

Are interdecadal sea level changes along the Indian coast influenced by variability of monsoon rainfall?

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Abstract. The Mumbai (Bombay) tide gauge data, the only century-long record in the Indian Ocean, show that interdecadal changes in sea level mimic those in rainfall over the Indian subcontinent. We propose that the link between rainfall and sea level arises from changes in salinity in coastal waters. Rivers fed by southwest monsoon (June–September) rainfall bring a large fraction of the runoff to the Bay of Bengal, from where it is transported to the west coast of India by an equatorward East India Coastal Current, which is triggered partly by the withdrawal of the southwest monsoon. The West India Coastal Current carries the low-salinity water from the bay as well as the runoff from local rivers northward. The advection of the riverine inflow to Mumbai occurs within a season, but the slow mixing in the ocean forces changes in the cross-shore density gradient on longer timescales. This density gradient forces a two-layer geostrophic circulation, with a surface current, which flows with the lighter water on its right, and an undercurrent. Lower (higher) salinity at the coast implies higher (lower) coastal sea level and a rise (fall) of the pycnocline at the coast. Thus the interdecadal variability of sea level along the Indian coast can be linked directly to the variability of the monsoon, the major aspect of the climate of the region, but by a mechanism that is different from those generally proposed to link sea level to climate change; these hypotheses usually invoke a change in volume because of global warming.

1. Introduction

Sea level has been recognized as an excellent marker of climate change. “Global” sea level appears to be rising relative to land at the rate of 1 mm yr^{-1} , and this is attributed, in part, to the warming of the globe due to the greenhouse effect [Gornitz *et al.*, 1982]. The warming is expected to affect sea level in two ways. First, it is expected to raise the temperature in the upper ocean; second, it is expected to melt polar ice caps, releasing the large quantities of water trapped in them. Both processes would increase the volume of water in the world oceans, thereby raising sea level all over the globe.

The issue, however, is more complicated, and it is difficult to obtain a single global index for long-term sea level changes [Barnett, 1984]. While there is an apparent increase in sea level in most of the world oceans, there are regions that show a different trend, southeast Asia, for example. Apart from this, there is the problem of separating natural, low-frequency sea level variability from that caused by anthropogenic effects; this calls for long sea level time series and such records are not many. Moreover, the available records are not distributed uniformly, most being from

Scandinavia. Between the tropics of Cancer and Capricorn, only three records, those at Mumbai (Bombay), Honolulu, and Balboa, go back at least to the beginning of this century. Of these, only Mumbai is in the Indian Ocean (Figure 1), and its tide gauge has been considered to be representative of the low-frequency sea level variability in the basin [Gornitz *et al.*, 1982; Barnett, 1984].

Given that the ocean and the atmosphere form a coupled system, it is unquestionable that changes in sea level must reflect changes in climate. A major aspect of the climate of the Indian subcontinent, in particular, and of Asia, in general, is the monsoon, which is linked to global climate by planetary-scale processes in the atmosphere and the ocean. What is the relation between the variability of the Indian monsoon and the interdecadal changes in sea level at Mumbai on the Indian west coast? This question forms the subject of this paper.

2. Observations

We use monthly sea level data from the archives of the Permanent Service for Mean Sea Level (PSMSL). The time series of annual mean and extrema of sea level over a year at

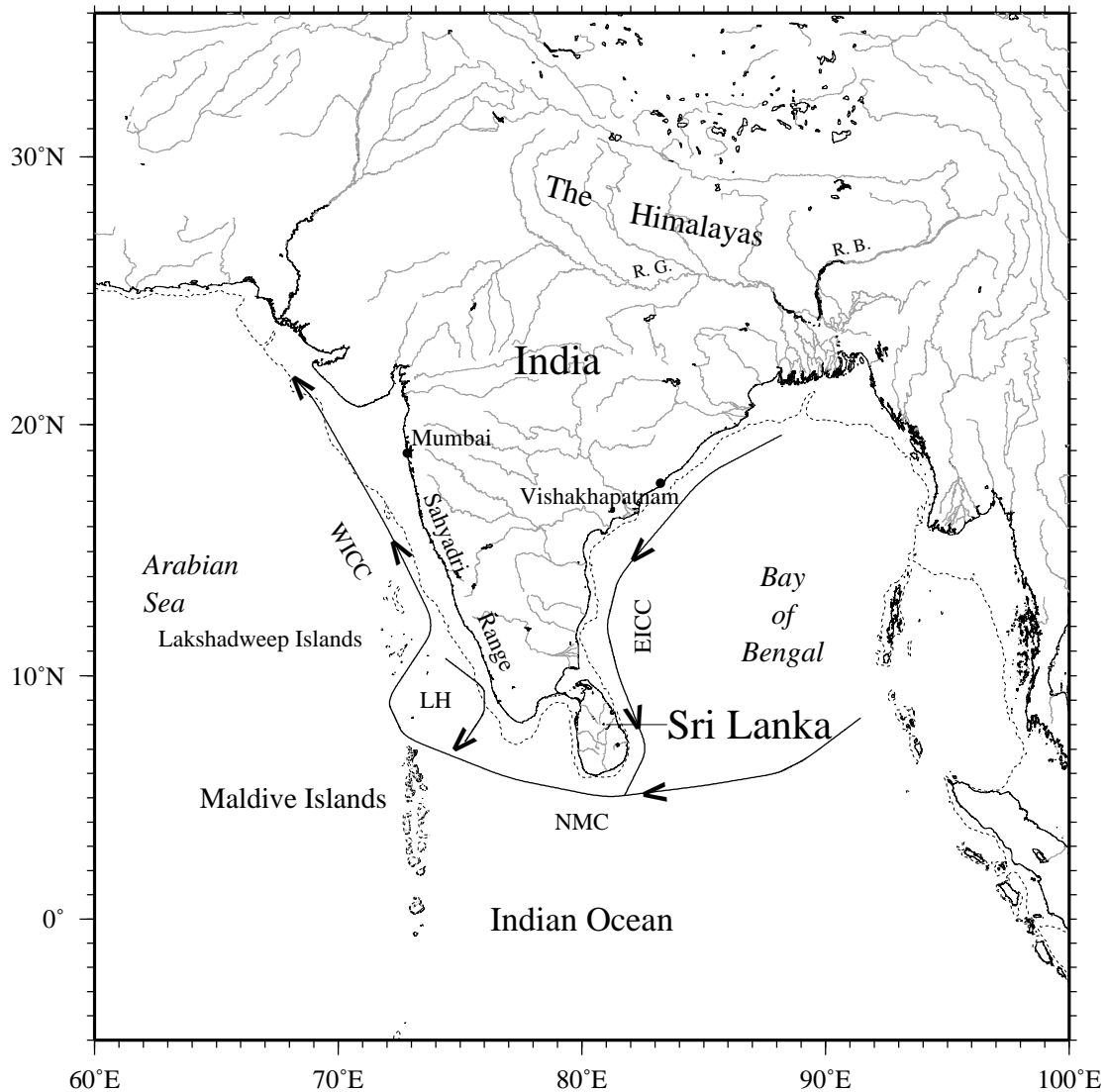


Figure 1. The geography of the Indian subcontinent. The major rivers are marked on the map; the dotted line marks the 200-m isobath. The abbreviations are as follows: EICC, East India Coastal Current; WICC, West India Coastal Current; NMC, Northeast Monsoon Current; LH, Lakshadweep High; R.G., River Ganga; and R.B., River Brahmaputra. Most of the major rivers of the subcontinent originate in the Himalayas and flow into the Bay of Bengal. Some of them originate in the Sahyadri Range (Western Ghats) and flow eastward into the bay; smaller rivers flow westward from the Sahyadris into the Arabian Sea. The EICC flows equatorward during October–January, and the WICC flows poleward during November–March. Together, these two coastal currents transport the runoff into the northern bay to the Arabian Sea and spread this freshwater, as well as the runoff into the Arabian Sea, along the west coast of India.

