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Surface freshwater from Bay of Bengal runoff and Indonesian throughflow in the tropical Indian Ocean.

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Abstract. According to recent estimates, the annual total continental runoff into the Bay of Bengal (BoB) is about 2950 km^3, which is more than half that into the entire tropical Indian Ocean (IO). Here we use climatological observations to trace the seasonal pathways of near surface freshwater from BoB runoff and Indonesian throughflow (ITF) by removing the net contribution from precipitation minus evaporation. North of 20°S, the amount of freshwater from BoB runoff and ITF changes with season in a manner consistent with surface currents from drifters. BoB runoff reaches remote regions of the Arabian Sea; it also crosses the equator in the east to join the ITF. This freshwater subsequently flows west across the southern tropical IO in the South Equatorial Current.

1. Introduction

Five of the world's 50 largest rivers, the Brahmaputra, Ganga, Irrawaddy, Godavari and Mahanadi, flow into the Bay of Bengal (BoB) north of 15°N. Seasonal monsoon rain and river runoff create shallow pools of freshwater in the north Bay in summer (*Gopalakrishna et al.* [2002]; *Vinayachandran et al.* [2002]) and autumn. Observations often show an intense, shallow halocline at 5-20 m depth, and another halocline at about 30 m (Figure 1). Freshwater from the north Bay flows down the east coast of India in October-December in the East India Coastal Current (EICC; *Shetye et al.* [1996]). Some of this water appears to flow around Sri Lanka, carrying nutrients into the near surface eastern Arabian Sea (*Prasanna.Kumar et al.* [2004]). Later in winter, water from the southern Bay flows past Sri Lanka in the Northeast Monsoon Current (NMC) (*Gopalakrishna et al.* [2005]). In addition to the western pathway in winter identified from observations (*Donguy and Meyers* [1996]; *Rao and Sivakumar* [2003]), model simulations (e.g. *Han et al.* [2001]) and experiments using tracers/drifters suggest that BoB water enters the tropical IO to the east and south in the second half of the year, crossing the equator mainly in the east (*Jensen* [2003]; *Miyama et al.* [2003]).

Surface freshwater influences the upper ocean via several dynamic and thermodynamic processes: It can change the upper ocean velocity field by creating dynamic height gradients (*Gordon et al.* [2003]; *Han et al.* [2001]); salinity stratification influences the mixed layer depth and vertical penetration of surface momentum flux into the upper ocean, modifying velocity shear and entrainment rates (*Howden and Murtugudde* [2001]; *Han et al.* [2001]); if the mixed layer is very shallow, penetrative shortwave flux can be comparable to (or even exceed) surface net heat flux (*Godfrey et al.* [1998]; *Sengupta and Ravichandran* [2001]); sub-mixed layer water, isolated from evaporative and longwave cooling, is warmed by penetrative sunlight (*Sengupta et al.* [2002]). Models agree that BoB river runoff shallows the mixed layer in many regions of the tropical IO, and thickens the barrier layer, although there is no consensus on the effect of runoff on sea surface temperature (*Howden and Murtugudde* [2001]; *Han et al.* [2001]). This could be due to differences in the treatment of mixing and penetration of sunight between models, as well as surface fluxes. Although model results are instructive, models generally have large mixing, and cannot produce sufficiently strong and shallow haloclines in the BoB. If a shallow halocline resists diapycnal mixing, BoB surface water should continue to stay relatively fresh as it travels to remote regions of the tropical IO. Thus it is likely that most present day models underestimate the true reach and impact of BoB freshwater.

