen 2006; Li and Clarke 2007); North Pacific (Fu and Qiu 2002; Zhang et al. 2010); Atlantic (Cabanes et al. 2006)]. While these mechanisms reproduce major qualitative features of interannual sea level patterns in the tropical and subtropical Pacific and the subtropical Atlantic, they are less successful in capturing the low levels of observed variability in the tropical Atlantic (cf. Cabanes et al. 2006). In addition, general circulation model studies show that interannual sea level patterns in the tropical Atlantic are driven mainly by atmospheric forcing and not related to intrinsic ocean variability (Cabanes et al. 2006; Penduff et al. 2011), suggesting that aspects of the ocean’s direct response to the atmosphere are being overlooked. Absent in most recent linear-modeling studies is consideration of ocean dynamical adjustment to surface buoyancy exchanges.1 Earlier works considered the problem of Rossby wave generation by buoyancy fluxes (e.g., Magaard 1977), but paucity of observations (e.g., Willebrand et al.

1 But see Thompson and Ladd (2004) for an exception to this rule for the case of the North Pacific.
1980) prevented a meaningful test of such models. Recently, Piecuch and Ponte (2012) used a data-constrained ocean general circulation model to demonstrate the important role of the ocean’s dynamic response to buoyancy forcing in producing interannual sea level changes in the southeast tropical Pacific. In this region, buoyancy-driven sea level anomalies act to compensate wind-forced changes and exhibit a strongly nonlocal character, made manifest in broad westward-propagating features, which were hypothesized by Piecuch and Ponte as representing forced linear first baroclinic mode Rossby waves.

In this paper, we explore a number of outstanding issues relating to South Atlantic sea level variability in particular and the nature of anomalous sea level generated by surface buoyancy exchanges in general. First, using the general circulation model approach of Piecuch and Ponte (2012) described in section 2, we investigate whether consideration of the dynamical effects of surface buoyancy exchanges helps to explain patterns of interannual sea level variability in the tropical South Atlantic (section 3) and also determine what are the relative influences of local atmospheric forcing and ocean dynamics on buoyancy-driven steric sea level changes (section 4). Next, employing forced linear models of baroclinic adjustment, we use this regional case study to test the hypothesis of Piecuch and Ponte (2012) that buoyancy-driven tropical sea level anomalies can be interpreted as forced linear first baroclinic mode Rossby waves (section 5) and inquire into the nature of opposing behavior between wind-forced and buoyancy-driven sea level changes (section 6). A summary of results and general discussion is provided in the final section.

2. General approach

We consider a physically consistent ocean state estimate produced by the Estimating the Circulation and Climate of the Ocean (ECCO) consortium (Wunsch et al. 2007). ECCO estimates represent solutions to the Massachusetts Institute of Technology (MIT) general circulation model (Marshall et al. 1997) constrained to observations using nonlinear least squares (Wunsch and Heimbach 2007). The particular solution used here is defined on a coarse 1° horizontal grid with 23 vertical levels and has been used in several previous investigations of interannual-to-decadal changes in sea level patterns (Wunsch et al. 2007; Piecuch and Ponte 2011, 2012). Interested readers are referred to Wunsch et al. (2007) for a more complete description of the solution and its caveats. This investigation is based on monthly averaged model output over 1993–2004. All time series (and quantities derived from them) considered in this study represent interannual anomalies, which are computed by first removing the time mean, linear trend, and mean seasonal cycle, then passing the resulting time series through a low-pass filter to remove any remaining signals with periods shorter than 1 year. To motivate the analysis that follows, patterns of interannual tropical sea level variability from the ECCO solution are shown in Fig. 1b. These distributions correspond well to the satellite-derived patterns shown in Fig. 1a. Important for our purposes, the solution qualitatively captures the characteristically low variability in the tropical Atlantic, and its detailed consideration can provide valuable insight into the observed variability.

