Annual and interannual sea level variations in the Indian Ocean from TOPEX/Poseidon observations and ocean model simulations

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Abstract. Sea level variations relative to a 4-year mean in the Indian Ocean north of 10°S are examined during 1993–1996 using both a numerical reduced gravity model with realistic coastline geometry and wind stress and sea level measurements from the TOPEX/Poseidon altimeters. The annual signal is found to be composed of propagating as well as nonpropagating features. The propagation speeds in the model and altimetry generally agree to within 25% or less. Complex empirical orthogonal function (CEOF) decomposition yields a separation between the annual and semiannual cycles (46 and 30% of the respective variance for the model, and 40 and 26% for the altimetric measurements, respectively). The propagation of these signals across the ocean basin is indicated by the spatial phase functions. Both temporal phase functions are steady for the annual cycle, though the amplitudes are modulated in time. The results for the semi-annual cycles are similar, but the temporal phase functions are disrupted for ~12 months starting in 1994. This may be due to an unusually strong monsoon during that time. The correlation between model sea level variation and those measured by altimetry is highly variable in both space and time. Low-frequency filtering of the sea level anomalies, obtained by summing the two largest CEOF modes (the annual and semiannual cycles), improves the correlation. The filtered anomalies correlate in time as high as 0.9 in the western Arabian Sea and as high as 0.7 south of the equator and in the eastern Bay of Bengal. There are pockets of poor correlation (as low as -0.4) in the eastern Arabian Sea, central Bay of Bengal, and central equatorial region. These areas tend to contain recurring Rossby wave interactions as represented by the 1.5-layer model. Each area is associated with a “phase nexus” (analogous to an amphidromic point in tidal theory) or a strong gradient of the model spatial phase functions. The spatial correlation between the filtered anomalies is typically 0.6 over much of the observation period but contains unexplained declines as low as 0.2 during a few months in both 1995 and in 1996.

1. Introduction

The Indian Ocean (IO) is characterized by seasonal changes of its surface circulation due to the two opposite northeast and southwest monsoon regimes. The Somali Current flows northeastward during the southwest monsoon and flows southwestward during the northeast monsoon. Gyre circulations in the Arabian Sea (AS) and in the Bay of Bengal (BB) also reverse semiannually. The seasonal variation of the wind-driven circulation pattern in the IO is greater than that in any other ocean.

Previous modeling studies of this ocean have generally been concerned with the Somali Current reversal and the generation of the gyres like the Great Whirl and the Southern Gyre in the northwestern IO during the onset of the summer monsoon (Luther and O'Brien,
In situ observations of this ocean are typically confined to small regions, as was done by Schott [1987], although there are studies describing the seasonal variations across the entire IO [Rao et al., 1989; Molinari et al., 1990]. Rao et al. [1989] generated mean monthly mixed layer depth, sea surface temperature, and surface current climatologies for the tropical IO, while Molinari et al. [1990] used three different satellite-tracked drifting buoy data sets for generating mean monthly climatology of surface currents in the tropical IO. They also computed phase and amplitude of the first two harmonics of the annual cycle of the surface currents.

Satellite altimeters are very useful for studying the ocean circulation as they provide nearly global observations of sea level. These observations can be used for validating model results and also for improving the models. Perigaud and Delecluse [1992, 1993] analyzed Geosat satellite altimeter data for comparison with numerical simulations from a reduced gravity model forced by observed winds. After decomposition into complex empirical orthogonal functions (CEOF's) they found that the low-frequency anomalies are largely described by the first two modes for observations as well as simulations. It was also found that basin-averaged observed and simulated sea level changes are moderately correlated over 1985–1988. However, the accuracy of Geosat data is questionable because of the absence of an onboard radiometer and also because of its poor orbit accuracy.

Stammer et al. [1996] compared the sea level response of the 0.25° global ocean model of Semtner and Chervin [1992] to observations by the TOPEX/Poseidon altimeter (TP) over ~2 years. They found that on scales greater than 2° the model sea level variations associated with the seasonal cycle in the IO were similar to those in TP observations. However, they did not do a detailed comparison within the IO. Their model amplitude was roughly half the TP amplitudes. This was attributed to incomplete model surface boundary conditions at the large scale, but at smaller scales it was speculated to result from spatial resolution that was too coarse or artificial damping terms.

