Why is Bay of Bengal Warmer than Arabian Sea During the Summer Monsoon?

By

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Abstract: The near-surface Bay of Bengal remains significantly warmer than the Arabian Sea during summer monsoon (June-September). Analysis of the heat budgets of the near-surface Arabian Sea and Bay of Bengal shows significant differences between them during the summer monsoon. In the Arabian Sea, the winds associated with the summer monsoon are stronger and favour the transfer of heat to deeper layers owing to overturning and turbulent mixing. In contrast, the weaker winds over the bay force a relatively sluggish oceanic circulation that is unable to overturn, forcing a heat-budget balance between the surface fluxes and diffusion and the rate of change of heat in the near-surface layer.

Key words: North Indian Ocean, Sea Surface Temperature, Heat budget, Air-sea interaction, Monsoon

1. Introduction

Arabian Sea and the Bay of Bengal are similar in several aspects such as their location on the same latitude band, semi-enclosed nature, opening on the southern side and exposure to the changing monsoon winds etc. Both receive similar amounts of solar radiation at the top of the troposphere. In spite of this, there are striking dissimilarities between the two. First, the winds over the two basins are different especially during the summer monsoon as a consequence of the low-level jet, Findlater Jet, which occurs over the Arabian Sea (Findlater, 1969). Second, the precipitation exceeds evaporation in the bay while the evaporation exceeds precipitation in the Arabian Sea. Annual rainfall over the bay varies from 1 m off southeast India to more than 3 m in the Andaman Sea and the coastal region north of it (Baumgartner and Reichel, 1975) while it is barely 1 m over the Arabian Sea. In addition, the bay receives an annual runoff of 1.5×10^{12} m^3 from rivers flowing into it. Therefore, the surface layer in the bay is much fresher than that in the Arabian Sea; the average salinity of the top 50 m in the Arabian Sea exceeds that in the bay by nearly 3 PSU. The large inflow of freshwater from precipitation and runoff results in strong near-surface stratification in the Bay.

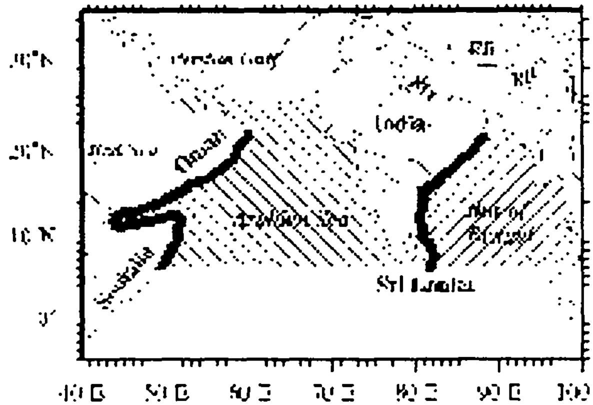

Figure 1. The geography of the north Indian Ocean. The two major rivers, Ganga and Brahmaputra, that debouch into the northern Bay of Bengal, are indicated by RG and RB respectively. The hatched areas show the two control volumes used for estimating the heat budgets; the southern boundary is fixed at 6^0 N, which roughly marks the southern tip of Sri Lanka, and the northern boundary at 25^0 N. The dark strips along the western boundary of the basins represent the coastal strips used for computing the contribution of coastal pumping to the heat budget.

These differences extend to the evolution of sea surface temperature (SST) in the two basins (Figure 2). Prior to the onset of the summer monsoon, during April-May, the north Indian Ocean becomes the warmest area among the world oceans (Joseph, 1990). Soon after the onset of the monsoon in June, the winds strengthen and SST decreases. The Arabian Sea cools rapidly, but SST in the bay remains higher than 28 ^{0}C, the threshold for deep convection in the atmosphere over tropical oceans (Gadgil et al.,1984; Graham and Barnett, 1987; Sud et al., 1999). This difference in SST is reflected in the convective activity in the atmosphere, with the convection over the bay being perhaps the largest in the tropics during summer. The number of low pressure systems (LPSs) that form there by far exceeds that in the rest of the north Indian Ocean (Mooley and Shukla, 1989). These LPSs move westward over India (Sikka:1977; Goswami, 1987; Mooley and Shukla, 1989) bring rainfall to central and northern India.

In this paper we investigate the processes that ensure that bay remains warm (SST > 28 ^{0}C) during the summer monsoon through an analysis of the heat budget of the upper ocean. This is done using climatologies of air-sea fluxes, radiative fluxes, winds, and hydrography.

