

Upper Ocean Spatial Structure and Evolution along MISO Pathways in the Indian Ocean

N00014-17-1-2718

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Long-Term Goals

This project built on previous efforts that to understand monsoon variability in the region surrounding Sri Lanka and in the western Indian Ocean/Arabian Sea. Specific objectives include:

- Characterize the spatial and temporal variability of the mixed layer along MISO propagation pathways.
- Assess the dynamics that govern mixed layer evolution along the propagation pathways of MISO. Where and when do one-dimensional processes dominate (indicating atmospheric control)? How do advection and diapycnal mixing impact the interplay between atmosphere and upper ocean?
- Quantify exchange between the Bay of Bengal and Arabian Sea. How do these advective fluxes of heat and salt impact upper ocean evolution and MISO?

Work Completed

Efforts align with MISOBOD objectives and the broader, long-term goals pursued by past DRIs focused on variability in the Bay of Bengal and Arabian Sea. These include:

- Physics. Quantify key processes that govern both local and remote forcing of the circulation within the Bay of Bengal. Develop proxies/indices representing the seasonal and interannual variability of the currents and their relationship to atmospheric forcing at multiple scales. Investigate the continuity of features associated with the monsoon circulation system with regard to the interior circulation.
- Predictability. Improve the depiction of key ocean processes within different models. Construct and evaluate data proxies and indices for use as indicators of predictability.

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14. ABSTRACT A multi-year effort using long-endurance autonomous Seagliders and SGX gliders maintained continuous occupation of a section that sampled the boundary current off the southern tip of Sri Lanka, designed to understand exchange between the Arabian Sea and Bay of Bengal. Sampling spanned the period between April 2016 to December 2019, after which the COVID pandemic interrupted operation. The nearly four-year record demonstrates that the direct pathway along the Sri Lankan coast plays a significant role in exchange between the Arabian Sea and Bay of Bengal, with important contributions from both surface and subsurface flows.						
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Glider Operations

Long-term glider operations extended the time series of boundary current sections south of Sri Lanka, which aimed to quantify exchange between the Bay of Bengal and the Arabian Sea. This sampling built on experience gained through the ASIRI and NASCar DRIs.

A total of five glider missions have sampled the exchange section south of Sri Lanka, along 80.5°E to provide nearly continuous coverage from April 2016 to December 2019. Gliders provided roughly 75% coverage of this time period (998 glider-days over a period of 1297 days), collecting over 7700 profiles (Fig 2). These missions collected 36 sections of temperature, salinity, and geostrophic velocity (e.g., Fig 1) during all phases of the annual cycle. Long-term observations allow us form seasonal averages, reflecting specific phases of the monsoon cycle (e.g., Figs 3 and 4), and facilitate quantification of variability in water mass properties and transports.

The COVID-19 pandemic halted glider operations in early 2020, forcing us to pause sampling the southern Sri Lanka section. Personal safety issues along with travel restrictions make it impractical to resume during the remaining MISOBOD performance period, though we used the remaining port-pandemic time to negotiate arrangements that would allow south Sri Lanka operations within the framework of the next (EKAMSAT) DRI.

As part of the ship-supported studies in the interior of the Bay of Bengal, we also deployed 2 Seagliders during the July 2019 MISOBOD process cruise on R/V Sally Ride. These gliders, equipped with passive acoustics (to provide rain rate and wind speed estimates) and turbulence microstructure, sampled with the drifting buoys for several days, before heading to the long-term ONMI buoys. The two Seagliders then occupied a section through the central Bay of Bengal before positioning for recovery off the southern tip of Sri Lanka, completing six months of sampling. Analysis of these data is ongoing.

Analysis

Using four years of Argo data (2016-2019; overlapping with the sampling of the Southern Section), the properties of all profiles collected in the Arabian Sea overlaps slightly with those from the Bay of Bengal (Fig 5). Generally speaking, floats stayed in the basin where they were deployed - only 60 floats out of 455 floats with unique identification numbers moved across 80°E. While there is some overlap, the Bay of Bengal is generally much less salty than the Arabian Sea. Characteristics of 3 distinct water masses originating from both basins are defined by the density and salinity limits (Fig. 5a). Geographically, the fresh Bay of Bengal waters (B1, $32 < S < 33.5 \text{ g kg}^{-1}$ and B2, $33.5 < S < 34.5 \text{ g kg}^{-1}$) are mostly present at the northern end of the Bay of Bengal (Fig. 5), with connections on both sides of the Bay. These waters are detected in a few profiles in the Arabian Sea, with distributions that suggest connections immediately south of Sri Lanka and along the Equator (black arrows in Fig. 5b,c). Similarly, the high salinity waters (A1, $S > 35.25 \text{ g kg}^{-1}$) are present everywhere in the Arabian Sea, and gradually get deeper towards the east side of

the Indian Ocean along the Equator and South of Sri Lanka. In the Bay of Bengal, this water mass has been identified as the High Salinity Core (HSC), with various direct and indirect links to the Arabian Sea High-Salinity Waters. Locations of Argo profiles that contain these water masses (Fig 5d) suggest that the high salinity waters spread in two distinct bands, along the Equator and south of Sri Lanka, entering the Central Bay of Bengal mostly south of 8°N, below the surface.

A geographical representation of the seasonal transport of these water masses across the glider sections, overlaid with seasonal averages of SSH (Fig 6) shows that while the mean transports (solid bars) agree well with the expected circulation, particularly for the fresh water masses, blue, at the surface, and the modified Arabian Sea water, red, during the SW monsoon, individual sections can easily have water masses flowing the opposite way than the average.

These results are detailed in Rainville et al. (2022).

Training

As part of MISOBOP we continued training of NARA and Ocean University personnel (Sri Lanka) in general oceanographic field operations and INCOIS researchers and technicians (India) on glider operations and data analysis. This included lectures and short courses given at INCOIS and other academic institutions in India.

Products

Anutaliya, A., and Coauthors, 2022: Seasonal and Year-To-Year Variability of Boundary Currents and Eddy Salt Flux along the Eastern and Southern Coasts of Sri Lanka Observed by PIES and Satellite Measurements. *J. Phys. Oceanogr.*, 52, 3015–3031, <https://doi.org/10.1175/JPO-D-22-0030.1>.

Rainville L., C.M. Lee, P. Jinadasa, K. Arulananthan, S.U.P. Jinadasa, H.J.S. Fernando, W.N.C. Priyadarshani, and H. Wijesekera, 2022. Water mass exchanges between the Bay of Bengal and Arabian Sea from multi-year sampling with autonomous gliders. *J. Phys. Oceanogr.*, 52(10), 2377-2396. <https://doi.org/10.1175/JPO-D-21-0279.1>

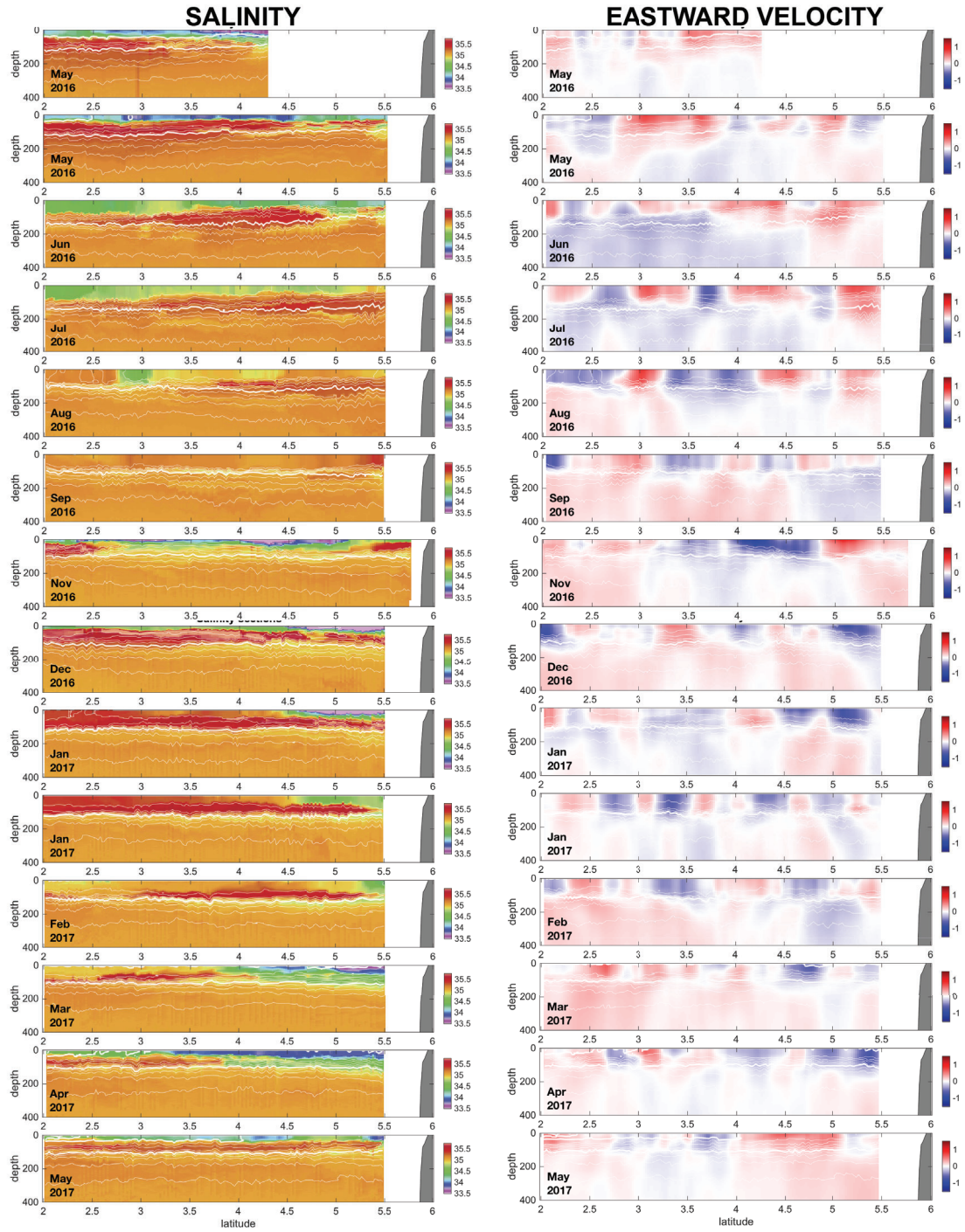


Figure 1. Meridional sections of (left) salinity and (right) zonal geostrophic velocity from the first 14 sections from 2°N to the coast of Sri Lanka, from SG526 and sg530.

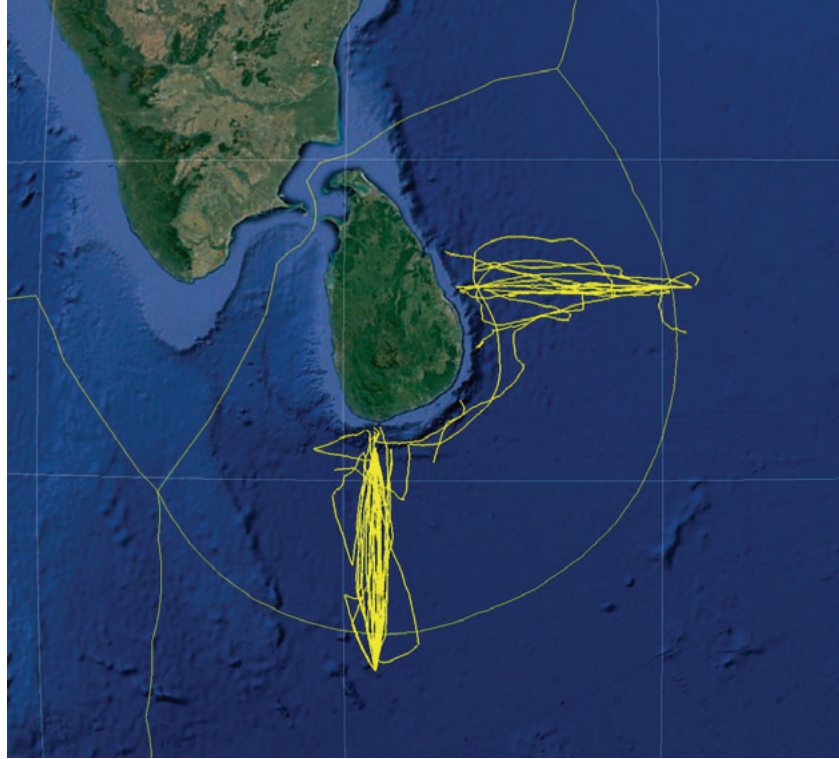


Figure 2. Map of the Seaglider tracks sampling the circulation around Sri Lanka from December 2013 to December 2019 (yellow). Boundaries of Exclusive Economic Zones are shown in white.