In the present study altimetric data from TP are used for detecting mesoscale to basin-scale sea level variations over the IO north of 10°S. TP altimetry provides an opportunity to study the contribution of planetary waves to sea level variations because of its rapid 10-day repeat orbit and high accuracy [Chelton and Schlax, 1996].

The observed sea level variations are compared with simulations derived from a reduced gravity model forced by observed winds. Vertically integrated ocean models, as used here, exchange decreased vertical resolution for increased horizontal resolution. The latter is essential for an accurate numerical representation of wave processes [Grotjahn and O'Brien, 1976]. A 1.5-layer version is used for its simplicity and to facilitate comparison to the results of Perigaud and Delecluse [1993]. This is the first comparison between TP sea level variations with this class of models in the IO. Therefore it is useful to start with the simplest model and gradually increase the model complexity in later studies in order to discern what dynamical mechanisms are responsible for the sea level changes.

Section 2 provides a brief description of the TP data. Section 3 presents the model. Section 4 describes the complex CEOF technique, and section 5 presents the results. Section 6 is a summary and discussion.

2. Altimeter Data

The observational data used in the study are the sea level anomaly heights from the TP altimeter. This database contains 1° x 1° gridded sea level anomaly heights observed by the TP every 10 days. Detailed information about data processing and error analysis are given by Tapley et al. [1994]. We summarize the salient features.

The usual media and instrument corrections (ionosphere, wet and dry troposphere, and electromagnetic bias) and geophysical corrections (tides and inverted barometer) have been applied to the measurement. Subsequently, the altimeter measurements have been reduced to sea surface topography heights using a precise orbit computed at the University of Texas Center for Space Research using the available satellite laser ranging and Doppler tracking. Data are interpolated to a common grid center along a high-resolution mean sea surface computed from 4 years of TP data and then averaged into 1° x 1° grids and smoothed over 600 km wavelengths with a Gaussian weighted filter with a roll-off of three degrees. The estimated accuracy of the sea level anomalies in this data is 3–4 cm over the entire ocean.

The sea level anomaly (SLA) examined throughout this article is the deviation from the 4-year mean surface height. We extract the data for 4 years (1993–1996) consisting of 144 cycles in our region of interest, which is from 10°S to 23°N and from 40–119°E. The starting cycle corresponds to January 10, 1993.

3. Model

The ocean model is a fully nonlinear, 1.5-layer reduced gravity system formulated in spherical coordinates. The dynamical equations are
The wind data used for forcing the model are derived from the monthly pseudo stress data from the Florida State University. These data cover the entire IO with a resolution of $1^\circ$ [Legler et al., 1988]. They are converted into wind stress utilizing a constant drag coefficient of $1.3 \times 10^{-3}$. The model is spun up from rest for 6 years using the 1977–1992 mean of monthly wind fluctuations. After this initial spin-up the model is run for 4 years starting January 10, 1993. The 4-year mean interface depth field is subtracted for consistency with the TP data. The model results are stored for 4 years starting from January 10, 1993, at 10-day intervals.

4. CEOF Technique

This technique is generally applied to the analysis of two-dimensional data sets for resolving the dominant propagating modes of variability [Horel, 1985]. Propagating structures are extracted from a data set by decomposing the data into CEOFs and by sorting the dominant modes in order of decreasing variance. This technique has been applied to the detection of low-frequency propagating Rossby wave signals off the eastern Pacific coast [Shriver et al., 1991] and IO [Perrigaud and Delecluse, 1992, 1993].

Denote a real time-varying data field by $d(x, y, t)$, where $(x, y)$ indicates spatial position and $t$ is time. Denote a complex data field by $D(x, y, t)$, the real part of which is the original data field, $d$, and the imaginary part is the quadrature (Hilbert transform) of $d$. The latter is the data field phase advanced by $\pi /2$ in time. More specifically, if the data field $d(x, y, t)$ has the Fourier representation

$$d(x, y, t) = \sum_\omega a(x, y, \omega) \cos \omega t + b(x, y, \omega) \sin \omega t,$$

then the Hilbert transform $d_H(x, y, t)$ can be represented as

$$d_H(x, y, t) = \sum_\omega b(x, y, \omega) \cos \omega t - a(x, y, \omega) \sin \omega t.$$

Then $D(x, y, t)$ can be written as

$$D(x, y, t) = d(x, y, t) + id_H(x, y, t),$$

where $i = \sqrt{-1}$.