Anomalous sea level $\xi$ comprises forced components, relating to the ocean’s response to atmospheric changes, as well as an intrinsic component, resulting from internal
ocean variability not directly tied to external forcing mechanisms (e.g., Penduff et al. 2011). To distinguish the various generating mechanisms, we run a suite of numerical experiments wherein the surface boundary conditions are modified (cf. Piecuch and Ponte 2012). The atmospheric forcing fields (wind stresses, radiative and turbulent heat exchanges, evaporation and precipitation, etc.) used to produce the full ECCO solution (shown in Fig. 1b) derive from six-hourly products from the National Centers for Environmental Prediction (NCEP)-National Center for Atmospheric Research (NCAR) reanalysis described by Kalnay et al. (1996), iteratively adjusted and optimized as part of the ECCO optimization procedure (Wunsch and Heimbach 2007). The respective forcing fields are cast in terms of surface fluxes, not bulk formulae, and therefore they are effectively uncoupled from one another [e.g., one can “turn off” the direct effects of one forcing variable (e.g., mechanical input by surface wind stresses) while retaining any implicit and indirect effects it may have on another forcing variable (e.g., the influence of wind speed on turbulent heat fluxes)]. Specifically, we run four experiments such that prescribed wind stresses $W$ and surface buoyancy exchanges $B$ are either fully variable $V$ or set to a climatological mean seasonal cycle $C$. Initial conditions are the same for all model runs. The experiments allow us to decompose sea level changes $\zeta$ (depicted in Fig. 1b) into components relating to the direct sea level response to wind forcing $W$, direct response to buoyancy forcing $B$, intrinsic variations including any transient adjustments to initial conditions $I$, as well as any nonlinear effects represented by differences between sea level produced under the combined influence of both winds and buoyancy and the linear superposition of sea level components separately generated by either winds or buoyancy $N$.

\[
\zeta = \zeta^{WWVB},
\]
\[
\zeta_W = \zeta^{WWCB} - \zeta^{CWCB},
\]
\[
\zeta_B = \zeta^{WWVB} - \zeta^{CWCB},
\]
\[
\zeta_I = \zeta^{CWCB},
\]
\[
\zeta_N = [\zeta - \zeta_I] - [\zeta_W + \zeta_B].
\]

(Roman superscripts denote time series from the respective experiments, e.g., $\zeta^{WWVB}$ is anomalous sea level generated by climatological wind stresses and variable buoyancy fluxes.)

Finally, for correctly interpreting the experiments, it is necessary to explain one technical aspect of the model optimization: while the model solution is fit to data, it is generated by an unconstrained model run. First, initial conditions and boundary conditions are iteratively adjusted, following Wunsch and Heimbach (2007); thereafter, using the adjusted parameters, the model is freely run forward in time to produce the present results. This approach is distinct from optimization methods commonly used, for example, in numerical weather prediction. For more discussion on this topic, see Wunsch et al. (2007).

3. Generating mechanisms of $\zeta$ variability

An expanded view of $\zeta$ patterns in the tropical South Atlantic is given in Fig. 2a. Changes in $\zeta$ are relatively small equatorward of 20°S but can be somewhat larger at higher latitudes. To reveal underlying causal mechanisms, patterns produced by the various forcing scenarios are shown in the remaining panels of Fig. 2. In general, $\zeta_I$ and $\zeta^N$ do not constitute the dominant contributions to $\zeta$ changes (Figs. 2a,d,e), though they can be important at higher latitudes. We note that similar patterns of $\zeta_I$ variability in the tropical South Atlantic have been found in eddy-admitting general circulation model studies (see Fig. 13 of Cabanes et al. 2006 and Fig. 3 of Penduff et al. 2011). In contrast, components of the ocean’s direct response to the atmosphere make primary contributions at these tropical latitudes (Figs. 2a–c). In regions equatorward of 20°S, variations in $\zeta_W$ and $\zeta_B$ are larger than changes in $\zeta_I$, suggesting that $\zeta_W$ and $\zeta_B$ are out of phase. Figure 3 confirms this suggestion, showing anticorrelation between $\zeta_W$ and $\zeta_B$ in these low-latitude areas. Previous analyses in the North Pacific have noted similar instances of “destructive interference” between wind-driven and diabatically forced sea level changes (Thompson et al. 2002; Thompson and Ladd 2004).

To give a sense of spatial and temporal scales, in Fig. 4 we show basinwide Hovmöller plots of $\zeta_I$, $\zeta_W$, and $\zeta_B$ at four different latitudes. Qualitatively, $\zeta$ variability occurs predominantly on relatively short interannual time scales (periods of 2 years or so) and rather large zonal scales (thousands of kilometers), though there are cases in which changes occur at longer periods and/or shorter spatial scales. Moreover, the semicoherent striations of positive or negative $\zeta$ anomalies suggest westward propagation, the speed of which appears to decrease with increasing latitude. Considering $\zeta_W$ and $\zeta_B$ separately, westward propagation is more clearly evident, and $\zeta$ changes are seen to represent the partially canceling effects of the forced components. Some $\zeta_W$ signals appear to be produced along the eastern boundary. Observing similar boundary generation of propagating signals in the eastern South Pacific, Vega et al. (2003)
suggest the remote influence of equatorial pycnocline disturbances that are transmitted east along the equator and then poleward along the western coast of South America via Kelvin waves; perhaps similar processes underly the boundary generated signals seen here. Excepting latitudes close to the equator, variability in $\zeta_B$ is relatively low along the immediate vicinity of the eastern boundary and changes appear to be generated over the basin interior. Visually inspecting instances of interior generation of propagating $\zeta_W$ and $\zeta_B$ anomalies, it appears that in certain cases the respective forcing mechanisms may be collocated in time and space.