Southwest Monsoon JJA

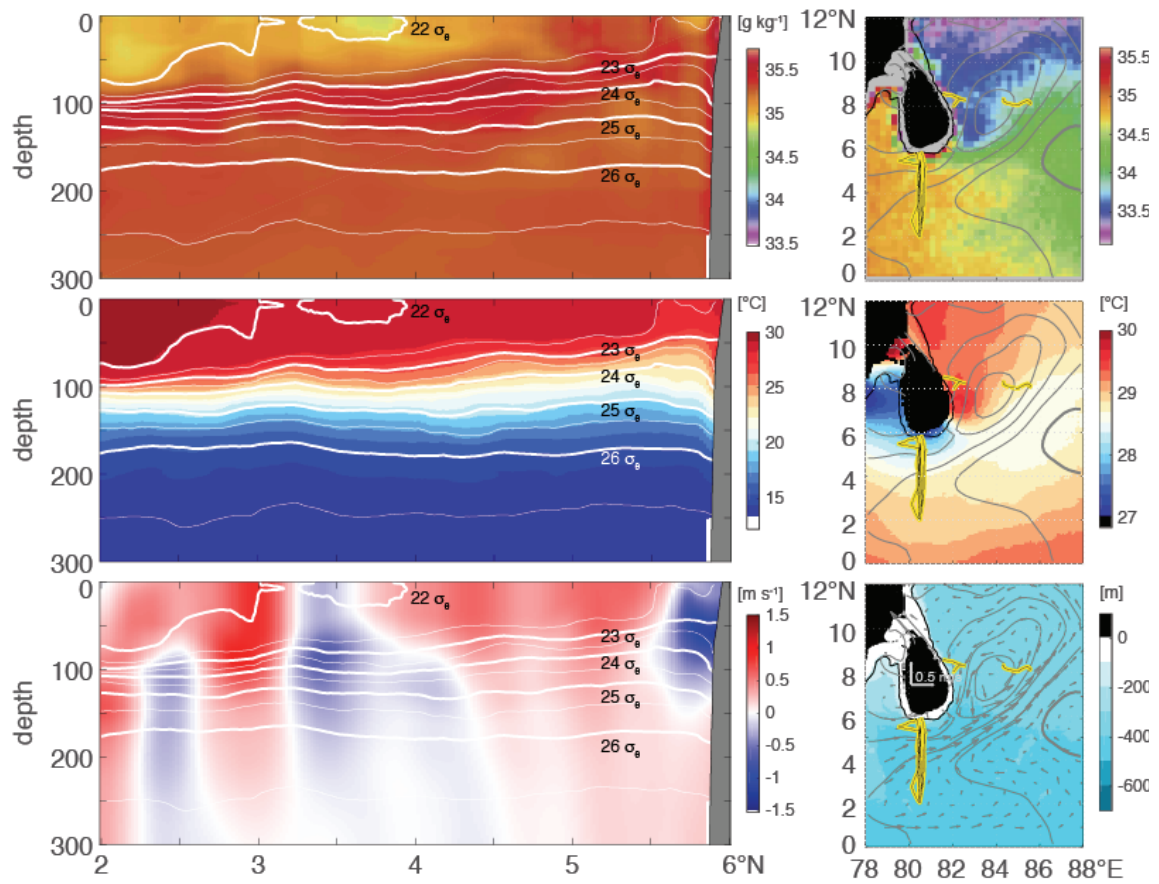


Figure 3. Salinity, Temperature, and eastward Geostrophic velocity during the Southwest Monsoon (June, July, and August) as functions of latitude and depth on the left, from gliders, and their surface values from remote sensing as functions of longitude and latitude on the right.

Northeast Monsoon DJF

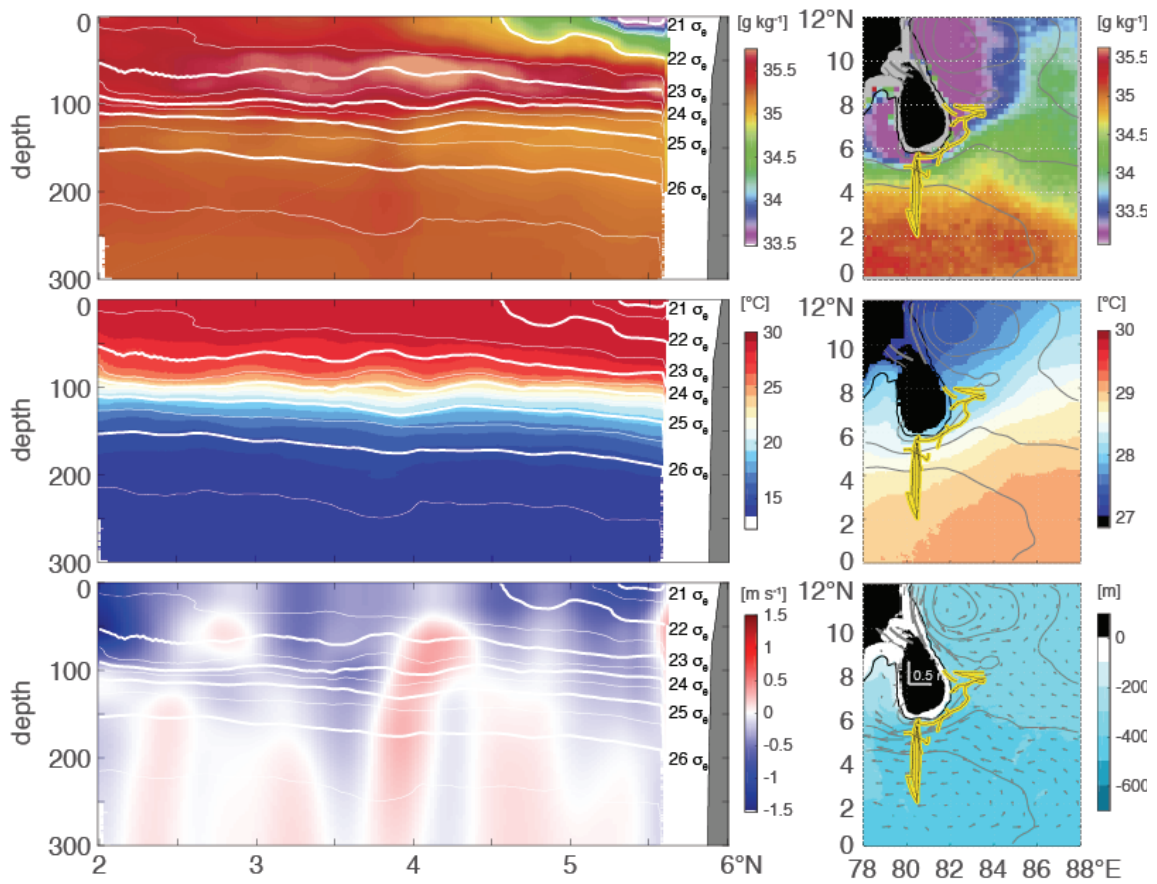


Figure 4. Salinity, Temperature, and eastward Geostrophic velocity during the Northeast Monsoon (December, January, and February) as functions of latitude and depth on the left, from gliders, and their surface values from remote sensing as functions of longitude and latitude on the right. Color scales are the same as for Fig. 3.

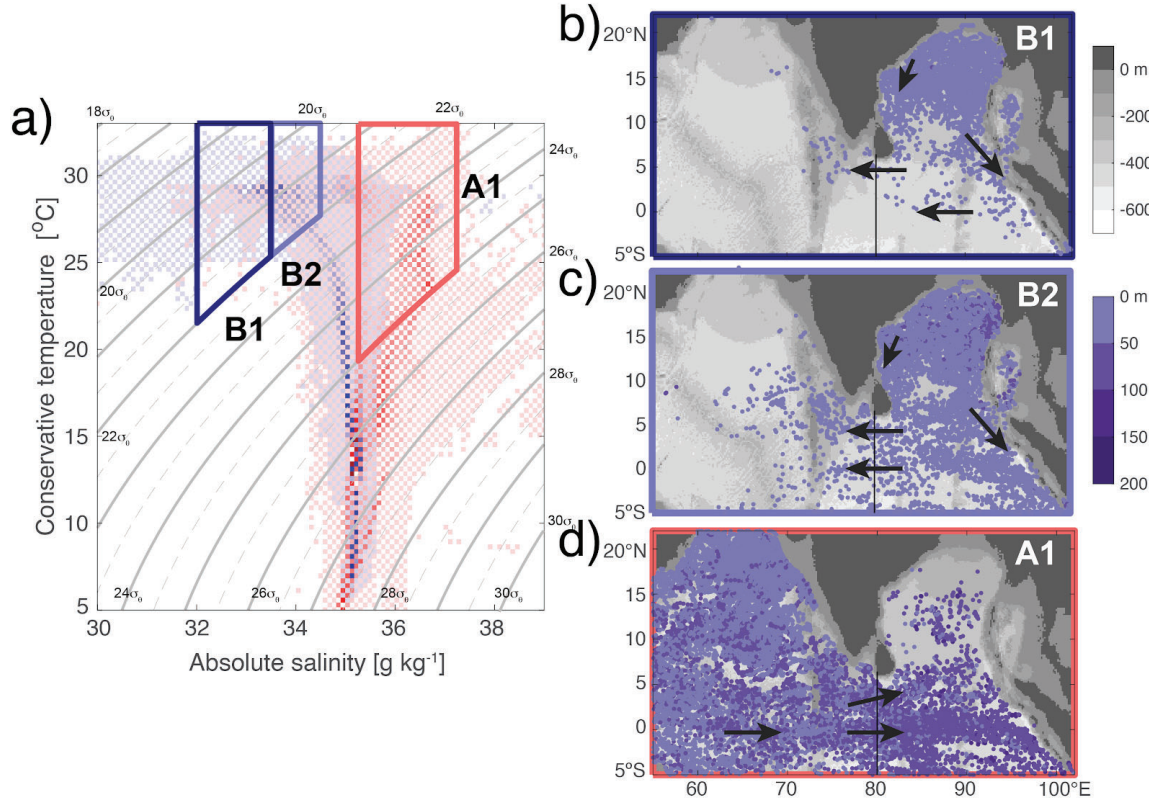


Figure 5. (a) Temperature-salinity diagram of all Argo float profiles (2016-2019) in the Indian Ocean north of 5°S in the Arabian Sea (red) or Bay of Bengal (blue). Each bin is split in two to show the overlap of both density distributions. (b-d) Maps of the locations of every float profile with given water mass criteria (boxes in (a)). Color of the dots indicate the depth where the water mass is found (purple colorbar). Depth is shaded in gray.

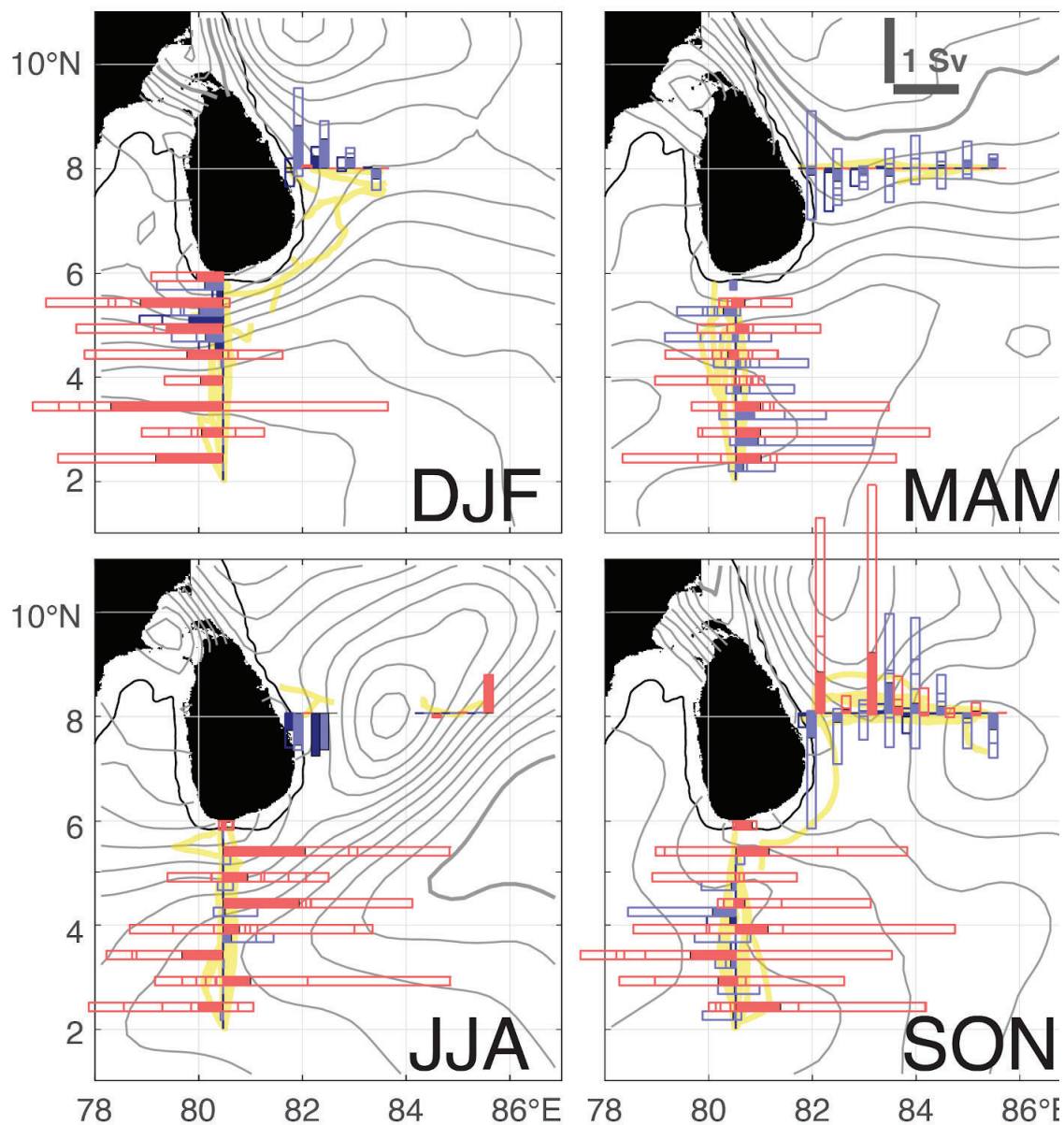


Figure 6. Seasonal circulation of important water masses (salty waters from the Arabian Sea [A1], in red; freshwaters from the Bay of Bengal [B1 and B2], in blue) captured by the gliders over several years, overlaid on the mean SSH contours (2.5 cm intervals) for each season. 1000-m isobath is shown in black. Seasonal mean transports are shown by the solid bars, and open bars represent individual sections. Yellow lines show the glider sampling for each season.