The next step in the analysis is the formation of the covariance matrix of the data field $D(x, y, t)$. The covariance matrix of $D$ is Hermitian and positive definite by construction. It thus possesses real positive eigenvalues and a set of orthogonal complex eigenvectors $s_n(x, y)$. Hence $D(x, y, t)$ can be represented as
Plate 1. Spatial patterns of the first CEOF for the model simulated and TP observed sea level anomaly. (a) Model Amplitude, (b) Model Phase, (c) TP Amplitude, and (d) TP Phase. The color bar indicates both amplitude (lower numbers) and phase in degrees (upper numbers).
Plate 2. Spatial patterns of the second CEOF eigenvector for the model simulated and TP observed sea level anomaly. (a) Model Amplitude, (b) Model Phase, (c) TP Amplitude, and (d) TP Phase. The color bar indicates both amplitude in cm (lower numbers) and phase in degrees (upper numbers).
where the asterisk implies complex conjugation. The functions $s_n(x, y)$ will be referred to as spatial functions. The principal components $r_n(t)$ are obtained as

$$r_n(t) = \sum_{x,y} D(x, y, t) s_n(x, y)$$

and will be referred to as temporal functions. In practice, these are finite-dimensional vectors because of the discrete representation of the data set.

From each eigenmode $n$, 4 measures can be defined for representing features in the data. The spatial amplitude function $S_n(x, y)$ represents the spatial distribution of variability associated with each eigenmode, and the spatial phase function $\theta_n(x, y)$ shows the relative phase of fluctuation among the various spatial locations where the data field $d$ is defined. The spatial gradient of $\theta_n(x, y)$ provides the local wavenumbers. Similarly, the temporal amplitude function $R_n(t)$ and the temporal phase function $\phi_n(t)$ provide measures of temporal variability associated with each eigenmode. The time-derivative of $\phi_n(t)$ is a measure of frequency.

In the present study, the CEOF technique is applied to 10 day intervals of TP and model SLA in the region bounded zonally between 23°N and 10°S and meridionally between 40°E and 119°E and from January 10, 1993 to November 28, 1996. This excludes the dampening layer in the southern part of the model domain (29°S to 10°S).

5. Results

5.1. Model

It is well known that reduced gravity models can simulate the semiannual reversal of the Somali Current and the gyre circulations in the AS as well as BB [Luther and O'Brien, 1985; Potemra et al., 1991; McCreary et al., 1993]. The model used in this study also produces an annual cycle with these circulation features. The seasonal cycle of SLA is examined in summer and winter.

Figure 1. The sea level anomalies simulated by the model on the indicated dates. Units are centimeters.
for both model and TP. Interannual variability is also found. The prominent features are now highlighted.

Some model structures recur in the boreal summer circulation of the upper IO in all years examined. Among these are the anticyclonic (as indicated by a maximum of SLA) Great Whirl and a lesser anticyclonic eddy in the AS, though their relative strengths vary interannually (Figure 1a). For example, in 1993 the Great Whirl has a peak amplitude over 5 cm, and the AS circulation has a peak amplitude over 3 cm. In 1996 the Great Whirl is ~6 cm, but the AS circulation achieves 8 cm. The BB shows little recurring structure, with a mixture of cyclonic and anticyclonic circulation in 1993, generally cyclonic in 1994, predominantly anticyclonic in 1995, and generally cyclonic in 1996. There is a weak tendency for the western BB to contain cyclonic vorticity. The eastern equatorial region has a declining sea level toward the east in 1994 and 1995, suggestive of an upwelling Kelvin wave, but a relatively flat surface in 1993 and rising SLA in 1996. Between the equator and 10°S the SLA is usually comprised of several positive and negative recirculation regions with amplitudes of roughly ±6-8 cm. The relative number, position, and strength of these recirculations are different each year. Similar results are noted in the boreal winter circulation (Figure 2), though often the direction of circulation is reversed as compared to summer. This reversal is particularly prominent in the AS, which is dominated by cyclonic circulation during this season.

