Mechanisms of Mixed-Layer Salinity Seasonal Variability in the Indian Ocean

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Abstract Based on a joint analysis of an ensemble mean of satellite sea surface salinity retrievals and the output of a high-resolution numerical ocean circulation simulation, physical processes are identified that control seasonal variations of mixed-layer salinity (MLS) in the Indian Ocean, a basin where salinity changes dominate changes in density. In the northern and near-equatorial Indian Ocean, annual salinity changes are mainly driven by respective changes of the horizontal advection. South of the equatorial region, between 45°E and 90°E, where evaporation minus precipitation has a strong seasonal cycle, surface freshwater fluxes control the seasonal MLS changes. The influence of entrainment on the salinity variance is enhanced in mid-ocean upwelling regions but remains small. The model and observational results reveal that vertical diffusion plays a major role in precipitation and river runoff dominated regions balancing the surface freshwater flux. Vertical diffusion is important as well in regions where the advection of low salinity leads to strong gradients across the mixed-layer base. There, vertical diffusion explains a large percentage of annual MLS variance. The simulation further reveals that high-frequency small-scale eddy processes primarily determine the salinity tendency in coastal regions (in particular in the Bay of Bengal) and shear horizontal advection, brought about by changes in the vertical structure of the mixed layer, acts against mean horizontal advection in the equatorial salinity frontal regions. Observing those latter features with the existing observational components remains a future challenge.

1. Introduction

In general terms, both temperature and salinity control the density of sea water. This holds especially near the surface, where buoyancy fluxes are active as part of the exchange of heat and freshwater between the ocean and atmosphere. Heat and freshwater fluxes at the surface influence the interaction between ocean and atmosphere; at the same time, they control the dynamics in the ocean’s surface mixed layer. Detailed knowledge about the surface and mixed-layer dynamics and their forcing in terms of salinity and temperature is therefore of high interest.

In the past, it was difficult to approach the problem of surface buoyancy forcing of the ocean and resulting mixed-layer variations in a holistic manner because neither surface freshwater fluxes nor high-quality surface salinity data were available with sufficient spatial and temporal coverage. Due to the lack of both, it was therefore assumed that temperature essentially controls the evolution of the mixed layer. However, through modern salinity observations, the notion emerges that this is not universally true and that salinity might instead play a dominant role at least in some regions. This holds especially in low latitudes where intense precipitation can control the haline stratification and its seasonal and interannual change (e.g., Ando & McPhaden, 1997; Lukas & Lindstrom, 1991; Maes & O’Kane, 2014; Sprintall & Tomczak, 1992). Based on hydrographic climatologies, Donguy and Meyers (1996), Boyer and Levitus (2002), and Rao and Sivakumar (2003) investigated the seasonal variability of the mixed-layer salinity (MLS) and compared its amplitude and phase with evaporation minus precipitation (E-P) and horizontal salinity advection. Along the same lines, Yu (2011) investigated processes contributing to seasonal salinity variability and identified E-P as a dominant process. However, from an analysis of 13 year long ECCO ocean synthesis fields, Vinogradova and Ponte (2013a) concluded that salinity is regionally an adequate proxy for surface freshwater flux, but not on global scales.

To illustrate where salinity dominates the seasonal changes of the mixed-layer density, Figure 1 shows the ratio, as in Dong et al. (2009), between the annual amplitudes of the mixed-layer density due to the seasonal...
In agreement with Figure 1, previous studies (e.g., Nyadjro & Subrahmanyam, 2014; Rao, 2015; Sakova, 2010; Schott et al., 2009) indicated that the tropical Indian Ocean shows enhanced spatial and temporal variability in the near-surface salinity compared to the Atlantic and Pacific Oceans. This is caused by the seasonally varying monsoonal climate, which dominates the northern Indian Ocean, leading to large seasonal variations in ocean currents. In this context, the Southwest (or summer) monsoon determines the climate of the northern Indian Ocean during boreal summer (July/August). Asian landmasses heat up and generate a low-pressure area over the continent resulting in strong winds from the Southwest, which drive the eastward Southwest Monsoon Current. In boreal winter, the Indian Ocean dynamics are determined by the Northwest or winter monsoon—winds blowing from the northeast—which is followed by the current reversal from the Southwest Monsoon Current to the westward Northeast Monsoon Current. These seasonal current reversals dominate the water mass change north of 4°N.

Low-salinity waters flow from the Bay of Bengal—where precipitation and river runoff exceed evaporation—into the Arabian Sea, characterized by high salinity due to enhanced evaporation, during the winter monsoon and reduce the salinity amplitudes there (e.g., Rao, 2015; Rao & Sivakumar, 2003). During the summer monsoon reversed conditions are present, when the Southwest Monsoon Current brings high-salinity waters from the Arabian Sea to the east. Brandt et al. (2002) and Shankar (2004) showed that the large annual amplitude in the southern Arabian Sea between 6°N and 10°N is due to the westward propagation of annual Rossby waves, which originate from the western coast of India, forced in the Bay of Bengal as well as by the action of wind stress curl over the Arabian Sea. Upwelling-favorable winds along the west coast of the Arabian Sea (Somalia, Yemen, and Oman) start with the onset of summer monsoon in May and end in September. Along the equator, high-salinity waters extend from the Arabian Sea to the eastern Indian Ocean and, south of approximately 10°S, low-salinity waters originating from the Indonesian Throughflow extend westwards toward Madagascar.

Most previous studies of the Indian Ocean focus on the analysis of climatological salinity fields, model output, and measurements made during observational campaigns. Da-Allada et al. (2015), for example, analyzed the dominant contributions to seasonal mixed-layer salinity (MLS) variability in the highly variable regions in the western tropical Indian Ocean using in situ MLS data showing the dominance of horizontal advection in the southern Arabian Sea and the importance of surface freshwater flux south of the equator. There, the salinity budget could not be closed exactly due to unexplained processes like vertical mixing. Zhang et al. (2016) analyzed the MLS variability in the southeast tropical Indian Ocean using in situ measurements as well as model data concluding that surface freshwater fluxes are the dominant contribution to seasonal MLS changes, but also advection of low-salinity waters within the South Java Current, Indonesian Throughflow and the Leeuwin Current contribute to the observed seasonal salinity changes. The authors found also that on interannual time scales, local precipitation events and transport anomalies by the currents contribute to MLS anomalies.

Only a few studies took advantage of the newly available satellite-retrieved salinity observations for calculating MLS budgets, which potentially can reveal many more details than heavily smoothed in situ data (this...
holds also for Argo data). Using in situ and satellite measurements together with several model simulations, D’Addezio et al. (2015) examined the seasonal salinity and salt transport variability in the monsoonal northern Indian Ocean, concluding that satellite data capture finer salinity structures than Argo does, but that satellite errors and the low resolution of Argo data in coastal regions make model simulations necessary when analyzing near-coastal current systems. Further satellite-based studies of sea surface salinity (SSS) in the Indian Ocean focused on the variability during Indian Ocean Dipole (IOD) events (Du & Zhang, 2015; Grunseich et al., 2011; Nyadjo & Subrahmanyam, 2014). All these studies show that the agreement of satellite-retrieved and in situ salinity fields is encouraging, but large salinity budget imbalances remain due to unexplained processes or errors in the contributing terms (approximately 0.5 month$^{-1}$ during positive IOD in 2012; see Nyadjo & Subrahmanyam, 2014, their Figure 5). One reason for large MLS budget imbalances using satellite SSS could be due to near-surface salinity stratification (Song et al., 2015), which happens in regions of strong and highly variable precipitation events. Boutin et al. (2013) showed that the 20% of the bias between 10 day averaged satellite and uppermost in situ SSS can be explained by the effect of precipitation. Boutin et al. (2014) analyzed the precipitation signature using monthly averaged SMOS fields and interpolated drifter measurements near the ITCZ and found that even 40% of the bias comes from rain and the resulting near-surface stratification.

Besides unrepresented physical processes, satellite SSS errors due to land contamination and RFI also contribute to MLS balance residuals in near-coastal areas (D’Addezio et al., 2015; Köhler et al., 2015). Sampling errors are likely to occur due to the large spatial and temporal SSS variability near the coast and off major river outlets. Associated processes can be linked to small-scale variability and again to near-surface vertical salinity gradients (e.g., Köhler et al., 2015; Song et al., 2015; Vinogradova & Ponte, 2013b).

In the present study, we build mainly upon the analyses of Yu (2011) and Vinogradova and Ponte (2013a) and further investigate details of the Indian Ocean mixed-layer salinity changes as revealed by satellite-retrieved salinity fields and in situ data and compare results with the output from a high-resolution ocean circulation model. We show to what extent satellite SSS fields are usable as a proxy for MLS variations and help quantifying the errors coming from unconsidered processes in past observational MLS budget studies. Our specific focus is on processes neglected in previous observational studies of the seasonal cycle of MLS including vertical diffusion, eddy advection, and vertical variations of the mixed-layer structure. We study these last two processes at submonthly time scales. We focus on a larger region than the previously discussed studies to provide a more comprehensive view on the MLS variability and its seasonal cycle in the Indian Ocean. We furthermore employ a high-resolution eddy-resolving model, which allows us to discuss the contribution of small-scale and high-frequency processes. This is in contrast to previous modeling studies (e.g., Akhil et al., 2014; Moon & Song, 2014), which rely on eddy-permitting simulations and therefore cannot be used to estimate the full effect of the mesoscale circulation, especially in near-coastal and equatorial Indian Ocean regions.