The use of upper ocean salinity observations to study the movement of freshwater is problematic because salinity is influenced by seasonally varying rain (P) and evaporation (E) (*Donguy and Meyers* [1996]; *Rao and Sivakumar* [2003]). Here we propose a simple use of observations to trace the seasonal pathways of BoB surface freshwater by removing the influence of P-E in the upper tropical IO. Our main results are based on four climatological datasets - seasonal temperature and salinity from the World Ocean Atlas 2001 (WOA01)(*Boyer et al.* [2002]); monthly 1°by 1°precipitation from the Global Precipitation Climatology Project (GPCP), based on blended satellite and gauge data (*Huffman et al.* [1997]); monthly evaporation from the Southampton Oceanography Centre (SOC) data (*Josey et al.* [1998]), and monthly 2°by 2°surface currents in the tropical Indian Ocean, based on 1985-2002 trajectories of drogued WOCE drifters (*Shenoi et al.* [1999]), which represent depth averaged currents in the upper 15 m. We also use a monthly 1°by 1°dataset on continental runoff into the ocean, based on gauge data and a river network and water balance model (*Dai and Trenberth* [2002]), to examine BoB freshwater balance.

2. Bay of Bengal Freshwater

In addition to several major rivers, numerous smaller streams discharge into the Bay of Bengal. The total annual continental runoff into the Bay is 2950 km^3, obtained by integrating the *Dai and Trenberth* [2002] data based on *Fekete et al.* [2002], along the BoB coastline north of 6°N. This is about 60 percent of the total runoff into the entire tropical IO north of 30°S. Integration over the outflow regions gives total annual runoff of 1300 km^3 for the

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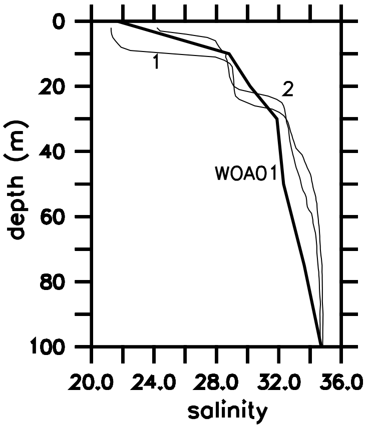

Figure 1. Salinity profiles in the north Bay at two stations from October 2003 cruise of ORV Sagar Kanya (thin lines marked 1 and 2), and October profile from WOA01 (bold line). All profiles are from 89°-90°E, 19°-20°N.

Brahmaputra-Ganga-Meghna river system, and 1000 km^3 for the Irrawaddy and Salween together (Supplementary Figure S1). Discharge at river mouths can be considerably larger than purely gauge-based estimates because some tributaries may be unmonitored, and gauges are often located far upstream of river mouths. For example, the sum of the Brahmaputra and Ganga flow from the gauges at Bahadurabad (north Bangladesh) and Farakka (India) is about 1000 km^3/year (*Dai and Trenberth* [2002]). There appears to be no available long term gauge data for the upper Meghna or the Salween. We estimate the amount of freshwater in the upper 30 m of the tropical IO from the WOA01 seasonal dataset, using a reference salinity S_o(x,y) defined as the spatially varying annual mean salinity of the upper one kilometre. The upper ocean freshwater content, i.e. the total freshwater (in meters) from the surface to H=30 m, FW(x,y,t), is then computed with respect to the reference salinity S_o :

$$FW(x, y) = \int_0^H \alpha dz, \tag{1}$$

where $\alpha = (\rho_f - \rho)/(\rho_f - \rho_s)$; ρ is the density of seawater with observed temperature (T) and salinity (S); ρ_f is the density of water with S=0 and temperature T; and ρ_s is the density of water with salinity S_o and temperature T. H is chosen as 30 m in order to account for most of the freshwater in the upper ocean (see Figure 1). The seasonal evolution of FW is shown in Supplementary Figure S2. We first examine upper ocean freshwater content integrated over the Bay of Bengal, in response to forcing by P-E and runoff R (Figure 2). The annual total GPCP precipitation over the Bay is 4700 km^3, and SOC evaporation is 3600 km^3. Precipitation and runoff are largest in June-October. Although surface windspeed is highest in July-August, evaporation does not have a maximum at this time because the sea-air difference of temperature and specific humidity is small (e.g. *Josey et al.* [1998]). Net freshwater input to the Bay $\int\int(P+R-E)dxdy$ is negative in January-March, and largest in June-August (Figure 2b) in agreement with the estimates of *Rao and Sivakumar* [2003]. The annual total freshwater input is 4050 km^3 (about 1.6 m/year). WOA01 data shows that in response to the forcing, storage of freshwater in the upper 30 m ($\int\int(FW)dxdy$) increases from about 4200 km^3 in May to a maximum of 5700 km^3 in November, three months after the peak input (Figure 2b). In order to balance the net input of 4050 km^3/year, the net southward transport of freshwater out of the Bay of Bengal (across 6°N, 80°-100°E) must be 0.13 Sverdrup. This large transport of freshwater, of possible climatic significance (*Wijffels* [2001]), has not received much attention (*Gordon* [2001]). Note, that the depth of the freshwater transport, or its seasonal cycle, cannot be inferred from available climatological observations.