Time-longitude plots of model SLA at 15°N, 10°N, and 10°S (Figure 3) show westward propagation, indicating Rossby waves dominate in the eastern AS and in the BB. Woodberry et al. [1989] noted the Seychelles-Mauritus Ridge presented a propagation barrier for Rossby waves in the southern tropical IO. There is some evidence for a change in the propagation along 10°S near 60°E in Figure 3d.

The estimated phase speeds in the AS are 9 cm s⁻¹ at 15°N, 21 cm s⁻¹ at 10°N, and 32 cm s⁻¹ at 10°S. Along the equator the Kelvin propagation is estimated to be ~1.1 m s⁻¹.

5.2. Altimetry

The altimetric map on July 17, 1993 (Figure 4) often shows 6–9 cm anomalies near the Somali coast between the equator and 10°N. The contours are, however, not
Figure 3. The model simulated sea level anomalies as a function of longitude and time at the indicated latitudes. Units are centimeters.

closed, and the negative anomalies seen in the model near the coast are not picked up by this TP data set, probably because of the filtering of the altimeter data (see section 2). Recurring patterns of SLA during 1993–1996 include the −14 to −6 cm cyclonic circulation anomaly around the southern tip of India, the +4–6 cm and 10 cm anticyclonic eddies near 10°N, 60°E and 8°S, 100°E, respectively, and −12 to −4 cm SLA near the southwest corner of the domain. The wintertime case (Figure 5) shows even less obvious seasonality. There is a recurring anticyclonic circulation anomaly about the southern tip of India, but the strength of this feature ranges from 2 cm in 1994 to 14 cm in 1995. There are cyclonic recirculation regions near (4°N, 60°E) and (5°S, 60°E), also with a wide range of amplitudes. There is another SLA minimum east of India, around (13°N, 85°E).

A time-longitude plot shows that the SLA is, in fact, strongly seasonal, particularly in the AS and northern BB (Figure 6). At 10°N there is evidence of wave propagation in the eastern AS with an estimated propagation speed of 17 cm s^−1. At 10°S there are seasonally varying patterns of wave propagation across the entire AS. The estimated speed of propagation is ~25 cm s^−1. Along the equator there are eastward propagating Kelvin waves. These waves propagate with a speed of ~1.3 m s^−1.

5.3. CEOF1: Annual Cycle

The seasonal cycle is not easily distinguished in the “snapshot” portraits of SLA. The time-longitude diagrams more clearly reveal the seasonal cycle. Obscuration of the seasonal cycle in snapshot figures may be due to domination by interannual dynamics or by strong
subannual events. The latter is largely removed by examining the largest CEOF modes.

The first CEOF mode contains 46% of the variance for the model and 42% for TP. Temporal phases for both model and TP indicate an annual frequency (Figure 7). The phase rate of change is fairly constant with little high-frequency fluctuation. The correlation between temporal amplitudes of the model results and TP observations is 0.45. Both contain variability at interannual periods, suggesting the CEOF technique does not perfectly isolate the annual cycle. (An alternative explanation is that the annual cycle is not periodic.) The spatial amplitudes for the model as well as TP (Plates 1a and 1c) show relatively large signals in the southern and far western AS and the eastern and western BB. Relatively weak amplitudes are found in the central BB. Additionally, a local maximum of amplitude is found in observations as well as in simulations between 8–10°S and 80–100°E.

The orientations of the phase lines are mainly meridional in the eastern AS and BB (Plates 1b and 1d). In the western BB the TP phases are relatively uniform, indicating a lack of propagation. The signs of the gradients of the spatial and temporal phase functions generally imply westward propagation, except near the equator. In the equatorial waveguide eastward propagation consistent with the propagation of Kelvin waves is implied.

The SLA of CEOF1 along 15°N, 10°N, the equator, and 10°S for simulation and observation are presented in Figures 8 and 9. At 15°N both the model and TP show alternating bands of positive and negative SLA at the western parts of the AS as well as the BB, though the latter has a much weaker amplitude in the model results than for TP. This is consistent with the results in Figures 1 and 2, where there is a relatively weak annual cycle in the BB. The anemic simulation might result from the diffusion term in the model equations, which would suppress the large gradients required for large-amplitude signals. There are many other unrealistic attributes of the model, as discussed below. The exact cause and solution to this problem remain unclear.