The remainder of the paper is organized as follows: in section 2, a brief description of the ensemble-mean satellite SSS product and all auxiliary data sets used in this work is provided together with a description of the model. In section 3, we compare the satellite and model SSS fields with the aim of validating the simulation. In section 4, the Indian Ocean MLS complete budget is computed from the model output and the main forcing factors acting regionally are identified. Furthermore, observational fields are used to derive those budget terms which can be retrieved with in situ and satellite measurements, at the same time checking if the observation-based reconstruction of the terms matches in spatial patterns those from the model. Section 5 discusses a few Indian Ocean regions in detail, emphasizing the novel important role played by small-scale and high-frequency processes (and budget terms) not suitable to be diagnosed from observations. Finally, a summary and concluding remarks are provided in section 6.

2. Data and Methods

2.1. Satellite and In Situ Data

Three satellite missions designed to retrieve SSS were launched since 2009. The first mission was the European Space Agency’s (ESA) “Soil Moisture and Ocean Salinity” (SMOS; Font et al., 2004), which started measuring in November 2009, followed by the American National Aeronautics and Space Administration (NASA) “Aquarius/SAC-D” mission (Lagerloef et al., 2008) 2 years later. The latter mission ended on 7 June 2015 due
to an unrecoverable hardware failure. The third mission retrieving soil moisture and SSS is the NASA SMAP (Soil Moisture Active Passive; Fore et al., 2016) mission launched in January 2015. The satellite sensors onboard those satellites measure the incoming microwave radiation in the L-Band (approximately 1.4 GHz), where the sensitivity of the permittivity to salinity in the first centimeter of the ocean is at its highest.

Various studies (e.g., Banks et al., 2016; Köhler et al., 2015; Tong et al., 2015) showed that satellite-retrieved SSS is corrupted by systematic errors coming from ice and land contamination (Font et al., 2013; Gabarro et al., 2012; Oliva et al., 2013), radio frequency interference (RFI; Oliva et al., 2012), seasonal biases, or latitudinal biases mainly due to thermal drift of the instrument (e.g., Boutin et al., 2012; Gourrion et al., 2011; Reul et al., 2012). Those biases are difficult to characterize and several institutions provide satellite-retrieved SSS products based on different correction methods (e.g., Kolodziejczyk et al., 2016). It is not clear, however, which product is most accurate. Therefore, a monthly ensemble-mean SSS was here constructed by averaging four different satellite SSS products derived from SMOS, Aquarius, and SMAP data (see Table 1).

The ensemble-mean SSS product has a spatial resolution of $1^\circ \times 1^\circ$ and covers the period 2011–2016. The spread within the ensemble is taken as the uncertainty of the retrieved SSS. To validate the remotely sensed salinity, a gridded in situ salinity product is used comprising mostly Argo float data. The EN4.2.0 data set of the Met Office Hadley Centre provides global, quality-controlled ocean temperature and salinity profiles and corresponding monthly objective analysis fields. In the EN4 product, data from all types of ocean profiling instruments measuring temperature and salinity were merged into one single profile at each grid location. A detailed description of data sources, quality control procedures, processing, and the method of analysis can be found in Good et al. (2013). This product (hereafter referred to as “in situ”) has a monthly resolution and is optimally interpolated to a $1^\circ \times 1^\circ$, with covariance scales of 300 and 400 km in zonal and meridional directions, respectively (Good et al., 2013).

To assess the role of atmospheric freshwater fluxes on salinity variability, Objectively Analyzed air-sea FLUXes (OAFLUX, http://oaflux.whoi.edu/data.html) evaporation estimates with a monthly $1^\circ \times 1^\circ$ resolution were used. Precipitation was obtained from the monthly $0.25^\circ \times 0.25^\circ$ Tropical Rainfall Measuring Mission’s (TRMM) 3B43 V7 data set and gridded onto a $1^\circ \times 1^\circ$ grid. River runoff was obtained from the monthly Dai and Trenberth global river flow and continental discharge data set as described in Dai (2016). Zonal and meridional velocities from the Earth and Space Research (ESR) Ocean Surface Current Analysis Real-time (OSCAR) NASA project are used to calculate the MLS advection. The currents are calculated as a combination of quasi-steady geostrophic flow, viscous wind-driven ageostrophic flow, and a thermal wind adjustment using, respectively, satellite sea surface height, wind, and temperature. The velocities are given as monthly filtered averages on a $1^\circ \times 1^\circ$ grid (ftp://podaac-ftp.jpl.nasa.gov/allData/oscar/L4/oscar_1_deg) and are representative of the upper 30 m of the water column (Johnson et al., 2006).

All of the observational analyses in our work concern the period 2011–2016.

2.2. Numerical Simulation

The model used is the Parallel Ocean Program 2 (Smith et al., 2010) component of the National Center for Atmospheric Research Community Earth System Model configured on a global tripolar grid with a horizontal resolution of approximately $0.1^\circ$. The model consists of 62 vertical levels with a resolution of 10 m in the
upper 160 m, increasing to 250 m in the deepest layer. The equation of state is computed using the method of McDougall et al. (2003). Surface mixed layer and interior diapycnal mixing processes are parameterized using the k-profile parameterization (Large et al., 1994), which represents turbulent shear mixing and double diffusive convection by increasing vertical diffusivity above the given background diffusivity.

The forcing at the sea surface is provided by the Coordinated Ocean Ice Reference Experiments (COREv2) interannually varying forcing data set, which is derived from a combination of NCEP reanalysis and remote sensing products, as well as gauge measurements in the case of river runoff. River runoff data, containing river discharges at discrete river mouth locations varies at a monthly frequency using the Dai and Trenberth (2002) climatology. Some of the input data are adjusted using more reliable in situ and satellite measurements to address some known biases and data limitations. The forcing is 6-hourly for 10 m wind vectors, 10 m air temperature, specific humidity, sea level pressure, and air density, daily for longwave and shortwave radiation and monthly for solid (snow) and liquid precipitation (see Griffies et al., 2013, Table 1).

The model is initialized from rest with temperature and salinity distributions provided by the World Ocean Circulation Experiment climatology (Gouretski & Koltermann, 2004) interpolated to the high-resolution model grid, integrated forward for 15 years in a spin-up phase and then integrated subject to the COREv2 atmospheric state in the period 1977–2009 (Bryan & Bachmann, 2014; Johnson et al., 2016). The model employs global weak salinity restoring, damped to the Polar Science Center Hydrographic Climatology 2.0 (Steele et al., 2001) using a relaxation time scale of 4 years. The global mean of restoring flux is subtracted at each time step to preserve the global salinity budget.

In our study, we use the model output from the 30 year period 1980–2009 and compute vertically integrated (surface to the depth of the mixed layer) monthly averaged quantities related to the MLS and its budget.

### 3. SSS Variation as a Proxy for MLS Variation

Individual temperature and salinity profiles (EN4 database and model) were first used to determine the ocean mixed-layer depth (MLD), which defines a layer of almost constant potential density adjacent to the surface. The MLD is the depth at which potential density $\sigma$ changes by $\Delta \sigma = 0.03$ kg m$^{-3}$ relative to the surface (de Boyer Montégut et al., 2004). The MLD was averaged over the domain of the grid box and the volume-averaged salinity within that mixed layer is taken as the MLS. In the model case, MLD is taken as the shallowest depth of the maximum buoyancy within the water column.

The average MLD for the model and observations is presented in Figures 2a and 2b, reaching values around 70 m south of 12°S and off Somalia (in the model case). The mixed layer is shallow around the coast of India and south of the equator up to 12°S, where river runoff and/or enhanced precipitation is present. Both the model and observations show corresponding spatial patterns but model MLD is overall around 10 m smaller. The MLD varies especially in the southern part of the study area, in the Arabian Sea and in the east equatorial Indian Ocean (not shown). The model MLD variability is enhanced compared to the observations. From the amplitude of the annual MLD cycle (Figures 2c and 2d), it can be seen that the annual cycle dominates the variability mainly south of 12°S and in the Indonesian Throughflow influenced region. In summary, the MLD shows corresponding mean and annual variability patterns among observations and model simulations.

Moon and Song (2014) focused on the controlling processes of the western tropical Indian Ocean near-surface seasonal salinity variability and how far vertical stratification leads to differences between Argo and Aquarius measurements. The authors argue that stratification is important in precipitation and river runoff dominated regions and can explain the absolute differences between Argo and Aquarius. Also described by, e.g., Boutin et al. (2013), Vinogradova and Ponte (2013b), and Song et al. (2015), it is crucial to compare satellite SSS with in situ uppermost salinity or MLS in regions of large near-surface salinity stratification. The mean bias between satellite and uppermost Argo salinity can reach values between 0.1 in tropical ocean regions and 1 in the high latitudes (Boutin et al., 2013, 2014; Drucker & Riser, 2014; Köhler et al., 2015). However, Drucker and Riser (2014) came to the conclusion that different measurement depths and vertical salinity stratification contribute less to the differences between in situ and satellite measurements than L-Band radiometric accuracies.
But to take advantage of the high spatial and temporal resolution of satellite SSS for the analysis of MLS budgets, it is necessary that SSS and MLS present the same variability (the mean bias not being crucial). Shown in Figure 3 is the correlation of the ensemble-mean SSS and in situ MLS over the 2011–2016 period, reaching values larger than 0.8 in the southern Arabian Sea, the central and southeast Indian Ocean and also in most of the central Bay of Bengal. These are regions where MLS and SSS show a significant annual signal (Figure 4). The correlation is, on the other hand, low in the western Arabian Sea, the western equatorial Indian Ocean and south of 24°S in the subtropical salinity maximum region. Those are all regions where rain events are rare and salinity stratification is not important. Many factors can cause the low correlation in those areas, like weak variability in SSS and/or MLS, measurement errors in the satellite SSS (e.g., land contamination, galactic noise, and RFI) and sampling errors in the in situ MLS fields. The SSS and MLS variance in those regions is also low, which leads to a computed low signal-to-noise ratio (not shown).