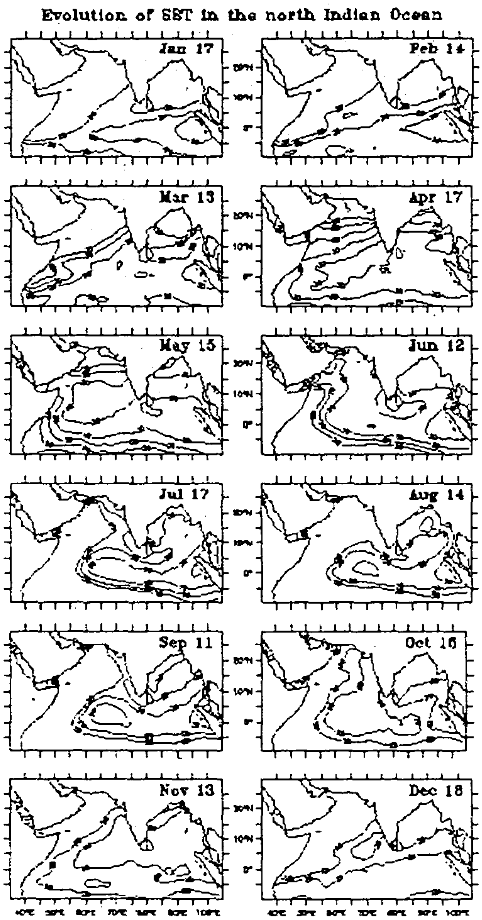

Figure 2. Evolution of SST in the North Indian Ocean. Weekly SST data Reynolds and Smith (1994) for 1982-1998 were combined to form a weekly climatology. The data set has a spatial resolution of $1^0 x 1^0$. Contours are drawn only for SST greater than the deep-convection threshold of 28 ^{0}C; contour interval is 1 ^{0}C.

2. Heat budget of the upper ocean

To determine the causes of the higher SST in the Bay of Bengal, we analyze the heat budget of the upper ocean. We define the region of interest (Figure 1) to exclude the area south of 6^0 N, which marks the southern tip of the Indian subcontinent and the area north of 25^0 N. To define the control volume completely, we have to fix its lower boundary; we fix this at 50-m based on the observed mixed-layer depths in the two basins. The details of the computation of heat budget are available in Shenoi et al. (2002). Essentially, the heat budget is computed based on the assumption that the rate of change of heat in the control volume is balanced by the fluxes of heat through its boundaries owing to advective and non-advective processes. The control volume is heated by the surface fluxes. The diffusion of heat through the bottom together with the advective fluxes is called oceanic processes because these are intrinsic to the ocean.

A refined version of the Comprehensive Ocean-Atmosphere Data Set (COADS) compiled by Josey et al. (1996) having a resolution of 1^0x1^0 is used to estimate the heat and radiative fluxes at the surface. This data set is preferred to a similar climatology based on COADS compiled by Oberhuber (1988) and to a climatology derived from the NCEP/NCAR Reanalysis programme (Kalnay et al.,1996) because it is more recent, uses better methods for generating the climatology, and has been compared extensively with observations (Josey et al., 1999; Weller et al., 1999).