In the eastern AS there is evidence of westward wave propagation in both simulation and observation. At 15°N the estimated phase speed is 10 cm s⁻¹ from simulation and 9 cm s⁻¹ from observation. At the equator...
and 10°S, there is similarity between model and observations, with evidence of westward phase propagation across the basin at 10°S in both model and TP. At the equator the speed of the model Kelvin waves is \( \sim 1.1 \) m s\(^{-1}\). The same estimate for TP is 1.3 m s\(^{-1}\). At 10°S the estimated phase speed is 21 cm s\(^{-1}\) in both the simulation and the observation. It therefore appears the first CEOFs are closely related and capture similar dynamical behavior. However, there are definite regions where this correlation does not hold, as will be shown in more detail in Section 5.5.

### 5.4. CEOF2: Semianual Cycle

The second CEOF contains 30% of the variance for the model and 24% for TP. The amplitudes of the temporal functions for the second CEOF for the model and observations (Figure 10) have a correlation of 0.68. Both amplitude functions achieve a strong maximum in 1994. This will be considered in more detail below. The temporal phase functions are similar. Both model and TP contain indications of semianual frequency during much of the period of study, but from early 1995 to early 1995 the dominant mode of variation is a roughly 1-year mode for both model as well as TP.

The spatial amplitude functions (Plates 2a and 2c) have some similarity. For instance, both possess relative maxima in the southwestern part of the domain and in the eastern equatorial region. The relative amplitude is generally greater in the TP data. Also, a small region of large amplitude is found just east of Saudi Arabia in both analyses. The spatial phase functions (Plates 2b and 2d) show dominant westward propagation in the latitude band from equator to 10°N and in the southern basin for both the model and observations.

### 5.5. Correlation Analysis

The temporal correlation between the model and TP (Plate 3) is high in some parts of the basin, but low, even negative, in others. The correlation reaches \( \sim 0.7 \) in the central AS and 0.5 in the southern AS. It is \( \sim 0.6 \) in the northern BB and 0.8 in the southern part of the basin. Areas of poor agreement (correlations as low as -0.4) include the eastern AS, central BB, and central equatorial region.
Plate 3. Map of correlation in time between TP sea level anomalies and model. (a) unfiltered. (b) filtered as CEOF1+2 mode. (c) as in (a), but using a 3.5 layered model (M. Luther, personal communication, 1999).
Secondary fluctuations can degrade the match between the model and TP results. Summing only the first two CEOF modes helps eliminate these secondary, higher-frequency fluctuations. After such filtering the correlation generally improves (Plate 3b). The correlation now reaches as high as 0.9 in the Arabian Sea and as high as 0.7 in the eastern BB. However, the regions of lowest correlation mentioned above are not improved by this procedure, suggesting a more fundamental problem with model accuracy.

The correlation in space between the model and observations (Figure 11a) is typically between 0.4 and 0.8, with a decline after 1994. After CEOF filtering (Figure 11b), the correlation generally improves, and the decline is essentially eliminated. However, two brief declines below 0.2 occur, one in late 1995 and the other in early 1996. The reason for these is unclear but is likely due to strong, high-frequency events not captured by the model.

6. Summary and Discussion

A reduced gravity model of the IO is driven with the FSU winds for 1993–1996. Model sea level anomalies relative to the 4-year mean surface height field are compared to altimetry measurements from TP at 10-day intervals. Both show a strong seasonal cycle. Model results south of 10°S are not examined because of distortions from the boundaries of the limited domain and damping of the winds below this latitude. It is likely that recirculation features that penetrate into this damped region, such as those on the southern bound-
The seasonal cycle in the northern IO is, to a large extent, controlled by the reversal of the western boundary currents and westward propagation of planetary waves. A common summer model circulation is composed of anticyclonic circulation anomalies in the western AS and off Somalia and a westward current between the equator and 10°S (Figure 1). A common winter SLA pattern includes cyclonic circulation anomalies in the AS and anticyclonic SLA filling the BB (Figure 2). There is significant summertime interannual variability in the BB and east of ~80°E south of the equator. Wintertime interannual variability appears strong throughout the IO, particularly south of the tip of India. Altimetry from TP appears to show greater interannual variability than the model. However, this may also be due to the relatively strong high-frequency variations in the TP data.