These results prompt several questions: Does the buoyancy-driven variability represent dynamical ocean adjustment or simply the local effects of atmospheric forcing? What accounts for the striking anticorrelation between wind-driven and buoyancy-forced sea level anomalies? In the remainder of this study, we consider these issues in depth.

4. Closed budgets of buoyancy-driven steric sea level

We begin investigating the nature of $\zeta_B$ changes by considering spatial variability patterns of buoyancy-driven steric sea level $\zeta_{p,B}$ (Fig. 5a). Distributions of $\zeta_B$ and $\zeta_{p,B}$ are nearly identical (cf. Figs. 2c and 5a), demonstrating that the effects of ocean bottom pressure changes driven by surface buoyancy fluxes (not shown) are negligible. To determine whether changes in $\zeta_{p,B}$ implicate local atmospheric forcing or ocean dynamics, we exploit the consistency of the state estimate. Following the methodology of Piecuch and Ponte (2011), steric sea level $\zeta_p$ can be completely partitioned into contributions from advection by ocean currents $A$, parameterized mixing $M$, and local surface buoyancy forcing $F$.

$$\zeta_p = A + M + F.$$  

For general discussion of this budget and its formulation, see Piecuch and Ponte (2011).

In regions of anticorrelation noted previously (Fig. 3), changes in $\zeta_{p,B}$ represent the joint action of local forcing $F_B$ and advection $A_B$ (Figs. 5a–c). Parameterized mixing processes $M_B$ mostly do not contribute in these areas, but along the coast of South America and south of 20°S these effects can be important (Fig. 5d). The dissimilarity between patterns of $\zeta_{p,B}$ and $F_B$ demonstrates that integration of the local surface buoyancy fluxes is not a good indicator of the total sea level response to anomalous buoyancy forcing.

Hovmöller diagrams of major budget terms (Fig. 6) elucidate the dynamical nature of buoyancy-driven
FIG. 4. Hovmöller diagrams of (top) $\zeta$, (middle) $\zeta_B$, and (bottom) $\zeta_W$, computed along (left) $5^\circ$S, (middle left) $10^\circ$S, (middle right) $15^\circ$S, and (right) $20^\circ$S in the South Atlantic. Black contour represents zero crossing. Scale bar has units of cm.
steric sea level changes. We observe that the various budget terms in Eq. (2) can be characterized by variability on distinct scales. For example, dominant changes in steric height $z_r$ occur mainly over shorter interannual periods, whereas important changes in local forcing $F_B$ sometimes occur over relatively long time scales. This contrast highlights the varying action of the advection term: evident are longer period “local” changes in $A_B$, particularly at the lowest latitudes, which act mainly to balance $F_B$, as well as shorter period “nonlocal” $A_B$ variations, which produce much of the apparent westward propagation. [Although the atmospheric dynamics are not pursued, note that there are some cases in which the forcing term seems to contribute to the westward propagation as well (e.g., Fig. 6e).]

This budget analysis demonstrates the role of advective ocean transports, yet since it incorporates full ocean dynamics described by the nonlinear primitive equations, it does not follow that this behavior can be understood within the framework of linear Rossby wave theory. Explicit demonstration of the action of Rossby waves is the target of the next section.

5. Baroclinic response to surface buoyancy exchanges

To interpret the budget analysis above, we consider the first baroclinic mode sea level response to surface buoyancy forcing described by the linear equation (see appendix),

$$\frac{\partial \eta_B}{\partial t} - \zeta_r \frac{\partial \eta_B}{\partial x} = -\frac{D_{\text{mix}} B_s}{D^2} \rho_0,$$

