Seasonal cycle of sea level and currents along the coast of India

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A framework involving long baroclinic waves, which treats the north Indian Ocean as a single dynamical entity, has been remarkably successful in simulating the low-frequency, large-scale circulation in the ocean. It is, however, unable to simulate the observed seasonal cycle of sea level along the Indian coast. The reason appears to be the absence of salinity variation in these models. The large inflow of freshwater into the seas around India forces large changes in salinity, and hence, in coastal sea level.

THE surface of the sea deforms continuously. Its level, measured relative to an arbitrary datum, is called sea level, which changes with time, and is the most obvious indicator of change in oceans. Changes in sea level are greater in shallow waters in the vicinity of a coast than in the open sea, and since a large fraction of the human population resides in coastal areas, variations in sea level have aroused interest for a long time. A reasonably accurate prediction of sea level was necessary for safe navigation of boats and ships in harbours. This, and the relative ease of measuring sea level compared to measuring, say, temperature or currents, has led to it being one of the best-documented oceanic variables. Hourly measurements of sea level are available at several places over the globe, some of the records stretching back to the last century.

A tide gauge measures sea level relative to land; an increase in sea level may either be due to an absolute rise in sea level, or due to the sinking of land. The largest changes evident in a typical sea-level record (Figure 1) are those due to astronomical tides. These regular, periodic patterns are modified by the effects of weather; exchange of energy between the atmosphere and the ocean occurs at all space and time scales, from the generation of short-period wind-waves to the slow transfer of energy, over a century, from the tropics to the poles by the global ocean circulation. The sea-level record also includes the effect of secular changes forced by changes in the volume of ice locked in major ice caps and glaciers.

In the north Indian Ocean, which we define as extending from 5°S to the Asian land mass in the north, the major signals in the coastal tide-gauge records are the semi-diurnal and diurnal tides and the seasonal cycle forced by the monsoon winds, the most significant aspect of the weather of the region. Since these are the most prominent signals, it is necessary to understand them

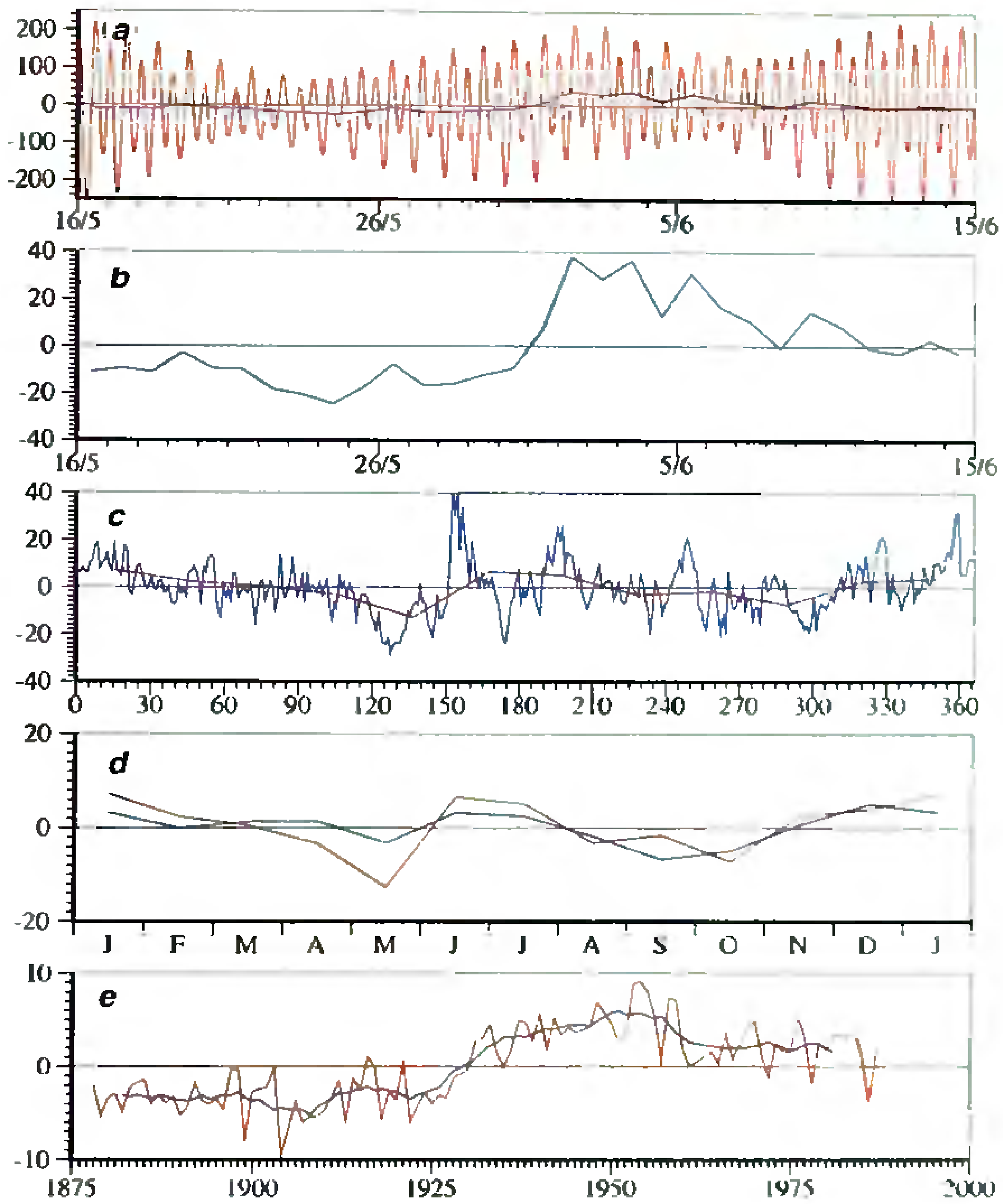

Figure 1. Sea-level variability at Mumbai, which has the only more-than-century-long sea-level record in the Indian Ocean. Hourly, daily, monthly, and annual sea level (cm) are plotted as a function of time to reveal the variability over a range of frequencies. ***a***, Hourly sea level (red) and daily sea level (blue) during 16 May–15 June, 1976. The residual signal (daily sea level) after removing the tide is much smaller than the tidal oscillation. The detided signal is computed using a low-pass filter; ***b***, Daily sea level as in ***a***, but with an increase in the scale of the ordinate. Note the storm surge of about 70 cm in the beginning of June. It is also seen in ***a*** and perturbs the tidal cycle; ***c***, Daily sea level (blue) and monthly sea level (red) for 1976. Monthly sea level is computed by averaging the hourly sea level over a month; ***d***, Monthly sea level for 1976 (red) as in ***c***, but with the ordinate changed; superimposed on this is the climatological monthly sea level (blue); ***e***, The annual sea level over the length of the data record, 1878–1988 (red), showing the interannual variability in sea level at Mumbai. Filtering annual sea level with a 10-year boxcar filter (blue) reveals interdecadal variations.