As discussed by Calla et al. (2013), strong RFI is present in the western Arabian Sea, which contributes to satellite SSS data errors there. At the equator in the western Indian Ocean, eddy kinetic energy is enhanced due to the extension of the equatorial jets to the eastern coast of Africa (e.g., Nagura & McPhaden, 2010; Sikkakolli et al., 2013). Due to the strong currents transporting the floats, less Argo measurements exist in this particular region. This sampling deficiency leads to a misrepresentation of MLS variability. Only approximately 10 Argo measurements within a 1° grid cell and 6 years exist in that region, leading us to conclude that the Argo data set is not adequate to resolve small-scale variations in MLS. Eddy activity is also strong in the subtropical gyre, where Argo data distribution is again reduced and SSS and MLS show in general less variability. Lower correlations can also be observed off Sumatra and Java, where the near-surface stratification due to strong rain events is an additional source of the low agreement between MLS and SSS variability. In summary, Figure 3 presents where satellite SSS could in principle be best used as a proxy for the MLS variability.
To further investigate the match between satellite SSS, in situ MLS, and model MLS, the amplitude and phase of the respective annual cycles are shown in Figure 4. In case of the model, amplitude and phase were calculated over the complete 30 year period. It can be seen from the figure, the patterns of annual amplitudes for the different data sets match spatially. As discussed in previous studies, winds and ocean current patterns completely reverse north of 10°S due to the monsoon cycle. South of 10°S, the southeast trade winds are present during the entire year, although with a strong annual cycle (Schott & McCreary, 2001). The northern edge of the southeast trade winds shifts northward during summer and retreats poleward during winter, which is represented by the large annual SSS (or MLS) amplitude around 10°S. There, the amplitude of the annual SSS cycle reaches values of approximately 0.5 with the salinity maximum occurring in June (Figures 4b, 4e, and 4h). The model MLS maximum occurs approximately 1 month earlier in this region. Large annual amplitudes can be found in the coastal regions of the Bay of Bengal, where, besides other processes, high precipitation and river runoff govern the variability. The salinity maximum occurs in boreal spring, when river runoff and precipitation are low. In the central Bay of Bengal, the salinity maximum occurs in September/October, which is represented in all data sets, but the amplitude is reduced in the case of the satellite product.

To conclude, satellite SSS show corresponding spatial patterns of annual variability to in situ and model MLS. Therefore, SSS is a fairly good proxy for MLS variations in large areas of the Indian Ocean and we can take advantage of the high spatial and temporal resolution of satellite SSS to analyze the underlying processes of MLS variance in the regions of strong salinity signals. On the other hand, the positive comparison between in situ and model MLS annual signals gives confidence in the good representation of salinity variations in the model, which motivates its use to study the full MLS budget in the following section.

Figure 4. Amplitude, phase, and total variance explained by the annual cycle for (a) satellite SSS (period 2011–2016); (b) in situ MLS (period 2011–2016) and model MLS (period 1980–2009).
4. Processes Causing MLS Changes

The temporal development of MLS in the Indian Ocean is governed by processes such as evaporation, precipitation, river runoff, horizontal and vertical advection, and mixing. Despite many existing studies (e.g., Akhil et al., 2014; Da-Allada et al., 2015; Moon & Song, 2014; Nyadjro & Subrahmanyam, 2014; Rao & Sivakumar, 2003), the relative importance of those processes to the MLS budget, their time change, and regional variation are not well known, certainly not with the spatial details that are being pursued in the present work. To identify the dominant processes causing the salinity field to vary seasonally, complete MLS budgets (including all possible terms) were computed with the model. We further test if satellite SSS can be used as MLS in the budgets, with the advantage of having a higher horizontal resolution, as the satellite measurements provide. The model is of crucial importance in the analysis of the driving mechanisms, specially in coastal areas where in situ measurements are sparse and satellite-retrieved salinity data have to be treated with caution.

4.1. Mixed-Layer Salinity Balance Equation

Let us define the monthly time average of a parameter \( x \) as

\[
\bar{x} = \frac{1}{1 \text{mo}} \int_{1 \text{mo}} x \, dt
\]

and the vertical average as

\[
[z] = \frac{1}{h} \int_{-h}^{0} z \, dz
\]

The corresponding decomposition into temporal and vertical “mean” contribution and deviations from the mean (from here on termed “eddy” and “shear” contributions, respectively) is assumed:

\[
\bar{x} = \bar{x} + x'
\]

\[
x = [x] + x''
\]

With these definitions, the integral form of the MLS budget can be written as follows:

\[
\frac{\partial \bar{S}}{\partial t} = -\nabla \cdot \bar{u} \bar{S} + \frac{1}{h} \bar{w} \bar{S} \bigg|_{-h}^{0} - \nabla \cdot \bar{u}' \bar{S}' - \frac{1}{h} \bar{w}' \bar{S}' \bigg|_{-h}^{0} + \frac{1}{h} S_{\text{ref}} (E - P - R) - \frac{1}{h} \left( \frac{\partial \bar{S}}{\partial z} \right)_{-h}^{0} - [\kappa \nabla ^{2} \bar{S}]
\]

The left-hand side of equation (1) is the mixed-layer average of the monthly averaged salinity tendency term. The first two terms on the right-hand side of equation (1) represent, respectively, the contributions to the MLS tendency from advection of mean salinity \( \bar{S} \) by the mean horizontal \( \bar{u} \) and vertical \( \bar{w} \) flow (henceforth termed mean advection). The subscript \(-h\) means that the quantities are evaluated at the base of the monthly averaged mixed layer. The third and fourth terms on the right-hand side correspond to the advection of salinity temporal fluctuations \( S' \) by the time-varying horizontal \( u' \) and vertical \( w' \) flow (henceforth termed eddy advection), respectively. The fifth term on the right-hand side corresponds to the surface external forcing (SEF), representing the combined effects of surface freshwater flux due to evaporation, precipitation, and river runoff \( (E - P - R) \), with \( S_{\text{ref}} \) being a constant reference salinity used to convert freshwater flux into virtual salinity flux. The last two terms on the right-hand side of equation (1) are the contribution to the MLS tendency from vertical and horizontal diffusion processes. \( \kappa \) is the salinity vertical diffusivity and the horizontal diffusion is formulated as a high-order hyperdiffusivity with coefficient \( B \). Equation (1) is solved in the model and closes exactly but cannot be diagnosed using observations alone.

The observational database is insufficient to describe the full three-dimensional structure of the oceanic flow in the mixed layer and only sparse measurements of the three-dimensional salinity distribution exist. To diagnose the MLS budget using the sparse observational data, the tendency and mean advection terms
Figure 5. Average (a) model vertical velocity at the mixed-layer depth, (b) vertical velocity calculated from the divergence of the mixed-layer averaged horizontal model velocities, and (c) vertical velocity calculated from the divergence of the horizontal OSCAR velocities.
divergence. Motivated by the above agreement, the observational vertical velocity is therefore diagnosed in our work from the divergence of the mixed-layer averaged horizontal velocity field, in this case the OSCAR velocities. The obtained vertical velocity shows promising corresponding patterns and magnitudes compared to the model case (Figure 5c).

The vertical salinity difference $\frac{\partial S}{\partial z}$ in the ENTR term is calculated as the difference between MLS and the salinity at the mixed-layer base, both evaluated from Argo profiles. The vertical salinity gradient $\frac{\partial S}{\partial z}$ in VDIFF is calculated from the salinity difference 10 m above and below the MLD. The salinity vertical diffusivity $\kappa$ is taken from the model, from which a monthly climatology was calculated and gridded at $1^\circ \times 1^\circ$ resolution.

The two eddy advection terms and the shear advection cannot be computed from observations because measurements (1) of the three-dimensional structure of currents in the mixed layer and (2) of high-frequency correlations of salinity and currents are not available. Moreover, HDIFF is not attempted to be diagnosed from observations due to the highly unknown horizontal eddy diffusion and since it was seen to be rather small in the model. Therefore, the observational budget residual includes horizontal and vertical eddy advection, shear horizontal advection, and HDIFF (plus estimation errors in all other terms: SEF, mean HADV, ENTR, and VDIFF).

In the following analysis, the relative importance of the different terms of equation (2) (SEF, mean HADV, ENTR, and VDIFF) will be quantified from satellite and in situ data. As stated, in the case of the model all terms in the budget are computed exactly, allowing to quantify the contribution of eddy terms and the validity of the approximations done when using observations. To analyze the magnitude of the different contributions to the seasonal cycle of the MLS tendency, an annual harmonic was fitted to each budget term. From the comparison between observational and model budgets, we test which terms can be retrieved from observations and where residual terms, not available from the limited observations, become important.