FIG. 5. (left) Spatial distributions of closed $\zeta_{r,B}$ budget terms. Standard deviations of (a) steric height $\zeta_{r,B}$; (b) local forcing $F_B$; (c) advection $A_B$; and (d) parameterized mixing $M_B$. (right) Spatial distributions of linear buoyancy forced Rossby wave model solutions. Standard deviations of (e) interior solution $\eta_B^{\text{int}}$; (f) static solution $\eta_B^{\text{static}}$; (g) dynamic solution $\eta_B^{\text{dynam}}$; and (h) boundary-forced solution $\eta_B^{\text{bound}}$. We do not consider linear solutions within 3° of the equator. Scale bar has units of cm.
FIG. 6. Hovmöller diagrams of (top) $\zeta_{\rho B}$, (middle) $F_B$, and (bottom) $A_B$, computed along (left) 5°S, (middle left) 10°S, (middle right) 15°S, and (right) 20°S in the South Atlantic. Black contour represents zero crossing. Scale bar has units of cm.
where $\eta_B$ represents the sea level response, $c$, the phase speed of baroclinic Rossby waves, $B_s$ surface buoyancy exchanges, $\rho_0$ a reference density, $D_{mix}$ the mixed layer depth, and $D^F$ the equivalent forcing depth (e.g., Gill 1982). We use the variable $\eta$ to denote linear model solutions, distinguishing them from ECCO-produced sea level $\zeta$. This framework assumes a low Rossby number and neglects the influences of background mean flows and bottom topography. Moreover, by focusing solely on the first baroclinic mode, we implicitly assume that higher baroclinic modes are unimportant (cf. Thompson and Ladd 2004). The solution to Eq. (3) is

$$\eta_B = \eta_B^{[\text{int}]} + \eta_B^{[\text{bound}]} ,$$

(4)

where the interior solution $\eta_B^{[\text{int}]}$ is computed by accumulating the surface buoyancy forcing along Rossby wave characteristics from the eastern boundary $x_r$,

$$\eta_B^{[\text{int}]}(x,y,t) = \frac{D_{mix}}{D^F \rho_0 c_r} \int_{x_r}^x B_s(x',y,t + \frac{x-x'}{c_r}) \, dx' ,$$

(5)

and the boundary generated solution $\eta_B^{[\text{bound}]}$ represents westward propagation of signals produced at the eastern boundary,

$$\eta_B^{[\text{bound}]}(x,y,t) = \eta_B(x_r,y,t + \frac{x-x_r}{c_r}) \exp \left( \epsilon \frac{x-x_r}{c_r} \right) .$$

(6)

We draw attention to the exponential decay term, characterized by the Newtonian dissipation rate $\epsilon$, appended to the boundary solution. While this is not described explicitly by the dynamics in Eq. (3), it is included as a simple means of capturing the tendency for boundary-generated signals to be dissipated, for example, as found by Vega et al. (2003) in the eastern South Pacific. Our reason for including dissipation in the boundary solution but not in the interior solution is that dissipation time scales in the open ocean are longer than those at the boundaries (see Qiu and Chen 2006 and references therein), hence we assume such effects are unimportant for the interior case. Finally, the interior solution can be decomposed into a static component, defined as the time integral of the local forcing (cf. Cabanes et al. 2006),

$$\eta_B^{[\text{static}]}(x,y,t) = -\frac{D_{mix}}{D^F \rho_0} \int_0^t B_s(x,y,t') \, dt' ,$$

(7)

as well as a dynamic component, equal to the difference of interior and static terms,

$$\eta_B^{[\text{dyn}]} = \eta_B^{[\text{int}]} - \eta_B^{[\text{static}]} .$$

(8)

We evaluate linear model solutions on a 1° horizontal grid. The forcing term $B_s$ is set equal to the monthly ECCO surface buoyancy forcing. The phase speed $c_r$ is defined as a function of latitude, computed using the Liouville–Green approximation (e.g., Gill 1982),

$$c_r(y) = \frac{1}{f^2} \frac{d}{dy} \left( \frac{1}{\pi} \int_0^H N(x,y,z) \, dz \right)^2 ,$$

(9)

where $\langle \rangle$ denotes zonal average at constant latitude (hence the term within the square brackets is the zonal mean gravity wave phase speed). $H$ is the ocean depth, and $N$ represents the Brunt–Väisälä frequency computed at each grid cell from the time-mean ECCO density structure. We assume $D_{mix}$ and $D^F$ are comparable (cf. Wunsch and Gill 1976; Doi et al. 2010) and set their ratio equal to unity. In the place of $\eta_B$ on the right-hand-side of Eq. (6), we use the corresponding eastern boundary values of the ECCO buoyancy-driven solution $\xi_B$. Assuming the representativeness of their results, we follow Vega et al. (2003), setting $\epsilon = (300 \text{ days})^{-1}$, but note that our results are relatively insensitive to the value of $\epsilon$.