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this location radiate phase lines of all values. In the model they maintain a fairly zonal structure, but in the TP they spread almost evenly in all directions, indicating a rotating feature as shown by Mizoguchi et al. [1999].

The CEOF analysis of Perigaud and Delecluse [1993] yielded results similar to those presented here for the model output but very different results for the altimetric analysis. Their model CEOF1 spatial amplitude function also had a broad low in the BB that spread southward and eastward into the equatorial IO. The corresponding phase also showed high phase in the central BB and eastern AS, with lower phase in the equatorial region. However, their first temporal phase was not a steady cycle and appeared to be poorly related to the annual cycle. Their second CEOF was a relatively steady annual cycle. Apparently, the models generate similar dynamics, but their behavior in time is different, probably because of the different years being studied. This change from the 1985–1988 simulation of Perigaud and Delecluse [1993] to our 1993–1996 simulation is evidence of near-decadal variability in the IO.

The Geosat CEOF1 spatial function by Perigaud and Delecluse [1993] is unlike that found for TP. They found a relatively uniform amplitude north of 10°S, except for a maxima in the northern BB. In contrast, the TP results have a maximum arching across the southern AS, with several other maximum of lesser spatial extent. The spatial structure of CEOF1 appears to be broader for the TP (Plate 1c) than for Geosat [Perigaud and Delecluse, 1993, Figure 2). Again, this may be due to the different times examined or to the different instrument accuracies.

The semiannual signal is captured by CEOF2 (Figure 10). There is agreement between the model and TP temporal functions. Both model and TP show heightened amplitude in 1994, roughly coincident with a shift in the phase from a 6-month cycle to something slightly slower than a 12-month cycle, lasting into 1995. The reason for this remains uncertain. However, a change
in the wind stress is a likely candidate since the model is strictly wind-driven.

A vector EOF (VEOF) analysis [Legler, 1983] of the Florida State University pseudo stress in the model region yields a predominantly seasonal cycle, representing more than 77% of the variance, with the strongest loading in the northwestern part of the basin (Figure 12a). The semiannual cycle of the temporal amplitude (Figure 12b) and the annual cycle of the phase function (Figure 12c) represent the flip-flop of the summer and winter monsoon. The summer monsoon is generally greater in amplitude than the winter monsoon. The latter appears to have a longer duration, on the basis of the relative length of time the temporal phase function is in the winter state. The second VEOF mode accounts for 5.5% of the variance and is an unlikely source of the disturbance in the 6-month cycle.

The VEOF seasonal cycle is fairly steady over the years examined, although in 1994 the temporal amplitude function was ~25% stronger than the typical peak value. This increase may have overshadowed the semiannual cycle in the variance decomposition, resulting in the unusual behavior seen in Figure 10. The first VEOF spatial mode has significant loadings along the eastern land boundaries and across the basin between the equator and 10°S. These same areas have large magnitude in the spatial amplitude functions of CEOF2 (Plates 2a and 2c). Changes in the cycle of the local winds in

Figure 9. The TP sea level anomalies contained in the first CEOF as a function of longitude and time at the indicated latitudes. Units are centimeters.
Figure 10. Temporal function of the second CEOF for the model and TP observed sea level anomaly. (a) Model amplitude, (b) model phase, (c) TP amplitude, and (d) TP phase. Amplitudes are normalized and phases are in degrees.

This region could alter the normal oceanic circulation in these regions and influence the loading of the sea level response as captured by CEOF2.

There is high correlation between the model and TP in the AS and south of the equator (Plate 3a). The central BB does not contain high correlations, nor does the equatorial region. The reason for the lack of correlation in certain regions might be due to limitations of the model or the wind stress forcing. The model is a 1.5-layer reduced gravity model. As such, it is an "equivalent barotropic" model with a modified phase speed $c = \sqrt{g/H}$ representing only the first baroclinic mode (see section 3). Higher modes (more layers) may be required for a better simulation. Also, the wind stress is composed of monthly averages and, as such, does not contain higher-frequency components. This limitation may be important in regions of fast wave activity such as the equatorial region. Another source of disagreement between the two data sets might be the preprocessing of the TP data mentioned in section 2.