In the course of the computation, we use satellite-retrieved SSS as a proxy for MLS variations on different time scales. That means that satellite SSS is taken as variable $\frac{\partial S}{\partial z}$ in equation (2) for the observational budget case. Variable $\frac{\partial S}{\partial z}$ is still computed from Argo profiles.

4.2. Mean MLS Budget Terms

Figure 6 shows the spatial distribution of the mean budget terms calculated from the model and observations. In a first view, the patterns agree spatially and represent primarily the summer monsoon conditions. SEF (Figures 6a and 6b) is negative in the eastern Indian Ocean, where precipitation is dominant and the MLD is small. Positive SEF can be observed in the western Indian Ocean and south of 12°S. SEF is mainly balanced by the sum of mean HADV and VDIFF.

During the summer monsoon, the Southwest Monsoon Current brings high-salinity waters from the Arabian Sea to the east, which is mainly reflected in the positive contribution of mean HADV in near-equatorial regions (Figures 6c and 6d). Around India model and observational results differ, the model showing mainly a negative mean HADV contribution in the near-coastal Arabian Sea and Bay of Bengal, whereas the observational mean HADV is positive. The same can be observed at the equator in the eastern Indian Ocean. This is attributed to differences in the horizontal velocity deriving from the coarser resolution and sampling errors in OSCAR, but also to the different study periods. Several IOD events occur between 2011 and 2016 mainly influencing horizontal advection in the Indian Ocean (e.g., Nyadjo & Subrahmanyan, 2014).

VDIFF (Figures 6e and 6f) is positive in regions where precipitation and/or river runoff and the mean horizontal advection of low-salinity waters leads to a shallowing of the MLD and a strong vertical salinity gradient across it. This can be observed in the coastal Bay of Bengal, eastern Arabian Sea, and in a zonal band slightly south of the equator. Therefore, VDIFF acts against the freshening from SEF and mean HADV by mixing high-salinity waters from underneath into the mixed layer. The observational VDIFF presents similar patterns as the simulation.

The magnitude of ENTR (Figures 6g and 6h) is small compared with the other terms, with positive contribution slightly south of the equator and negative contributions to the north of that positive maximum and in near-coastal regions of the Bay of Bengal. This alternating pattern of positive and negative ENTR around the equator is a combination of upwelling of high-salinity waters to the south of the wind-induced equatorial
roll (e.g., McCreary, 1985; Miyama et al., 2003; Philander & Deelecluse, 1983; Wang & McPhaden, 2017) and downwelling to the north of the equatorial roll and the consequent horizontal induction at the MLD. Also here model and observations agree, showing similar patterns and amplitudes.

Figure 7 shows the mean distribution of the remaining budget terms calculated from the model output. Horizontal diffusion was found to be negligible and is therefore not presented. Shear advection and the eddy advection terms show larger amplitudes than entrainment and are maximal in the highly stratified ocean regions in the eastern Indian Ocean around the equator and the near-coastal regions of the Bay of

Figure 6. (a, b) Mean surface external forces (SEF), (c, d) mean horizontal salinity advection (mean HADV), (e, f) vertical salinity diffusion (VDIFF), and (g, h) entrainment (ENTR) for (left) the model over the period 1980–2009 and (right) the observations (period 2011–2016). Terms were computed according to equation (2) (see text).
Bengal and eastern Arabian Sea. The contribution of shear HADV is mainly positive and shows opposing patterns compared to the mean HADV, indicating that shear horizontal advection reduces the effect of mean horizontal advection.

The eddy horizontal advection is positive in a small stripe off the northern and western coasts in the Bay of Bengal and negative immediately offshore, indicating that on scales less than 1 month high-salinity waters of the interior Bay of Bengal are being mixed with low-salinity waters of the near-coastal regions. Eddy vertical advection is negative throughout the coastal Bay of Bengal and southwestern Indian coast, which reflects, again on time scales smaller than 1 month, downward flux of low-salinity waters into the subsurface.

To conclude, SEF is mainly balanced by the sum of mean HADV and VDIFF, whereas the contributions of ENTR and shear and eddy terms are small compared with the former terms, but being nevertheless regionally important. The observational results are encouraging, showing corresponding spatial patterns to the model. The observational mean HADV is the term which differs the most, particularly in the eastern Indian Ocean.

4.3. Seasonal Cycle of MLS Budget Terms

The annual amplitude of model ST (Figure 8, 1st row) agrees well with the annual SSS amplitude previously discussed, with maxima in the southern Arabian Sea, Bay of Bengal, and in the tropical Indian Ocean south of the equator. The main patterns of annual cycle amplitude agree also well to the results in Vinogradova and Ponte (2013a), although some differences can be observed. As an example, our simulation shows an annual amplitude maximum around 14°N and 73°E, which is not present in ECCO (compare with Vinogradova & Ponte, 2013a, their Figure 2) and coincide with regions of enhanced small-scale variability. A larger annual amplitude can be found in the central Indian Ocean, where ECCO reaches values less than 0.2 month⁻¹ and our simulation values up to 0.3 month⁻¹. Furthermore, enhanced annual variability can be observed around Sumatra and Java, not captured in ECCO. Outside the described regions, variability of ST is less than 0.1 month⁻¹ on annual time scales. Differences between the annual amplitude of simulated and observed ST can be found in coastal regions in the eastern Arabian Sea, in the southern Bay of Bengal, and off Sumatra and Java, where the observations show less variability than the model.
Figure 8. Amplitude of the annual cycle of (a, b) salinity tendency (ST), (c, d) surface external forces (SEF), (e, f) mean horizontal salinity advection (mean HADV), (g, h) vertical salinity diffusion (VDIFF), and (i, j) entrainment (ENTR) for (left) the model over the period 1980–2009 and (right) the observations in the period 2011–2016. Terms were computed according to equation (2) (see text).
From Figure 8, it can be seen that mainly the mean HADV (Figures 8e and 8f) resembles the spatial variability of ST in the northern Indian Ocean, showing clear annual amplitude maxima in the eastern and southern Arabian Sea and in the Bay of Bengal, reaching values of approx. 0.3 month$^{-1}$. The observations show similar patterns with discrepancies in the western Arabian Sea and south of the Bay of Bengal, where the model presents a zonal coherent behavior that might not be resolved by the observations. The simulated mean HADV term is enhanced south of 12°S in the eastern Indian Ocean around 90°E, not seen in the observations.

South of the equator, both SEF (Figures 8c and 8d) and the mean HADV balance the ST variability at annual time scales. SEF also shows a large amplitude in regions where E-P has a strong seasonal cycle in the northern Bay of Bengal and in the eastern Arabian Sea, induced by enhanced monsoonal precipitation and river runoff. The model presents a stronger SEF annual cycle in the northern Indian Ocean and the observations, on the other hand, more intensification south of the equator.

In the model, VDIFF (Figures 8g and 8h) shows a strong annual cycle in the northern Bay of Bengal and in the eastern Arabian Sea in near-coastal areas. The amplitude is also enhanced to the east of Sri Lanka and in SEF-dominated regions south of the equator, reaching values up to 0.3 and 0.1 month$^{-1}$, respectively. The observational VDIFF is consistent with the simulation and only significantly different in the western Arabian Sea and around Sri Lanka.

Figure 9. Amplitude of the annual cycle of the three components of entrainment (computed according to equation (2), see text) for (left) the model over the period 1980–2009 and (right) the observations in the period 2011–2016.
The ENTR term (Figures 8i and 8j) is enhanced in regions of strong vertical salinity gradients, as in the Bay of Bengal and in the southeastern Arabian Sea. ENTR reaches values up to 0.1 month$^{-1}$ in the eastern Indian Ocean slightly north and south of the equator in the model case and around 0.05 month$^{-1}$ in the observational case. A larger ENTR annual amplitude in the observations can be seen in the central Indian Ocean between 3°S and 12°S, where open ocean upwelling takes place (Schott et al., 2009). The discrepancies between model and observational ENTR south of the equator are attributed to differences in the temporal change of the MLD (first term of ENTR, Figures 9a and 9b) and differences in the vertical flux at the mixed-layer base (third term of ENTR, Figures 9e and 9f), showing both larger amplitudes in the observations due to a shallower MLD in that case (compare with Figures 2a and 2b).

From Figure 9, it can be seen that the annual variance of the ENTR term is primarily determined by lateral induction at the mixed-layer base (second term in ENTR, Figures 9c and 9d) and by the vertical term of ENTR. The large amplitudes in the eastern Indian Ocean in a narrow band around the equator reflects the upwelling (downwelling) at the southern (northern) edge of the near-surface wind-induced equatorial roll prominent during the summer monsoon season (Wang & McPhaden, 2017). As discussed by Philander and Delecluse (1983), McCreary (1985), Miyama et al. (2003), and Wang and McPhaden (2017), the northward cross-equatorial surface wind stress and the southward Ekman transport at the subsurface introduces a shallow roll limited to the mixed layer within a few degrees of the equator. The upwelling-induced elevation of the mixed layer to the south and the downwelling-induced deepening of the mixed layer to the north leads to a steeper slope of the mixed layer, which again results in enhanced lateral induction at the mixed-layer base. These processes are limited to a narrow band around the equator and the coarser resolution of the observational data sets makes it difficult to observe these processes.