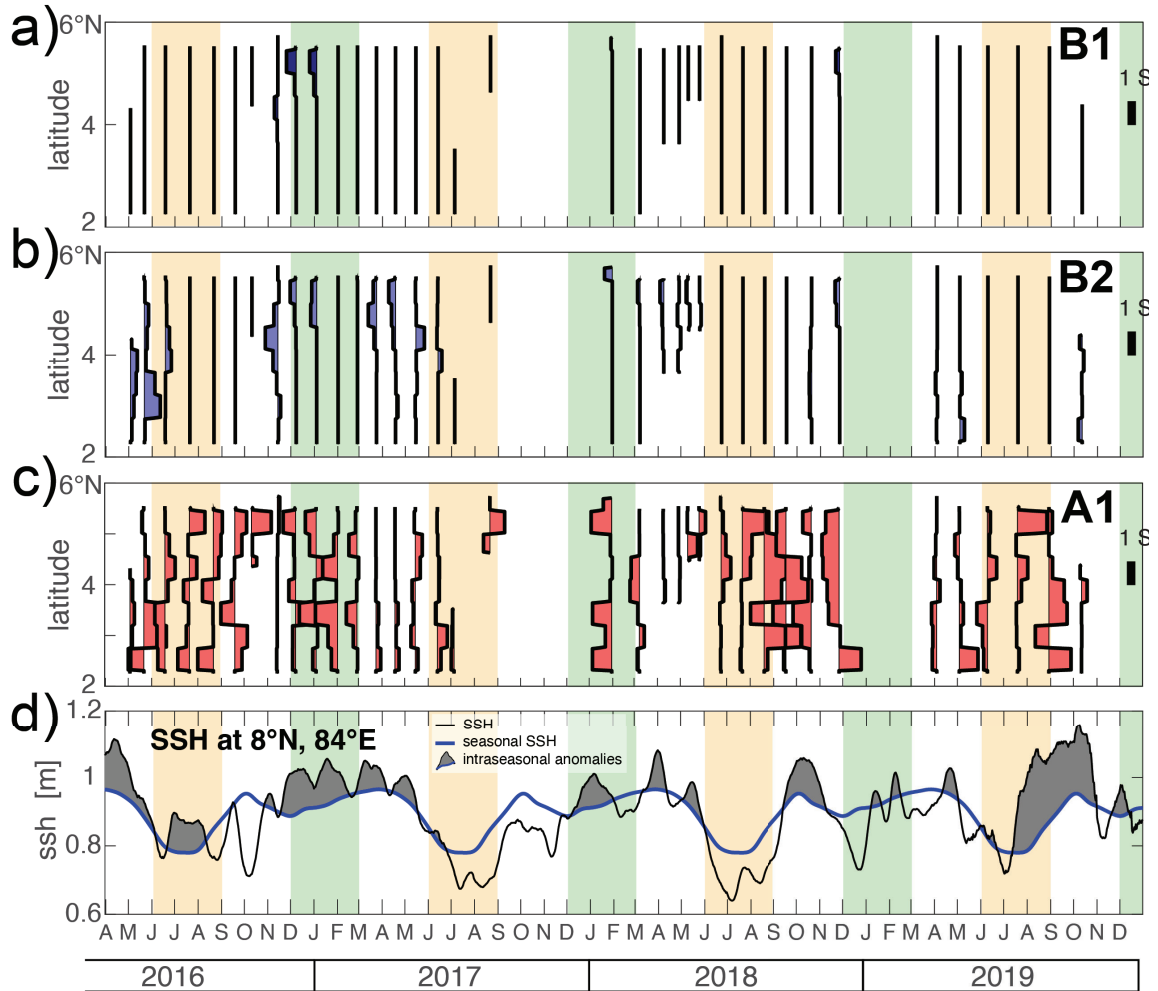


Figure 7. Transport of water with the properties of water masses B2, B1, and A1 (a-c) for all sections across 80°E as a function of yearday and latitude (in 50 km bins). (d) Sea-surface height time series in the middle of the Sri Lanka Dome, with the 10-year seasonal average (blue) and positive intra-annual anomalies highlighted.



Water Mass Exchanges between the Bay of Bengal and Arabian Sea from Multiyear Sampling with Autonomous Gliders

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ABSTRACT: We present high-resolution sustained, persistent observations of the ocean around Sri Lanka from autonomous gliders collected over several years, a region with complex, variable circulation patterns connecting the Bay of Bengal and the Arabian Sea to each other and the rest of the Indian Ocean. The Seaglider surveys resolve seasonal to interannual variability in vertical and horizontal structure, allowing quantification of volume, heat, and freshwater fluxes, as well as the transformations and transports of key water mass classes across sections normal to the east (2014–15) and south (2016–19) coasts of Sri Lanka. The resulting transports point to the importance of both surface and subsurface flows and show that the direct pathway along the Sri Lankan coast plays a significant role in the exchanges of waters between the Arabian Sea and the Bay of Bengal. Significant section-to-section variability highlights the need for sustained, long-term observations to quantify the circulation pathways and dynamics associated with exchange between the Bay of Bengal and Arabian Sea and provides context for interpreting observations collected as “snapshots” of more limited duration.

SIGNIFICANCE STATEMENT: The strong seasonal variations of the wind in the Indian Ocean create large and rapid changes in the ocean’s properties near Sri Lanka. This variable and poorly observed circulation is very important for how temperature and salinity are distributed across the northern Indian Ocean, both at the surface and at depths. Long-term and repeated surveys from autonomous Seagliders allow us to understand how freshwater inflow, atmospheric forcing, and underlying ocean variability act to produce observed contrasts (spatial and seasonal) in upper-ocean structure of the Bay of Bengal and Arabian Sea.

KEYWORDS: Indian Ocean; Monsoons; In situ oceanic observations; Interannual variability; Intraseasonal variability; Seasonal variability

1. Introduction

Upper-ocean processes occurring in the Bay of Bengal have strong impacts on Indian Ocean circulation. Large riverine freshwater (FW) discharge and excess precipitation into the Bay of Bengal produce strongly stratified surface layers, confining the impact of solar warming to relatively shallow mixed layers and thus producing elevated sea surface temperatures (Rao and Sivakumar 2003; Murty et al. 1992). Flow south of Sri Lanka, modulated by the monsoons and

equatorial dynamics, exports FW from the Bay of Bengal, linking the basin with the rest of the Indian Ocean and balancing the large evaporative fluxes of the Arabian Sea (e.g., Schott and McCreary 2001; Vinayachandran et al. 2013).

There is strong meridional contrast in upper-ocean structure across the Bay of Bengal (Fig. 1; e.g., Vinayachandran et al. 2012). Riverine FW input plays a greater role in the north, where the barrier layer is more prominent than in the south. The formation and evolution of barrier layers, linked to the modulation of vertical mixing and entrainment at the base of the mixed layer, play a major role in setting SST and modulating ocean–atmosphere interactions on a variety of time scales (Vinayachandran et al. 2002). Thin surface mixed layers due to the presence of freshwater enhance the oceanic response to atmospheric forcing. Strong atmosphere–ocean feedback in this region likely plays a

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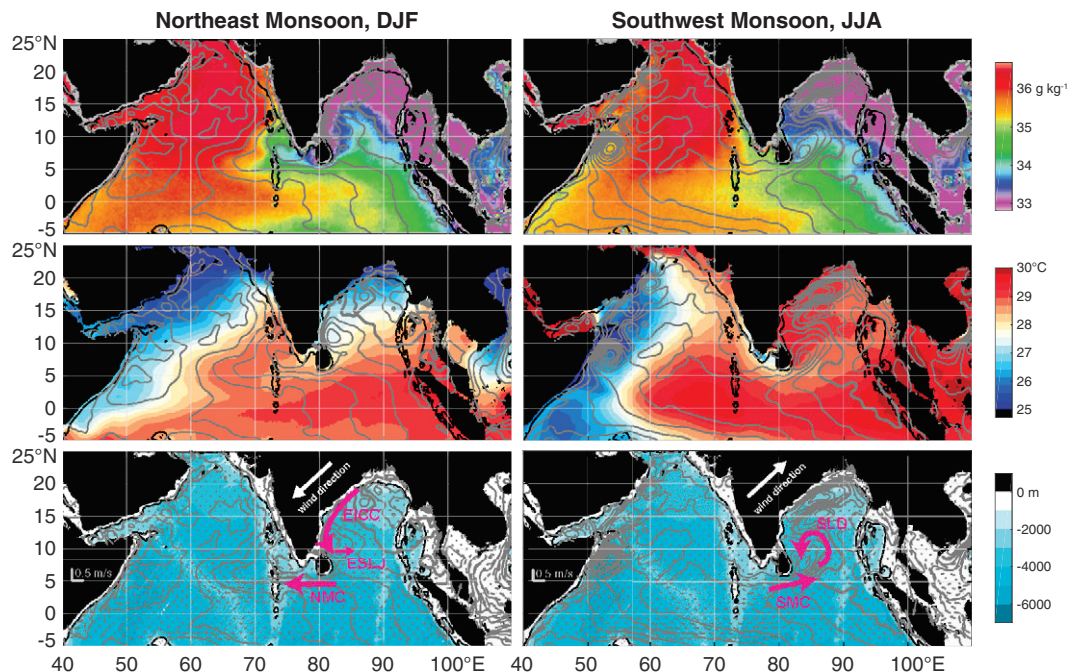


FIG. 1. (top) Maps of surface absolute salinity (from SMAP; Meissner et al. 2018), (middle) temperature (from optimally interpolated microwave and infrared satellites; Wentz et al. 2000), and (bottom) dynamical height and surface currents (from AVISO; Ducet et al. 2000) for the (left) northeast and (right) southwest monsoons. Maps are the seasonal averages for 4 years from 2016 to 2019. In winter, the Northeast Monsoon Current (NMC), East India Coastal Current (EICC), and East Sri Lanka Jet (ESLJ) are labeled. In summer, the Southwest Monsoon Current (SMC) and Sri Lanka Dome (SLD) are labeled.

critical role in determining the properties of water masses exported from the Bay of Bengal and the dynamics of atmospheric disturbances, in particular the monsoon intraseasonal oscillations (MISO) (e.g., Pirro et al. 2020a; Roman-Stork et al. 2020b).

Because it is so variable, the average seasonal circulation of the northern Indian Ocean has been difficult to describe (Shankar et al. 2002; de Vos et al. 2014; Su et al. 2021). Monsoon forcing produces distinctive, seasonal circulation patterns within the Bay of Bengal (Fig. 1). During the northeast monsoon (winter), northeasterly winds drive equatorward flow, carrying fresh Bay of Bengal water south within the East India Coastal Current (EICC). Argo floats, ship-based measurements, SSH, and models show that several eddies are typically positioned off the Sri Lankan coast, deflecting the bulk of the EICC eastward (Fig. 1) to form the East Sri Lanka Jet (ESLJ; Vinayachandran et al. 2005). This forces much of the Bay of Bengal freshwater to take a circuitous eastward path that eventually turns south and west to join the Northeast Monsoon Current [NMC, sometimes referred to as the Winter Monsoon Current (WMC)], the main wintertime communication pathway between the Bay of Bengal and the Arabian Sea. The NMC is further modulated by both coastal Kelvin waves propagating around the Bay of Bengal, and by Rossby waves propagating westward across the Bay of Bengal from the Sumatran coast after the collapse of the southwest (SW) monsoon (Rao et al. 2010; Pirro et al. 2020b; Wijesekera

et al. 2016a). Regardless, the properties of fresher water masses exported out of the Bay of Bengal depend strongly on their exit pathways—whether it is the direct route following the Sri Lankan coast or the more convoluted path through the Bay of Bengal.