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before we begin to study other frequencies evident in sea-level records – storm surges, intraseasonal oscillations, interannual and interdecadal changes, etc.

All these frequencies are seen in the sea-level record at Mumbai (Figure 1). The change in sea level due to tides is much greater than those related to weather (Figure 1 *a*); the latter is obtained by filtering the hourly sea-level record with a low-pass filter. Even the surges due to storms (Figure 1 *b*) are almost an order of magnitude smaller than the tide at Mumbai. Even at other locations along the coast of India (Figure 2), the tides are larger than the weather-forced changes.

Tidal and non-tidal sea-level oscillations are usually studied separately because of the vastly different ways in which they are forced. In this article, we ignore the tides, focusing instead on the seasonal cycle of sea level at the Indian coast. Though this is much smaller than the tides, the related movement of the thermocline is important (see Boxes 1 and 2).

Coastal sea level and large-scale coastal circulation

To obtain the seasonal cycle of sea level, it is necessary to filter the tides and other high-frequency oscillations out of the sea-level data. This is most easily achieved by averaging the hourly sea level over a month to obtain 'monthly sea level'. Monthly sea-level data are compiled by the Survey of India from hourly measurements and are archived at the Permanent Service for Mean Sea Level (PSMSL), UK, along with a history of the datum with respect to which the sea level was measured. A climatology is obtained from the monthly sea level by averaging the data for a particular month, say January, over all the years of the record. The climatological seasonal cycle is then obtained by removing the annual mean from the climatology of monthly sea level. (See Figure 1 *d* for a comparison between the climatological seasonal cycle and monthly sea level for 1976 at Mumbai.)

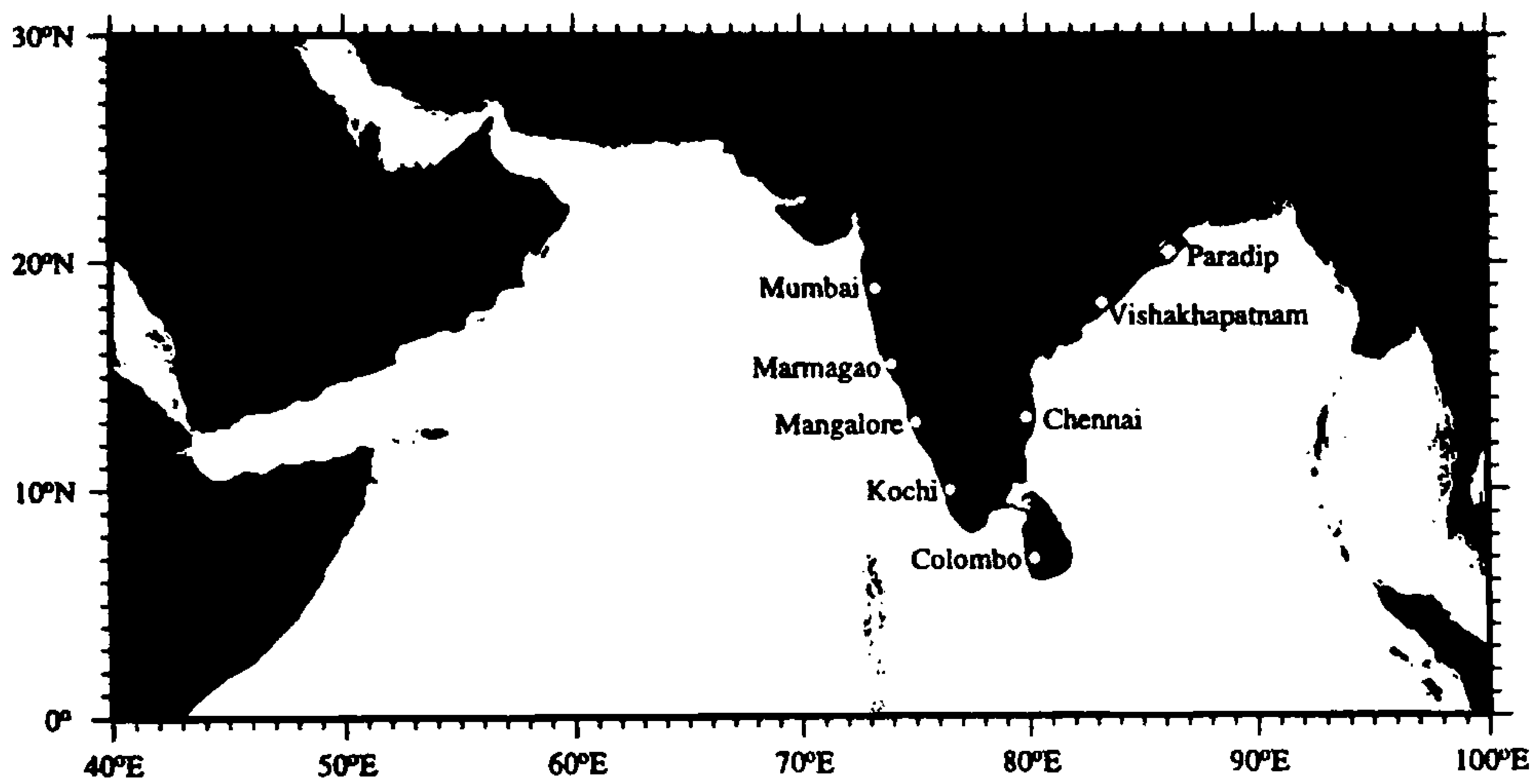


Figure 2. Tide gauge stations along the coast of the Indian subcontinent.

Box 1. Geostrophy

The rotation of the earth on its axis makes it a non-inertial frame of reference for measurements made relative to it, giving rise to a pseudo-force called the Coriolis force. This, along with the large 'aspect ratio' (the ratio of the horizontal scale to the vertical scale) of the fluid stretched out on its spherical surface, has profound implications for the dynamics of the ocean.

For large-scale motions, the force due to the pressure gradient tends to balance the Coriolis force – this is called *geostrophy*. A consequence is that currents tend to flow along isobars, rather than across them; in the northern hemisphere, a geostrophic current flows with high pressure on its right. Another consequence of the Coriolis effect is that the stress exerted by the wind on the surface of the sea does not set up a downwind current; the current at the surface flows at an angle to the wind, and the transport in the surface boundary layer, called the Ekman layer, is to the right of the wind in the northern hemisphere.