Mumbai are shown in Figure 2, which also shows the low-frequency variability, obtained by subjecting the data to a low-pass filter with a 10-year running mean. In general, both the maximum and minimum sea level over the year, as well as the seasonal averages, rise and fall in tune with the annual mean; this is particularly true on the interdecadal timescale (Table 1). Annual mean sea level at Mumbai exhibits considerable interannual and interdecadal variability. Sea level was generally low through 1880–1920, rising thereafter; it peaked in the late 1950s, falling thereafter. This change in

the trend in mean sea level at Mumbai during the second half of this century has been noted before [Emery and Aubrey, 1989]. The rise in the first half of this century was ~ 11 cm, the fall after that being about ~ 4.5 cm.

Figure 2 also shows the variation in all-India annual rainfall [Parthasarathy *et al.*, 1993, 1995; Shankar, 1998], a popular index for monsoon variability. Annual sea level at Mumbai is correlated with all-India and local annual rainfall. The correlation is significant even when the rainfall and sea level are low-pass filtered; it increases when the data are

truncated to the post-1900 period (Table 1). Though the filtered rainfall is noisier than the filtered sea level, it also undergoes low-frequency, interdecadal oscillations, increasing from a low early this century to a peak in the 1950s, falling thereafter. The lower correlation between the complete time series is because sea level in the late 18th century was just 2 cm higher than in the early 19th century, but rainfall during the late 18th century was as high as during the 1950s.

3. Hypothesis

Our hypothesis is that the seasonal inflow of the monsoon rainfall into the seas around India and the dynamics of currents along the Indian coast provide the link between the rainfall over the Indian subcontinent and the sea level along the coast of India, with coastal salinity playing an intermediate role. There is considerable spatial and temporal variation in rainfall over the Indian subcontinent and the surrounding ocean, and this leads to large variations in salinity along the coast. The Indian west coast, including Mumbai, receives very high rainfall, $\sim 200 \text{ cm year}^{-1}$, almost 90% of which is during the southwest monsoon (June–September). The high rainfall is because of the Sahyadri range (or Western Ghats) running parallel to the coast $\sim 100 \text{ km}$ inland (Figure 1); this range blocks the moisture-rich monsoon winds. The resulting runoff is carried to the Arabian Sea by numerous swift, seasonal streams. Over the rest of the subcontinent, there is considerable spatial variability in the amount of rainfall, but about $\sim 80\%$ is received during the southwest monsoon. Most of this rain falls in the catchment areas of eastward flowing rivers (Figure 1), whose inflow into the Bay of Bengal peaks in August, a month after the peak in rainfall. About 70% of this inflow comes from the Ganga and the Brahmaputra, which discharge $\sim 7.2 \times 10^{11} \text{ m}^3$, the fourth largest discharge in the world, into the northern bay during June–October [Martin *et al.*, 1981; Shetye, 1993]. Assuming that 80% of the rainfall over the west coast of India reaches the Arabian Sea, the runoff into the Indian west coast is about a third of that of the Ganga and the Brahmaputra [Shankar, 1998]. Thus the quantum of freshwater inflow from the rivers into the seas around India is highly seasonal.

Equally seasonal are the coastal currents around India. The weakening of the southwest monsoon winds after July, coupled with remote forcing from the eastern Bay of Bengal and the equatorial Indian Ocean, forces an equatorward East India Coastal Current (EICC) in the northern bay; the EICC is poleward along the rest of the Indian east coast [Shetye *et al.*, 1991a; McCreary *et al.*, 1993, 1996]. Together, these currents trap the runoff in the northern bay. As the southwest monsoon withdraws and the northeast monsoon sets in, the equatorward EICC expands southward, forcing coastal downwelling and advecting the riverine inflow as a coastally trapped low-salinity plume that is nearly 60-m deep [Shetye *et al.*, 1996]. By November the EICC is equatorward all along the east coast, and there is a sharp drop in the salinity along the coast. The EICC flows into the westward Northeast Monsoon Current (NMC), which bends around Sri Lanka

and flows along the western flank of the Lakshadweep high [Shankar and Shetye, 1997] into the poleward West India Coastal Current (WICC); the riverine inflow into the Bay of Bengal is thus transported into the Arabian Sea. This inflow is spread along the west coast of India by the WICC and is spread offshore off southwest India by the westward propagating Rossby waves that constitute the Lakshadweep high [Shenoi *et al.*, 1999]. By February the EICC reverses to flow poleward, forcing coastal upwelling and raising salinity along the east coast [Shetye *et al.*, 1993]. The WICC along the northern part of the coast, however, continues to flow poleward till March, spreading the low-salinity water from the bay, as well as the inflow from local rivers during the southwest monsoon, along the west coast.

During the southwest monsoon, when the freshwater actually enters the Indian coastal regime, the EICC and WICC favor upwelling, and hence the low-salinity water is trapped at the surface and pushed offshore by the Ekman flow, except in the northern bay, where the remotely forced equatorward EICC traps the river runoff. It is only after the southwest monsoon, when these currents reverse, that the low-salinity water is pushed toward, and advected along, the coast. Even along the west coast, the lowest salinities at the surface are during the southwest monsoon, but the drop in salinity is restricted to a shallow surface layer [Shetye *et al.*, 1990]. As along the east coast, it is during the northeast monsoon, when the WICC favors downwelling and there is inflow of freshwater from the Bay of Bengal, that the low-salinity layer deepens [Shetye *et al.*, 1991b], this being more pronounced off southwest India where the low-salinity water spreads offshore because of the Lakshadweep high. The EICC and WICC, forced by the low-frequency, large-scale winds over the north and equatorial Indian Ocean, have a speed of 25 km day^{-1} [Shetye *et al.*, 1991b, 1996]; therefore the low-salinity water from the bay is advected to Mumbai by December, when the climatological seasonal cycle of sea level peaks there [Shankar, 1998].

A simple model for the seasonally reversing EICC and WICC is an annually forced Kelvin wave [Shankar and Shetye, 1997], which is upwelling favorable during the southwest monsoon and downwelling favorable during the northeast monsoon. The current and sea level associated with such a wave are shown in Figure 3a. Under the influence of such a wave the current and sea level at the coast undergo a seasonal cycle; a downwelling (upwelling) favorable coastal current implies an increase (decrease) in coastal sea level. The currents associated with this cycle distribute the riverine inflow along the coast.