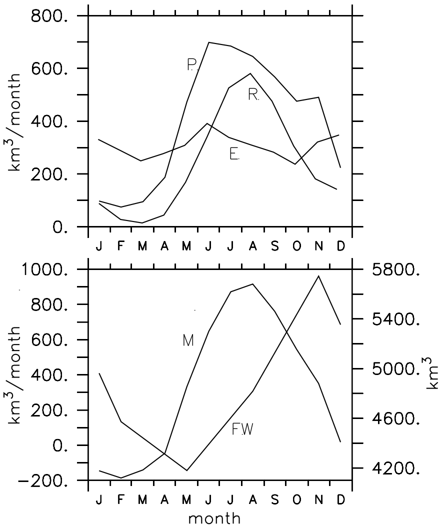

Figure 2. (a) Seasonal cycle of precipitation (P, from GPCP), evaporation (E, from SOC), and continental runoff R integrated over the BoB, in km^3 per month, and (b) seasonal cycle of net freshwater input (M = P-E+R, left axis) in km^3 per month, and mixed layer freshwater content in the Bay (FW) in km^3 (right axis).

3. Bay of Bengal runoff in the tropical Indian Ocean

Seasonal changes in freshwater content FW are due to runoff, local as well as advected P-E, and flux of saltier water across the 30 m level (see below). In order to remove the contribution from local as well as remote P-E, we take the FW field for each month and subtract from it the time integral of P-E, and horizontal advection of P-E, over the previous three months. Drifter currents are used in the calculation of horizontal advection. Our recipe generates monthly fields of upper ocean freshwater from runoff, RW(x,y,t), which represents freshwater “imported” to the tropical IO across lateral boundaries. The choice of three-month integration of P-E is based on the response time of the Bay to freshwater forcing (Figure 2b). The spatial patterns of RW are not sensitive to the use of four-month rather than three-month integrals of P-E. Apart from unknown errors arising from the sparse sampling of WOA01 salinity, the greatest uncertainty in RW comes from differences between rainfall climatologies (*Wijffels* [2001]; *Yu and McCreary* [2004]). We have therefore compared RW maps based on CMAP (*Xie and Arkin* [1997]) and GPCP rain (which is considered a more reasonable estimate over the BoB than CMAP, in *Yaremchuk et al.* [2005]). Although magnitudes differ by over 20%, the broad spatial patterns of P in the two climatologies are similar, leading to an uncertainty in monthly RW within 0.2 m over most of the tropical IO. We have also considered evaporation from COADS climatology (*DaSilva et al.* [1994]), but the difference between COADS and SOC annual evaporation integrated over the tropIO is less than 2%. In other words, the