Climatological seasonal cycle of coastal sea level

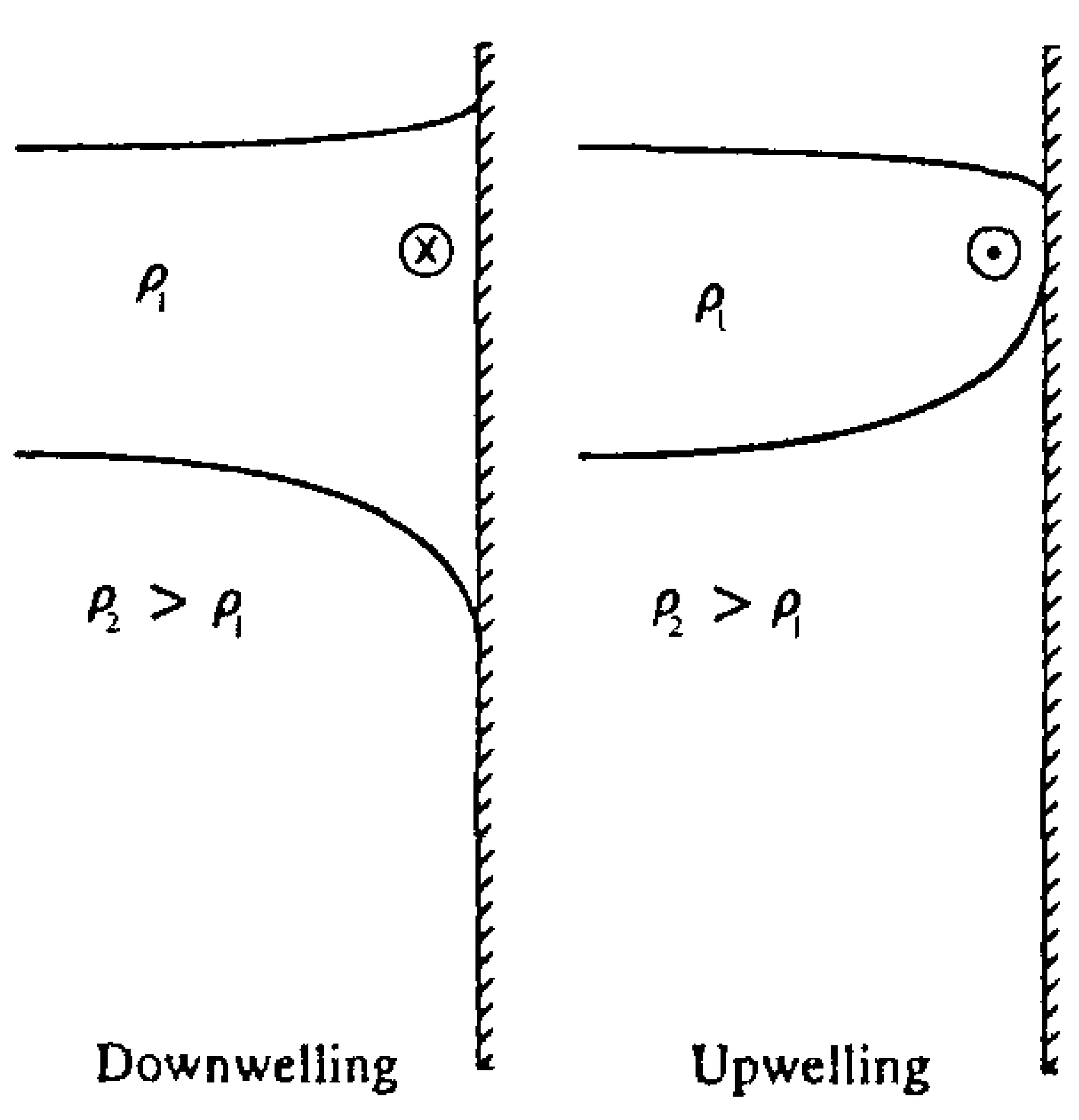
The seasonal cycle of sea level at a coast is affected by three factors: astronomic tides, atmospheric pressure, and 'steric oscillations', which are due to changes in specific volume[1]. The steric sea level is low (high) when the water is cold (warm), or has high (low) salinity. Changes in steric level are linked to surface fluxes and horizontal and vertical advection of heat and salt. Even in the absence of horizontal advection and of fluxes across the boundaries of the ocean, coastal currents forced by winds cause changes in the local mass field by forcing coastal upwelling or downwelling, thereby influencing the steric level (see Box 2). We shall distinguish between this and the changes in steric level caused by horizontal advection of heat and salt and by the exchange of heat or salt across the boundaries of the ocean; the former shall be considered the effect of wind-forced coastal currents, the latter that of thermohaline forcing. Therefore, we have four factors forcing changes in monthly sea level at a coast: astronomic tides, atmospheric pressure, wind-driven coastal currents, and thermohaline effects. Of these, the astronomic tides contribute less than 1 cm, which is an order of magnitude less than the meteorological effects; hence, we ignore their effect.

At the low frequencies of interest, the effect of atmospheric-pressure variations in the tropics can be 'corrected' by using the Inverse Barometer (IB) approximation[2]; the corrected sea level is lower (higher) than the measured sea level when the local atmospheric pressure is lower (higher) than the mean atmospheric pressure over the world's oceans. Since this mean is a function of time because of a shift in air mass towards Siberia in winter[1], the IB approximation involves an adjustment of sea level to both local and nonlocal pressure fluctuations; it is isostatic, i.e. it does not lead to pressure changes in the deeper ocean[2].

Along the coast of India, seasonal sea-level changes due to atmospheric pressure variations vary from about 3 cm at Colombo to 13 cm at Paradip. The corrected sea level is lower (higher) than the measured sea level during the south-west (north-east) monsoon, when the local atmospheric pressure is low (high). The seasonal changes

Box 2. The two-layer ocean model

In the oceans, temperature usually is vertically well mixed in a shallow upper layer about 100 m deep; this is called the mixed layer. Below this, it drops rapidly over a depth of about 500–1000 m, this region being called the thermocline. The change in density with depth in the ocean is much smaller than the average density of sea water. If we consider the ocean to be composed of two layers, the upper layer about 100 m deep overlying a much deeper lower layer, the ratio of the difference in density between the layers to the average density is about 0.003. This results in a much larger oscillation of the thermocline for a small change in sea level; for example, a 10 cm change in sea level, as seen even in the interannual variability of annual sea level at Mumbai (Figure 1 *e*), is associated with a 30 m vertical movement of the thermocline, a rise (drop) in sea level being associated with a fall (rise) of the thermocline. This is a large, observable change of major consequence for ocean dynamics. Hence, the 'noise', which the weather-related changes appear to be in Figure 1, is as important as the 'signal' due to the tides.



Such a change in the ocean may be forced by winds too, as shown in the schematic here which is of a section normal to the coast. It shows a two-layer model ocean, as in the $1\frac{1}{2}$-layer reduced-gravity model used in the simulations described in this article; the model ocean has a flat bottom and the coasts are treated as vertical walls. The lower layer of the model ocean is denser and much deeper than the shallow upper layer. The interface between the two layers is the model pycnocline, the density-equivalent of the thermocline.