The annual amplitude of the sum of the budget terms not diagnosed with observations—the remaining terms in equation (2)—is presented in Figure 10a for the model case. It is composed of the sum of shear advection (shear HADV), eddy advection (eddy HADV+ eddy VADV), and horizontal diffusion and is enhanced in the near-coastal regions of the Bay of Bengal, in the eastern and southern Arabian Sea, off Somalia, and at the northern tip of Sumatra, reaching values larger than 0.2 month$^{-1}$, therefore comparable with other budget terms. In near-equatorial regions, values up to 0.1 month$^{-1}$ are present.
The observational budget residual includes processes not accounted for (shear advection, all eddy advection terms, and horizontal diffusion), but also errors in all the budget terms computations (instrument, sampling, and gridding errors; errors due to the approximations taken, etc). The spatial patterns are however positively correlated (compare Figures 10a and 10b), with a tendency for maxima in coastal regions of the Bay of Bengal, off Somalia, and the northern tip of Sumatra. Largest differences to the model can be observed in the northern Arabian Sea, where the observational residual shows annual amplitudes of those small-scale high-frequency terms around 0.25 month$^{-1}$ and the model less than 0.1 month$^{-1}$. We believe that in that region the observations are less resolving, and what appears in the residual are mostly errors in the estimation of the largest terms.

Figure 10 also shows the annual amplitude of the model eddy and shear salinity advection. It can be seen that both the eddy and shear advection terms have a strong annual cycle in the coastal Bay of Bengal (approx. 0.3 month$^{-1}$), whereas shear advection shows additionally a large amplitude (approx. 0.1 month$^{-1}$) in a pair of narrow bands south and north of the equator. The latter are again attributed to the equatorial roll and will be discussed in detail ahead when discussing specific Indian Ocean regions. All in all, the remaining terms not diagnosed with observations are obviously relevant in regions of enhanced small-scale high-frequency variability (coastal Bay of Bengal and eastern Arabian Sea) and in the eastern equatorial Indian Ocean, where the unique vertical structure of the mixed layer gives rise to additional small-scale advection processes.

Seasonal changes in SEF and oceanic processes from Vinogradova and Ponte (2013a) and our model estimates show similar spatial patterns. However, the present model shows finer structures (compare Bay of Bengal and eastern Arabian Sea) due to the higher spatial resolution and larger magnitudes (e.g., in the eastern Arabian Sea, in the central Indian Ocean, and off Sumatra and Java). Differences between our and their results are attributable to differences in the mixed-layer depth, in the average salinity, in circulation patterns and different surface flux forcing fields. Furthermore, differences can come from significant differences in the model parametrizations. For instance, double diffusion parametrization included in the CESM configuration is not present in ECCO. The different temporal length of the analyzed model estimates (13 years in the case of Vinogradova and Ponte (2013a) and 30 years here), as well as the higher resolution of the present model including spatial small-scale variability not addressed with 1° gridded ECCO fields, all add to the different results.

4.4. Variance Explained by Each MLS Budget Term

We now turn our attention to the proportion of MLS tendency explained by each particular budget term (Figure 11). The model and the observational results are once again consistent with each other.

SEF (Figures 11a and 11b) explains a large proportion of ST in regions of strong E-P annual cycle (western and central south Indian Ocean and off Java). In the northern Bay of Bengal and around Sri Lanka, ST is also largely explained by SEF. The model shows ST largely explained by SEF south of 20°S, a feature not present in the observations and attributable to the larger model MLD variability in this region (see Figure 2c). There is a significant proportion of observed annual salinity tendency explained by SEF in the western Arabian Sea, which can not be found in the model due to minor differences in the surface freshwater flux fields and lower model salinity variability in this particular region.

The mean HADV (Figures 11c and 11d) largely explains ST in the southern Arabian Sea and in off-equatorial regions in the western and central Indian Ocean (between 12°S and 5°S). The patterns are consistent among the two data sets, however, the observations point to larger explained ST variance by mean HADV in Indonesian Throughflow-dominated regions (between 12°S and 24°S) in the southeastern Indian Ocean.

VDIFF explains ST variance in the eastern Indian Ocean south of the equator and in the central Bay of Bengal (Figures 11e and 11f). ENTR explains significant proportions of ST in noncoastal areas of the Arabian Sea and in the central Bay of Bengal (Figures 11g and 11h). Other regions include the off-equatorial Indian Ocean south of 6°S.

The proportion of annual ST explained by the sum of SEF, mean HADV, ENTR, and VDIFF is presented in Figures 11i and 11j. In the model case, the sum explains more than 90% in the southern Arabian Sea and in the off-equatorial regions in the eastern and central Indian Ocean. In the observational case, a smaller proportion in reduced ocean areas is obtained. From Figures 8a and 8b, it is obvious that annual ST is enhanced.
Figure 11. Percentage of the MLS tendency annual variance explained by (a, b) surface external forces (SEF), (c, d) mean horizontal advection (mean HADV), (e, f) vertical diffusion (VDIFF), (g, h) entrainment (ENTR), and (i, j) the sum of the considered terms (SEF + mean HADV + ENTR + VDIFF), using (left) model data and (right) satellite and in situ observations. Budget terms are computed according to equation (2). Locations with explained variance less than 5% are masked white.
in the northern Bay of Bengal and in the coastal regions of the eastern Arabian Sea, but the sum of the considered processed cannot explain annual ST variance, indicating that the unconsidered eddy and shear processes are of importance for closing the budget in particular regions.

In summary, the spatial patterns of annual variance explained by SEF, mean HADV, ENTR, and VDIFF for model and observational cases show an overall good agreement. However, the patterns of explained ST by oceanic processes (mean HADV, ENTR, and VDIFF) differ in magnitude and in some cases location. There are contributing deficiencies from using mixed-layer averaged velocities in the MLS mean HADV term. This holds also for the assumptions regarding the vertical velocity and horizontal induction made when calculating ENTR. It is nevertheless surprising the large extent to which there is consistency between the MLS advection retrieved from OSCAR velocities and from the model.

5. A Closer Look at Processes in Selected Regions

In this section, selected Indian Ocean regions will be studied where it was seen that satellite SSS is well correlated with MLS. Furthermore, these will be regions where strong agreement or disagreement exists between model and observational budgets. Some of these are also regions where the existing observations are not enough to close the MLS balance since they do not capture specific small-scale high-frequency processes.

Figure 12 shows the salinity tendency ($ST_{lhs}$) and the sum of the contributing terms ($ST_{rhs}$) from the observational budget for each selected ocean box. Shaded areas indicate the errors in $ST_{rhs}$ and $ST_{lhs}$ estimated via error propagation of the standard errors of all contributing terms (including the measurement error). For reference, the right-hand side of the budget equation (2) for the model case is shown, since in the model we know the balance closes. In the second row of Figure 12, the main terms contributing to the $ST_{rhs}$ are presented for the model and observational cases. To help having a general view on the processes, Figures 13 and 14 show the climatological monthly mean of all model-based budget terms for the months of February, May, and September.

At first glance, it is noticeable that observations-based and model-based ST show a corresponding behavior. The satellite errors in near-coastal boxes (NWIO, EIO, and BOB, Figures 12a, 12c, and 12d) are larger (up to 0.2 month$^{-2}$), attributable in part to the large uncertainties in satellite SSS in coastal areas.

5.1. Northwest Indian Ocean (NWIO)

In the NWIO region (Figure 12a), observational $ST_{lh}$ shows a seasonal cycle corresponding to the $ST_{lh}$ with a correlation of 0.96. A maximum (positive) tendency is reached in boreal summer between May and August and a minimum in boreal winter (compare with Figures 13a–13c). Satellite-based $ST_{lh}$ and $ST_{lh}$ lie within the error bars. All in all, the $ST_{lh}$ is overestimated with a RMSD of 0.11 month$^{-1}$. The seasonal cycle of model ST (brown curve) corresponds to the observed $ST_{lh}$.

Mean horizontal advection dominates the seasonal cycle of ST in the NWIO (Figure 12e, note that the mean HADV term is multiplied by 0.5 in the NWIO region), being positive between May and October and negative during the rest of the year. During the summer monsoon strong southward Ekman transport is present (attributable to the strengthening of southwesterly winds) transporting high-salinity waters from the northern Arabian Sea (Chowdary et al., 2006; Da-Allada et al., 2015). The dominance of mean HADV in this ocean region was already mentioned in the previous chapter. Additionally, the Southwest Monsoon Current brings high-salinity waters from the Arabian Sea into the Bay of Bengal. During the winter monsoon, low-salinity waters are advected from the Bay of Bengal into the NWIO region and the strong northward Ekman transport brings the low-salinity waters to the north leading to a decrease in salinity advection (Schott et al., 2009) and negative ST between November and March/April. This can be seen in Figures 13g–13i, where mean HADV shows a large negative contribution in February and positive contribution in May and September. Using the output of the HYCOM model, Nyadjro et al. (2010) studied the salinity advection in the uppermost 5 m layer of the monsoon-influenced Indian Ocean and found that the Ekman component of advection is more important than geostrophic advection for the exchange of water masses in the northern Indian Ocean.
In the observational case (Figure 12e), surface freshening due to SEF is highest in May, coinciding with a maximum in ENTR (see also Figures 13e and 13n). Enhanced precipitation in May leads to an increasing vertical salinity gradient and together with the increasing MLD contributes to larger ENTR. This relative SEF minimum is not present to the same extent in the model SEF, which explains the differences in the ENTR term. Both observational and model SEF are otherwise small except from January to April, when evaporation is enhanced. ENTR is small throughout the year, with the aforementioned contribution peaking during May. Further to the east of the chosen NWIO region, close to the Indian Coast (Figure 13m) a strengthened negative contribution of ENTR in February can be observed, which results mainly from a shallowing of the MLD (temporal MLD development and ENTR1) as well as the lateral induction at the mixed-layer base due to the steeper MLD with respect to the deeper surrounding mixed layer.