The seasonal cycle of currents alone, however, cannot explain the link between all-India rainfall and the sea level at Mumbai on the much longer, interdecadal timescale (Figure 2). Our contention is that this link arises as a consequence of the spreading of the river runoff along the coast by the seasonal EICC and WICC. The result of the spreading and mixing of these low-salinity waters is a salinity gradient normal to the coast; the timescale associated with the tem-

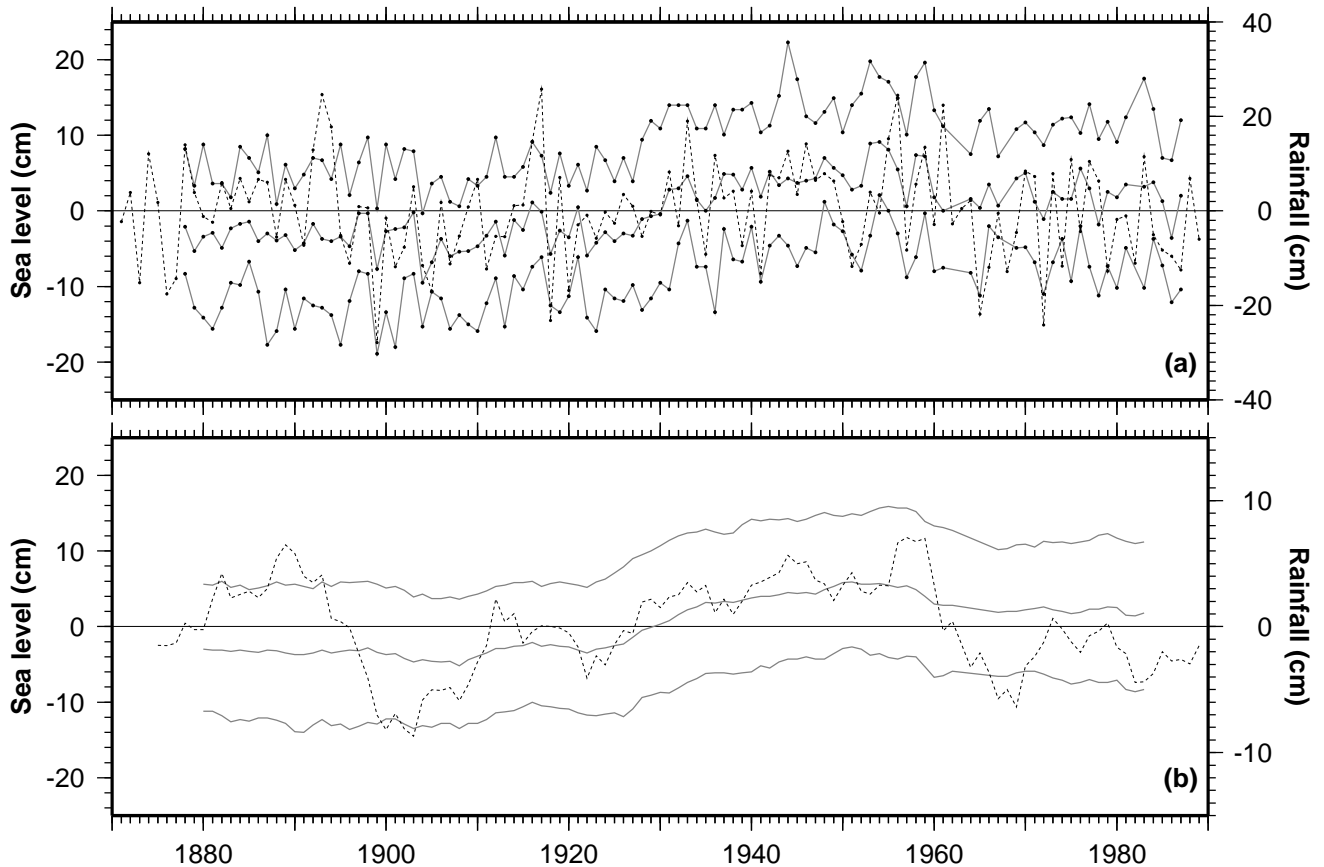


Figure 2. (a) Anomalies of annual all-India rainfall (centimeters; dotted line) and of the annual maximum (upper solid line), mean (middle solid line), and minimum (lower solid line) sea level (centimeters) at Mumbai (the mean rainfall and sea level have been removed from the time series). For sea level the year is defined from July to June; for rainfall, the year is defined from from January to December. (b) As in Figure 2a, but low-pass filtered with a 10-year running mean.

poral variability of this gradient is much longer than a year. We propose that it is this gradient that drives a weak, quasi-steady circulation (Figure 3b), providing the link between all-India rainfall and sea level at Mumbai.

In section 4 we put together a simple dynamical framework to justify the above proposition. We begin with a model that has been found useful in understanding the seasonal cycle of wind-forced currents around India and then constrain it to reflect the weak, quasi-steady forcing that arises because of the cross-shore salinity gradient. The predicted circulation has a surface current and an undercurrent, and the sea level variation is consistent with that depicted in Figure 2.

4. Mechanics of Interdecadal Sea Level Change Along the Coast of India

To a good approximation the EICC and the WICC, as well as the wind-forced circulation in the north Indian Ocean, can be simulated with reduced gravity models. Most of these models have a constant density in the active model layer(s) [Yu *et al.*, 1991; Potemra *et al.*, 1991; Bruce *et al.*, 1994;

Shankar and Shetye, 1997]. The $2\frac{1}{2}$ -layer model of McCreary *et al.* [1993] allows horizontal variation of temperature within a layer; Han *et al.* (Unpublished manuscript, 1998) also allow horizontal variation of salinity. In models that allow density to vary horizontally within a layer the vertical average of the pressure force in each layer is used to compute the horizontal accelerations; this ensures that there is no variation in the vertical, within a layer, of the density and of the horizontal current [Ripa, 1992, 1995]. In the following discussion, we use reduced gravity models to show how a cross-shore density gradient affects coastal sea level. Our primary aim is to examine how the currents and sea level associated with an annual coastal Kelvin wave are modified by a cross-shore variation in density; as noted earlier, an annual Kelvin wave is a good description of the conditions along the coast of India. We do not, however, model the processes that lead to this cross-shore gradient in salinity.

4.1. The $1\frac{1}{2}$ -Layer Model

Consider a $1\frac{1}{2}$ -layer, variable density reduced gravity model. The density being prescribed in the active upper layer

Table 1. Linear correlations. R and η denote rainfall and sea level; the subscripts “I”, “V”, and “M” denote all-India, Mumbai, and Vishakhapatnam, “m”, “min”, and “max” denote annual mean, minimum, and maximum, and “MAM”, “JAS”, and “NDJ” denote seasonal averages for March–May, July–September, and November–January. N_{yr} implies an N -year running mean has been used to filter the data. ($N = 1$ implies no running mean.) The superscript “T” implies that the data have been truncated; only the post-1900 data are included. r is the linear correlation coefficient between the variables in the first two columns; r_s is the correlation at the $P\%$ significance level. For sea level, the year is defined from July–June, for rainfall from January–December. For computing the correlation between all-India rainfall and sea level at Vishakhapatnam, the data for 1961 have been dropped. The method of computing correlations follows Shankar [1998].

		N_{yr}	r	r_s	P
$\eta_{M,m}$	$\eta_{M,max}$	1,5,10	0.879,0.974,0.990	0.247,0.450,0.609	99,99,99
$\eta_{M,m}$	$\eta_{M,min}$	1,5,10	0.862,0.968,0.985	0.247,0.450,0.609	99,99,99
$\eta_{M,m}$	$\eta_{M,MAM}$	1	0.907	0.247	99
$\eta_{M,m}$	$\eta_{M,JAS}$	1	0.845	0.247	99
$\eta_{M,m}$	$\eta_{M,NDJ}$	1	0.863	0.247	99
R_I	$\eta_{M,m}$	1,5,10	0.321,0.328,0.464	0.247,0.299,0.424	99,95,90
R_I	$\eta_{M,m}$	1 ^T ,10 ^T	0.416,0.654	0.278,0.557	99,95
R_M	$\eta_{M,m}$	1,5,10	0.492,0.666,0.752	0.247,0.450,0.609	99,99,99
$\eta_{V,m}$	$\eta_{M,m}$	1	0.401	0.316	95
R_I	$\eta_{V,m}$	1	0.297	0.264	90

of the model may vary in space (because of a variation in salinity); it is not allowed to evolve in time, implying that the timescale associated with density changes is much longer than the transient motions such as Kelvin waves that are supported by the model. This is because our interest is in the long-term effects of changes in coastal salinity: how is the annual Kelvin wave modified by a quasi-steady cross-shore salinity gradient? We solve for a free Kelvin wave propagating poleward along a north-south coast on an f plane; since there is no cross-shore motion for a Kelvin wave, there is no zonal velocity ($u = 0$).