As a consequence of the Coriolis effect, wind blowing with the coast on its right (left) forces Ekman flow into (out of) the coast, leading to a coastward rise (fall) in sea level and to a coastward fall (rise) in the pycnocline. This is called coastal downwelling (upwelling). It sets up a cross-shore pressure gradient that drives a geostrophic alongshore current (see Box 1). In the picture, the current shown by 'X' ('·') within a circle is into (out of) the plane of the paper. Thus, when there is downwelling (upwelling) at the coast, the current flows with the coast on its right (left), sea level rises (falls) at the coast, and the pycnocline slopes down (up) towards the coast. The change in sea level is given by $\Delta\eta = \frac{\rho_2 - \rho_1}{\rho_2} \Delta H$, where H is the thickness of the upper layer; $-\Delta H$ is the vertical movement of the interface.

observed in the corrected sea level are due to the two remaining factors: wind-driven coastal currents and thermohaline effects. The corrected sea level at different stations along the coast reveals a close relationship (Figure 3).

1. The range of the seasonal cycle is greater on the east coast (45 cm) than on the west coast (30 cm). The drop in the range from Chennai to Kochi is due to a decrease in the maximum; the minimum does not change (much) from the east to the west coast.
2. Along the east coast there are two maxima, a minor peak during the south-west monsoon and a major peak during November; the winter peak is broad (in time) at Paradip and narrows equatorward along the coast. The minimum occurs during April.
3. Along the west coast, sea level peaks during December, the peak being broader than along the east coast, making the seasonal cycle more symmetric; there is also a minor secondary peak at Kochi and Mumbai during the south-west monsoon, when the minimum occurs along the west coast.

The causes of the observed variability, seasonal or interannual, are not clear. Earlier studies (see Shankar[3] for a survey of the literature) have noted a similarity between the variation in local rainfall, and hence runoff, and that in sea level. Most crucial, however, is the striking resemblance of the seasonal cycle of coastal currents to that of the sea level along the Indian coast. This, and the significant correlation between interannual variations in sea level along the coast[4], suggests that the low-frequency variability of sea level along the Indian coast is linked to the large-scale circulation in the north Indian Ocean. This is not surprising, given that a similar relationship has been noted elsewhere too[1].

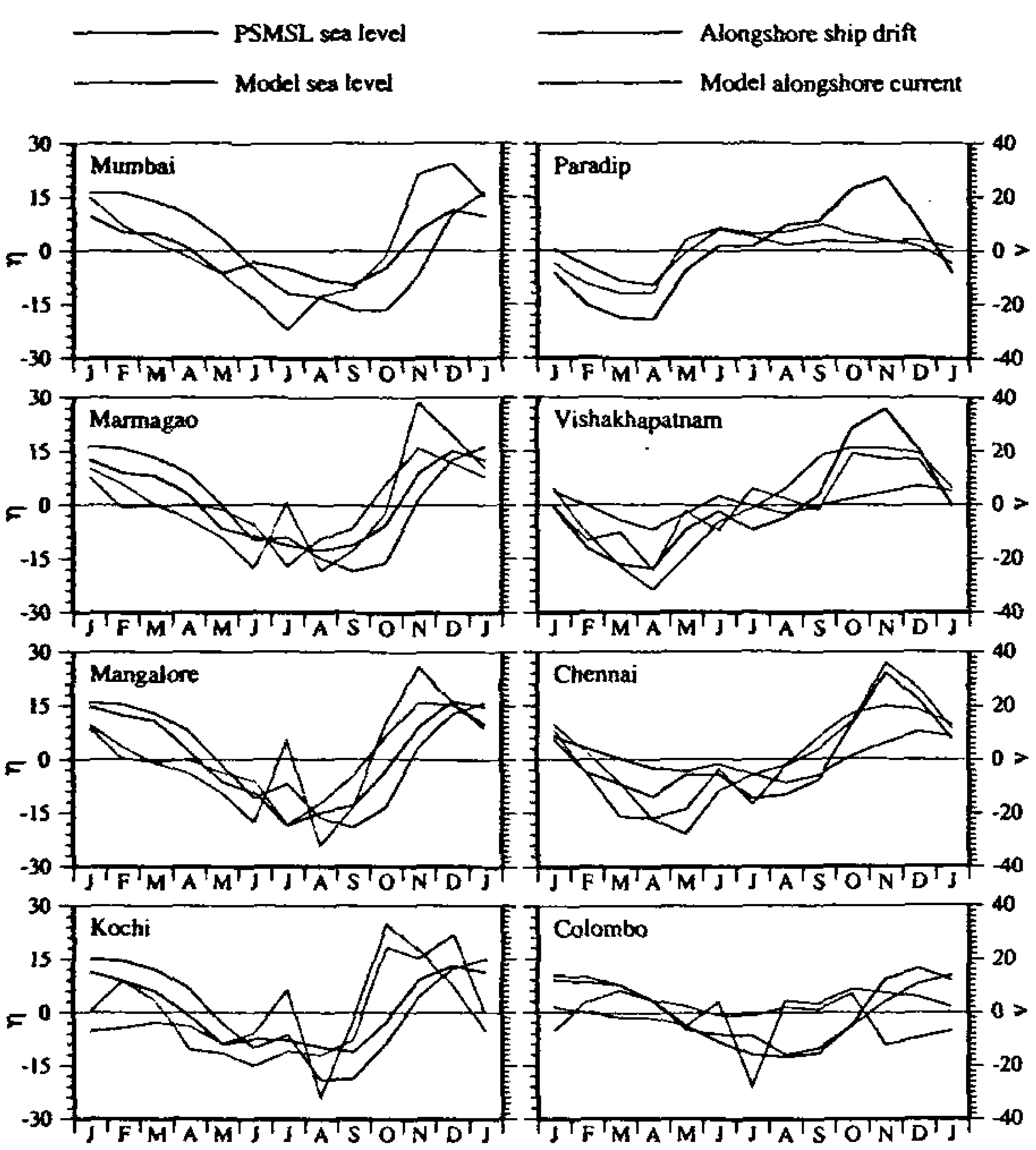


Figure 3. Climatological seasonal cycle of sea level (cm) and alongshore currents (cm s^{-1}) along the coast of the Indian subcontinent. The figure shows the corrected sea level (sea level corrected for the effect of atmospheric pressure using climatological sea-level pressure data from the Comprehensive Ocean–Atmosphere Data Set), alongshore currents computed from ship drifts[12], and model sea level and alongshore currents. All variables plotted are anomalies; the annual mean has been removed.