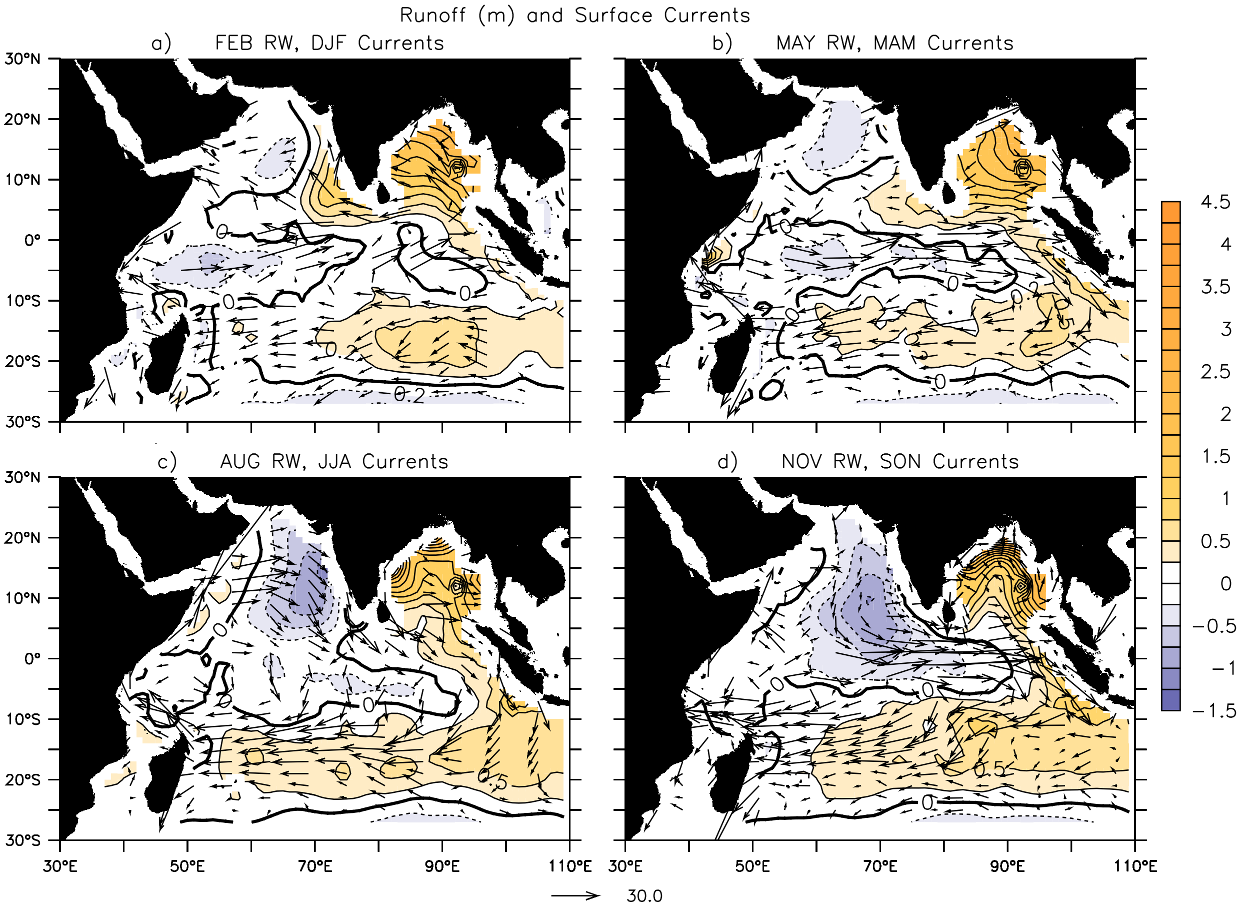

Figure 3. Monthly maps of RW in metres, and upper 15 m mean currents averaged over three months. (a) February (b) May (c) August and (d) November. December-February mean currents are overlaid in (a), and so on. Positive RW is shaded with the largest values being darkest; zero RW contour is bold. The arrow represents current speed of 30 cms^{-1}

spatial patterns of RW do not appear to be highly sensitive to the combination of rainfall and evaporation products chosen. We also find that the broad patterns of RW do not change if H is taken to be mixed layer depth (surface density plus 0.5 kgm^{-3} criterion, with 30 m minimum), rather than uniform 30 m. Note that uncertainty in runoff does not influence the RW estimates.