Model VDIFF (Figure 12e) is enhanced during the first half of the year, when low-salinity waters were advected from the eastern Indian Ocean into the Arabian Sea, and peaks in May, when precipitation leads to a decreasing MLD and increasing vertical salinity gradient (see Figures 13j and 13k). Observational VDIFF (Figure 12e) has two maxima, in February and in June, showing an amplitude two times larger than model VDIFF. The conclusion here is that freshening due to enhanced precipitation in May is attenuated by VDIFF.
The model sum of the remaining terms (Figure 12m, summed up in the solid magenta curve in Figure 12e) varies between 0.01 and 0.03 month$^{-1}$, whereas the residual in the observational case ($\text{Res} = \text{ST} - \text{SEF} - \text{mean HADV} - \text{VDIFF} - \text{ENTR}$) reaches a maximum value of $-0.18$ in June, when mean HADV is at its highest. Compared to the model, mean HADV is clearly overestimated (blue curves in Figure 12e), which can be to some extent explained by the overestimation of the zonal velocity component (Figure 12i).

Figure 13. Climatological mean of (a–c) ST, (d–f) SEF, (g–i) mean HADV, (j–l) VDIFF, and (m–o) ENTR for the months of (left) February, (middle) May, and (right) September.
5.2. Central Indian Ocean (CIO)

In the CIO region (Figure 12b), observational and model cases show a corresponding annual cycle, the first with a correlation coefficient between $ST_{rhs}$ and $ST_{lhs}$ of 0.95. $ST_{rhs}$ and $ST_{lhs}$ match within the error bars except in February, with RMSD around 0.06 month$^{-1}$. The model annual ST cycle matches the observational $ST_{lhs}$ with a correlation coefficient of 0.94 and a RMSD of 0.04 month$^{-1}$. Both observations and model $ST_{rhs}$ have their maximum in May (see also Figure 13b). The southeastward advection of high-salinity waters into the CIO region leads to the observed ST maximum in May/June (Figure 13h). The mean HADV and ST decrease from June to August (Figures 12b and 12f), reaching a second minimum in September (Figures 13c and 13i), when westward equatorial surface currents prevail advecting low-salinity waters to the west.

Model and observational SEF are negative throughout the year (Figure 12f), with strongest freshening between December and February leading to a shallow mixed layer and strong vertical salinity gradient between the mixed layer and the ocean beneath. The enhanced contribution of SEF is clearly shown in Figure 13d, being negative between 12°S and the equator.

Model VDIFF shows a behavior corresponding to the observational VDIFF with maximal values between January and May (Figure 12f) attenuating the freshening effect of SEF by providing a salt flux into the mixed layer. The maximum VDIFF south of the equator is also well shown in Figures 13j and 13k. VDIFF starts decreasing when the mean HADV of saltier water into the CIO region is strengthened, leading to an increasing MLD and a decreasing vertical salinity contrast.

The contribution of ENTR is generally very low in the CIO region (Figures 12f and 13m–13o), but enhanced to the north of it in a narrow band around the equator, reflecting again the aforementioned upwelling and downwelling at the southern and northern edge of the equatorial roll during the summer monsoon season.

Figure 14. Climatological mean of (a–c) shear HADV, (d–f) eddy HADV, and (g–i) eddy VADV for the months of (left) February, (middle) May, and (right) September.
The sum of the remaining terms in the model is small but has a maximum in April and minimum in June (Figure 12f), which is the period of largest high-salinity advection and reflects mainly the presence of shear horizontal advection (Figure 12n). The shear HADV term is positive throughout the year except in May and June (see also Figure 14b). That minimum can be explained by the vertical structure of velocity and salinity within the mixed layer. Wind stress with a northward component is present in May bringing high-salinity surface waters from the CIO region to the north, whereas the meridional velocity in the subsurface is southward bringing low-salinity waters into the CIO region. On average, this results in a negative contribution of shear HADV acting against the front in the CIO region which is enhanced due to the precipitation-induced shallow MLD. In June, the conditions are slightly different. At the equator low-salinity waters from the precipitation-dominated regions are advected westward, whereas the southward advection of high-salinity waters in the CIO region is enhanced (compare the meridional velocity minimum in Figure 12j and the HADV maximum in Figure 12f). Again there is a vertical distribution anomaly (shear) of meridional velocity within the mixed layer, with a transport of high-salinity waters northward at the surface and a transport of low-salinity waters southward at the subsurface, resulting in a negative contribution of shear advection to ST in the CIO region.

Also shown in Figure 12f is that mean HADV is underestimated from May to September in the observational case. Bentamy et al. (2008) compared wind fields from QuickScat (included in the model COREv2 forcing fields) and ASCAT (OSCAR input) with the result that ASCAT tends to underestimate high winds (above 20 m s\(^{-1}\)) present in near-equatorial regions. This is obvious in the meridional velocity between May and August, the period of maximum southward Ekman advection.

In April, the vertical structure of meridional velocity corresponds to that observed in May and June with northward velocity at the surface and southward velocity at the subsurface. However, the salinity signal being advected is opposite. The surface northward flow brings low-salinity waters to equatorial regions, whereas the subsurface flow brings higher-salinity waters to the CIO region, resulting in a positive shear HADV contribution.

### 5.3. Eastern Equatorial Indian Ocean (EEIO)

The EEIO region features a seasonal ST cycle with minima in February and August and maxima in May and December (Figure 12c). Observational ST\(_{\text{rhs}}\) match the ST\(_{\text{lhs}}\) within the error bars but with a correlation of only 0.78 and a RMSD of 0.09 month\(^{-1}\). Unlike the other regions, the model ST differs more pronounced from the observed ST\(_{\text{obs}}\). The model shows also a clear semiannual cycle but shifted by 1 month. Mean HADV dominates the semiannual ST cycle (Figure 12g), but a large difference is seen between model mean HADV and observational mean HADV, which is dominated by the zonal velocity differences (Figure 12k), ultimately explaining the shift in ST\(_{\text{obs}}\).

Observational and model SEF are negative throughout the year reflecting a precipitation-dominated region. Largest differences between model and observations are seen in October and November (Figure 12g).

VDIFF is large throughout the year due to strong vertical gradients and shallow MLD. Both VDIFF from the model and observations show a corresponding small semiannual cycle, with maxima in December to February and in May (Figure 12g). The first is attributable to the advection of low-salinity waters into the EEIO region as well as to the freshening due to SEF (compare Figures 13d, 13g, and 13j). The advection of high-salinity waters from the northwest in May (Figure 13h) is weakened by the shear horizontal advection (Figure 14b), the latter acting against the meridional front of low-salinity waters to the north and the high-salinity waters advected by the equatorial jets.

ENTR is small except for September to November in the model case, when the equatorial roll is strongest. The model sum of the remaining terms varies around zero with minima in May and October and maxima in June and September (reflecting mainly shear HADV, Figure 12o) when vertical stratification is enhanced.

Figure 15 shows the model budget terms along the 10°S–5°N latitudinal section (location in Figures 14a–14c) averaged between 83°E and 85°E for February (winter monsoon), May (transition period) and September (summer monsoon). In February (Figure 15a), the ST decreases to the north, resulting from the advection of low-salinity waters with the South Java Current coming from the southeast slightly south of the equator and with the Northeast Monsoon Current north of the equator until 4°N. Mean HADV is mainly balanced by the positive contribution of shear HADV and ENTR, reaching values between 0.1 and 0.2 month\(^{-1}\).
The ENTR term is mainly determined in February by the vertical velocity (upwelling-related) contribution (Figure 15b and ENTR3 in equation (2)).

The shear HADV term depends significantly on the vertical structure of the meridional velocity and salinity (not shown). During the winter monsoon, the equatorial roll changes the direction due to reversed cross-equatorial wind stress, with upwelling north of the equator and a southward flow at the surface and a northward flow in the subsurface. This leads to a transport of low-salinity waters to the south at the surface and reversed conditions in the subsurface, resulting in a positive shear advection acting against the stratified salinity front. SEF is negative along the section but increases northward. VDIFF is large in regions of enhanced precipitation and has maxima at the latitudes where currents advect low-salinity waters and the MLD is shallow. North of 4°N, the sum of the described processes cannot explain the ST. There, mainly eddy advection contributes to the model ST, resulting from flow instabilities and eddy activity.

During the monsoon transition period in May (Figures 15c and 15d), the mean HADV mainly drives the ST south of the equator, with SEF and shear HADV acting against it. North of 4°N the ST is mainly due to a balance between the large contribution of VDIFF acting against the mean HADV of low-salinity waters. ENTR is in generally low, but slightly enhanced in a narrow band around the equator with a negative contribution resulting mainly from equatorial downwelling caused by the convergence of the flow field (Figure 15d and ENTR3 in equation (2)). North of 1°N, the temporal fluctuation of the MLD (ENTR1 term in equation (2)) is

Figure 15. Latitudinal variation at 84°E (between 10°S and 5°N) of MLD and all MLS budget terms (see legend) for the months of (a) February, (c) May, and (e) September. The ENTR term is further split into its components in (b, d, and f).
consistently positive and the lateral induction due to the steep MLD slope (ENTR2 term in equation (2)) is important south of the equator.