The temporal correlation between the model and TP SLA generally increases after filtering with CEOF1+2, but the overall spatial pattern of correlation remains unchanged (Plate 3b). Perigaud and Delecluse [1993] examined temporal correlations of model SLA and Geosat SLA without filtering. The locations of correlation extrema [Perigaud and Delecluse, 1993, Fig. 3f] have little similarity to those in Plate 3a. This may be due to their use of Geosat altimetric data, which are less accurate than the TP used here. It may also reflect changes in the circulation of the IO.

Each region of low correlation in Plate 3b generally corresponds to a phase nexus or a strong phase gradient in the largest model spatial phase function (Plate 1b). Near these regions small displacements of model features result in large changes in their structure. Discrepancies between model and TP data, such as those involving the phase of propagating features or the position of currents and eddies, would tend to produce low correlations in these areas.

A phase nexus is present off the southeastern tip of India/Sri Lanka in both model and TP (Plates 1b and 1d), though the TP nexus is weaker. This nexus feeds a strong phase gradient into the western BB, which is
Figure 11. Correlation between spatial maps of SLA from model and TP. (a) Unfiltered, (b) filtered as CEOF1+2 mode.

Figure 12. The first mode (77% of the variance) of the vector EOF analysis of the FSU wind stress over the model basin. (a) Spatial function, (b) amplitude of the temporal function, (c) phase of the temporal function. Vertical lines indicate transition between summer and winter monsoon states.
related to the annual cycle of Rossby waves. As the first half of the Rossby wave (e.g., positive SLA) impacts the east coast of India, the opposite phase signal is generated in the eastern BB. The opposite phase wraps around the previous Rossby signal as a coastally trapped Kelvin wave to the north and west and as a Rossby wave to the east. The first phase is then annihilated by the progression of the second, opposite phase. Examples of this can be seen in Figures 13 and 14. Similar model dynamics are found in the far western AS.

The model phase nexus near 3°S, 75°W is unmatched in the TP data and is related to a region of wave mean flow interaction. When the eastward traveling annual Kelvin wave impacts the eastern boundary, its signal is partially reflected as a westward propagating Rossby wave. The southern component of this wave is advected to the south by the South Equatorial Counter Current, yielding a net southwestward propagation. These combined dynamics produce the nexus at this location. This nexus is not present in the TP analysis (Plate 1d), suggesting this region is not well represented by the model.

The model phase nexus near 7°S, 55°W is matched in the TP data with comparable strength (each cycles from −π to π). This area is one of relatively high correlation since the model and TP structure are similar.

The negative correlations in Plates 3a and 3b are partly attributable to the lack of higher baroclinic modes in the 1.5-layer model. A similar correlation analysis between SLA from TP and a 3.5-layer model (M. Luther, personal communication, 1999) improves these regions (Plate 3c), raising the correlation from ~−0.3 to ~0.2, though the regions of low-correlation remain. The similarity of the low correlation patterns in Plate 3a and Plate 3c suggests some ocean dynamics are not represented accurately even with more degrees of freedom in the numerical model. It is not clear whether the inclusion of additional layers will further improve the model or whether new processes must be introduced. The complete presentation of the 3.5 layer model is left for a future manuscript.

Physical processes not in our model include fresh water fluxes from evaporation and precipitation (E/P) and

![Figure 13](image-url)
in 1986 to a local minimum of ~0.1 in 1988, though they did not discuss this result. The degradation is largely eliminated after CEOF1+2 filtering, indicating that semiannual to interannual variability of the IO is represented more accurately by the model than higher-frequency variations. There remains, however, short-term decreases of the spatial correlation below 0.2 (Figure 11b). These may be due to brief events in the IO not captured by the model. A likely source of this is the lack of high-frequency components in the wind stress forcing.

Further work is required on the interannual variations of the IO. This might include extending the model to include more vertical layers and fresh water fluxes. Vertical exchange has also been found to influence IO model circulation [McCready et al., 1993]. Finally, the use of monthly winds was frequently indicated as a possible source of model error, particularly below the monthly timescale. Using a near-daily wind product would likely improve the simulation accuracy.

Figure 14. The model sea level anomalies contained in the first CEOF on the indicated dates. Units are centimeters.
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