We scale the variables and use a perturbation expansion to obtain the zeroth-order, linearized perturbation equations; these are given by

$$-fv = -g\bar{\Gamma}\frac{\partial h}{\partial x} - \frac{g\bar{H}}{2}\frac{\partial \gamma}{\partial x}, \quad (1a)$$

$$\frac{\partial v}{\partial t} = -g\bar{\Gamma}\frac{\partial h}{\partial y} - \frac{g\bar{H}}{2}\frac{\partial \gamma}{\partial y}, \quad (1b)$$

$$\frac{\partial h}{\partial t} = -\bar{H}\frac{\partial v}{\partial y}, \quad (1c)$$

$$\eta = \bar{\Gamma}h + \bar{H}\gamma, \quad (1d)$$

where \bar{H} and $\bar{\Gamma}$ are the upper layer thickness and reduced gravity parameter in the basic state of no motion, about which the system is perturbed. \bar{H} and $\bar{\Gamma}$ are constants; v , h , η , and γ are the perturbation meridional (alongshore) velocity, upper layer thickness, sea level, and reduced gravity parameter; and γ is not a function of time t . The layer thickness

and reduced gravity parameter are $H = \bar{H} + h$ and $\Gamma = \bar{\Gamma} + \gamma$. $\Gamma = \frac{\rho_2 - \rho_1}{\bar{\rho}}$, where ρ is the density and the subscripts 1 and 2 refer to the two layers; $\bar{\rho}$ is an average density that is representative of the ocean; g is the acceleration due to gravity; and f is the Coriolis parameter.

We define an “equivalent layer thickness” as

$$H' = \bar{H} + h', \quad (2a)$$

where

$$h' = h + \left(\frac{\gamma}{2\bar{\Gamma}}\right) \quad (2b)$$

is the “equivalent perturbation layer thickness”. Equations (1a)–(1d) and (2a)–(2b) yield the free wave equation

$$\frac{\partial h'}{\partial t^2} = c^2 \frac{\partial h'}{\partial y^2}, \quad (3)$$

where $c = (g\bar{\Gamma}\bar{H})^{\frac{1}{2}}$ is the speed of the first-baroclinic-mode Kelvin wave. The solution to (3) and (1a) is

$$h' = \tilde{h}' e^{x/a + i(ty - \sigma t)}, \quad (4)$$

where \tilde{h}' is the equivalent amplitude of the Kelvin wave at the coast and $a = \frac{c}{f}$ is the internal Rossby radius of deformation. Equations (1a)–(1d), (2a)–(2b), and (4) yield the required zeroth-order solution

$$h = \tilde{h}' e^{x/a + i(ty - \sigma t)} - \left(\frac{\gamma}{2\bar{\Gamma}}\right), \quad (5a)$$

$$v = \left(\frac{g\bar{\Gamma}}{\bar{H}}\right)^{\frac{1}{2}} \tilde{h}' e^{x/a + i(ty - \sigma t)}, \quad (5b)$$

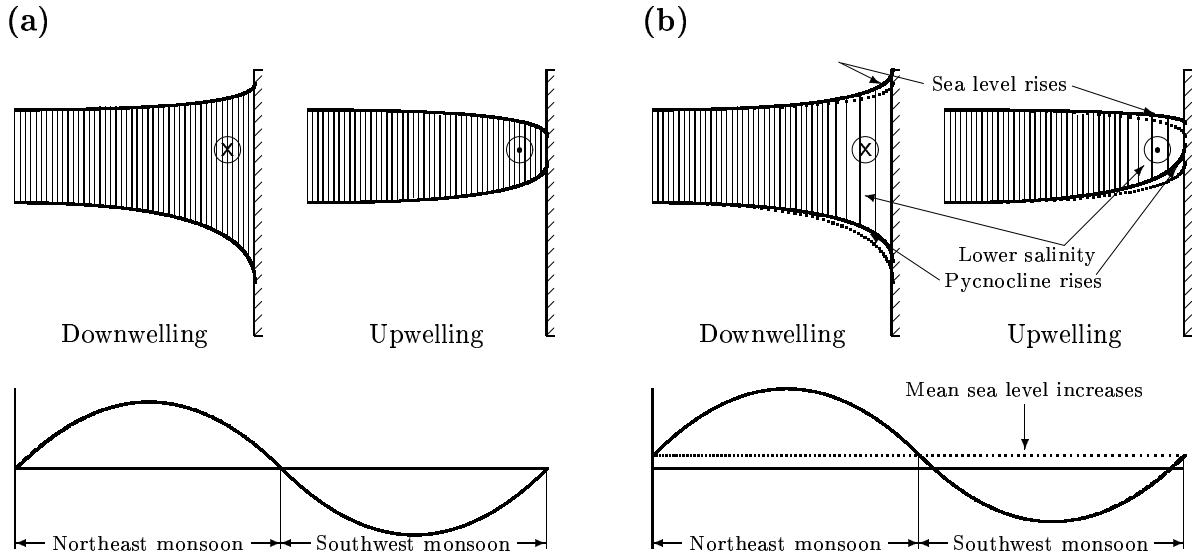


Figure 3. (a) Schematic of a section normal to the coast showing the relation between currents and sea level along the coast of India. The current shown by \times (dot) within a circle is into (out of) the plane of the paper. When there is downwelling (upwelling) at the coast, the current flows with the coast on its right (left), sea level rises (falls) at the coast, and the pycnocline (the top of which is shown by the line marking the bottom of the hatched area) slopes down (up) toward the coast. At Mumbai, there is downwelling (upwelling) during the northeast (southwest) monsoon. (b) Schematic to show how, according to our hypothesis, the conditions depicted in Figure 3a are modified when salinity near the coast is lower than that farther offshore. Wider spacing of the hatched lines indicates lower salinity. We propose that the salinity (hence density) gradient normal to the coast drives a weak, quasi-steady surface current that moves with the coast on its right, introducing a quasi-steady rise in sea level at the coast.

$$\eta = \bar{\Gamma}' h' e^{x/a + i(\gamma y - \sigma t)} + \left(\frac{\gamma}{2}\right) \bar{H}. \quad (5c)$$

In the absence of spatial variation in density ($\Gamma = \bar{\Gamma}; \gamma = 0$), we obtain the classical Kelvin wave solution. When density varies in space ($\gamma \neq 0$), the Kelvin wave amplitude is modified (it becomes the equivalent amplitude), but there is another change that is significant for sea level at the coast. When salinity, and hence density, decreases (increases) from the open sea toward the coast, i.e., when $\frac{\partial \gamma}{\partial x} > 0$ ($\frac{\partial \gamma}{\partial x} < 0$), sea level at the coast increases (decreases). This change in sea level depends only on γ and hence is independent of time. Since the low-salinity water is spread along the coast by currents that decay exponentially away from the coast, i.e., they are coastally trapped, the perturbation in salinity, and hence γ , decays exponentially away from the coast ($\gamma \propto e^{-kx}$, $k > 0$). Therefore the time-independent change in sea level predicted by the model also varies exponentially cross shore.