During the SW monsoon (summer), southwesterly winds (Fig. 1) reverse the flow between the Arabian Sea and Bay of Bengal (Summer Monsoon Current, SMC) and drive coastal upwelling south of Sri Lanka forming a “mini cold pool” (Vinayachandran et al. 2004; Rao et al. 2006; Pirro et al. 2020a). The SMC entrains this upwelled water, carrying it northeastward into the Bay of Bengal interior. Wind stress curl associated with SW monsoon winds drives Ekman pumping east of Sri Lanka, forming a cyclonic gyre, the Sri Lanka Dome (SLD; Cullen and Shroyer 2019). While the summer monsoon winds drive poleward flow east of India, the western side of the SLD produces equatorward currents along the Sri Lankan coast. Our knowledge of the summer monsoon circulation has recently significantly improved from studies based on direct observations from ship transects (Wijesekera et al. 2015), glider observations in the middle of the Bay of Bengal (Webber et al. 2018), long-term moorings (Wijesekera et al. 2016a), and surface drifters (Hormann et al. 2019), all building on past estimates based on sparse observations. The work presented here complements these snapshots.

Dynamics in the Bay of Bengal and northern Indian Ocean are clearly not one dimensional. Regional numerical simulations

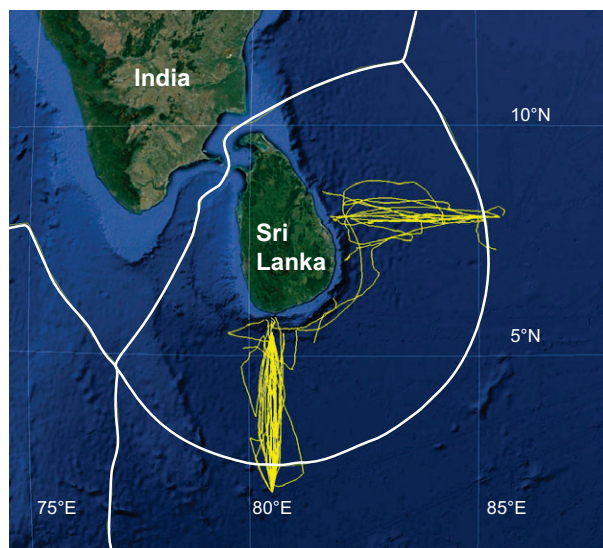


FIG. 2. Map of the Seaglider tracks sampling the circulation around Sri Lanka from December 2013 to December 2019 (yellow). Boundaries of Exclusive Economic Zones are shown in white. Background image from Google Earth.

highlight the interactions of variable winds and currents with local topography, in particular related to enhanced primary productivity in the wake of landmasses (Su et al. 2021). The observations described in this paper describe the variability of the ocean near Sri Lanka, a critical exchange point between the Arabian Sea and the Bay of Bengal. Multiyear glider observations (section 2) are used to describe the seasonal circulation around Sri Lanka including quantification of the mean, seasonal and intra-annual variations of volume, freshwater, and heat transport (section 3).

Discussion (section 4) focuses on understanding the connections between the Bay of Bengal, Arabian Sea, and greater Indian Ocean by tracing the dominant water masses. In particular, we focus on the freshwater exported from the Bay of Bengal, mostly during the northeast monsoon (Fig. 1), which is believed to play an important role in monsoon onset by creating a freshwater pool between the Maldives and India, modulating upper-ocean stratification, forming a barrier layer and a “mini warm pool” that amplifies air–sea feedbacks (Rao et al. 2015; Wijesekera et al. 2016a; Roman-Stork et al. 2020a). Similarly, high-salinity waters from the Arabian Sea entering the Bay of Bengal during the southwest monsoon influences interior stratification and plays a critical role in modulating atmosphere–ocean exchanges (Murty et al. 1992; Vinayachandran et al. 1999, 2018).

2. Data

a. Seaglider

Through a partnership between the Applied Physics Laboratory at University of Washington (APL-UW) and Sri Lanka’s National Aquatic Resources Research and Development Agency (NARA), 15 Seagliders were deployed in

joint operations that began in December 2013 (Fig. 2), conducted as part of the ASIRI (Air–Sea Interactions in the Northern Indian Ocean; Wijesekera et al. 2016b), NASCar (Northern Arabian Sea Circulation–Autonomous Research; Centurioni et al. 2017) and MISO-BOB (Monsoon Intraseasonal Oscillations over the Bay of Bengal; Shroyer et al. 2021) Department Research Initiatives from the Office of Naval Research.

Seagliders are small, reusable, long-range autonomous underwater vehicles designed to glide from the ocean surface to as deep as 1000 m and back while collecting profiles of temperature, salinity, and other oceanic variables with a vertical resolution of about 1 m in the upper part of the water column (Eriksen et al. 2001). Gliders steer through the water by controlling attitude (pitch and roll) and can thus navigate between waypoints to execute survey patterns. In the missions presented here, successive dives (each with a down and up profile) from a glider are generally separated by 4 km and 5 h. Typical horizontal speed is about 20 km day^{−1}. Mission durations depend largely on ambient stratification, profile depth, and instrument power, sometimes extending to nearly a year. Because the vehicles are relatively small and light, special handling gear is not required and field teams typically consist of one or, at most, two individuals. Standard sensor suites include pressure, temperature, and conductivity. Some of the missions near Sri Lanka also included sensors to measure dissolved oxygen, chlorophyll fluorescence, and optical backscatter, but these variables are not discussed here. Seagliders surfaced at the end of every dive cycle, downloading new commands and uploading data to a base station located at the University of Washington via Iridium satellite telemetry.

Initial processing is performed in near-real time. The different responses of temperature and conductivity sensors are accounted for and corrected through an analytical physical model (C. Eriksen 2018, personal communication; Morison et al. 1994; Lueck and Picklo 1990) integrated into the base station. A hydrodynamical flight model (Bennett et al. 2019) uses data from the glider’s attitude sensors and from the environment to estimate glider speed through the water, and thus location during the dive. The hydrodynamical model provides an estimate of the horizontal distance traveled through water in an ocean at rest, which, when compared to the actual positions at the beginning and end of the dive, provides a good estimate of the depth-averaged current (or, more accurately, ocean current averaged along the underwater trajectory of the glider). Repeated GPS fixes obtained during the surface drift, before and after every call to the base station, provides an estimate of ocean surface velocity.

Glider deployments and recoveries have taken place near the coasts from R/V *Samuddrika*, NARA’s research vessel, and small local chartered vessels.

b. Satellite remote sensing

Sea surface height products derived from altimetry (Ducet et al. 2000), along with optimally interpolated sea surface temperature (SST) and sea surface salinity (SSS) products, are used for situational awareness during operations and for the

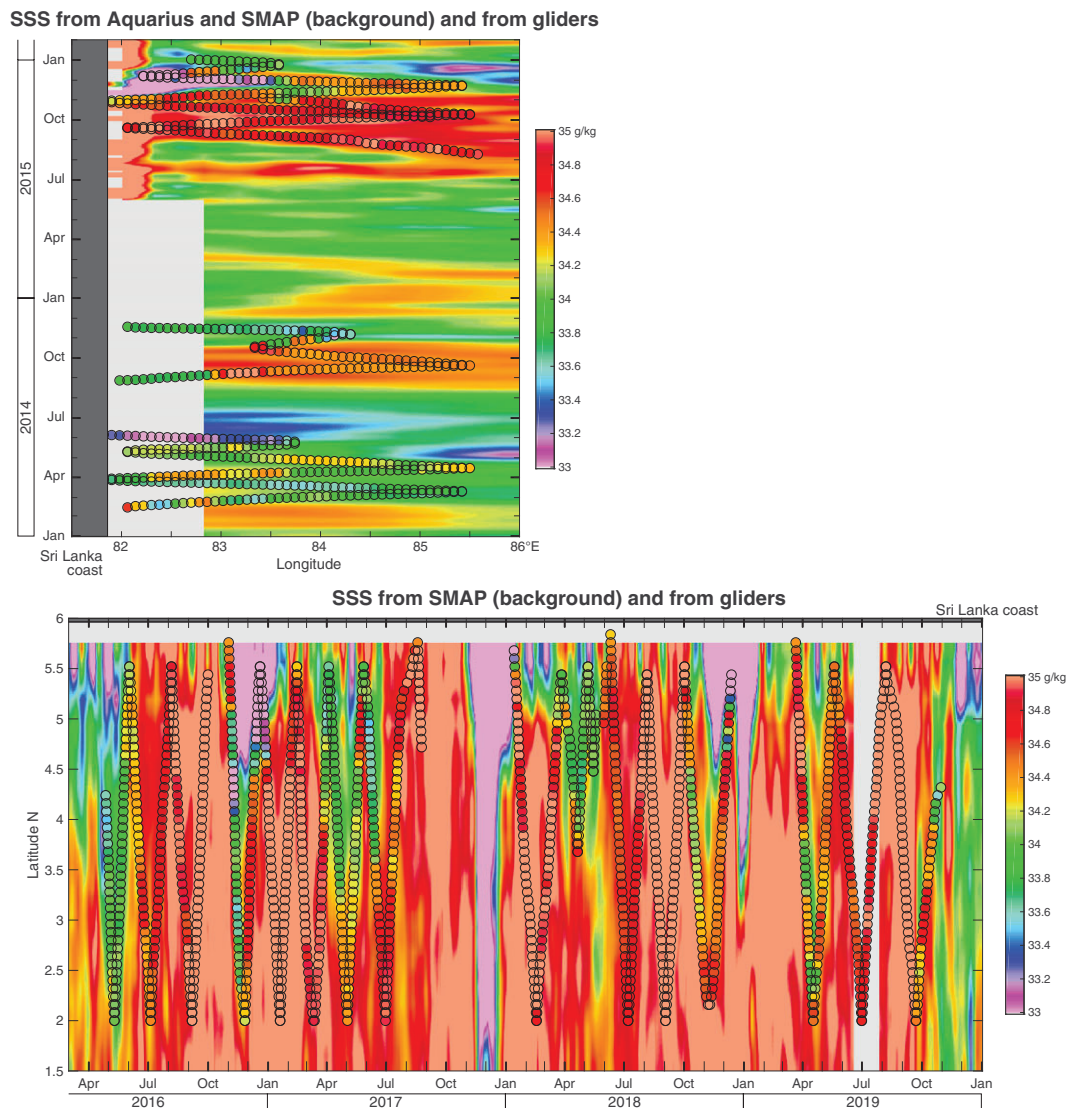


FIG. 3. (a) Map of SSS as a function of longitude and time along 8°N, from the coast of Sri Lanka to 86°E. Background colors are from satellite SSS (Aquarius until June 2015; SMAP after) and from the Seagliders in the black circle. Satellite data are effectively a 120-km running average. (b) As in (a), but SSS as a function of time and latitude along 80°30'E.

analysis presented here. Absolute dynamical topography is obtained from averaging the sea level anomalies over a prescribed period and adding the MDT_CNES_CLS18 mean dynamical topography (Mulet et al. 2021; also included in the files from Copernicus). SST data are a blended microwave and infrared optimally interpolated product (Wentz et al. 2000). Objectively mapped sea surface salinity from Aquarius resolving features at scales of about 150 km (Melnichenko et al. 2016) is used before June 2015, and from SMAP (Meissner et al. 2018) after April 2015. For SMAP, we use the 40-km v4 product, with an additional running mean filter of 120 km to create a product with similar scales as the Aquarius objective maps. Neither satellite captures the salinity near land, but improvement in the retrieval algorithm for SMAP v4

allows estimates within 30–40 km from the coast (Fig. 3a; Meissner et al. 2019). The general agreement between glider estimates of near-surface absolute salinity (top 5 m, on a 10-km horizontal scale) and satellite products (100-km scale) is encouraging. Direct comparison of the glider surface salinity (top 10 m) and SSS from satellite remote sensing (Fig. 4a) show little systematic bias (the linear regression, in black, has a slope of 0.88 ± 0.01), with a standard deviation of 0.32 g kg^{-1} .

c. Temperature, salinity, and velocity sections

Data collected by Seagliders operating near Sri Lanka can be organized into two sets of sections:

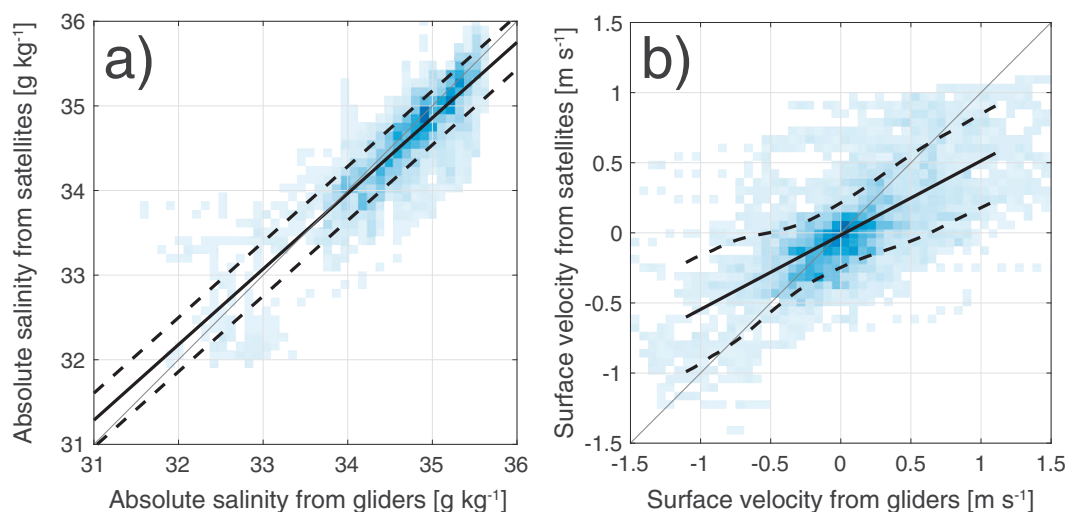


FIG. 4. Scatterplots of (a) glider vs satellite SSS and (b) glider vs satellite surface velocity. Colors provide a relative measure of the number of points. Solid lines are linear regressions and dashed lines show the standard deviation.