Variability distributions characterizing the interior solution $\eta_B^{[\text{int}]}$ and buoyancy-driven steric sea level $\xi_{B,p,B}$ show good correspondence, the former capturing major characteristics of the latter (cf. Figs. 5a and 5e). Between 5°S and 25°S, the correlation coefficient of the two spatial patterns is 0.73. Differences in regional detail are apparent, however. For example, magnitudes of $\eta_B^{[\text{int}]}$ are comparatively small in the northeast corner of the domain; here, boundary generated signals are important (Fig. 5h). Amplitudes of $\eta_B^{[\text{int}]}$ are also small along lowlatitude eastern interior regions; such discrepancies may signal a need to consider slower moving, higher-order baroclinic modes to capture satisfactorily the buoyancy-driven sea level behavior. In contrast, at higher latitudes and along the coast of South America, $\eta_B^{[\text{int}]}$ magnitudes are relatively large; because parameterized mixing is important in roughly the same regions in the ECCO solution, these differences suggest places where it is necessary to incorporate dissipative effects, which are not included in the interior solution.

Linear model solutions also provide a good description of other major characteristics of the buoyancy-driven steric components from ECCO. For example, the linear solution $\eta_B$ reproduces (to lowest order) the timing, scales, and westward propagation associated with $\xi_{B,p,B}$ variations (cf. Figs. 6 and 7). Considering the four latitude bands, the correlation coefficient between
longitude–time distributions of $\eta_\B$ and $\xi_\B$ is maximum along $5^\circ$S (0.71) and minimum along $10^\circ$S (0.44), while intermediate values of 0.53 and 0.60 describe their relationship along $15^\circ$ and $20^\circ$S, respectively. Also worthy of mention, the rich variability characteristics associated with the buoyancy-driven advection from the general circulation model solution—including all resolved ocean transports described by the nonlinear primitive equations—are nicely captured by the relatively simple dynamical component of the linear baroclinic model (cf. third rows of Figs. 6 and 7). Strong correspondence between these two terms can also be inferred from their spatial variability patterns, which are very similar in most places (cf. Figs. 5c and 5g). Discrepancies between the two terms do occur, but are mainly restricted to areas where parameterized mixing is important.

6. Understanding the relationship between $\xi_W$ and $\xi_B$

Why are $\xi_W$ and $\xi_B$ anticorrelated over much of the tropical South Atlantic (Fig. 3)? To test whether such behavior is to be expected within a linear Rossby wave model framework, we first consider the gravest baroclinic mode sea level response to wind forcing $\eta_W$ described by the linear equation (see appendix),

$$\frac{\partial \eta_W}{\partial t} - c_5 \frac{\partial \eta_W}{\partial x} = -\frac{D}{D^F} \left( \nabla \times \frac{\tau}{f \rho_0} \right),$$

where $D$ is the equivalent depth, $D^F$ the equivalent forcing depth, $f$ the Coriolis parameter, and $\tau$ the wind stress vector. The full solution to Eq. (10), as well as its various components (boundary and interior, static and dynamic), can be computed using the general methodology outlined in the previous section, making the necessary substitutions of forcing terms. Linear wind-driven solutions are computed on a 1° horizontal grid with the same Rossby wave phase speeds used in the previous section. Wind stresses from ECCO are used for the forcing term $\tau$ and the ratio $D/D^F$ is set to a constant value of $2.5 \times 10^{-3}$ (e.g., Wunsch and Gill 1976).

Hovmöller plots of linear wind-forced solutions are shown in Fig. 8. Overall, the full solution $\eta_W$ roughly reproduces the general behavior of the ECCO solution $\xi_W$ (cf. first row of Fig. 8 and second row of Fig. 4). Considering boundary and interior terms (second and third rows of Fig. 8), eastern boundary generation is important nearer to the equator at $5^\circ$S and $10^\circ$S whereas away from the equator at $20^\circ$S interior wind stress curl predominates; interior and boundary mechanisms both can make contributions along $15^\circ$S, depending on the longitude. The influence of the boundary generated Rossby waves can be understood partly by considering the ratio of the phase speed $c_5$ to the dissipation rate $\epsilon$ [cf. Eq. (6)], which represents the distance traveled in one eddy folding time period: at lower latitudes, the South Atlantic basin width $\ell$ is less than $c_5/\epsilon$ (not shown), hence boundary generated signals can cross the basin quickly and more-or-less undamped, exerting a basin wide influence (Fig. 8); at higher latitudes, in contrast, $\ell$ is larger than $c_5/\epsilon$ (not shown), so boundary-generated anomalies propagate slowly and are strongly dissipated as they cross the basin (Fig. 8).

Correlations between linear Rossby wave solutions $\eta_W$ and $\eta_B$ are shown in Fig. 9. These values can be directly compared to the correlation coefficients between $\xi_W$ and $\xi_B$ given in Fig. 3. Although there are quantitative differences between these two figures, the patterns are broadly similar, and negative correlation values are found in most places equatorward of $20^\circ$S (cf. Figs. 3 and 9). This suggests that the simple mechanisms considered here capture at least some rough qualitative aspects of the general opposing $\xi_W$ and $\xi_B$ behavior.