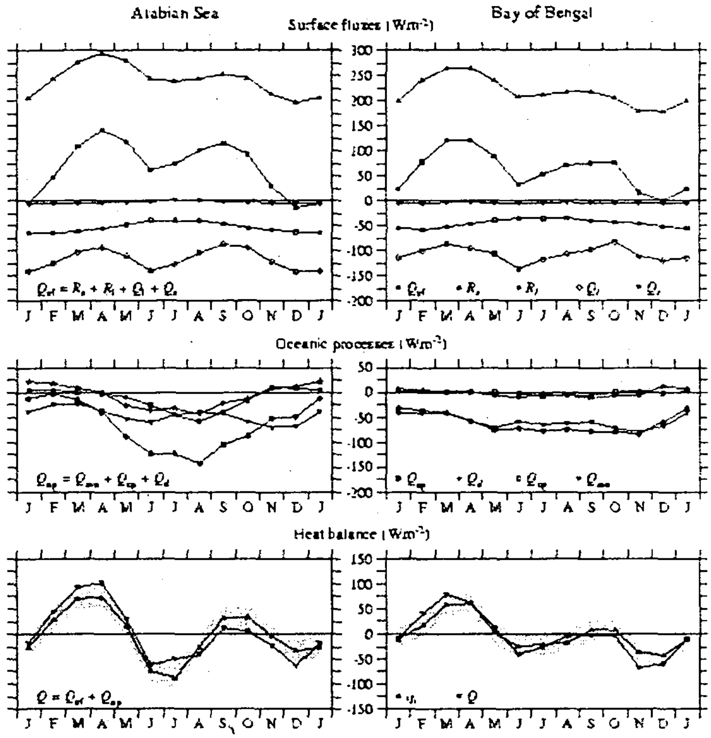

Figure 3. The heat budget of the upper ocean in the Arabian Sea (left) and the Bay of Bengal (right). The top panel shows climatological monthly-mean net shortwave radiation (R_s), net longwave radiation (R_l), latent heat flux (Q_l), and sensible heat flux (Q_s), and their sum, the net surface flux (Q_{sf}). The middle panel shows the fluxes due to oceanic processes: meridional overturning (Q_{mo} = Q_{moe}+Q_{mog}), coastal pumping (Q_{cp} =

$Q_{cpe}+Q_{cpg}$), and diffusion (Q_d) the sum of the fluxes due to these processes (Q_{op}) is also shown. The bottom panel shows the rate of change of heat (q_t) in the control volumes and the net flux of heat ($Q = Q_{sf}+Q_{op}$) into or out of them. The estimated error on (q_t is 23 Wm^{-2}), based on the standard error of (0.3 ^0C) on the annual-mean temperature (Levitus and Boyer, 1994), is also shown (grey shade). The southern (northern) boundary of the control volumes is at (6^0 N) (25^0 N) and their depth is 50 m. The units are Wm^{-2} and a positive flux implies that the control volumes gain heat.

Advective processes considered include the meridional overturning cell that exports heat out of the north Indian Ocean (Lee and Marotzke, 1997; Garternicht and Schott, 1997) and the cross-shore flow at the western boundary layer resulting in upwelling or downwelling in the western-boundary regime (coastal pumping). The coastal pumping also causes overturning and removes heat from the control volume, but, unlike meridional overturning, it does not remove heat from the north Indian Ocean. Vertical mass flux at the northern and eastern boundaries is ignored in comparison to that at the western boundary. Also ignored is the contribution of the negligible flow between the Red Sea and Persian Gulf and the control volume defining the Arabian Sea (Duing and Leetmaa, 1980).

The 10-day repeat-cycle data of sea-level anomalies from the TOPEX/Poseidon altimeter during 1993-1997 (Le Traon et al.. 1998) together with annual mean profiles of temperature (Levitus and Boyer. 1994) and salinity (Levitus et al., 1994) are used to estimate the climatological, monthly-mean geostrophic current at the southern boundary. The zonal component of wind stress (Josey et al., 1996) is used to estimate the transport due to Ekman processes. Similarly, the alongshore component of the wind stress and the sea-level anomalies from the altimeter are used to estimate the coastal pumping along the coastal strip (represented by the dark coastal strip in Figure 1). The monthly mean profiles of temperature (Levitus and Boyer, 1994) are used to compute the heat flux through the bottom of the control volume owing to diffusion. Again, the same temperature profiles are used to estimate the rate of change of heat content in the control volume.

The heat budget of the Arabian Sea and Bay of Bengal on a monthly mean time scale is shown in Figure 3. In the Arabian Sea, the dominant oceanic processes are coastal Ekman pumping, meridional overturning due to Ekman flow, and diffusion. Of these, the first two, which are directly influenced by the winds and cause overturning, are important primarily during the summer monsoon, when they remove heat from the control volume. Diffusion, the only oceanic process not directly influenced by the winds, is important throughout the year. In the bay, these wind-forced processes have a minor impact on the heat budget; diffusion overwhelms other oceanic processes.

The reason for this lies in the asymmetry of the wind field in the north Indian Ocean. The weaker winds over the bay force a relatively sluggish oceanic circulation there, making it difficult for overturning or coastal pumping to remove heat from the control volume there. Thus, in the Arabian Sea, the heat gained at the surface is balanced by the rate of change of heat in the control volume and by overturning, coastal pumping, and diffusion. In the bay, the heat gained at the surface is balanced by the rate of change of heat in the control volume and the loss of heat by diffusion. It is the difference in the structure and magnitude of winds that keeps the mean temperature of the top 50 m of the bay warmer than that of the Arabian Sea.

Acknowledgments

We thank Simon Josey for making available the Southampton Oceanography Centre climatology and the CLS (Collecte Localisation Satellites) Space Oceanography Division, Paris for the altimeter products. Ferret, GMT, and the GIMP were used for analysis and graphics. The Department of Ocean Development, New Delhi provided financial support.

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