- Eastern section (8°N): A zonal section extending offshore at 8°N from the east coast of Sri Lanka. Six missions, conducted between December 2013 and January 2016, completed 17 full and partial occupations of this section, totaling 2243 dives, 433 glider-days (a complete day when a glider is actively sampling) distributed over 751 calendar days, and 7312 km total track distance.
- Southern section (80°30'E): A meridional section extending southward from the southern tip of Sri Lanka at 80°30'E. Five missions, conducted between April 2016 and December 2019, completed 36 full and partial occupations of this section, totaling 3863 dives, 998 glider-days (distributed over 1297 calendar days), and 12 964 km total track distance.

Both sections are roughly 400 km in length, and nominally require 3 weeks for a glider to traverse. Slow occupation speed makes it likely that both temporal and spatial variability contribute to patterns observed along any individual section, necessitating careful choice of scales when interpreting sections for temporal and spatial patterns of circulation and water mass variability.

To facilitate analysis, standard eastern and southern sections are defined as regular grids, with horizontal spacing of 2 km and vertical spacing of 1 m, extending between the nominal end points of each line. For each occupation, temperature and salinity measured along the glider's flight path are mapped onto the standard grid by averaging all observations collected within 10 km horizontally and 5 m vertically of each grid point, effectively smoothing with decorrelation length scales of roughly 10 km (horizontal) and 5 m (vertical).

Geostrophic shear of the horizontal velocity component perpendicular to sections is estimated from the thermal wind equation, i.e., from the horizontal slope of density surfaces. To estimate the absolute geostrophic velocity across the section, careful considerations must be taken in deriving the

reference depth-averaged current. Since small-scale structures are not expected to be in geostrophic balance, the density fields are further smoothed with a 40-km running mean. The uncertainty of individual estimates of depth-averaged currents, typically about 0.1 cm s^{-1} in the cross-track component (much less affected by the parameters derived as part of the flight model calculations, like drag or lift; Bennett et al. 2019) is likely reduced by the spatial average of several dives. Surface drifts, estimated from the glider positions recorded during the 15–20 min period when the glider is at the surface and transmitting its data, and glider-estimated depth average currents from the flight model are smoothed using the same horizontal scale. As the mixed layer can be deep and have significantly different speed than the stratified water, the reference absolute geostrophic in the top 1000 m is calculated by removing (smooth) surface drift over the depth of the mixed layer from the (smooth) depth-averaged currents. The final velocity sections include that same estimate of the surface drift over the depth of the surface mixed layer (assuming no vertical shear in the mixed layer), smoothly transitioning to the interior geostrophic velocity over a depth of 50 m. The velocity estimates are therefore not truly “geostrophic velocity,” but rather an estimate of the “subinertial” velocity across the sections. This technique is similar to adding an Ekman velocity component to the surface (e.g., Centurioni et al. 2009) but does not depend on knowledge of the wind or reanalysis products. Surface velocities from these sections agree relatively well with satellite altimetry products (Fig. 4). Glider estimates are often larger than the satellite estimates, consistent with larger smoothing scales used for the altimetry products. In Fig. 4b, the slope of the regression is 0.53 ± 0.01 for speed less than 1 m s^{-1} , but closer to 0.6 or even higher near zero (not shown), indicating that satellite altimetry often fails to capture large velocities because of the temporal and spatial scales used in the processing (Chelton et al. 2011). Standard deviations (calculated over sliding windows of width 0.5 m s^{-1}) are large,



FIG. 5. (top) Upper-ocean temperature, (middle) absolute salinity, and (bottom) northward velocity as functions of longitude and depth for the 17 sections along 8°N. Only the top 200 m (out of 1000 m) are shown.

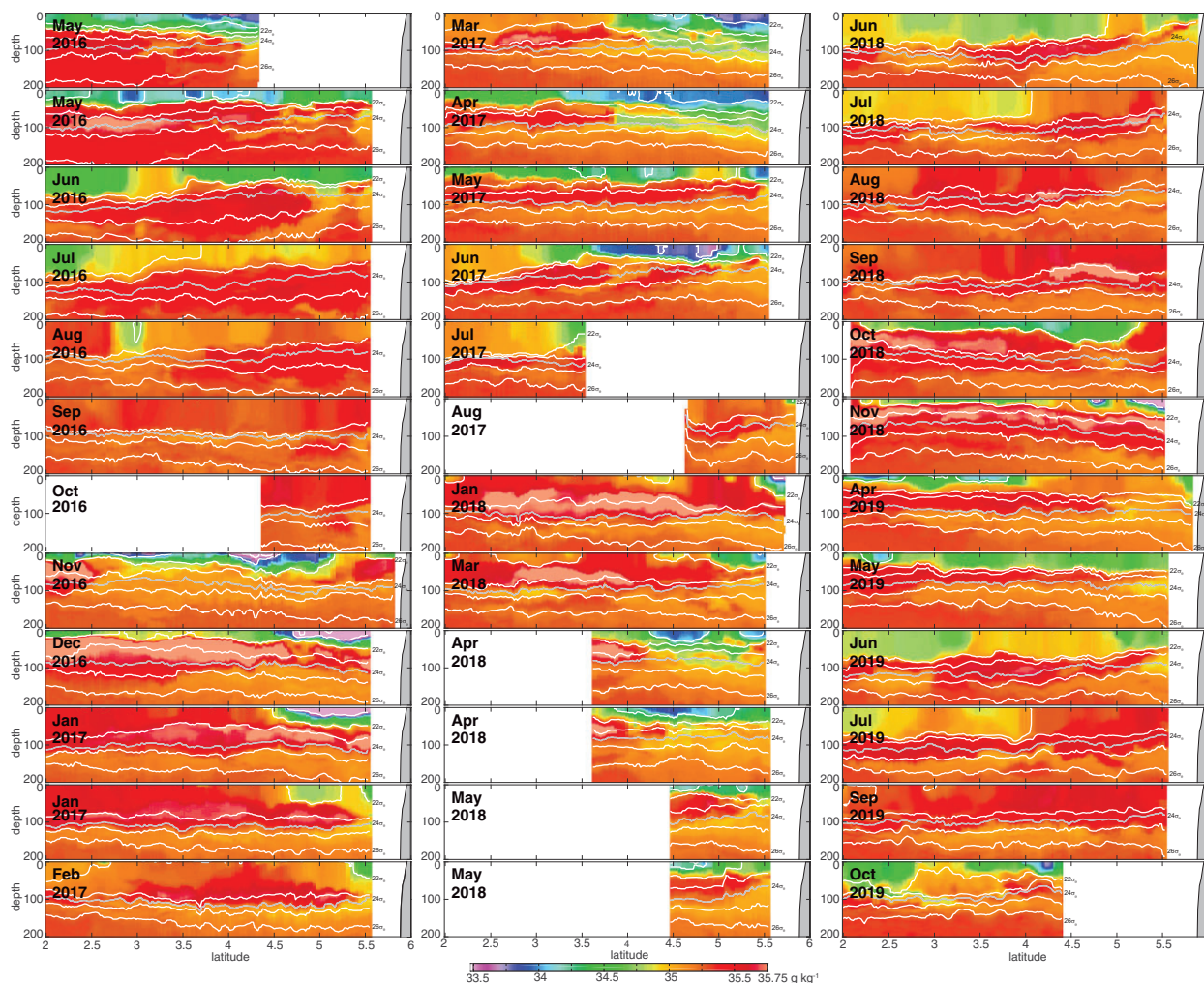


FIG. 6. Upper-ocean absolute salinity as a function of latitude and depth for the 36 sections for the southern section along $80^{\circ}30'E$. Only the top 200 m (out of 1000 m) are shown. Isopycnals are plotted in white (every 1 kg m^{-3} , $24\sigma_{\theta}$ in light gray).

on the order of 0.35 m s^{-1} . Comparisons with the Ocean Surface Current Analyses Real-Time (OSCAR) surface velocity product (Bonjean and Lagerloef 2002), which include a wind-driven component and arguably a better representation of the ocean surface velocity, lead to very similar results. We note that when comparing the OSCAR velocities evaluated at the glider locations with the same for AVISO, they agree much higher with a slope of 0.9 and a standard deviation is 0.21 m s^{-1} . The mismatch between the gliders and both altimetry-based estimates are therefore likely due to intrinsic differences on how the fields are sampled by altimetry and gliders.

1) EASTERN SECTION

Sections along $8^{\circ}N$ (Figs. 3a, 5) capture the complex circulation near Sri Lanka, including the narrow southward advection of freshwater near the coast (e.g., September 2014 section) and the subsurface high-salinity water from the Arabian Sea entering the Bay of Bengal (the salty subsurface northward current in fall of each year). Importance

of sampling near the coast is particularly obvious when comparing surface glider observations with estimates from satellites (Fig. 3), where the freshwater near the coast is often missed by remote sensing from space.

Between December 2013 and August 2015, a subsurface mooring was deployed by the United States Naval Research Laboratory at $8^{\circ}N$, $85^{\circ}30'E$, capturing upper-ocean velocity, temperature, and salinity (Wijesekera et al. 2016a). While the mooring provides great temporal resolution of the properties at this location (Pirro et al. 2020a), it minimally contributes to the spatial sections presented here (mooring data appear as a line in Fig. 5, May 2014, October 2014, and November 2014, mostly in temperature). Mooring data are included here for completeness.

The long sampling gap in early 2015 is a direct consequence of the difficulty to get sustained measurements in regions with difficult access. Operations for the $8^{\circ}N$ were complicated by the availability of R/V *Samudrika* and access to the port of Trincomalee for political reasons.