The Kelvin wave is superimposed on the time-independent increase in sea level. In the linear system described by (1a)–(1d), we can ignore the Kelvin wave to look at the steady state. It consists of an increase in sea level and an upsloping of the pycnocline toward the coast (Figure 3b); that is, when there is no Kelvin wave and density decreases coastward, sea level rises at the coast, the pycnocline upslopes, and there is no current in the layer. This is so because the

cross-shore pressure gradient at the middle of the layer vanishes. More insight into such systems can be gained by extending the model to include another active layer.

4.2. The $2\frac{1}{2}$ -Layer Model

Consider a $2\frac{1}{2}$ -layer reduced gravity model, but with no difference in density between the two active layers in the basic state, i.e., $\bar{\Gamma}_1 = \bar{\Gamma}_2 = \bar{\Gamma}$; the perturbation density in the two layers, however, is different, i.e., $\Gamma_1 \neq \Gamma_2$. The model now supports velocity shear in the vertical, permitting the existence of a “shear mode.” As with the $1\frac{1}{2}$ -layer model, the perturbation density varies in the horizontal in the active layers; the interface between the lower active layer and the deep ocean is the model pycnocline. A precise definition of the density field is required because this model differs from the usual $2\frac{1}{2}$ -layer reduced gravity model, in which $\bar{\Gamma}_1 \neq \bar{\Gamma}_2$, implying the existence of two baroclinic modes. Our interest is restricted to understanding why the $1\frac{1}{2}$ -layer model does not yield a quasi-steady velocity along with the quasi-steady rise in sea level, and for this the introduction of shear in the vertical is sufficient; the assumption that there is no difference in unperturbed density between the two shear layers ($\bar{\Gamma}_1 = \bar{\Gamma}_2$) simplifies the algebra.

We define the reduced gravity parameters as follows. Let ρ_1 , ρ_2 , and ρ_3 be the densities of the three model layers ($\rho_1 <$

$\rho_2 < \rho_3$). Then, by definition,

$$\Gamma_1 = \frac{\rho_3 - \rho_1}{\bar{\rho}},$$

$$\Gamma_2 = \frac{\rho_3 - \rho_2}{\bar{\rho}},$$

where $\bar{\rho}$ is an average density that is representative of the ocean. In the basic state, $\bar{\Gamma}_1 = \bar{\Gamma}_2 = \bar{\Gamma}$. Hence we have $\Gamma_1 = \bar{\Gamma} + \gamma_1$ and $\Gamma_2 = \bar{\Gamma} + \gamma_2$. Let

$$\bar{\Gamma} = \frac{\rho_3 - \bar{\rho}_{1,2}}{\bar{\rho}}, \quad (6a)$$

where

$$\bar{\rho}_{1,2} = \frac{\bar{H}_1 \bar{\rho}_1 + \bar{H}_2 \bar{\rho}_2}{\bar{H}}, \quad (6b)$$

$$\bar{\rho}_1 = \frac{1}{A} \iint_{\mathbf{A}} \rho_1 dx dy, \quad (6c)$$

$$\bar{\rho}_2 = \frac{1}{A} \iint_{\mathbf{A}} \rho_2 dx dy. \quad (6d)$$

\bar{H}_1 and \bar{H}_2 are the layer thicknesses in the basic state and \mathbf{A} is the area of the model domain. \bar{H}_1 , \bar{H}_2 , and $\bar{\Gamma}$ define the basic state about which the system is perturbed. Since $\rho_2 > \rho_1$, we have $\bar{\rho}_2 > \bar{\rho}_1$. Then

$$\gamma_1 = \frac{\bar{\rho}_{1,2} - \rho_1}{\bar{\rho}} = \frac{\bar{\rho}_1 - \rho_1}{\bar{\rho}} + \left(\frac{\bar{H}_2}{\bar{H}} \right) \left(\frac{\bar{\rho}_2 - \bar{\rho}_1}{\bar{\rho}} \right) \quad (7a)$$

$$\gamma_2 = \frac{\bar{\rho}_{1,2} - \rho_2}{\bar{\rho}} = \frac{\bar{\rho}_2 - \rho_2}{\bar{\rho}} - \left(\frac{\bar{H}_1}{\bar{H}} \right) \left(\frac{\bar{\rho}_2 - \bar{\rho}_1}{\bar{\rho}} \right). \quad (7b)$$

This yields $\gamma_1 - \gamma_2 > 0$, as expected. This definition of the density field and reduced gravity parameters ensures that \bar{H}_1 is a meaningful quantity: it is the depth up to which salinity at the coast is affected by the runoff from rivers. A consequence of the definition is that $\gamma_2 \neq 0$, but it may be constant.

The zeroth-order, linearized perturbation equations for this $2\frac{1}{2}$ -layer model, equivalent to (1a)–(1d) for the $1\frac{1}{2}$ -layer model, make the same assumptions as for the $1\frac{1}{2}$ -layer model.

$$-fv_1 = -g\bar{\Gamma} \frac{\partial}{\partial x} (h_1 + h_2) - \frac{g\bar{H}_1}{2} \frac{\partial \gamma_1}{\partial x} - g\bar{H}_2 \frac{\partial \gamma_2}{\partial x}, \quad (8a)$$

$$\frac{\partial v_1}{\partial t} = -g\bar{\Gamma} \frac{\partial}{\partial y} (h_1 + h_2) - \frac{g\bar{H}_1}{2} \frac{\partial \gamma_1}{\partial y} - g\bar{H}_2 \frac{\partial \gamma_2}{\partial y}, \quad (8b)$$

$$\frac{\partial h_1}{\partial t} = -\bar{H}_1 \frac{\partial v_1}{\partial y} \quad (8c)$$

are the equations for the first layer, and

$$-fv_2 = -g\bar{\Gamma} \frac{\partial}{\partial x} (h_1 + h_2) - \frac{g\bar{H}_2}{2} \frac{\partial \gamma_2}{\partial x}, \quad (9a)$$

$$\frac{\partial v_2}{\partial t} = -g\bar{\Gamma} \frac{\partial}{\partial y} (h_1 + h_2) - \frac{g\bar{H}_2}{2} \frac{\partial \gamma_2}{\partial y}, \quad (9b)$$

$$\frac{\partial h_2}{\partial t} = -\bar{H}_2 \frac{\partial v_2}{\partial y} \quad (9c)$$

are the equations for the second layer. The perturbation sea level is given by

$$\eta = \bar{\Gamma} (h_1 + h_2) + \bar{H}_1 \gamma_1 + \bar{H}_2 \gamma_2. \quad (10)$$

Let

$$h = h_1 + h_2, \quad (11a)$$

$$\bar{H} = \bar{H}_1 + \bar{H}_2, \quad (11b)$$

$$v = \frac{\bar{H}_1 v_1 + \bar{H}_2 v_2}{\bar{H}}, \quad (11c)$$

where \bar{H} is the ‘‘cumulative layer thickness’’ in the basic state, h is the ‘‘cumulative perturbation layer thickness,’’ and v is the mean velocity. Then (8a)–(8c), (9a)–(9c), and (11a)–(11c) yield the free wave equation

$$\frac{\partial h'}{\partial t^2} = c^2 \frac{\partial h'}{\partial y^2}, \quad (12)$$

where

$$h' = h + \left(\frac{\bar{H}_1^2 \gamma_1 + 2\bar{H}_1 \bar{H}_2 \gamma_2 + \bar{H}_2^2 \gamma_2}{2\bar{\Gamma}\bar{H}} \right) \quad (13)$$

is the ‘‘equivalent cumulative perturbation layer thickness’’ and $c = (g\bar{\Gamma}\bar{H})^{\frac{1}{2}}$. The solution is

$$h' = \tilde{h}' e^{x/a+i(l y - \sigma t)}, \quad (14)$$

where $a = \frac{c}{f}$. Then (13) and (14) imply

$$h = \tilde{h}' e^{x/a+i(l y - \sigma t)} - \left(\frac{\bar{H}_1^2 \gamma_1 + 2\bar{H}_1 \bar{H}_2 \gamma_2 + \bar{H}_2^2 \gamma_2}{2\bar{\Gamma}\bar{H}} \right). \quad (15)$$