The equatorial roll is strongest during the summer monsoon, which translates again into an anomalous vertical structure of the meridional velocity and therefore projects on the shear HADV in September (Figure 15e). On the other hand, enhanced upwelling (downwelling) immediately south (north) of the equator, whereas low-salinity waters from the northeast are advected southwestward in a band south of the equator (see mean HADV in Figure 15e and also Figure 13i). Despite the negative contribution of mean HADV between 2°S and 0°S and of enhanced precipitation there, the ST increases, which comes from upwelling-related ENTR (Figures 15e and 13o) and from shear HADV (Figures 15e and 14c). North of 0° where the Southwest Monsoon Current determines a strong positive mean HADV contribution, ST actually decreases, resulting mainly from a sum of two contributions: (1) the lateral induction in the strong MLD slope region and downwelling-related ENTR (see also Figure 13o) and (2) the also negative shear HADV due to the anomalous northward advection of low-salinity waters within the equatorial roll at the surface. Eddy HADV is strong (and negative) as well in the Southwest Monsoon Current and in the East India Coastal Current around 4°N.

5.4. Bay of Bengal (BOB)

Akhil et al. (2014) analyzed the major contributions to the dynamics of the seasonal cycle of near-surface salinity in the Bay of Bengal using a regional model simulation. They found that the minimum MLS between June and September is mainly explained by the freshwater input due to river runoff and precipitation as well as freshening by horizontal advection and that the positive salt flux into the near-surface layer comes mainly from vertical processes and not (like previously thought) from advection of high-salinity waters from the Arabian Sea. Using output from another model simulation, Wilson and Riser (2016) confirmed that vertical processes are important for balancing the near-surface ST. Here we extend the analysis to the complete mixed layer and assess in detail the importance of processes not discussed in the above-mentioned works.

As previously discussed, the Bay of Bengal shows a strong seasonal MLS cycle which is maximal in the northern Bay at the major river outflow regions and enhanced at the western rim of the Bay. In the northern BOB region (Figure 12d), observational $ST_{rhs}$ and $ST_{lhs}$ differ largely during boreal summer. $ST_{lhs}$ shows a clear annual cycle with a minimum in June/August, when river runoff is maximal. Model ST shows a corresponding annual cycle with a minimum in August and a maximum in November. $ST_{rhs}$ shows a divergent behavior after June, largely overestimating $ST_{lhs}$. Interestingly, the model shows the same behavior, i.e., the sum of mean HADV, SEF, ENTR, and VDIFF is not enough to explain ST after May/June (see blue curve in Figure 12d summing up the contributions possible to retrieve from observations). Besides the above main processes, the shear HADV and eddy processes making up the residual are of crucial importance for closing the budget.

In our model, the emerging picture is also that freshening due to river runoff and precipitation augmented by fresh advection is attenuated by enhanced vertical diffusion, as can be seen in Figures 13f, 13i, and 13l. However, the remaining processes have to be taken into account to explain the negative ST after June, which otherwise would be positive. A closer look into the remaining processes (Figures 12p and 14c, 14f, and 14i) shows a dominance of the eddy processes, which are strongest during the second half of the year. The contribution of shear horizontal advection is smaller than the eddy advection, reaching a maximum in September, when vertical stratification is maximal.

The differences between the model and observational mean HADV term are attributable to differences in the horizontal flow field in this particular box (Figure 12f). The horizontal OSCAR velocities are small during August/September compared with the model and in the case of the zonal velocity have a different direction. In the observational budget, VDIFF is only peaking by the end of the year and is much smaller from June to October than the model VDIFF.

Figure 16 shows the budget terms along a section crossing diagonally the Bay of Bengal (as marked in Figure 13i). The contribution of SEF and ENTR to ST is small along this section, whereas mean HADV, VDIFF, and the remaining terms are of importance in closing the budget. The mean HADV shows an oscillatory behaviour with maxima propagating to the northwest (not shown) from September onward. This is the
imprint of long Rossby waves with a phase velocity of approximately 0.05 m/s. Those waves have been previously described in the Bay of Bengal by Eigenheer and Quadfasel (2000). In Figure 16a, the oscillatory signal is evident in all advection-related terms. During both months, shear and eddy horizontal advection show an antiphase behavior compared to the mean HADV (counteracting it). The waves imprint an alternating meridional anomaly in the velocity field, in a certain phase advecting low-salinity waters from the more coastal region and in the opposite phase bringing higher-salinity waters from offshore.

Stratification is large close to the coast, where low-salinity waters are advected with the East India Coastal Current. These low-salinity waters and the resulting stratification favor an upward salt flux from below the mixed layer, which is expressed by the large VDIFF term in the near-coastal region around 85.2°E.

Last, we turn attention to the near-coastal region to analyze more in detail the impact of river runoff and precipitation on the coastal MLS budget. Figure 17 shows latitude-time plots of all budget terms averaged over a 220 km stripe along the western coast of the Bay of Bengal. ST (Figure 17a) shows a northern minimum between June and August migrating southward until September–October. Subsequently, a strong ST maximum can be observed between September and November followed by a slightly reduced salinification between January and May. As already discussed by Akhil et al. (2014), the salinity change is a result of the SEF by precipitation and river outflow (summer rainfall and the Ganges-Brahmaputra outflow to the north and the Godavari river at approximately 15°N), mean HADV within the East India Coastal Current and VDIFF due to strong stratification in the coastal region. The latter counteracts the surface freshening by SEF and mean HADV by providing an upward salt flux (Figure 17d). The contribution of entrainment is small compared with the other terms.

However, the budget cannot be closed without taking into account the small-scale high-frequency processes, namely without taking into account the shear HADV and the eddy HADV and VADV terms (Figures 17f–17h). These show magnitudes comparable with the other terms. The three terms show a patchy structure but propagation from the north to the south is still evident, with the largest contribution in the north between July and November, slightly later than the SEF onset (time needed for the circulation to respond to the freshwater input). The budget temporal evolution seems to follow the pattern: first a SEF signal starts in the north and stratification increases leading to increasing vertical fluxes and also to a vertical “shear” anomalies in the mixed layer. Then the increasing horizontal gradients between offshore and the coastal region enhances the mean HADV and associated small-scale mixing processes, ultimately translating into eddy HADV and eddy VADV enhancement.

Shear horizontal advection has a second maximum in the Godavari outflow region at about 15.5°N, when river discharge starts increasing and stratification enhances. Both shear and eddy horizontal advection are positive when SEF and mean HADV provide a negative contribution to the ST (June–September), which means they counteract the freshening by stirring the coastal fresher waters with saltier waters from the interior of the Bay of Bengal or from below the mixed layer. An observational study by Kumar et al. (2013) shows strong eddy activity in the near-coastal region in the western Bay of Bangal mixing low-salinity waters with high-salinity offshore waters. The eddy VADV is negative throughout the years with the maximum contribution to freshening within the mixed layer between June and December in the northern Bay.

Figure 16. Model MLS budget terms and MLD along the section marked in Figure 13i crossing the Bay of Bengal from east to west for the month of (a) September and (b) October.
of Bengal and also with a southward signal propagation until 11°N. The freshening pattern coincides with the salinification pattern of VDIFF, indicating a downward flow of fresher waters and an upward diffusion of saltier waters across the base of the mixed layer.

We therefore have shown that a high spatial and temporal resolution is essential to resolve all the processes contributing to the MLS variance and that especially in near-coastal regions, the small-scale high-frequency terms contribute to a large extent to the budget and that without them the balance can not be closed. Not only the dynamics in the coastal current region of the Bay of Bengal are of importance; also the processes in the interior are of interest showing a clear northeastward propagating signal in the mean horizontal advection, thus implying planetary wave dynamics.

All in all, the results of Akhil et al. (2014) can be confirmed that the freshwater input from rivers and precipitation is the dominant contribution to seasonal MLS variance in the northern part of the Bay of Bengal and that horizontal advection of low-salinity waters leads to the spread of fresher waters along the coasts to the east and west. Besides this, VDIFF is important in the strong stratified and shallow coastal regions by counteracting the surface freshening (Akhil et al., 2014). The large contribution of VDIFF can also be detected in the observational budget (compare Figures 6f and 8h), but the horizontal structures at the coast are heavily smoothed, which is mainly attributable to the scarcity of valid Argo data in this near-coastal region and the smoothing by objective mapping of salinity and MLD data. Additionally to the Akhil et al. (2014) conclusions, we found that the remaining budget terms play a determinant role in closing the budget, especially in the coastal regions where eddy and shear horizontal advection show mainly an opposite behavior.

Figure 17. Latitude-time plots of the MLS budget terms over the period 1988–1997 along a 220 km wide stripe adjacent to the western coast of the Bay of Bengal: (a) salinity tendency (ST), (b) surface external forces (SEF), (c) mean horizontal advection (mean HADV), (d) vertical diffusion (VDIFF), (e) entrainment (ENTR), (f) shear horizontal advection (Shear HADV), (g) eddy horizontal advection (Eddy HADV), and (h) eddy vertical advection (Eddy VADV).
compared to the mean horizontal advection. The contribution of eddy vertical advection is negative and decreases even more to the coast, where stratification is enhanced.