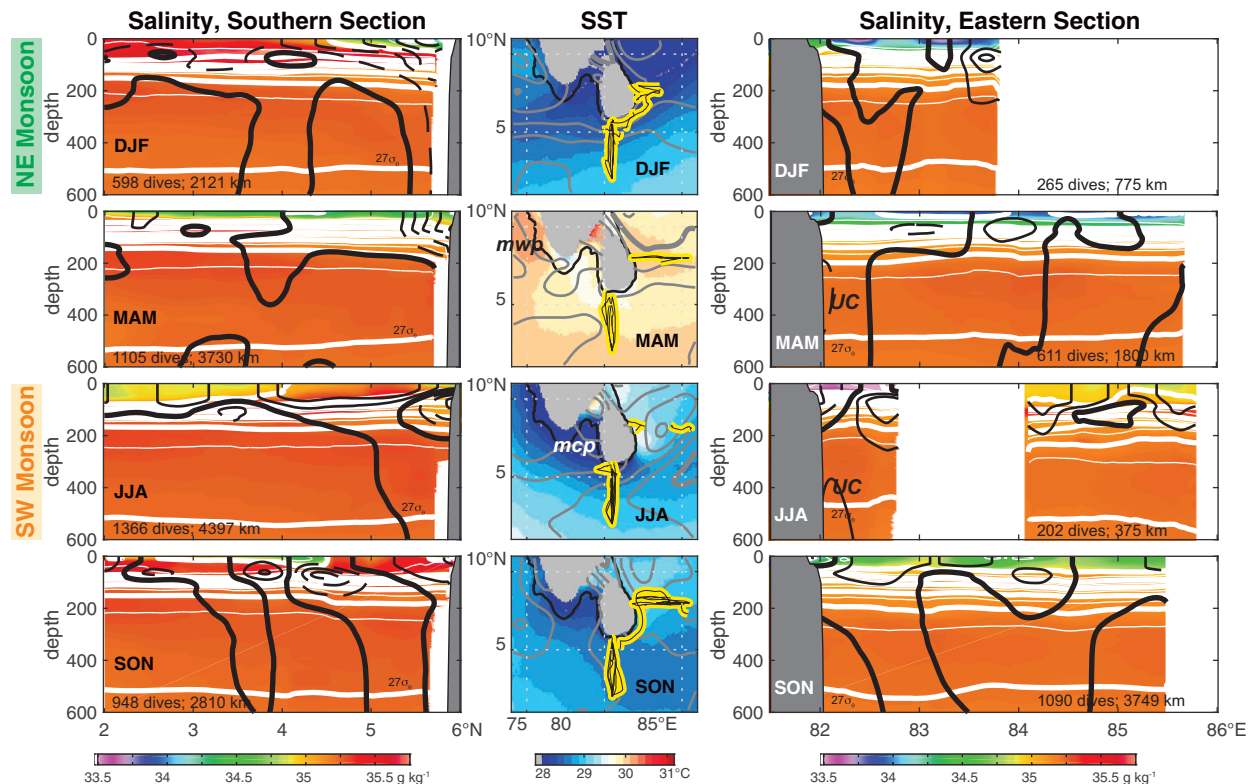


FIG. 7. Seasonal average of the upper-ocean absolute salinity as a function of (left) latitude and depth for the southern section and (right) longitude and depth for the eastern section. Only the top 600 m (out of 1000 m) are shown. The range for absolute salinity is the same for all salinity sections. Isopycnals are plotted in white (thin contours every 0.5 kg m^{-3} , 1 kg m^{-3} in bold). Cross-section velocity is contoured in black [0.5 m s^{-1} intervals; 0 m s^{-1} in bold, positive values (east or north) are solid, negative are dashed]. The Sri Lanka Undercurrent (UC) is indicated. (center) Averaged maps of SST corresponding to the same 3 months, with glider tracks (yellow) and SSH contours (5 cm intervals, gray) overlaid. In italic, the Arabian Sea mini warm pool (mwp) in the spring season is indicated, and the mini cold pool (mcp) in the SW monsoon.

Glider sampling, completing each section in less than a month, is well suited to capture the main modes of variability seen in satellite sea surface salinity, with the direct observations extending closer to the coast in a region important for freshwater export. There is significant seasonal and interannual variability in all fields.

2) SOUTHERN SECTION

Seasonal sampling on the southern section along $80^{\circ}30'E$ (Figs. 3b, 6) is more regular and spans a longer time period than that at $8^{\circ}N$. Some of the sections start near the shelf break at $5^{\circ}50'N$, but operationally the 40-km region between $5^{\circ}30'$ and $5^{\circ}50'N$ was avoided because of heavy ship traffic and fishing activity. Note that the continental shelf south of Sri Lanka is less than 10 km wide (coastline along $80^{\circ}30'E$ is located at $5^{\circ}56'N$).

The southern section exhibits considerable variability across individual occupations. For example, the 36 sections of absolute salinity collected along the southern section (Fig. 6) show strong interleaving of fresh and salty water masses, and narrow filaments of freshwater particularly near the coast, but also throughout the section. Changes

from section to section observed in the glider sections are not unlike the variability observed between a series of ship-based measurements obtained in the early 1990s (Schott et al. 1994).

3. Results

a. Seasonal average of upper-ocean sections

Glider sampling is not uniformly distributed in time, with significant gaps in some seasons. For the southern section, there are at least five complete sections during a given 3-month period, which allows us to calculate a good estimate of the seasonal average from close to the coast of Sri Lanka to $2^{\circ}N$ (Fig. 7). The seasonal cycle measured by the gliders shows the freshwater concentrated near the coast during the winter months, and a variable circulation comparable to that described by numerical simulations of Su et al. (2021) forced with reanalysis winds, emphasizing how reversing monsoon winds and regional bathymetry are responsible for setting up the seasonal circulation around Sri Lanka.

Sparsely sampling along $8^{\circ}N$ makes the 3-month average less representative and with more gaps (Fig. 7). Seasonal

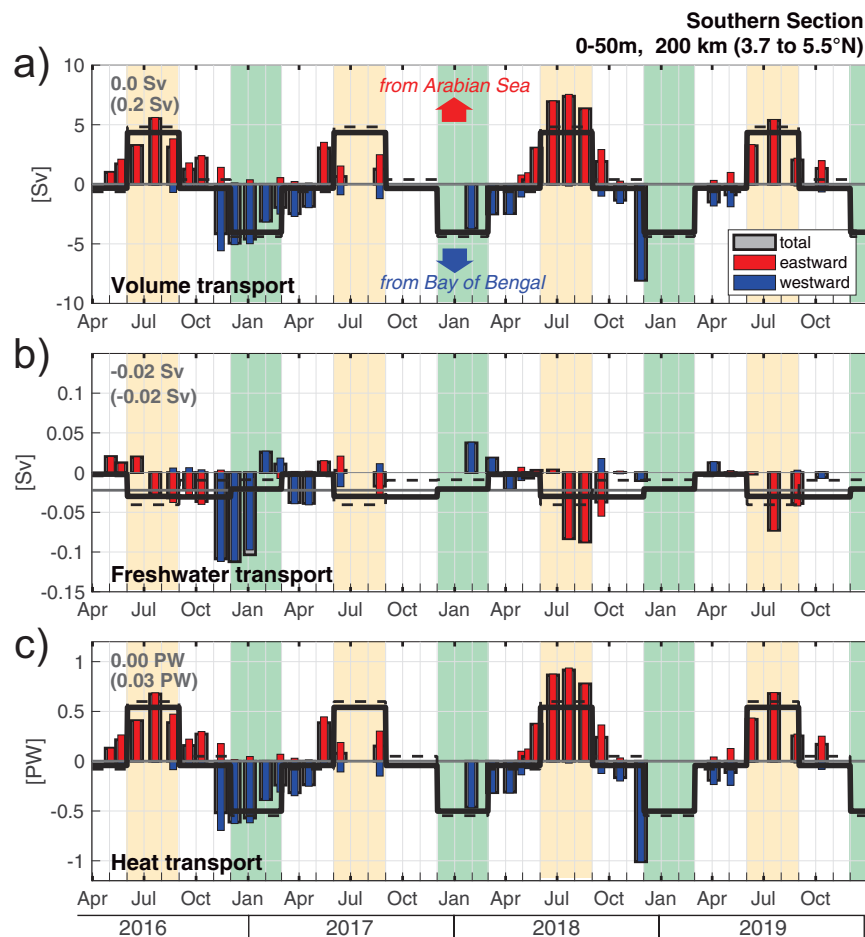


FIG. 8. Time series of the (a) volume, (b) freshwater, and (c) heat transport across the top 50 m and the portion of the southern section within 200 km of the coast, separated by the direction of the flow. Solid black lines indicate the seasonal average of all transports, and dashed lines the transports of the seasonal averages. Annual averages are listed on the top-left corner (transports of the mean sections in parentheses) in gray and illustrated by a constant gray horizontal line.

averages are consistent with previous observations and remote sensing. Freshwater is generally present near the coast of Sri Lanka, particularly in the winter months. Saltier waters are present in the summer months across the whole southern section and offshore in the eastern section, where the end of our observations (near 86°E) connect to where the sampling of Webber et al. (2018) begin—this is where the core of the high-salinity waters of the SMC are detected (Jinadasa et al. 2016).

Even in the seasonal average, sections show horizontal and vertical variability. Velocity and water properties below the surface make it difficult, or impossible, to infer the total transports from only surface properties. As with every individual section, we note that the velocities often reverse at depth. In particular, the undercurrent near the coast of Sri Lanka on the 8°N line, discussed in Anutaliya et al. (2017) based on pressure-inverted echosounders (PIES), models, and a subset of the glider observations presented here, is apparent in both

the seasonal average (particularly during the summer monsoon) and several sections.

Currents are strong in the upper 100–200 m. Low-salinity waters are often found near the coast of Sri Lanka, carried southward in the summer across 8°N and westward across 80°E during almost all seasons. Surface salinities agree well with seasonal averages from satellites, but the section emphasizes how the vertical profiles of salinity change at different places and times.

Signatures of the Sri Lanka Dome, described in more detail in de Vos et al. (2014) and Cullen and Shroyer (2019) and Lozovatsky et al. (2019), are not obvious in the averaged sections (Fig. 7, JJA on the eastern section) due to poor sampling. The upward doming of isopycnals associated with the cyclonic Sri Lanka Dome is a bit clearer in the individual sections bracketing the southwest monsoon (e.g., Fig. 5, June and September 2014 sections, with isopycnals being shallowest in the middle of the sections).

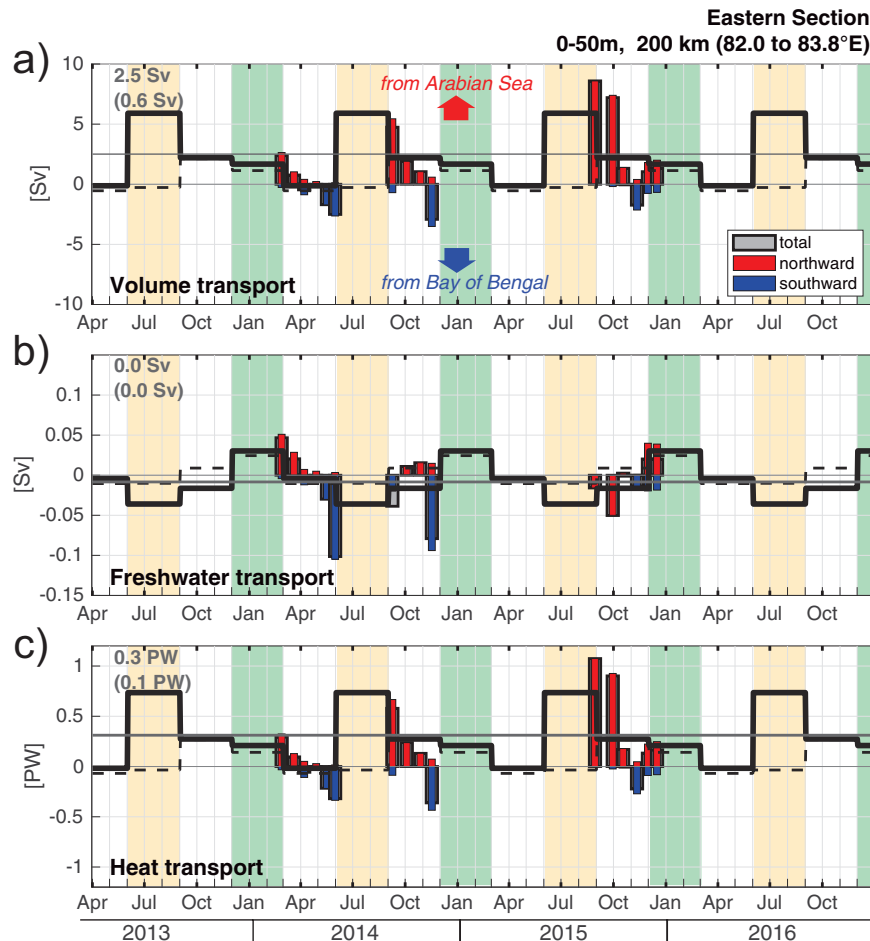


FIG. 9. As in Fig. 8, but for the eastern section within 200 km of the coast.

b. Ocean transport

With the estimates of velocity derived from the glider measurements, an observational estimate of volume, freshwater, and heat transports can be obtained from each section. While salt balance is a fundamental concept for any fixed volume of ocean (or the global ocean, e.g., Talley 2008), several past studies have discussed the concept of “freshwater transport”:

$$T_{FW} = \iint u_{\perp} \left(\frac{S_{ref} - S}{S_{ref}} \right) dx dz,$$

which effectively the freshwater fraction transport across a section (double integral) with salinity $S(x, z)$ and velocity component $u_{\perp}(x, z)$ perpendicular to the section, relative to a reference salinity S_{ref} . Reference here is 34.83 g kg^{-1} , corresponding to the same reference salinity for the Indian Ocean used by Roman-Stork et al. (2020a) for their freshwater flux component. The absolute salinity value referred to above is equivalent to 34.67 psu, the mean salinity of the basin. While we acknowledge the limitations and arbitrary nature of the freshwater transport (see, e.g., Schauer and Losch 2019), it is

a useful parameter to describe the net freshening impact of both the northeast and southwest monsoon currents. A more robust metric, namely, the volume transport of water matching specific salinity and density criteria, is presented in section 4.

Along the southern section, positive volume transport indicates water flowing eastward, from the Arabian Sea to the Bay of Bengal. The strongest surface layer transport (Fig. 8a) is negative during the NE monsoon (winter, JJA), and positive during the SW monsoon (summer, JJA), in line with previous observations. The transport estimates match previous estimates, for example the eastward transport of 8 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) estimated by Schott et al. (1994) during the SW monsoon north of $3^{\circ}45'N$ along the same meridional section. Other estimates, nicely summarized in Shankar et al. (2002), do vary quite a bit but are consistent with our estimates. Anutaliya et al. (2017) and Anutaliya et al. (2022) find great agreement between the direct (slow) estimates from gliders and those from PIES deployed concurrently to our surveys.