Monthly maps of RW (Figure 3) show the spread of BoB continental runoff, as well as near surface freshwater from the ITF (*Gordon* [2001]). In regions of negative RW, such as the Arabian Sea (AS), upper ocean salinity is higher than S_o(x,y) even after the influence of negative seasonal P-E is removed. The reasons are discussed later. Three-month mean (present month and previous two months) drifter current vectors (*Shenoi et al.* [1999]) are overlaid on RW. In February, there is about 1.5 m of RW in the upper BoB (Figure 3a), and 0.5 m off the west coast of India (WestCI). WestCI winter runoff is minimal, so this water must come from the Bay. Other observational studies find BoB surface water at least as far north as 15°N along the WestCI at this time, with surface salinity increasing gradually to the north (*Donguy and Meyers* [1996]; *Prasanna.Kumar et al.* [2004]). However, models have difficulty getting BoB water to move up the WestCI (*Han et al.* [2001]; *Jensen* [2001]). Bay runoff is carried far to the west in the southern AS by the NMC (Figure 3a, 3b), in agreement with some model results (*Han et al.* [2001]; *Jensen* [2001]; *Jensen* [2003]). Satellite sea surface height (SSH) and *in situ* observations (*Bruce et al.* [1998]) show one or more vortices with dimension of O(1000 km), generated in the southeastern AS in January-April, moving west at 15-20 cms^{-1}. Analyses of SSH also suggest the presence of several smaller (O(100 km)) eddies. It is possible that BoB freshwater is carried west in these vortices, in addition to the NMC. As much as 3 m of runoff accumulates in the north Bay by November. By this time, BoB runoff, carried equatorward in the EICC, has begun to flow past Sri Lanka and up the west coast of India (Figure 3d).

The signature of runoff exported from the Bay to the south and east is clear by August (Figure 3c); it appears to join ITF surface water and flow west in the surface South Equatorial Current (SEC). Back trajectories from the drifter velocity field (not shown) suggest that some water in the eastern SEC originates just south of the Bay. This pathway has not been reported in previous observational studies. Model results (*Jensen* [2001]; *Han et al.* [2001]; *Miyama et al.* [2003]) show that export of BoB mixed layer water to the south takes place via an eastern pathway in the second half of the year. However, BoB water appears to reach further south in our maps than in models (e.g. *Jensen* [2003]; Raghu Murtugudde, Personal Communication, 2005). The largest values of RW in the eastern SEC are found in September-November. This may be partly due to the (poorly known) seasonal cycle of near surface flow in the ITF.

The origin of negative RW in the Arabian Sea requires comment, because it clearly does not represent runoff (which is positive). The mean mixed layer depth in the AS is shallowest (about 30 m) in April-May due to thermal stratification, as well as salinity stratification in the southeastern sector. A subsurface salinity maximum (about 36.2) lies at 40-100 m depth in the central AS at this time (*Boyer et al.* [2002]). By August, when the mixed layer deepens to about 80 m, entrainment of saline water and advection from the north decreases RW to -0.5 m or less in this region. Thus negative RW is due to the flux of saltier water through the base of the upper layer (whether H is equal to 30 m or mixed layer depth). This flux is controlled by the combined action of vertical advection and mixing, as well as horizontal advection of salinity at deeper levels, both of which are unknown. In order to check if deep advection affects our results, seasonally varying reference salinity S_o was used instead of annual mean S_o. The patterns of RW are not sensitive to this change of S_o. If S_o is chosen to be 35, however, the values of RW do change significantly, specially in the AS where the differences are 0.2-0.5 m. Nevertheless, the only significant change in the spatial pattern is that the maximum westward extent of positive RW in the southern AS (Figure 3a,b) shrinks by about 10°of longitude. Note that flux through the base of the upper layer can mask the presence of freshwater from runoff. Therefore our estimates of runoff (positive values of RW) in the AS are likely to be too low. Outside the AS, negative RW does not seem to be associated with fluxes through the base of the upper ocean.

Advection of AS mixed layer water by southeastward drift and the eastward Summer Monsoon Current appears to push BoB runoff (zero RW contour) to its easternmost longitude off the southern tip of India in August-September (Figure 3c). It is likely that ingress of AS water into the southern Bay takes place mainly below 15 m depth, in accord with previous studies (*Rao and Sivakumar* [2003]; *Jensen* [2001]). Arabian Sea water moves far to the east along the equator in the Wyrtki jets (Figure 3b, 3d), in agreement with models (e.g. Figure 3 of *Jensen* [2003]; *Han et al.* [2001]).