Given the compensating relationship between the linear solutions, we would like to explore the relative influences of boundary generation and interior forcing. Based on Fig. 3, which shows that correlations between ECCO solutions $\xi_W$ and $\xi_B$ are mostly less than $-0.5$ equatorward of $15^\circ$S along the eastern boundary, one can infer broad compensation between boundary solutions $\eta_B^{[\text{bound}]}$ and $\eta_W^{[\text{bound}]}$ at low latitudes. To elucidate the relation between interior forcing effects, we show correlations between $\eta_W^{[\text{in}]}$ and $\eta_B^{[\text{in}]}$ in Fig. 10a. Interior solutions can be opposing across broad zonal regions, for example, between $15^\circ$ and $20^\circ$S. This is in contrast to correlation patterns between static components $\eta_W^{[\text{static}]}$ and $\eta_B^{[\text{static}]}$, which exhibit a smaller-scale structure with contiguous regions of correlation having limited spatial extent (Fig. 10b). The general patterns in Fig. 10 offer qualitative physical insight: apparent is a tendency for regions of compensating interior solutions (i.e., blue areas of Fig. 10a) to extend westward from areas of opposing static components (blue regions in Fig. 10b). Interpretation could be that—through the ocean’s dynamical adjustment—localized centers of anticorrelation between the integrated effects of wind stress curl and surface buoyancy forcing generate interior Rossby waves that lead to large scale anticorrelation “downstream.”

Putting the pieces together (Figs. 8–10), compensation between $\eta_W$ and $\eta_B$ at lower tropical latitudes results mainly from boundary generation whereas at higher tropical latitudes it is due more to interior forcing. But, of course, this suggestion elicits further questions: Why are interior forcing effects compensating? What is the origin of the compensating eastern boundary signals?
FIG. 7. Hovmöller diagrams of (top) $\eta_0$, (middle) $\eta_0^{\text{static}}$, and (bottom) $\eta_0^{\text{dyn}}$, computed along (left) 5°S, (middle left) 10°S, (middle right) 15°S, and (right) 20°S in the South Atlantic. Black contour represents zero crossing. Scale bar has units of cm.
FIG. 8. Hovmöller diagrams of (top) $\eta_W$, (middle) $\eta_W^{[\text{bound}]}$, and (bottom) $\eta_W^{[\text{int}]}$ computed along (left) 5°S, (middle left) 10°S, (middle right) 15°S, and (right) 20°S in the South Atlantic. Black contour represents zero crossing. Scale bar has units of cm.
One simple explanation might go as follows: suppose anomalous Ekman pumping driven by wind stress curl changes leads to a deepening thermocline, and a corresponding positive (steric) sea level anomaly and warming of the upper ocean; all else being equal, this upper-ocean warming would, in turn, drive sensible heat loss, cooling the ocean; this cooling due to heat flux would result in a negative buoyancy-driven (steric) sea level anomaly. Under such a scenario, the wind-forced dynamics are essentially damped by the effects of buoyancy. The generally weaker buoyancy-driven variability over large regions (e.g., Fig. 11b) is consistent with this hypothesis. Wind stresses and buoyancy exchanges can also be related through other mechanisms (e.g., high winds usually relate to strong heat fluxes) and more complex air–sea coupling (e.g., Xie 1997, 1999) could be at play in the observed compensation between $\eta_W$ and $\eta_B$. A full exploration of these issues is beyond our scope and left for future study.

7. Discussion and conclusions

Our goal was to elucidate low interannual sea level variability in the tropical South Atlantic and to understand the nature and action of buoyancy-driven sea level changes. We used solutions to a data-constrained general circulation model from ECCO as well as simple linear solutions of baroclinic ocean response. Based on results from forcing experiments performed with the ECCO model, we submit that low levels of variability (Fig. 2a) reflect in part the ocean’s direct response to the atmosphere, in particular the opposing action of buoyancy-driven and wind-driven changes (Figs. 2–4). To lowest order, buoyancy-driven sea level changes produced by the data-constrained general circulation model can be understood within the framework of a linear baroclinic Rossby wave model forced by the atmosphere and the eastern boundary (Figs. 5–7). What is more, rough characteristics of basinwide anticorrelation between sea level changes driven by winds and buoyancy apparent in ECCO (Fig. 3) can be reproduced qualitatively using the same linear model framework (Figs. 9–10).