These surface currents generally carry low-salinity waters westward ($S_{ref} - S > 0$, $u < 0$) during the NE monsoon, and high-salinity water eastward ($S_{ref} - S < 0$, $u > 0$) during the SW monsoon, resulting in both cases into a negative freshwater

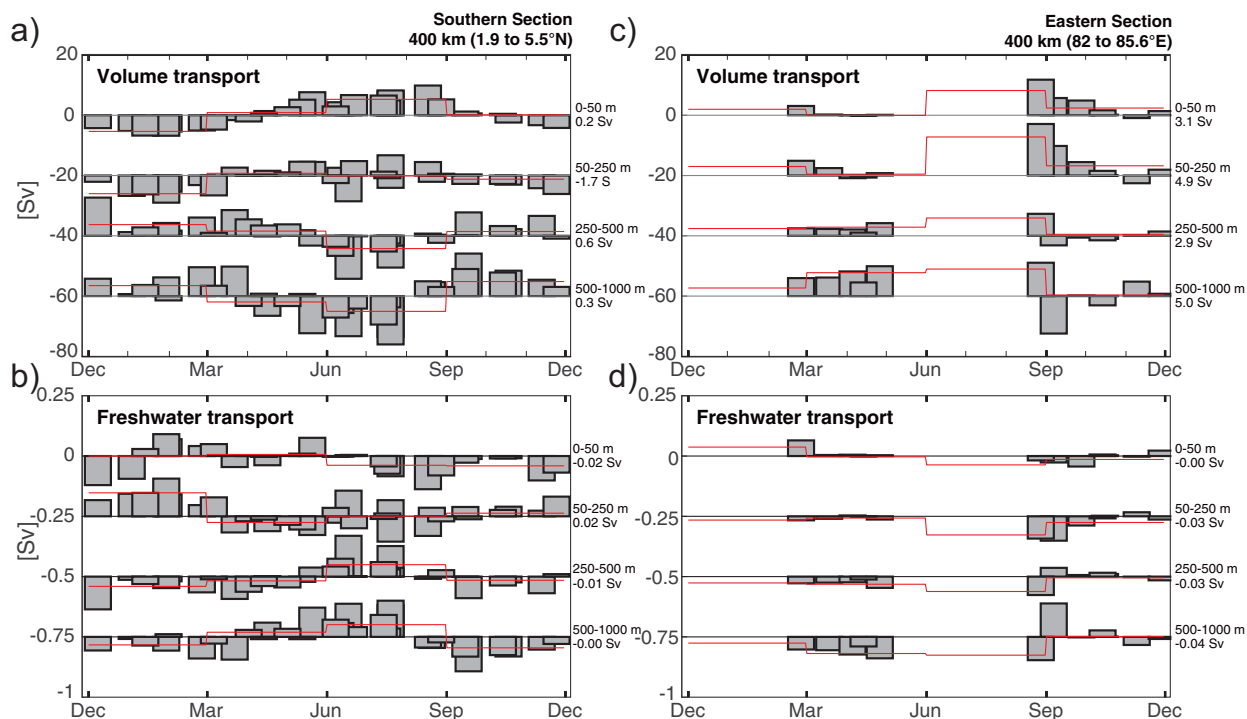


FIG. 10. Time series of the (a),(c) volume and (b),(d) freshwater transport for the (left) southern and (right) eastern sections for four different depth ranges, plotted in terms of year/day. Red lines show the seasonal average of transports from individual sections, with the annual average listed on the right.

transport (Fig. 8b). Heat transport (Fig. 8c) are large and follow the volume transports. They are included here for completeness. While interannual variations in the heat transport follows that of the volume transports (e.g., strong eastward volume and heat transport in JJA 2018 vs 2019), the interannual variability of freshwater transport does not follow the same pattern, therefore largely due to salinity anomalies as opposed to current anomalies. Note that transports shown in Figs. 8 and 9 are integrated only in the first 200 km from the coast and any individual section (displayed as a bar) is excluded if less than 75% of that distance is sampled.

Surface transports for the eastern section (Fig. 9) similarly are consistent with the expected seasonal circulation: averaged flow is northward and carries salty waters northward into the Bay of Bengal (negative FW transport, most evident in the average, black line) during the SW monsoon (JJA). Sections closer to the winter monsoon are generally southward and fresh, hence also associated with a negative FW transport. Note that the time span shown for the eastern section (Fig. 9) is the same as for the southern section (Fig. 8) but 3 years earlier, to illustrate how each line is sampled relative to the other and facilitate comparison.

The transports can be calculated for multiple layers, as the gliders provide continuous measurements of the upper 1000 m. Total volume and freshwater transport for several vertical layers, integrated across the entire sections (400 km), are plotted as functions of year/day in Fig. 10. Sections from different years often overlap, and the 3-month average for each layer is

shown as the red line. Collapsing the data onto year/day obscures interannual trends, instead emphasizing variability and the seasonal cycle.

The vertical structure of the transports is striking (Fig. 10). We note that these four layers have comparable transports despite vastly varying thickness. Transports at depths below 250 m are often opposite to those of the upper ocean. Deeper transports (500–1000 m) are generally in the same direction as those in the 250–500 m depth range.

Both sections (Figs. 8, 9) were sampled for several years, but sampling is uneven. As before, seasonal averages can be calculated for each variable, and these mean salinity and mean velocity sections are used to construct the “transport of the mean sections,” shown as dashed lines in Figs. 8 and 9. Arguably a more accurate measure of the mean transport is obtained by seasonally averaging the transports (solid lines in Figs. 8–10). Discrepancies are mostly due to poor sampling of several seasons (e.g., JJA along eastern section, Fig. 9), but also to the high variability of both salinity and velocity (i.e., the cross terms are important). The net annual volume and heat transports are small for the southern section (Figs. 8a,c, solid gray line and the 0.1 Sv value listed in the top-left corner), but large for the eastern section (Figs. 9a,c), likely due to poor sampling in both monsoon seasons. Throughout the paper, the seasonal mean and standard deviation are calculated from individual sections, weighing the portion of each section that fall within a given 3-month period.

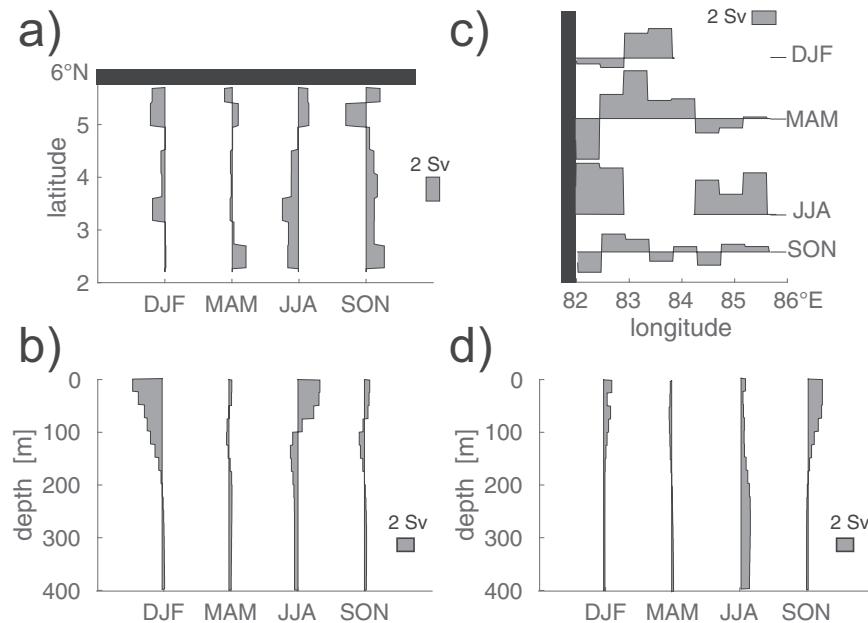


FIG. 11. Seasonal average of volume transports for the southern section as functions of (a) latitude in 50-km bins integrated over the top 250 m, and (b) of depth in 25-m bins integrated over the entire section. Similar seasonal average of volume transports for the eastern section as functions of (c) longitude in 50-km bins integrated over the top 250 m and (d) of depth in 25-m bins integrated over the entire section. Black bars in (a) and (c) represent the coast of Sri Lanka.

Perhaps a simpler way to look at the transport is to average all the transports from each season together. The seasonal average of the volume transports in the top 250 m of the water column in 50-km bins (Figs. 11a,c) show the surface circulation described earlier, where transports are generally westward near the coast of Sri Lanka (southward on the eastern line) except during the summer monsoon (JJA). Seasonal averages of the volume transports also generally reverse direction with depth and show a more complex vertical structure (Figs. 11b,d), most notably during transition periods. Implications of the vertical structure of the current for the import and export of water

masses from the Bay of Bengal and Arabian Sea will be discussed in section 4.

Finally, the total volume transport across the southern section, both for the 200 km closest to Sri Lanka and for the entire 400-km section, are listed in Table 1 for the top 250 m and the top 1000 m in each section. Both directions are integrated to obtain the total transport (positive eastward). The number of sections sampled in each season (over the entire 4-yr period) is listed on the first line. Standard deviation is a measure of the variability between these sections. While the average transports are largest during the monsoon seasons, particularly at the surface, the standard deviation is larger than the mean during the transition period. It is particularly noticeable that standard deviations grow significantly when the deeper depths are included, particularly when the offshore part of the section

TABLE 1. Seasonal average and standard deviation of depth-integrated transports (Sv), over a range of depths and lateral extent, for the southern section. Integral is from 3°42'N to 5°30'N for the 200-km extent and from 2° to 5°30'N for the 400-km values.

	DJF	MAM	JJA	SON
Southern section	5 sections	12 sections	12 sections	7 sections
0–250 m, 200 km	-7.4 ± 3.0	-1.4 ± 4.5	7.4 ± 4.2	-2.6 ± 5.9
0–1000 m, 200 km	-4.5 ± 2.9	-0.6 ± 5.9	0.7 ± 5.9	-1.1 ± 4.3
0–250 m, 400 km	-12.1 ± 3.4	1.6 ± 7.0	5.2 ± 6.1	-1.1 ± 5.6
0–1000 m, 400 km	-4.4 ± 8.2	1.2 ± 8.1	-4.2 ± 14.1	5.2 ± 7.9

TABLE 2. Seasonal average and standard deviation of depth-integrated transports over a range of depths and longitudes, for the eastern section. Integral is from 82.0° to 83°49'E for the 200-km extent and from 82.0° to 85°38'E for the 400-km values. Values are not listed if the coverage is not greater than 75%.

	DJF	MAM	JJA	SON
Eastern section	2 sections	5 sections	1 section	9 sections
0–250 m, 200 km	4.0 ± 1.5	0.5 ± 3.6	50.0	$50. \pm 8.8$
0–1000 m, 200 km	5.8 ± 2.6	11.8 ± 7.1	21.8	3.1 ± 10.4
0–250 m, 400 km	—	0.4 ± 3.9	21.0	5.6 ± 10.0
0–1000 m, 400 km	—	11.0 ± 5.3	35.8	6.5 ± 14.7

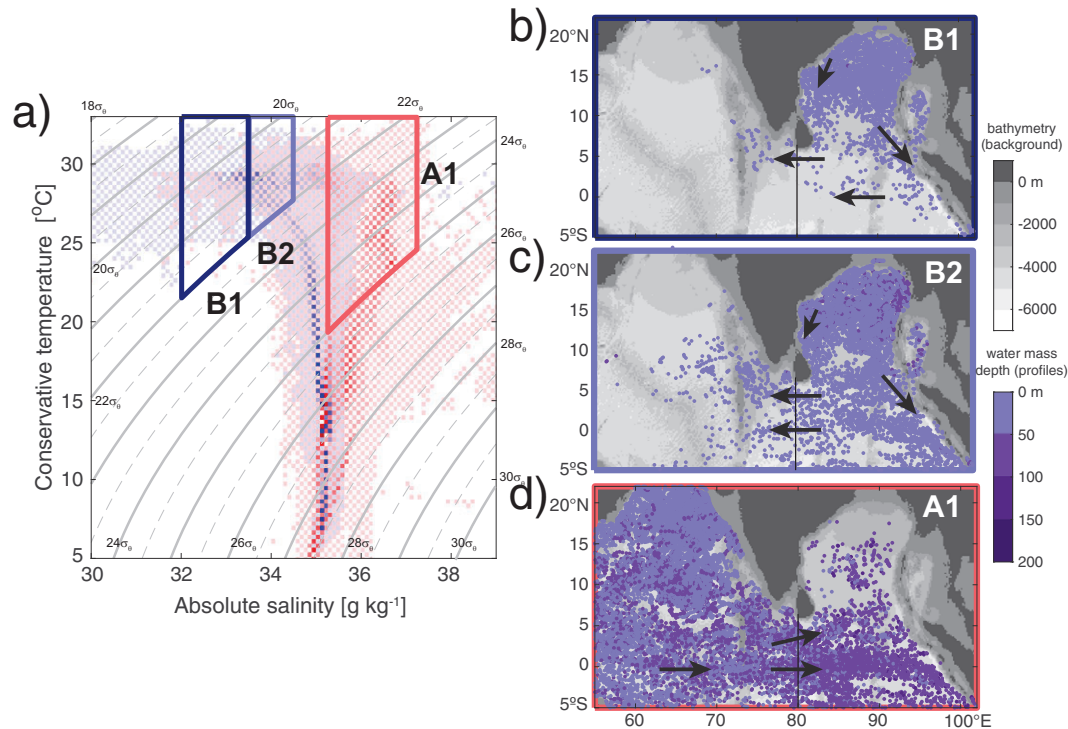


FIG. 12. (a) Temperature–salinity diagram of all Argo float profiles (2016–19) in the Indian Ocean north of 5°S in the Arabian Sea (red) or Bay of Bengal (blue). Each bin is split in two to show the overlap of both density distributions. (b)–(d) Maps of the locations of every float profile, for the entire 4-yr period, with given water mass criteria [boxes in (a)]. For (b)–(d), the color of the dots indicates the depth where the water mass is found (purple color bar) and bathymetry is shaded in gray (gray color bar).

is included. This points to the importance of equatorial dynamics in modulating the circulation.