The perturbation sea level is

$$\eta = \bar{\Gamma} \tilde{h}' e^{x/a+i(l y - \sigma t)} + \left(\frac{\bar{H}_1^2 \gamma_1 + 2\bar{H}_1 \bar{H}_2 \gamma_2 + \bar{H}_2^2 \gamma_2}{2\bar{H}} \right). \quad (16)$$

Since we are interested only in the quasi-steady part of the solution, we restrict our attention to it. It is the quasi-steady response that causes the interdecadal changes in sea level,

the Kelvin wave being superimposed on it. The quasi-steady component of h is

$$h_s = - \left(\frac{\overline{H}_1^2 \gamma_1 + 2\overline{H}_1 \overline{H}_2 \gamma_2 + \overline{H}_2^2 \gamma_2}{2\overline{H}} \right), \quad (17a)$$

and the quasi-steady sea level is

$$\eta_s = \frac{\overline{H}_1^2 \gamma_1 + 2\overline{H}_1 \overline{H}_2 \gamma_1 + \overline{H}_2^2 \gamma_2}{2\overline{H}}. \quad (17b)$$

Substituting (17a) into (8a) and (9a), we obtain the quasi-steady solutions for the alongshore velocity in the two layers. These are

$$v_{1,s} = \left(\frac{g\overline{H}_2}{2f\overline{H}} \right) \frac{\partial}{\partial x} (\overline{H}_1 \gamma_1 + \overline{H}_2 \gamma_2) \quad (17c)$$

$$v_{2,s} = - \left(\frac{g\overline{H}_1}{2f\overline{H}} \right) \frac{\partial}{\partial x} (\overline{H}_1 \gamma_1 + \overline{H}_2 \gamma_2). \quad (17d)$$

The introduction of the shear mode permits the existence of an undercurrent, but there is no mean velocity, i.e., $v_s = 0$, as in the $1\frac{1}{2}$ -layer model, and

$$\frac{v_{1,s}}{v_{2,s}} = - \frac{\overline{H}_2}{\overline{H}_1}.$$

From (17a)–(17d) we obtain

$$\frac{\partial \eta_s}{\partial x} \geq 0 \quad \forall \quad \frac{\partial \gamma_1}{\partial x} \geq -\tan \theta_\eta \frac{\partial \gamma_2}{\partial x}, \quad (18a)$$

$$\frac{\partial h_s}{\partial x} \leq 0 \quad \forall \quad \frac{\partial \gamma_1}{\partial x} \geq -\tan \theta_h \frac{\partial \gamma_2}{\partial x}, \quad (18b)$$

$$v_{1,s} \geq 0 \quad \forall \quad \frac{\partial \gamma_1}{\partial x} \geq -\tan \theta_v \frac{\partial \gamma_2}{\partial x}, \quad (18c)$$

$$v_{2,s} \leq 0 \quad \forall \quad v_{1,s} \geq 0, \quad (18d)$$

where

$$\tan \theta_\eta = \left(\frac{\overline{H}_2}{\overline{H}_1} \right)^2 \left(1 + \frac{2\overline{H}_2}{\overline{H}_1} \right)^{-1}, \quad (19a)$$

$$\tan \theta_h = \left(1 + \frac{2\overline{H}_2}{\overline{H}_1} \right)^{-1}, \quad (19b)$$

$$\tan \theta_v = \frac{\overline{H}_2}{\overline{H}_1}, \quad (19c)$$

with

$$\theta_\eta < \theta_v < \theta_h. \quad (19d)$$

Equations (18a)–(18d) and (19a)–(19d) imply that the velocities in the two active layers, $v_{1,s}$ and $v_{2,s}$, and the cross-shore gradients of η_s and h_s depend on the layer thicknesses in the basic state, \overline{H}_1 and \overline{H}_2 , and on the cross-shore gradients of both γ_1 and γ_2 . This is shown graphically in Figure 4. The solution shows that there exist conditions under which

sea level may rise at the coast; this need not, however, be accompanied by a rise of the pycnocline or by a surface current that flows with the higher sea level on its right, unlike in the simpler solution obtained for the $1\frac{1}{2}$ -layer model. For example, between the solid and dashed lines in the second quadrant in Figure 4, sea level falls coastward, but the surface current is still poleward (for an eastern boundary like the Indian west coast); between the solid and dotted lines in the fourth quadrant, sea level rises toward the coast but so does the pycnocline. An extreme case is that of sea level falling at the coast despite salinity decreasing coastward; this can happen if the density of the lower layer increases coastward at a sufficiently rapid rate (the region to the left of the solid line in Figure 4). Thus there exist a number of possible solutions, and it is the prevailing conditions, i.e., the thickness of the layer influenced by runoff (\overline{H}_1), the thickness of the layer below (\overline{H}_2), and the cross-shore gradients of the reduced gravity parameters in the two layers (γ_1 and γ_2) that determine whether sea level rises at the coast or not.

Despite the relative complexity of the $2\frac{1}{2}$ -layer solution it is similar to the $1\frac{1}{2}$ -layer model when there is a coastward decrease of salinity above the pycnocline. When this happens (first quadrant in Figure 4), sea level has to rise at the coast. These $1\frac{1}{2}$ -layer and $2\frac{1}{2}$ -layer solutions are depicted in Figure 5. As shown in Figure 4, unless there is a sufficiently large coastward increase in density in the second layer, sea level has to rise at the coast when salinity decreases coastward in the upper layer.

4.3. Application to the West Coast of India

At Mumbai, sea level rose by ~ 11 cm during 1900–1950 and fell by ~ 4.5 cm during 1950–1980. By how much does the salinity at the coast have to change to force these interdecadal changes in sea level? Our model domain has an eastern boundary and extends to infinity toward the west. The changes in salinity due to runoff from rivers affect only a layer of thickness, \overline{H}_1 , and are restricted to the vicinity of the coast, the slow mixing in the ocean acting over time to reduce the cross-shore density gradient. This cross-shore gradient is quasi steady because mixing in the ocean is slow; the more rapid changes forced by seasonal Kelvin waves are superimposed on it. In the lower active layer of thickness, \overline{H}_2 , density, and hence γ_2 , is assumed to be constant. Differentiating η_s in (17b) then eliminates the contribution from density variations in this layer. Integrating $\frac{\partial \eta_s}{\partial x}$ over $(-\infty, 0]$ for an exponential variation of density in x ($\gamma_1 \propto e^{kx}$, where k^{-1} is a length scale), we obtain

$$\eta_s|_{x=0} = \left(\frac{\overline{H}_1^2 + 2\overline{H}_1 \overline{H}_2}{2\overline{H}} \right) \gamma_1|_{x=0}. \quad (20)$$

For $\eta_s = 5$ cm and $\overline{H}_1 = \overline{H}_2 = 50$ m we obtain $\gamma_1 = 0.0013$. If we assume that the change in density is entirely due to a change in salinity, ignoring other processes (for example, temperature changes due to “global warming,” melting of ice caps, etc.), this translates to a decrease of 1.07 practical