6. Summary and Concluding Remarks
The goal of this study was to identify all the mechanisms responsible for the annual variability of MLS in the Indian Ocean. To this end, satellite salinity data were analyzed along with several auxiliary data sets required to compute MLS budget terms. The output of a high-resolution ocean model was used to test the results inferred from the observations and to identify the role of salinity budget terms not possible to compute from the measurements.

The model and observational MLS budgets show comparable spatial patterns and amplitudes of the terms dominating the salinity tendency on seasonal time scales. Depending on the geographic location, these terms involve the mean horizontal advection, surface freshwater fluxes, vertical diffusion, and to a minor extent entrainment. Surface freshwater fluxes dominate seasonal salinity variations in the western Indian Ocean south of the equator, where the ITCZ has a pronounced annual cycle. There the surface freshwater fluxes account for more than 60% of the variance of the salinity tendency. Large-scale dominance of mean horizontal advection was found in the southern Arabian Sea, near-equatorial and off-equatorial regions explaining around 60%–90% of the variance of the salinity tendency. In addition, the ITCZ influences advection patterns along equatorial regions. Our results confirm these findings with the dominant contribution of advection in the vicinity of SEF-dominated regions balancing the forcing at the sea surface. These two terms have been discussed in past studies and the conclusions here obtained agree well with the conclusions therein.

From the terms which can be estimated from observations, mean horizontal advection and vertical diffusion constitute the largest error sources. We found that the main differences between model and observational budget terms come from differences in horizontal velocity fields, resulting in a different representation of salinity advection. In the observational case, a misrepresentation of advection in near-equatorial regions is expected, mainly attributable to the underestimated ASCAT wind stress fields in those regions (Bentamy et al., 2008).

Vertical diffusion is of importance in regions of surface ocean freshening (precipitation under the ITCZ and coastal areas due to river runoff) as well as in all regions of strong horizontal advection of low-salinity waters. Both influence the vertical gradient across the mixed-layer base and provide a salt flux into the mixed layer. This term has been disregarded in some of the past studies, but its importance acknowledged in more recent works. Entrainment is of importance for the seasonal MLS variance in upwelling regions in the open ocean and in the Arabian Sea, and at the northern edge of the trade winds, slightly south of the equator. Entrainment shows a strong seasonal cycle at the southern and northern edge of the equatorial roll, but varying not in phase with the seasonal MLS tendency.

Large proportions of seasonal variance in the satellite ST are captured by the sum of the main contributing terms, but large differences appear at some locations studied. The salinity budgets are not completely closed when using observational fields. However, results are promising in ocean regions away from the coast, like the NWIO and CIO, where approximately 80% of the seasonal ST variance is explained by the sum of the contributing terms. On the other side, in near-coastal areas like the BOB and the EEIO region, we are far away from closing the budget with observations. In the EEIO region about 60% of the seasonal ST variance is explained by the sum of the contributing terms but nearly zero in the BOB region. In the case of the model, 87% (EEIO), 98% (CIO), and 99% (NWIO) can be explained by the sum of SEF, mean HADV, VDIFF, and ENTR, except in the BOB, where these processes are of minor importance and eddy and shear advection drive the observed seasonal ST cycle. Eddy advection is a dominant contribution to the seasonal ST in the near-coastal eastern Arabian Sea and the Bay of Bengal. Horizontal and vertical eddy advection are always important, on the one hand, when/where the stratification within the mixed layer and their horizontal gradients is enhanced (baroclinic instabilities) and, on the other hand, due to instabilities of horizontally sheared flows. The former takes place, when low-salinity waters are advected with the East and West India Coastal Currents from the Bay of Bengal along the coastal Arabian Sea meeting high-salinity waters there and also in the Southern Bay of Bengal around 10°N where horizontal gradients of stratification are large. The contribution of vertical eddy advection is mainly negative, which could result from upwelling of fresh
water or downwelling of saltier water. Shear advection plays a major role in corresponding regions and at
the equator in frontal zones acting against the meridional salinity fronts and being the main contribution in
closing the MLS balance. In addition, in these regions the MLD gradients are large and the vertical anom-
lies of salinity and velocity are huge.

The eddy processes cannot be estimated using observations with low temporal and spatial resolution and
there are no observations giving the three-dimensional structure of the oceanic flow, necessary for the esti-
mation of the shear advection term. In the past, many observational (e.g., Ren et al., 2011; Ren & Riser, 2009;
Sommer et al., 2015) and model (Mignot & Frankignoul, 2003) studies disregard vertical fluxes and diffusive
fluxes at the mixed-layer base. Our study shows that the vertical diffusion cannot be neglected, reaching
even larger amplitudes than entrainment.

In closing we note that, like for other tracers or budget computations, an integrated observation/model
study is still required to investigate salinity tendencies. The model is the only tool to derive the full budget.
Our results suggest, however, that the satellite SSS can be a reliable proxy for MLS variations in regions with
a large signal-to-noise ratio and regions with a certain distance to the coast. Clearly, the higher resolution of
satellite products compared to in situ data provides a reduction of uncertainty (Argo profiles lead to process
aliasing) in coastal regions where small-scale variability is abundant. However, those regions are where the
satellite retrievals exhibit enhanced uncertainties due to land contamination and RFI. In the future, better
resolved in situ and satellite data sets are needed to adequately examine the contributing processes solely
with observations, in particular those eddy terms which cannot presently be addressed. Coastal regions are
specially of interest, where to date the Global Ocean observational system is insufficient.

**Appendix A: Salinity Tendency and Mean Salinity Advection Budget Terms**

Using the definition of vertical average and the Leibnitz integral rule, the tendency term in equation (1) can
be rewritten:

\[
\frac{\partial S}{\partial t} = \frac{1}{h} \int_{-h}^{0} \frac{\partial \tilde{S}}{\partial t} dz = \frac{1}{h} \left\{ \frac{\partial}{\partial t} \int_{-h}^{0} S dz - \frac{\partial h}{\partial t} \right\} = \frac{1}{h} \left\{ \frac{\partial}{\partial t} \left( \bar{h} [\tilde{S}] \right) - \frac{\partial h}{\partial t} \tilde{S} \right\}
\]

(A1)

Using the chain rule in the first term on the right-hand side of (A1) and after rearranging terms we obtain

\[
\frac{\partial S}{\partial t} = \frac{\partial \tilde{S}}{\partial t} + \frac{1}{h} \left( \bar{h} [\tilde{S}] - \tilde{S} \right) \frac{\partial \bar{h}}{\partial t}
\]

(A2)

Similarly, the mean advection term can be rewritten:

\[
\nabla \cdot (\bar{u} \tilde{S}) = \frac{1}{h} \int_{-h}^{0} \nabla \cdot (\bar{u} \tilde{S}) dz = \frac{1}{h} \left\{ \nabla \cdot \int_{-h}^{0} \bar{u} \tilde{S} dz - \bar{u} \tilde{S} \tilde{S} \cdot \nabla \bar{h} \right\}
\]

(A3)

Using the decomposition into vertical average and its deviation, the integral on the right-hand side above
can be rewritten:

\[
\int_{-h}^{0} \bar{u} \tilde{S} dz = \int_{-h}^{0} \left( \bar{u} \tilde{S} + \tilde{S} \tilde{S} \right) dz = \bar{h} \left[ \bar{u} \tilde{S} \right] + \bar{h} \left[ \tilde{S} \tilde{S} \right]
\]

(A4)

Substituting (A4) in (A3) and applying the chain rule we obtain

\[
\nabla \cdot (\bar{u} \tilde{S}) = \frac{1}{h} \left\{ \bar{h} \left[ \bar{u} \tilde{S} \right] + \nabla \cdot \left[ \tilde{S} \tilde{S} \right] \cdot \nabla \bar{h} + \bar{h} \left[ \tilde{S} \tilde{S} \right] - \tilde{S} \tilde{S} \cdot \nabla \bar{h} \right\}
\]

(A5)

Integrating the mean continuity equation in the vertical over the mixed layer and using the Leibnitz integral
rule gives

\[
\int_{-h}^{0} \nabla \cdot (\bar{u} \tilde{S}) dz - \bar{w} \cdot \nabla \bar{h} = \nabla \cdot \int_{-h}^{0} (\bar{u} \tilde{S}) dz - \bar{w} \cdot \nabla \bar{h} - \bar{w} \cdot \nabla \bar{h} = 0
\]

(A6)

Using the vertical average decomposition and rearranging terms we obtain

\[
\nabla \cdot (\bar{h} \tilde{u}) = \tilde{u} \cdot \nabla \bar{h} + \tilde{w} \cdot \nabla \bar{h}
\]

(A7)
Substituting (A7) into (A5) and collecting terms gives

$$\nabla \cdot \mathbf{S} = \nabla \cdot \mathbf{S}^0 + (\mathbf{S} - \mathbf{S}^0) \cdot \nabla \mathbf{h} + \mathbf{w} \cdot \nabla \mathbf{h} + \frac{1}{h} \nabla \cdot \mathbf{h} \left[\nabla \cdot \mathbf{S}^0\right]$$

(A8)

Finally, substituting (A2) and (A8) in the budget equation (1) in section 4 and after rearranging terms the budget equation (2) is obtained.

References