Hydrography of the Indian coastal region

The first major hydrographic survey of the north Indian Ocean was made during the International Indian Ocean Expedition (IIOE)[5]. Though a lot of hydrographic surveys had been carried out in the north Indian Ocean since then, a coherent picture of the large-scale coastal circulation around India emerged only during the 1990s. We use the data collected during six hydrographic surveys during 1987–1994 in the Exclusive Economic Zone (EEZ) of India, in conjunction with ship drifts, to describe the large-scale circulation off the Indian coast. In the absence of direct current-meter measurements, these provide the only means of estimating currents off the Indian coast (see Box 3).

Three cruises surveyed the coastal waters off the Indian west coast during June–August 1987 (ref. 6), December 1987–January 1988 (ref. 7), and March–April 1994, and three cruises surveyed the coastal waters off the Indian east coast during July–August 1989 (ref. 8), March–April 1991 (ref. 9), and November–December 1991 (ref. 10). These cruises followed near-identical tracks, measurements being made along 'legs' perpendicular to the coast or perpendicular to the shelf-break, giving a coherent picture of the large-scale circulation in the coastal waters of India. Figure 4 shows the transports computed from dynamic topographies w.r.t. 1000 m based on the above six cruises. Interannual variability is expected to cause changes in transports of these current systems, but they still provide a rough picture of the climatological seasonal cycle of coastal currents around India.

The coastal currents around India change direction with season[3,6–10]. From November to January, the current off the Indian east coast, the East India Coastal Current (EICC), is equatorward all along the coast. It bends around Sri Lanka to flow along the Indian west coast as a poleward West India Coastal Current (WICC). Off south-west India, the poleward flow is along the western flank of an anticyclonic high in sea level[11] that forms in the region; a geostrophic current completes the circulation around the high, forcing an equatorward WICC off south-west India. The EICC reverses in February, and flows poleward during March–May, forming the western boundary current of a basin-wide anticyclonic gyre (sea level is raised in the interior of the Bay of Bengal). The anticyclonic high off south-west India persists through March–April, weakening thereafter and giving way to a cyclonic low during the south-west monsoon (June–September); during this period, the WICC flows equatorward along much of the Indian west coast. The EICC is weak during the south-west monsoon; it is poleward in the south, but is equatorward in the north.

Since the dynamic heights do not represent a climatology, we refrain from comparing them with tide-gauge sea level; this has been done elsewhere, and the agreement found to be good[1]. A better comparison is afforded by ship drifts because the far larger number of observations, spread over several years, makes it possible to define a climatology.

Coastal sea level, coastal currents, and local alongshore winds

We use a climatology of ship drifts[12] to compute the observed alongshore coastal current. The ship drifts are averaged over $2° \times 2°$ bins and the coastal currents computed from them represent the large-scale coastal circulation. Since these coastal currents are essentially geostrophic, we expect good agreement between the observed sea level and alongshore current. The seasonal cycle of the alongshore coastal current computed from ship drifts is shown in Figure 3. The variations in coastal current match those in coastal sea level, except at Colombo; the only major discrepancy is that the current peaks a month before the sea level does on the west coast.

Since the local alongshore coastal currents match tide-gauge sea level very well, it is possible that the local winds, which must play a role in forcing these currents, play a significant role in forcing the seasonal cycle of sea level. Figure 5 shows the monthly anomaly of local alongshore wind stress[13]. The annual mean of the wind stress at each location is removed to obtain the monthly anomalies; this is done because the sea-level and ship-drift data plotted are also anomalies. The mean wind-stress field, which is not negligible, can set up a steady alongshore sea-level gradient, but it cannot force time-dependent changes in sea level along the coast.

Though there are some similarities in the seasonal cycle of local alongshore winds and sea level, there are discrepancies too. There is a mismatch between the strength of local winds and the range of the sea-level cycle; the range of alongshore wind-stress on the east coast is about 5–6 times that on the west coast, but the range of sea level on the east coast is only about twice that on the west coast. At Colombo, the wind-stress cycle resembles that along the east coast of India, but the sea-level cycle resembles that along the west coast. This similarity in the sea-level cycle at Colombo and the stations on the Indian west coast suggests a common forcing mechanism, and points to the difficulty in explaining the seasonal cycle of sea level and currents along the Indian coast by invoking purely local causes.

Basin-scale dynamics of the north Indian ocean

The need to explain the seasonal cycle of circulation off the Indian coast led, during the 1990s, to the development of a framework involving long baroclinic waves, which had been invoked earlier to explain the sea-level variability along the Pacific coast of the Americas. Anomalies of monthly sea level along the eastern boundary of

Box 3. Measuring currents in the ocean

Direct measurement of currents is difficult in the oceans. In the absence of direct measurements, two methods have been used extensively to map the large-scale circulation in the world oceans.

The first method exploits geostrophy (see Box 1) to construct *dynamic topographies* of the sea surface relative to some level. Oceanographic surveys are used to collect vertical profiles of temperature and salinity (hydrography) in a region. Since the velocity of currents near the surface is much greater than that in the deep ocean, it is possible to use the geostrophic balance to compute the average current, at *a given depth relative* to some other depth, between two locations. When this is done for the surface in a region like the north Indian Ocean, we obtain the dynamic topography of the sea surface, which gives an idea of the transport w.r.t. some depth. Oceanographers measure transport in *sverdrups* ($1\ \mathrm{Sv} = 10^6\ \mathrm{m}^3\ \mathrm{s}^{-1}$) and a commonly used reference depth, also called the *level of no motion*, is 1000 m.

The second method makes use of ships. The deviation of a ship from its planned trajectory (due to winds and currents) is used to compute surface currents; these are called *ship drifts*.

Each of these two methods has its own drawbacks. Dynamic topographies constructed from hydrographic data give the geostrophic transport relative to an assumed level of no motion, typically 1000 m, and are therefore unreliable in the vicinity of a coast; also, they cannot account for the ageostrophic transport. Ship drifts, on the other hand, are influenced by the wind and are usually too noisy to yield a clear, coherent picture of the basin-scale circulation.

the Pacific are visually coherent and statistically well-correlated from Valparaiso in Chile to the Canadian border, these anomalies propagating poleward in both hemispheres. These anomalies are linked to El-Niño, whose domain extends across the equatorial Pacific Ocean; the link is provided by equatorial Kelvin and Rossby waves and coastal Kelvin waves (see Box 4). These waves propagate long distances, carrying information across the equatorial Pacific, and forcing changes in locations far removed from the winds that force these changes in the ocean. A dramatic example of this is the large El-Niño of 1982–1983, whose effect could be traced across the subtropical Pacific a decade later, forcing changes in the Kuro-Shio extension off Japan[14].