To provide a wider scope, we show in Fig. 11a correlations between anomalous wind- and buoyancy-forced sea level changes (from the ECCO solution) across the global tropical ocean. In addition to the South Atlantic region considered here, broad portions of the tropical Indian, Pacific, and North Atlantic Oceans are also characterized by strong anticorrelation as well as comparable levels of wind- and buoyancy-forced variability (Fig. 11b). The large contiguous region in the southeast tropical Pacific is the focus of Piecuch and Ponte (2012) who present qualitatively similar results regarding the nature and influence of buoyancy-driven ocean dynamics. Therefore, although our analysis here was geographically confined to the case of the South Atlantic, we argue that our results are general and that similar findings apply to other tropical basins.

While Vinogradov et al. (2008) hint at a dynamic sea level response to seasonal buoyancy forcing, most past studies of annual surface heating effects on sea level focus on the locally forced (or static) component, ignoring any dynamical adjustment (e.g., Wang and Koblinsky 1996; Stammer 1997; Gilson et al. 1998; Vivier et al. 1999). The assumption that annual heating does not involve a dynamic response is based on the notion that, because the forcing is coherent over large spatial scales, annual heating is unlikely to introduce significant zonal pressure gradients and related currents (Stammer 1997). Some recent investigations into interannual sea level changes implicitly make this same assumption, concentrating on local forcing and neglecting buoyancy-driven ocean dynamics (e.g., Qiu 2002; Cabanes et al. 2006; Zhang et al. 2010). The main
results of this study—as well as those of Piecuch and Ponte (2012) for the South Pacific and Thompson and Ladd (2004) for the North Pacific—contradict such notions, demonstrating that local effects do not constitute the total sea level response to buoyancy forcing.

A clarifying note on interpretation is necessary. While this study demonstrates that buoyancy forcing is capable of triggering baroclinic ocean response through linear Rossby waves, our observing this phenomenon in our model solutions is, in a sense, an artifact of having separated and considered in isolation the dynamical influences of momentum and buoyancy input. In the real world, these mechanisms do not act in isolation and, as we have found here, can be related. To the extent that generating mechanisms are collocated in space in time (e.g., Fig. 10), compensation between buoyancy-driven and wind-forced anomalies occurs locally, and, although it is mathematically equivalent to regard the resulting signal as a superposition of separate effects, the result is one signal driven by both momentum and buoyancy exchanges.

Finally, our main emphasis here has been on the relatively linear regime (apparent at lower latitudes) where sea level anomalies largely represent the ocean’s direct response to the atmosphere and where buoyancy-driven anomalies can be understood essentially in terms of linear wave theory. Although not the focus of much discussion, our results also point to a more complex regime (apparent at higher latitudes) where parameterized mixing processes, intrinsic variability, and nonlinear response to the atmosphere can all be important (Figs. 2 and 5). More broadly, the budget analysis of Piecuch and Ponte (2011) demonstrates that parameterized mixing contributes importantly to extratropical sea level changes generated by a coarse-resolution general circulation model. Moreover, a recent assessment by Penduff et al. (2011) reveals that intrinsic interannual sea level variability is substantial in many regions of the global ocean. Less well documented is the influence of nonlinear response to the atmosphere on observed sea level changes. Because it is important to understand such complex dynamics and to ensure that they are realistically simulated in ocean models, these topics should be the focus of future studies.

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**APPENDIX**

**Linear Ocean Response to Surface Winds and Buoyancy Exchanges**

Here we formulate a linear model of the oceanic response to momentum and buoyancy exchanges at the sea surface. Following Gill (1982), we consider the equations governing a linear and inviscid, hydrostatic and Boussinesq fluid near to steady state, assuming a rigid lid and neglecting both bottom topography and mean currents,

\[ -fv = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{1}{\rho_0} \frac{\partial x}{\partial z}, \quad (A1a) \]

\[ fu = -\frac{1}{\rho_0} \frac{\partial p}{\partial y} + \frac{1}{\rho_0} \frac{\partial y}{\partial z}, \quad (A1b) \]

\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad (A1c) \]
\[ \frac{\partial^2 p}{\partial t \partial z} + \rho_0 N^2 w = -g \frac{\partial B}{\partial z}. \quad \text{(A1d)} \]