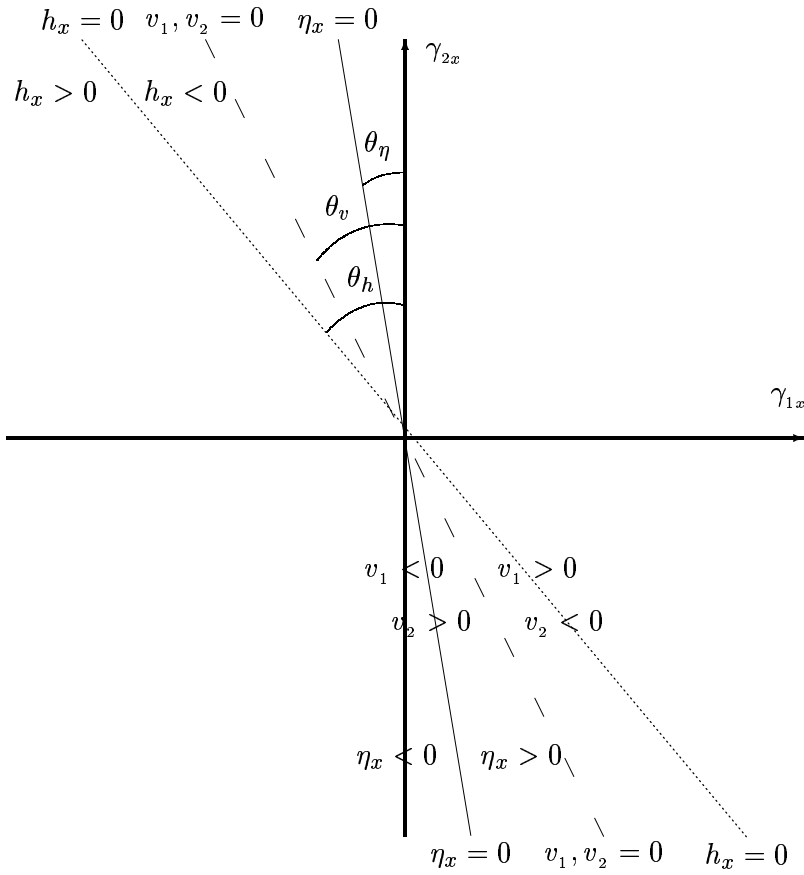


Figure 4. A graphical representation of the quasi-steady solution to the $2\frac{1}{2}$ -layer model equations. The possible quasi-steady solutions (the subscript s is dropped for convenience) are shown as a function of γ_{1x} (abscissa) and γ_{2x} (ordinate); the subscript x denotes a partial derivative with respect to x , the cross-shore coordinate. The lines $h_x = 0$ (dotted), $\eta_x = 0$ (solid), and $v_1 = v_2 = 0$ (dashed) mark the boundaries across which the cross-shore gradients of h and η and the velocities v_1 and v_2 change sign. To the right (left) of the respective lines, η_x and v_1 are positive (negative), and h_x and v_2 are negative (positive). Unless there is a sufficiently large coastward decrease in γ_2 ($\gamma_{2x} < 0$), a coastward decrease in the salinity of the upper active layer at an eastern boundary ($\gamma_{1x} > 0$) implies a rise in coastal sea level, a rise of the pycnocline toward the coast, and a surface current (undercurrent) that flows with the coast on its right (left).

salinity units (psu) in the salinity averaged over the top 50 m. A similar approach, using (17a) instead of (17b), gives the displacement of the pycnocline at the coast.

$$h_s|_{x=0} = - \left(\frac{\bar{H}_1^2}{2\bar{\Gamma}H} \right) \gamma_{1x}|_{x=0}. \quad (21)$$

For $\bar{\Gamma} = 0.0035$ [Shankar and Shetye, 1997; Shankar, 1998], we obtain $h_s = -4.8$ m at the coast. To compute the currents associated with the cross-shore density gradient, we need a measure of the length scale k^{-1} associated with the mixing of salt in the Arabian Sea. Horizontal eddy diffusivities in the Arabian Sea range from 1.3×10^6 (zonal) to 3.1×10^6 $\text{cm}^2 \text{s}^{-1}$ (meridional) [Somayajulu *et al.*, 1996]. Assuming an average value of 2×10^6 $\text{cm}^2 \text{s}^{-1}$, we obtain a length scale of 400 km for a timescale of 25 years, the period over which sea level at Mumbai drops by 4.5 cm. Then

(17c) implies $v_{1,s} = -v_{2,s} = 0.8$ cm s^{-1} ; thus the background thermohaline circulation forced by the cross-shore density gradient is much weaker than the seasonal WICC superimposed on it.

To summarize, a suitable model to explain the link between sea level and all-India rainfall is the following. An increase in rainfall over India leads to a decrease in salinity along the coast, resulting in a cross-shore density gradient to which the ocean responds as a two-layer system. When density decreases coastward in the surface layer, sea level rises at the coast, and this is accompanied by a geostrophic surface current that flows with the lighter water on its right; below the surface current is an undercurrent. The sum of the transport of the surface current and the undercurrent vanishes. Both the surface current and the undercurrent are weak compared to the seasonal surface current and undercurrent.

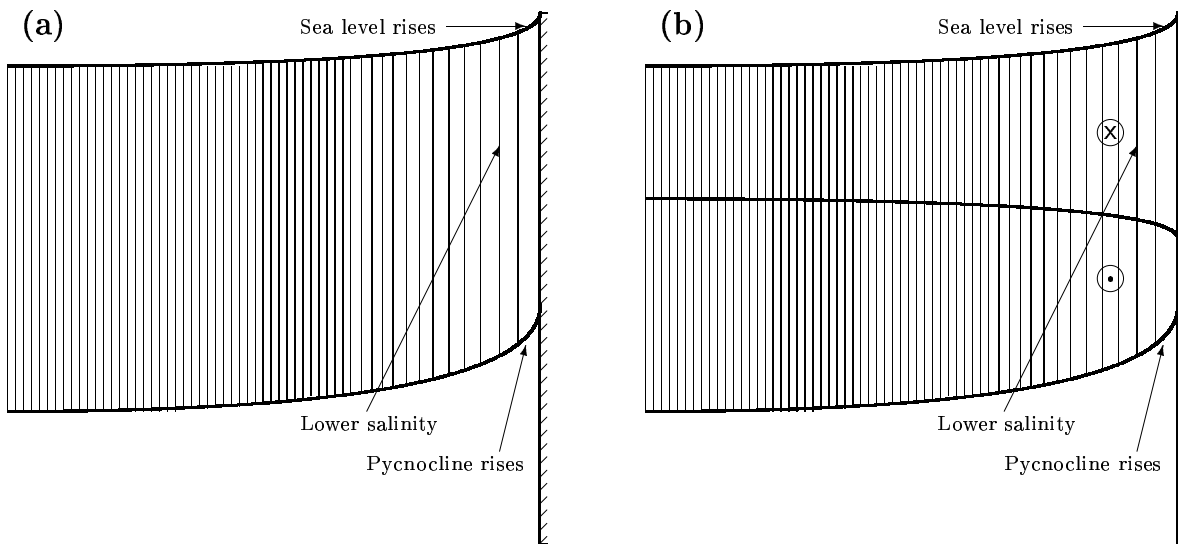


Figure 5. Schematic of a section normal to the coast showing the effect of a decrease in the salinity of coastal waters on sea level at an eastern ocean boundary in the Northern Hemisphere: (a) The response in a $1\frac{1}{2}$ -layer model and (b) the response in a $2\frac{1}{2}$ -layer model. The cross-shore time-independent variation in salinity forces a rise in coastal sea level and a time-independent geostrophic circulation with a surface current and an undercurrent. The surface current flows with the lighter water on its right, and the undercurrent flows in the opposite direction. The pycnocline (the interface between the undercurrent and the motionless deeper ocean) slopes up toward the coast. The transports of the two currents cancel to yield a zero net transport. The $1\frac{1}{2}$ -layer model yields only the mean of the surface current and the undercurrent, and hence predicts a zero current associated with the rise in sea level and the pycnocline at the coast. The current shown by x (dot) within a circle is into (out of) the plane of the paper, and wider spacing of the hatched lines indicates lower salinity.