The success of this theory in the Pacific Ocean appears to have prompted its application in the Indian Ocean; this was synchronous with the hydrographic surveys described earlier, and the theory survived because its predictions matched the observations. Since these waves communicate changes in sea level or currents forced by winds in one region to remote locations, they merge the equatorial Indian Ocean, the Bay of Bengal and the Arabian Sea into a single basin, the north Indian Ocean, which must be modelled as a whole even to simulate the circulation in its parts. The seasonally-reversing monsoon winds, which dominate the wind-stress signal in the north Indian Ocean, provide a mechanism to generate these waves. This, along with the small size of the basin and its essentially tropical

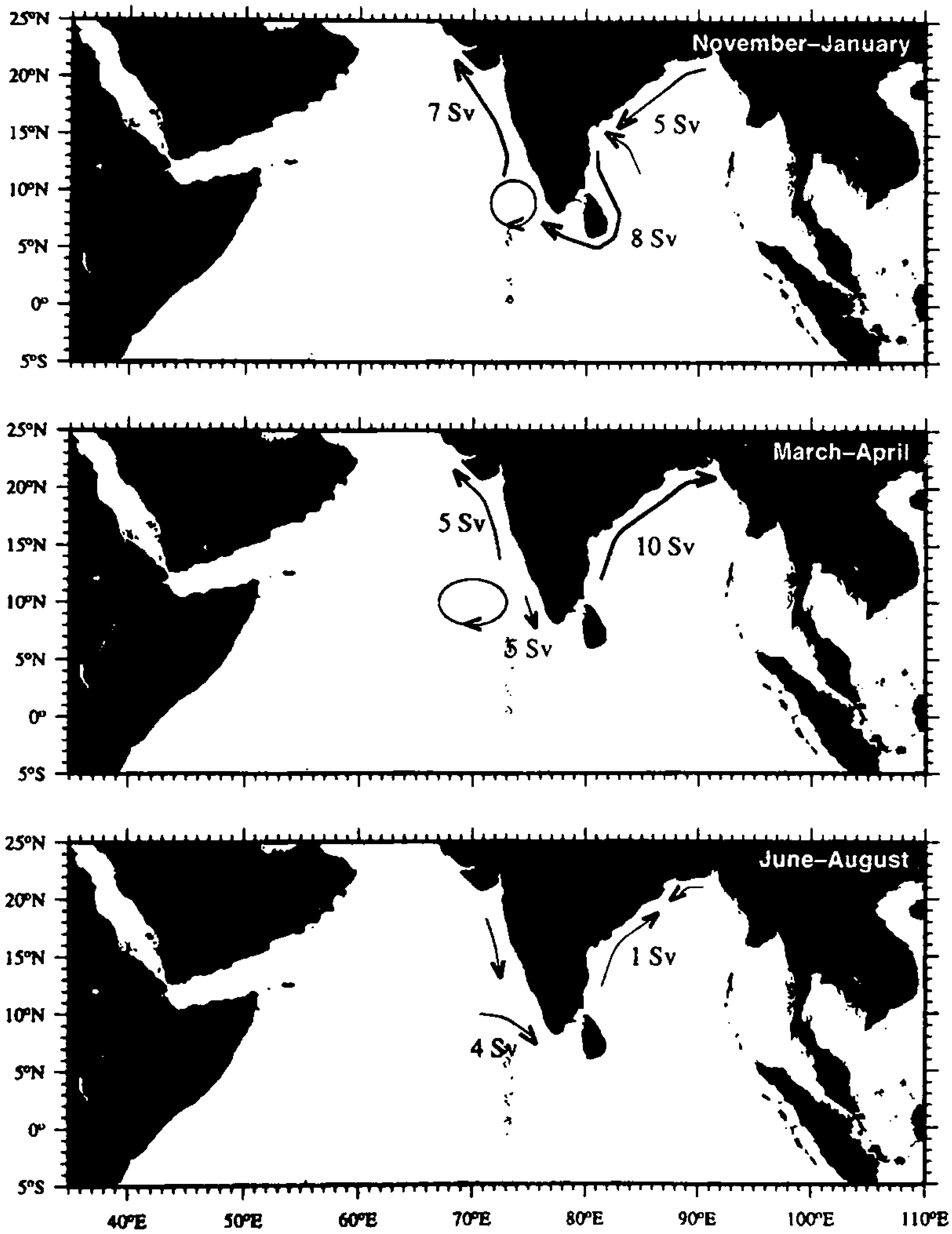


Figure 4. Schematic of the current systems along the Indian coast. The number against each current indicates the transport (Sv) associated with it in the top 1000 m of the water column (1 Sv = 10^6 m^3 s^{-1}). These estimates are based on the six cruises mentioned in the text, and provide an idea of the large-scale coastal circulation off India; interannual variability will change these estimates, and in the case of the weak currents, it may also force a change in direction.

nature, which means the Rossby waves have a high phase speed, implies that disturbances can cross the basin within a year. This is best seen in sea-level data from satellite altimetry, which provide a coverage of the basin that cannot be matched by the network of tide gauges. The TOPEX/POSEIDON sea-level anomalies[15] show how sea-level changes are communicated across the basin by Rossby waves, which propagate westward, and equatorial Kelvin waves, which propagate eastward along the equator. The coastal Kelvin waves, however, have an offshore