Here \( u, v, \) and \( w \) above are the zonal, meridional, and vertical components of the velocity, respectively; \( p \) is hydrostatic pressure and \( \rho_0 \) is a constant reference density; \( f \) and \( g \) are the Coriolis parameter and acceleration due to gravity, respectively; \( X \) and \( Y \) are horizontal stresses; \( N \) is the Brunt–Väisälä frequency; and \( B \) is a source of buoyancy (kg m\(^{-2}\) s\(^{-1}\)). Variables \( u, v, w, p, X, Y, \) and \( B \) represent deviations from a time mean, and can be written as linear combinations of vertical modes,

\[ [u; v] = \sum_{n=0}^{\infty} [\hat{u}_n(x, y, t); \hat{v}_n(x, y, t)] \hat{\psi}_n(z) / \rho_0 g, \quad \text{(A2a)} \]

\[ w = \sum_{n=0}^{\infty} \hat{w}_n(x, y, t) \hat{\psi}_n(z), \quad \text{(A2b)} \]

\[ p = \sum_{n=0}^{\infty} \hat{\xi}_n(x, y, t) \hat{\psi}_n(z), \quad \text{(A2c)} \]

\[ [\frac{\partial X}{\partial z}; \frac{\partial Y}{\partial z}] = \sum_{n=0}^{\infty} [\hat{X}_n(x, y, t); \hat{Y}_n(x, y, t)] \hat{\psi}_n(z) / g, \quad \text{(A2d)} \]

where \( \hat{\phi}_n \) and \( \hat{\psi}_n \) have units of height and pressure, respectively, satisfying the system,

\[ \hat{\psi}_n = \rho_0 c_n^2 \frac{d\hat{\phi}_n}{dz}, \quad \text{(A3a)} \]

\[ \frac{d\hat{\phi}_n}{dz} = -\rho_0 N^2 \hat{\phi}_n, \quad \text{(A3b)} \]

with \( c_{n} \) physically representing the mode \( n \) gravity wave phase speed. Omitting subscript \( n \) hereafter to simplify the notation, the modal sea level perturbation \( \eta(x, y, t) \) is defined as

\[ \eta = \frac{\hat{\psi}_0}{\rho_0 g}, \quad \text{(A4)} \]

where \( \hat{\psi}_0 = \hat{\psi}(0) \).

As a simplification, we assume that the buoyancy source is confined to the upper ocean, uniformly distributed within a mixed layer of depth \( D_{\text{mix}} \) and zero otherwise,

\[ B = \begin{cases} B_s & \text{if } 0 < z < -D_{\text{mix}} \\ 0 & \text{if } -D_{\text{mix}} < z < -H \end{cases}, \quad \text{(A5)} \]

where \( H \) is the full ocean depth, and \( B_s \) is the surface buoyancy flux. Roughly speaking, one can interpret this form as resulting from buoyancy anomalies due to surface exchanges that are subsequently redistributed by turbulent processes within the mixed layer. This assumption effectively precludes and bypasses interesting questions, beyond our scope, regarding the vertical structure of the buoyancy forcing term and the physical mechanism(s) responsible for setting it. Such questions are left for future study. We also assume that the stresses decay linearly over the mixed layer (Cabanes et al. 2006),

\[ \frac{\partial \eta}{\partial t} - c_r \frac{\partial \eta}{\partial x} = -\frac{D_{\text{mix}} B_s}{\rho_0} - \frac{D}{\rho_0} \left[ \frac{\nabla \times \tau}{f \rho_0} \right], \quad \text{(A7)} \]

where \( \tau_x \) and \( \tau_y \) are zonal and meridional surface wind stresses, respectively.

Equations (A1)–(A6) can be combined to yield a forced Rossby wave equation for the sea level response,

\[ D^F = \int_0^{-H} \frac{\hat{\psi}_n^2}{\phi_0} \int_{-D_{\text{mix}}}^{0} \hat{\psi} d z. \quad \text{(A8)} \]

The forcing terms on the right-hand-side of Eq. (A7) thus depend on the surface exchanges as well as the stratification. The effects of the forcing terms can be considered in isolation to study the effects of buoyancy forcing \( \eta_B \) and wind driving \( \eta_W \) separately,

\[ \frac{\partial \eta_B}{\partial t} - c_r \frac{\partial \eta_B}{\partial x} = -\frac{D_{\text{mix}} B_s}{\rho_0} \quad \text{(A9a)} \]

\[ \frac{\partial \eta_W}{\partial t} - c_r \frac{\partial \eta_W}{\partial x} = \frac{D}{\rho_0} \left( \nabla \times \frac{\tau}{f \rho_0} \right), \quad \text{(A9b)} \]

with \( \eta = \eta_B + \eta_W \). In this study, Eqs. (A9a,b) are examined for the case of the gravest baroclinic mode (\( n = 1 \),
which is analogous to assuming a 1.5-layer ocean, similar to other studies of interannual tropical sea level (e.g., Qiu and Chen 2006). The generally good correspondence between linear model solutions and ECCO solutions gives us confidence that this assumption is, at the very least, useful; however, we note that higher baroclinic modes probably are not entirely negligible (e.g., Thompson and Ladd 2004; Doi et al. 2010). Considering higher-order modes would be straightforward, but, given our scope here, we forego such consideration.

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