Geostrophy requires that the interface between the undercurrent and the motionless deeper ocean (the model pycnocline) slope up toward the coast (Figure 5). The slow mixing in the ocean makes this response quasi steady.

Thus a coastward decrease (increase) in salinity implies an increase (decrease) in both the annual maximum and the annual minimum of sea level and hence an increase (decrease) in sea level throughout the year, as seen at Mumbai. This change in sea level occurs on interannual and interdecadal timescales, much slower than the changes associated with the annual timescale and other Kelvin waves that are superimposed on it. Though it is the currents associated with these waves that spread the low-salinity water along the Indian coast, the long-term effect of spreading this water is to set up a weak, quasi-steady circulation.

5. Effect of Interannual Variability of Winds

One implication of the above finding is that similar changes in sea level must occur along the Indian east coast because a large fraction of the freshwater comes from the Bay of Bengal, and the salinity changes are much greater along the east coast. We find support for this contention in the sea level at Vishakhapatnam (Figure 6), which, however, has a much shorter record (1937–1988) than does Mumbai (1878–1988).

It is correlated with sea level at Mumbai and with all-India rainfall (Table 1); the short record, however, makes it difficult to use a low-pass filter to look at interdecadal changes in sea level.

In computing the correlation between all-India rainfall and sea level at Vishakhapatnam we have dropped one data point, the year 1961, from the 52-year data set. This data point was treated as an outlier because sea level anomalies for 1961 are the lowest recorded though the rainfall is the second highest recorded during the 52 years. This anomalous behavior, however, suggests that interannual variability associated with winds also forces significant changes in sea level, especially at Vishakhapatnam; 1961 was a year of unusual winds in the Indian Ocean [Reverdin *et al.*, 1986].

Both salinity and the large-scale winds over the north Indian Ocean are important forcing mechanisms for the seasonal cycle of sea level along the coast of India as well as for the alongshore variations in annual mean sea level [Shankar, 1998]. Numerical experiments with a reduced gravity model show that a uniform increase in winds over the entire north Indian Ocean raises annual mean sea level along the east coast of India but lowers it along the west coast, the sea level behaving like a see-saw with its pivot at the southern tip of the Indian subcontinent [Shankar, 1998]. Since an increase in monsoon rainfall reflects an increase in monsoon intensity,

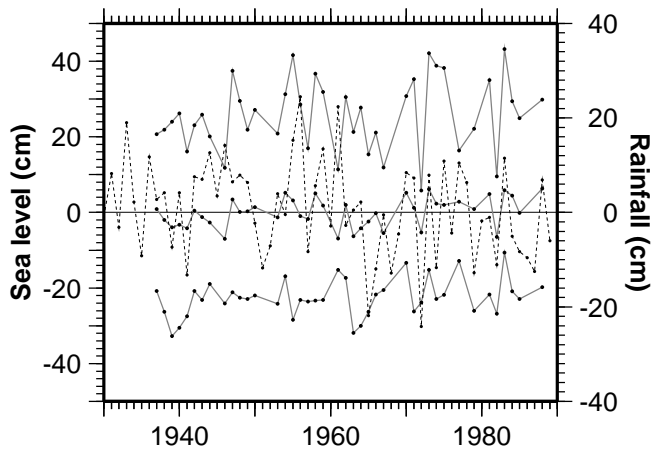


Figure 6. Anomalies of annual all-India rainfall (centimeters; dotted line) and of the annual maximum (upper solid line), mean (middle solid line), and minimum (lower solid line) sea level (centimeters) at Vishakhapatnam (the mean rainfall and sea level have been removed from the time series). For sea level the year is defined from July to June; for rainfall, the year is defined from January to December.

which is accompanied by an increase in the strength of the winds, this implies that winds and salinity act in concert to raise or lower sea level at Vishakhapatnam but oppose each other at Mumbai. Thus winds counter the effect of salinity (and monsoon rainfall) on the correlation between sea levels at Vishakhapatnam and Mumbai; winds are also the likely cause of the lower correlation between all-India rainfall and sea level at Vishakhapatnam.

The monsoon, however, is a complex phenomenon, and a strong monsoon is not necessarily accompanied by a uniform strengthening of winds over the basin. The linear model of *Clarke and Liu* [1994], which is forced by winds blowing over the equatorial Indian Ocean, shows that the interannual variability of sea level along the periphery of the north Indian Ocean is coherent, there being a similarity in the observed and simulated variability at several stations from southern Java to Mumbai. Though the data of *Clarke and Liu* [1994] are not sufficient to permit analysis at decadal timescales, it is possible that interdecadal changes in winds, either over the equatorial Indian Ocean or over the north Indian Ocean, also force interdecadal changes in sea level along the coast of India. Apart from salinity and winds, sea level is affected by ocean temperature and atmospheric pressure; these, however, vary too little in the tropics on these timescales to cause changes of the observed magnitude.

6. Discussion

We have shown that the interdecadal variability of sea level along the coast of the Indian subcontinent can be linked to the monsoon rainfall over the subcontinent, coastal salinity playing an intermediate role. The most striking feature of

this mechanism is the connection between two vastly different timescales, the seasonal and the interdecadal. The bulk of the rainfall over the Indian subcontinent occurs during June–September, and the runoff is carried to the seas around India within the same season. The equatorward EICC and the poleward WICC, which transport this low-salinity water along the Indian coast, are also seasonally varying currents. These seasonal phenomena act in concert with the low-frequency variability of monsoon rainfall and the slow mixing of salt in the oceans to force sea level changes on interdecadal timescales.

A change in the salinity of coastal waters caused by variations in monsoon rainfall, and hence in river runoff, creates a cross-shore density gradient to which the response is a two-layer geostrophic circulation with a surface current and an undercurrent. Though this background circulation, on which the seasonal circulation is superimposed, is very weak, the associated movements of the sea surface and the isopycnals are observable. When salinity decreases (increases) at an eastern ocean boundary like the Indian west coast, the current at the surface flows poleward (equatorward), there is an opposing undercurrent, sea level at the coast rises (falls), and the pycnocline, or the interface between the undercurrent and the deeper ocean, slopes up (down) toward the coast. Thus the interdecadal changes in sea level off the Indian coast can be linked to the dynamics of local coastal currents and to the variability in the rainfall over the Indian subcontinent.

This hypothesis, while linking interdecadal changes in sea level to those in the climate of the region, differs from those usually invoked in constructing scenarios of long-term changes in sea level. These also link interdecadal sea level variability to climate change [*Gornitz et al.*, 1982; *Wyrki*, 1990] but invoke the rise in temperature in the upper ocean and the melting of polar ice caps. It has been speculated, however, that global climate change is a complex process and will manifest differently in different regions; in the tropics, since ocean temperature is already high, it is expected to change evaporation and hence change atmospheric moisture, cloud cover, and rainfall [*Wyrki*, 1990]. The rainfall and sea level observations presented above highlight the complex relationship among the variables that define climate change. They do not, however, allow separation of natural, very low-frequency variability of monsoon rainfall and sea level from that caused by other effects, including the warming of the globe due to anthropogenic effects.

However, the absence of a reliable data set on spatial and temporal variability of salinity in the coastal region of India makes it difficult to verify the hypothesis. It is this that forces us to correlate monsoon rainfall over India, rather than coastal salinity, with coastal sea level.

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