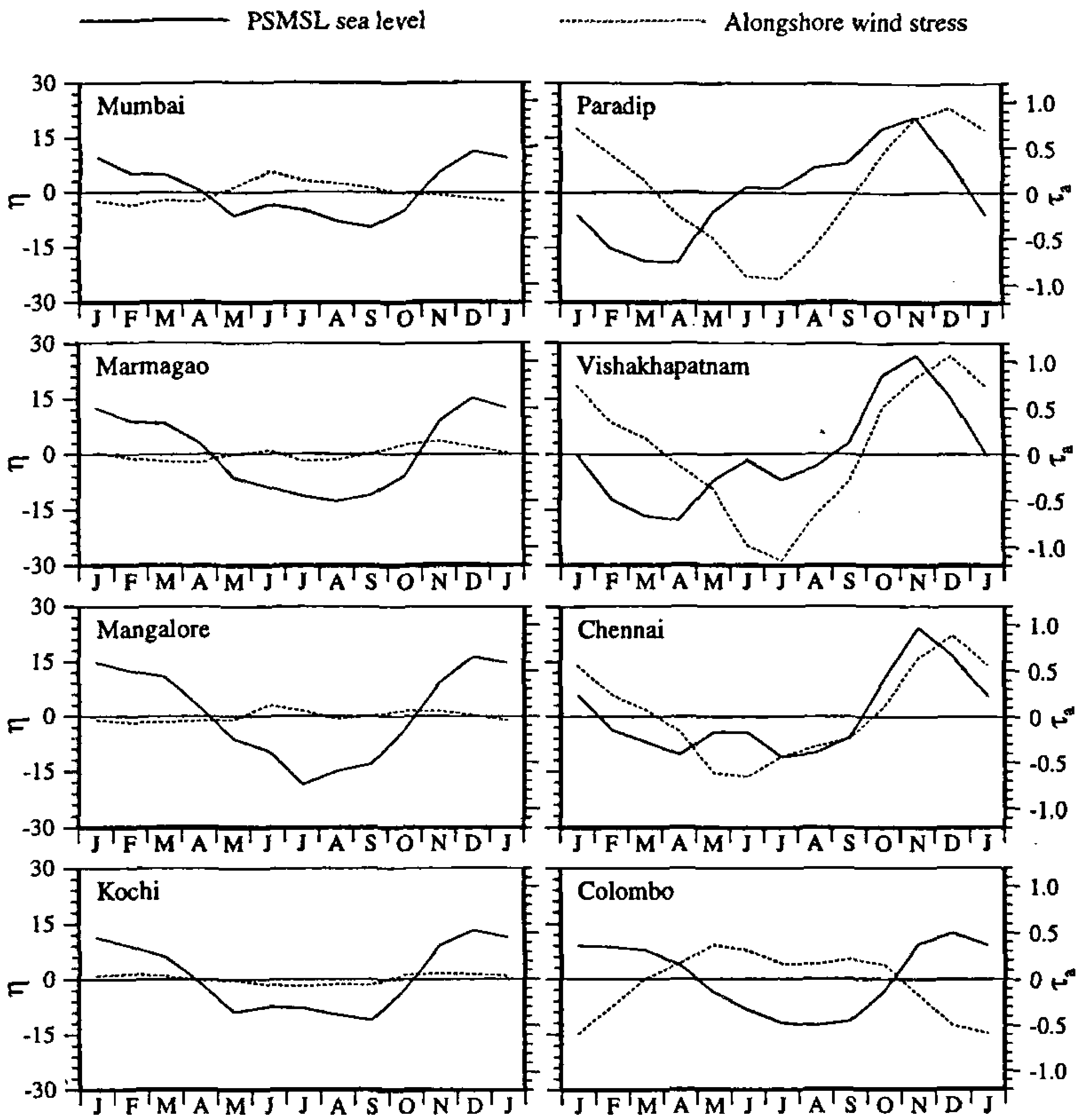


Figure 5. Climatological seasonal cycle of sea level (cm) and alongshore winds (dyne cm^{-2}) along the coast of the Indian subcontinent. The figure shows the corrected sea level (sea level corrected for the effect of atmospheric pressure using climatological sea-level pressure data from the Comprehensive Ocean–Atmosphere Data Set) and alongshore wind stress computed from the wind-stress climatology of Hellerman and Rosenstein[13]. All variables plotted are anomalies; the annual mean has been removed.

Box 4. Kelvin and Rossby waves

Theoretical studies of the dynamics of the north Indian Ocean during the last decade have used free and forced long baroclinic waves to describe the circulation in the basin. (Baroclinic means that the current varies with depth; the vertical mode in a $1\frac{1}{2}$ -layer reduced-gravity model is the first baroclinic mode.) For low-frequency dynamics, the most important of these waves are the equatorially-trapped Kelvin and Rossby waves and coastally-trapped Kelvin waves.

Kelvin waves propagate along a boundary. The vanishing Coriolis effect at the equator makes it a dynamical boundary, permitting the existence of Kelvin waves that propagate along the equator. These equatorially-trapped waves propagate eastward and have a meridional *e*-folding scale of about 200 km.

Equatorially-trapped Rossby waves propagate westward. At the semi-annual and annual frequencies, these waves exist throughout the north Indian Ocean.

Coastally-trapped Kelvin waves are like their equatorially-trapped counterparts; they propagate with the coast on their right (in the northern hemisphere). Their offshore *e*-folding scale varies with latitude, but is of the order of 100 km.