The tropical Indian Ocean gains freshwater from the ITF, continental runoff, and rain. Unlike rainfall, north IO runoff is highly localised in space (mainly BoB) and time (mainly summer). This enables the seasonal pathways of surface freshwater to be traced with reasonable clarity (Figure 3). Although our results are qualitative, they demonstrate that BoB freshwater reaches remote regions of the tropical IO on seasonal time scales. In particular, BoB runoff appears to join the ITF and flow west in the SEC. It is important to note that the signature of ITF and BoB "runoff" in the southern tropical IO is revealed only after the influence of P-E is removed (compare Figure 3 and Supplementary Figure S2). However, the present approach is not suited to study the transport of BoB or ITF water in the southwestern IO, or in the Somali Current (Figure 3c), because the relevant time scales are much longer than one season (*Jensen* [2001]; *Song et al.* [2004]). Finally, riverwater carries nutrients, and its isotopic composition differs from rainwater. Therefore some of our findings may be relevant to biogeochemistry and palaeoclimate.

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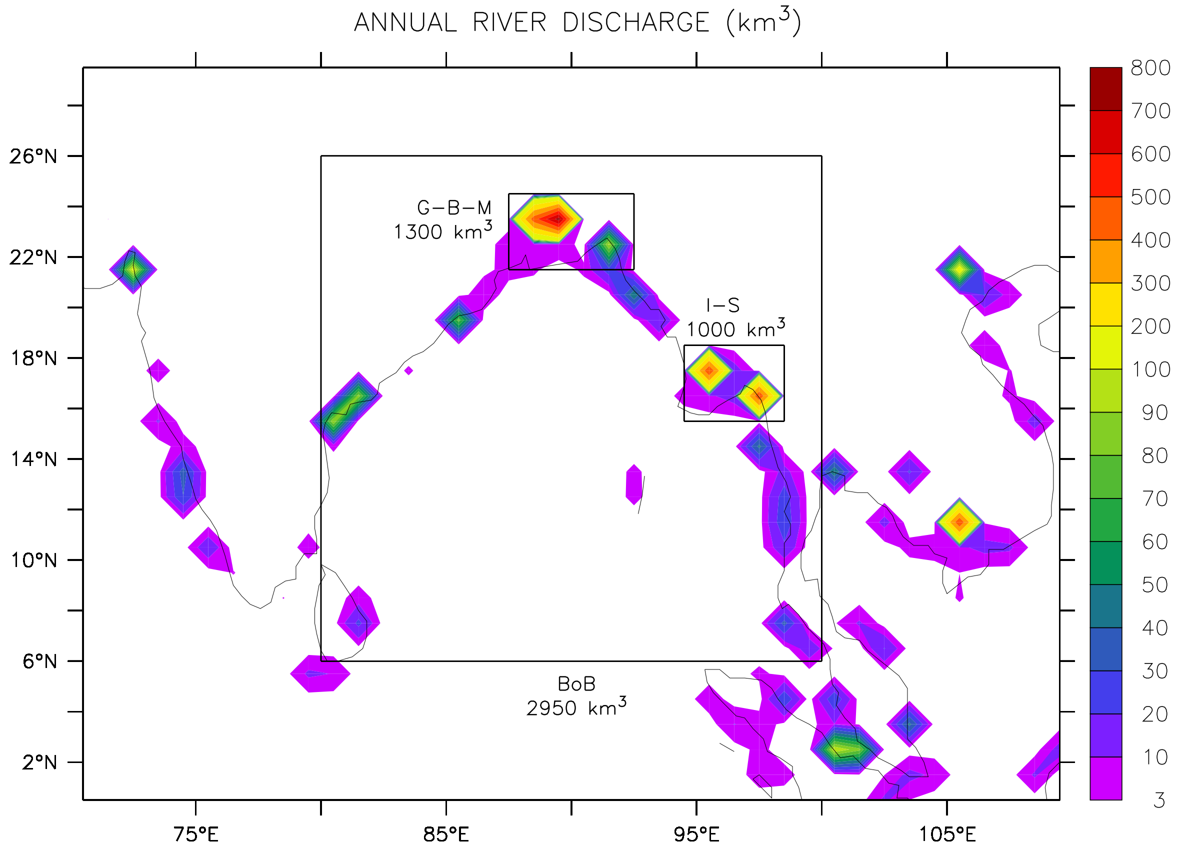
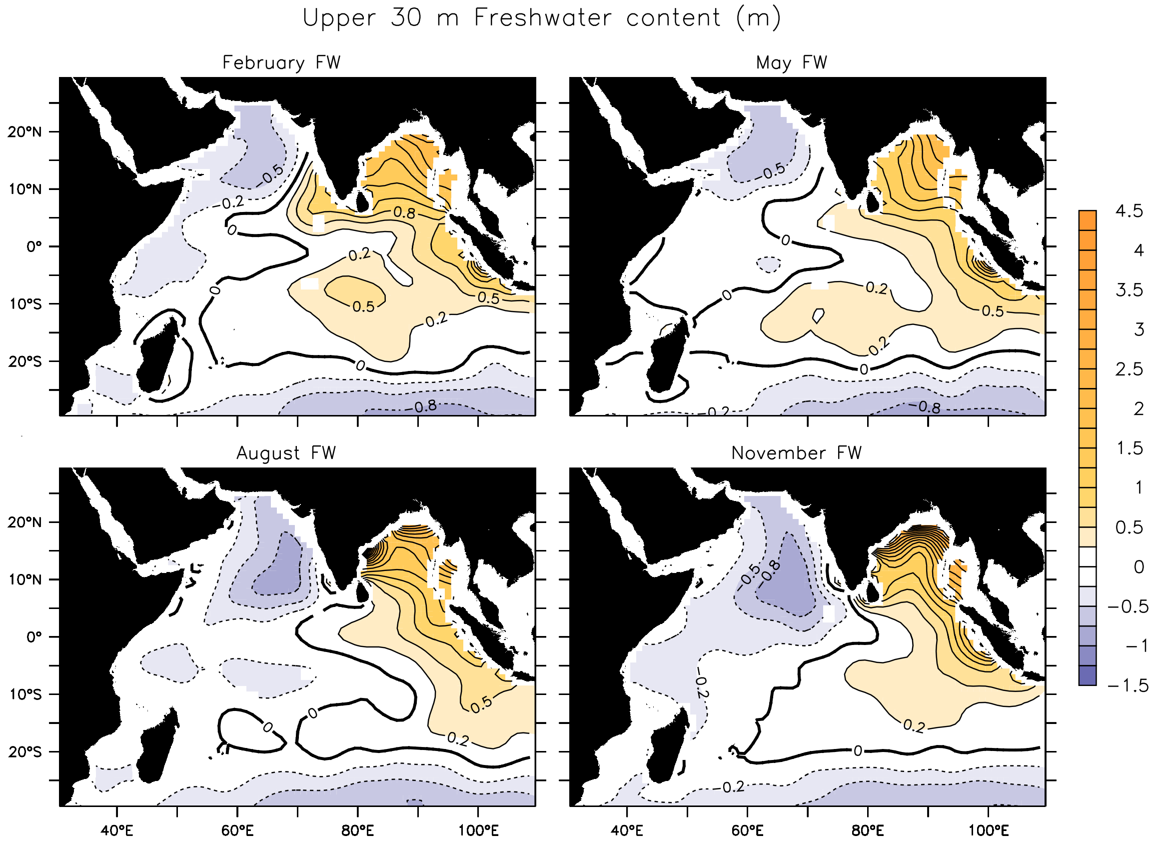


Supplementary Figure S1. Annual total continental runoff in km^3 based on estimates of Fekete et al (2002). BoB = Bay of Bengal, G-B-M = Ganga-Brahmaputra-Meghana, I-S = Irrawady-Salween. The annual total runoff, estimated by area integration over the respective boxes, is mentioned alongside each box.



Supplementary Figure S2. Freshwater content (in meters) in the upper 30 m of the tropical Indian Ocean, with a reference salinity S_o equal to the annual mean salinity averaged over the upper 1000 m.