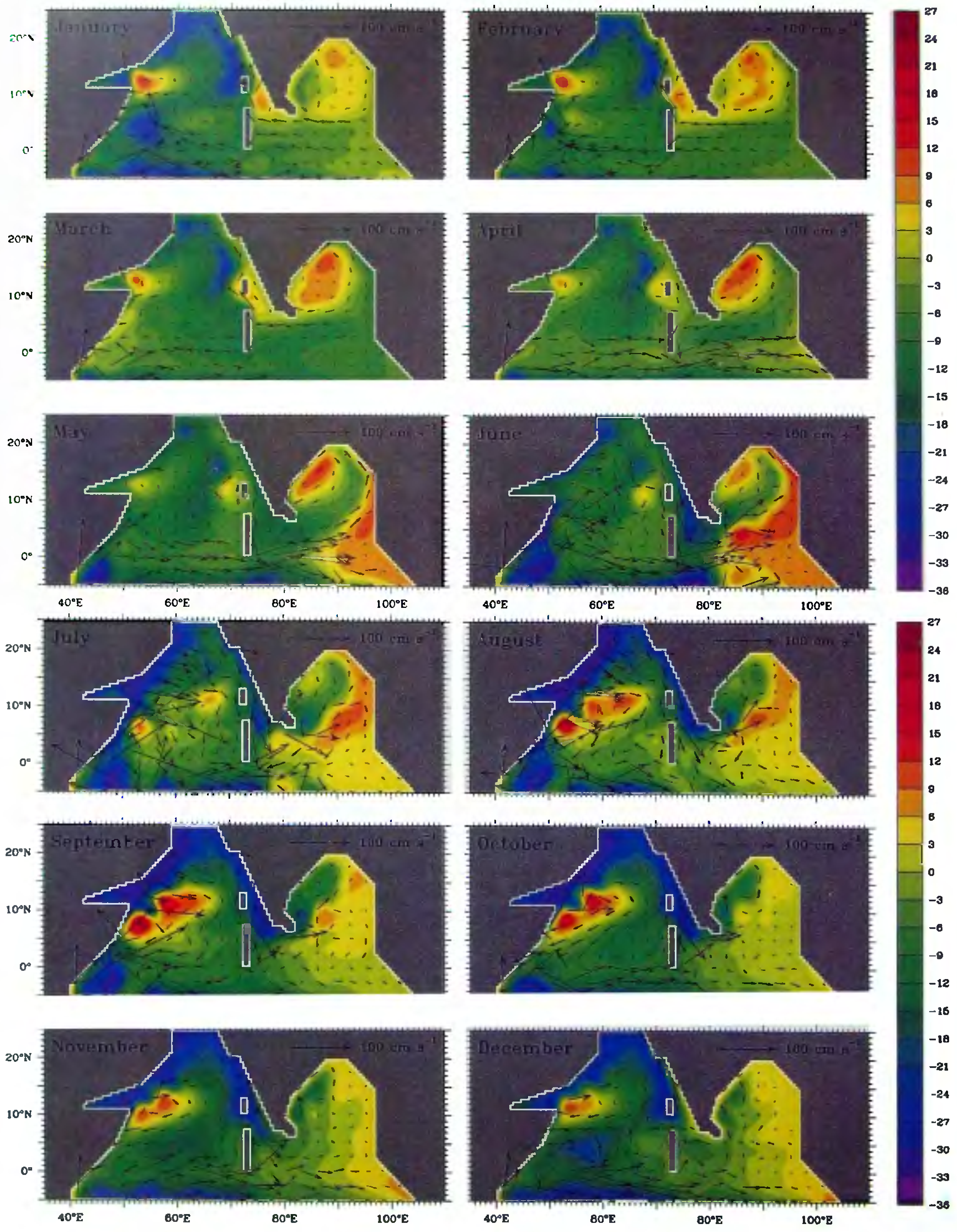


Figure 6. Annual cycle of surface circulation simulated by a $1\frac{1}{2}$-layer dynamical reduced-gravity model of the Indian Ocean. Plots of sea level deviation (cm) from the initial state and velocity (cm s^{-1}) are shown for the tenth year of simulation. The model ocean was forced by the climatology of Hellerman and Rosenstein[13].

e-folding scale of the order of 100 km, and they are not resolved by the gridded altimeter data.

Numerical models developed in the 1990s have been fairly successful in simulating the basic features of the observed circulation; their success implies that changes in currents and sea level off the Indian coast must, at least in part, be remotely forced. Most of the model studies are based on reduced-gravity models, which were earlier used successfully in studying the El-Niño, but there are some that use General Circulation Models (GCMs) (see Shankar[3] for a list of references).

Figure 6 shows the result of a simulation using a $1\frac{1}{2}$-layer dynamical reduced-gravity model for the Indian Ocean. The model ocean consists of two layers, the lower layer being denser and much deeper than the upper layer. The assumption that the pressure gradient vanishes in the lower layer eliminates the fast-moving surface gravity waves; this assumption is analogous to that of a level of no motion made in computing dynamic topographies from hydrographic data. Changes in the model sea surface are forced by the slower internal wave (or, in the context of the model, interfacial wave), which represents the first baroclinic mode. Shankar and Shetye[16] used this model to study the annual cycle of circulation off south-west India. This model, like the others mentioned above, is able to simulate the patterns of sea-level variability seen in altimeter data and in the surface circulation as revealed by ship drifts and hydrography. Its success is not restricted to the open sea; the model alongshore coastal currents are in excellent agreement with ship drifts (Figure 3).

The model sea level along the Indian coast, however, is at variance with that observed (Figure 3). First, the model fails to reproduce the winter peak along the east coast. Second, as a consequence of this failure, the model sea level has a higher range along the west coast than along the east coast. This failure is common to all the models listed in Shankar[3]. Therefore, it appears that all these models lack some element that is vital for simulating sea-level variability along the Indian coast.

Importance of salinity in the north Indian ocean

We have seen that the seasonal changes in corrected sea level (sea level corrected for the effect of atmospheric pressure) were due to two factors, wind-driven coastal currents and thermohaline effects. The success of the models in simulating the seasonal cycle of coastal currents rules out the former as a cause of the discrepancy. The latter consists of two factors, temperature and salinity. A $2\frac{1}{2}$-layer model[17] that incorporates thermodynamics, including surface fluxes, does a fair job at simulating the seasonal cycle of temperature in the Indian Ocean; it too, however, fails to simulate the seasonal cycle of sea level off India. This, and the low range of temperature variations in the north Indian Ocean, which is essentially a tropical basin, make temperature too an unlikely cause of this discrepancy. That leaves us with salinity.

The large inflow of freshwater into the seas around India, due to rainfall over the ocean and runoff from rivers, forces large changes in salinity, especially near the coast in the Bay of Bengal. Most of the major rivers in the Indian subcontinent flow into the bay. The maximum inflow of freshwater comes from the Ganga and the Brahmaputra, which together empty about 7.2×10^{11} m^3, the fourth largest discharge in the world, into the northern bay during June–October. This seasonal inflow of freshwater, coupled with the seasonal changes in coastal currents, is capable of forcing large changes in sea level. During the south-west monsoon, the EICC flows poleward, holding up the freshwater in the northern bay. The reversal of the EICC after the collapse of the south-west monsoon results in an equatorward freshwater plume along the east coast of India[10]. Shetye[18] speculated on the possible implications of this plume for the dynamics of the EICC, but McCreary *et al.*[19] showed that the plume was unlikely to contribute significantly to the EICC transport. This, however, does not rule out possible changes in sea level.

Since changes in river runoff and coastal salinity must be linked to monsoon rainfall over India, this hypothesis implies a relation between rainfall and coastal sea level. A recent study[3], showed that there is a high correlation between annual all-India rainfall and annual sea level along the Indian coast, whose variation is shown in Figure 1 *e*. Another empirical evidence lending credibility to this hypothesis is the nature of the seasonal cycle of sea level along the east coast. The peak is sharp towards the south, but is broad at Paradip, which is located close to the shelf-break in the northern bay. We expect the tide gauge at Paradip to respond to freshwater soon after it enters the bay; those at Vishakhapatnam and Chennai, however, will feel the influence of freshwater only when it is brought there by the equatorward EICC. Since this happens in November, the sea level peaks at this time along most of the east coast. The freshwater plume does not change the transport of the EICC, but there is a rise in sea level[3]. Freshwater is flushed down the coast rapidly by the equatorward current and hence, the sea level peak at Vishakhapatnam and Chennai is sharp. This explanation for the observed seasonal cycle of sea level along the Indian coast relies on simulations with an *ad hoc* manipulation of the coastal salinity field in a reduced-gravity model, but it is plausible, and the high correlation between monsoon rainfall and annual sea level at Mumbai offers empirical support for this hypothesis.

Summary

The seasonal cycle of sea level along the Indian coast matches that of the observed alongshore coastal currents computed from ship drifts. A comparison with the local alongshore winds, however, shows a mismatch, suggesting that remote forcing of the coastal circulation is important. Numerical models of the north Indian Ocean are able to simulate the seasonal cycle of coastal currents and surface circulation in the north Indian Ocean, but fail to reproduce the seasonal cycle of sea level along the Indian coast. This failure is due to their neglect of salinity, variations in this being large off the Indian coast, especially in the east.

Numerical models will have to include salinity effects to simulate correctly the seasonal cycle of sea level along the Indian coast. A major stumbling block in achieving this, however, is the paucity of data on salinity and freshwater inflow from rivers into seas around India. Even three decades after the International Indian Ocean Expedition (IIOE), it is difficult to construct an annual cycle of salinity along the coast. This makes it necessary to use proxies for salinity, say rainfall, but this approach leaves much to be desired[3].

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