The equivalent seasonal mean and standard deviation of transports across 8°N are listed in Table 2. Positive values indicate northward flow. The number of sections is significantly smaller, so both mean and standard deviations are less well defined, and some estimates are not possible to make. Generally speaking, the variability is larger than the mean, and individual estimates made from ship-based sections or short-term records are thus unlikely to be representative of the average. While the sampling during summer monsoon is poor, the JJA values are of the same order of magnitude as offshore transport measured during the 2016 Bay of Bengal Boundary Layer Experiment (BoBBLE) field campaign (a northward transport between 16.7 and 24.5 Sv; Webber et al. 2018). At that time the SMC was east of 85°30'E, the limit of the Sri Lankan Exclusive Economic Zone, where our measurements end and those described by Webber et al. (2018) begin. Unfortunately, the 8°N line was not sampled during the period covered by the BoBBLE field campaign. Although there is no good estimate for the northeast (NE) monsoon, transport magnitude and variability in late fall measured by the glider is consistent with the estimates from numerical simulations and the snapshot obtained from a ship in December 2013 at the beginning of the ASIRI program (Wijesekera et al. 2015).

4. Discussion: Transport of specific water masses

Unless sampling defines a closed volume, freshwater and heat transports are ill defined, with results highly dependent on the choice of reference value (e.g., Schauer and Losch 2019). An alternative, as shown in this section, is to characterize exchanges between the basins by quantifying the transport and property changes of individual water masses across the glider sections.

Using 4 years of Argo data (2016–19; overlapping with the sampling of the southern section), the properties of all profiles collected in the Arabian Sea (Fig. 12, in red, shaded by the density of points) overlaps slightly with those from the Bay of Bengal (Fig. 12, in blue). We consider all profiles in the Indian Ocean north of 5°S, and separate the two basins at 80°E (black line in Fig. 12). Note that each T – S pair of the T – S diagram has two juxtaposed colors—a shade of red and a shade of blue, to show the distributions in both basins. Generally speaking, floats stayed in the basin where they were deployed—only 60 floats out of 455 floats with unique identification numbers moved across 80°E. In total, there are 34 429 profiles in the Arabian Sea (from 340 floats), and 17 385 profiles in the Bay of Bengal (from 175 floats).

While there is some overlap, the Bay of Bengal is generally much less salty than the Arabian Sea. The temperature–salinity diagram (Fig. 12a) shows a clear distinction between high-salinity waters formed in the Arabian Sea (red) and the low-

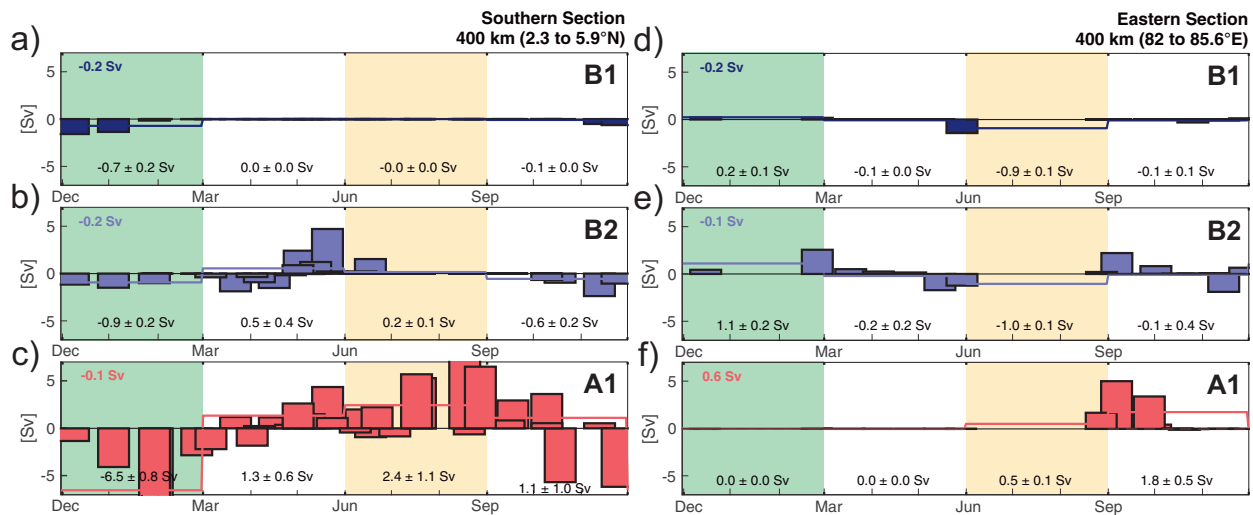


FIG. 13. Transports the B1, B2, and A1 water masses as a function of year day for the (a)–(c) southern section (positive is eastward) and (d)–(f) eastern section (positive is northward).

salinity waters of the Bay of Bengal (blue). Characteristics of three distinct water masses originating from both basins are defined by the density and salinity limits [Fig. 12a; note Hormann et al. (2019) that these are similar to those defined in Hormann et al. (2018), but they use practical salinity units instead of absolute salinity]. Geographically, the fresh Bay of Bengal waters (B1, $32 < S < 33.5 \text{ g kg}^{-1}$; B2, $33.5 < S < 34.5 \text{ g kg}^{-1}$) are mostly present at the northern end of the Bay of Bengal (Fig. 12), with connections on both sides of the Bay. These waters are detected in only a few profiles in the Arabian Sea, with distributions that suggest connections immediately south of Sri Lanka and along the equator (black arrows in Figs. 12b,c). B2 represent a more modified version of the very fresh waters originating from the large rivers of the northern Bay of Bengal, mixed during their transit, and eventually exported to the rest of the Indian Ocean. Similarly, the high-salinity waters (A1, $S > 35.25 \text{ g kg}^{-1}$) are present everywhere in the Arabian Sea, and gradually get deeper toward the east side of the Indian Ocean along the equator and South of Sri Lanka. In the Bay of Bengal, this water mass has been identified as the high-salinity core (HSC), with various direct and indirect links to the Arabian Sea High-Salinity Waters (ASHSW; Jensen et al. 2016; Vinayachandran et al. 1999; Webber et al. 2018; Todd 2020). Locations of Argo profiles that contain these water masses (Fig. 12d) suggest that the high-salinity waters spread in two distinct bands, along the equator and south of Sri Lanka, entering the central Bay of Bengal mostly south of 8°N , below the surface.

These three water masses can be identified directly in each glider section, and its transport is calculated by multiplying each water parcel (with vertical and horizontal extent) meeting these characteristics by their cross-section velocities. Integrating these transports over the whole 400 km sections, the seasonal cycle of the exchange circulation between the basins emerges (Fig. 13). Colors for plotting with each water mass are kept consistent through this section.

A geographical representation of the seasonal transport of these water masses, overlaid with seasonal averages of SSH (Fig. 14), shows that the mean transports obtained by averaging the transports over all sections (solid bars) agree well with the seasonal circulation described above. The fresher water masses (blue) at the surface are dominantly transported near the coast and follow the (wind-driven) geostrophic circulation inferred by altimetry, as does modified Arabian Seawater (red) particularly during the SW monsoon. Estimates from individual sections often show transport in the opposite direction than the average.

The vertical structure of the transport of these three water masses is plotted in Fig. 15. The fresh B1 and B2 water masses are confined to the surface (top 25 m), which has allowed previous studies to obtain accurate estimate of freshwater transports from remote sensing (e.g., Roman-Stork et al. 2020b) or surface drifters (e.g., Hormann et al. 2019). High-salinity waters from the Arabian Sea have a more complex vertical distribution, extending from the surface to over 100 m, often with subsurface maximum.

Based on the analysis of particle release experiments in an assimilation model, Sanchez-Franks et al. (2019) propose that the high-salinity core in the Bay of Bengal is linked to the Equatorial Undercurrent (EUC), originating in the western Indian Ocean. Similar connections have been made using data from surface drifters (Hormann et al. 2019) and gliders (Todd 2020). While no mechanism has been identified to export this water out of the EUC and into the Bay of Bengal, these studies suggest that the main pathways for Arabian Seawater entering the Bay is along the equator, i.e., not much high-salinity water transits north of 2°N along 80°E in either NEMO (Sanchez-Franks et al. 2019, their Fig. 2) or in surface drifters coming from the Arabian Sea (Hormann et al. 2019, their Fig. 3). In contrast, our observations, other models (e.g., Jensen et al. 2016), and earlier work (e.g., Vinayachandran et al. 1999) suggest that the pathway closer to Sri Lanka is



Seasonal and Year-To-Year Variability of Boundary Currents and Eddy Salt Flux along the Eastern and Southern Coasts of Sri Lanka Observed by PIES and Satellite Measurements

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ABSTRACT: Boundary currents along the Sri Lankan eastern and southern coasts serve as a pathway for salt exchange between the Bay of Bengal and the Arabian Sea basins in the northern Indian Ocean, which are characterized by their contrasting salinities. Measurements from two pairs of pressure-sensing inverted echo sounders (PIES) deployed along the Sri Lankan eastern and southern coasts as well as satellite measurements are used to understand the variability of these boundary currents and the associated salt transport. The volume transport in the surface (0–200-m depth) layer exhibits a seasonal cycle associated with the monsoonal wind reversal and interannual variability associated with the Indian Ocean dipole (IOD). In this layer, the boundary currents transport low-salinity water out of the Bay of Bengal during the northeast monsoon and transport high-salinity water into the Bay of Bengal during the fall monsoon transition of some years (e.g., 2015 and 2018). The Bay of Bengal salt input increases during the 2016 negative IOD as the eastward flow of high-salinity water during the fall monsoon transition intensifies, whereas the effect of the 2015/16 El Niño on the Bay of Bengal salt input is still unclear. The time-mean eddy salt flux over the upper 200 m estimated for the April 2015–March 2019 period along the eastern coast accounts for 9% of the salt budget required to balance an estimated 0.13 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) of annual freshwater input into the Bay of Bengal. The time-mean eddy salt flux over the upper 200 m estimated for the December 2015–November 2019 period along the southern coast accounts for 27% of that same salt budget.

SIGNIFICANCE STATEMENT: In the northern Indian Ocean, the highly saline Arabian Sea undergoes extreme evaporation while the Bay of Bengal (BoB) receives excess freshwater input. The focus of this study is the role of the observed time-variable circulation around Sri Lanka that permits the exchange between these basins to maintain their salinity distributions. The circulation fluctuates seasonally following the monsoon wind reversal and interannually in response to large-scale climate modes. The BoB freshwater export around Sri Lanka occurs during the northeast monsoon, whereas saline water import occurs during the fall monsoon transition of some years. However, rapid changes in both water volume transport and salt exchange can occur. The circulation over 0–200-m depth transports ~9%–27% of the BoB salt budget.

KEYWORDS: Thermohaline circulation; Boundary currents; ENSO; Mass fluxes/transport; Transport

1. Introduction

The Bay of Bengal (BoB), a semienclosed ocean basin in the northeastern Indian Ocean (IO), gains $\sim 0.13 \text{ Sv}$ ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) of freshwater annually from precipitation and river runoff minus evaporation as estimated from observations,

reanalysis products, and numerical simulations (Rao and Sivakumar 2003; Sengupta et al. 2006; Wilson and Riser 2016). The large freshwater input into the BoB often creates a barrier layer that occurs between a haline-controlled pycnocline and a deeper thermocline, promoting decoupling between the surface and deeper layer beneath (e.g., Sprintall and Tomczak 1992; Vinayachandran et al. 2002; Thadathil et al. 2007; Girishkumar et al. 2011). The shallow barrier layer results in a highly responsive SST that is found to directly affect rain events during the monsoon (Krishnamohan et al. 2019;

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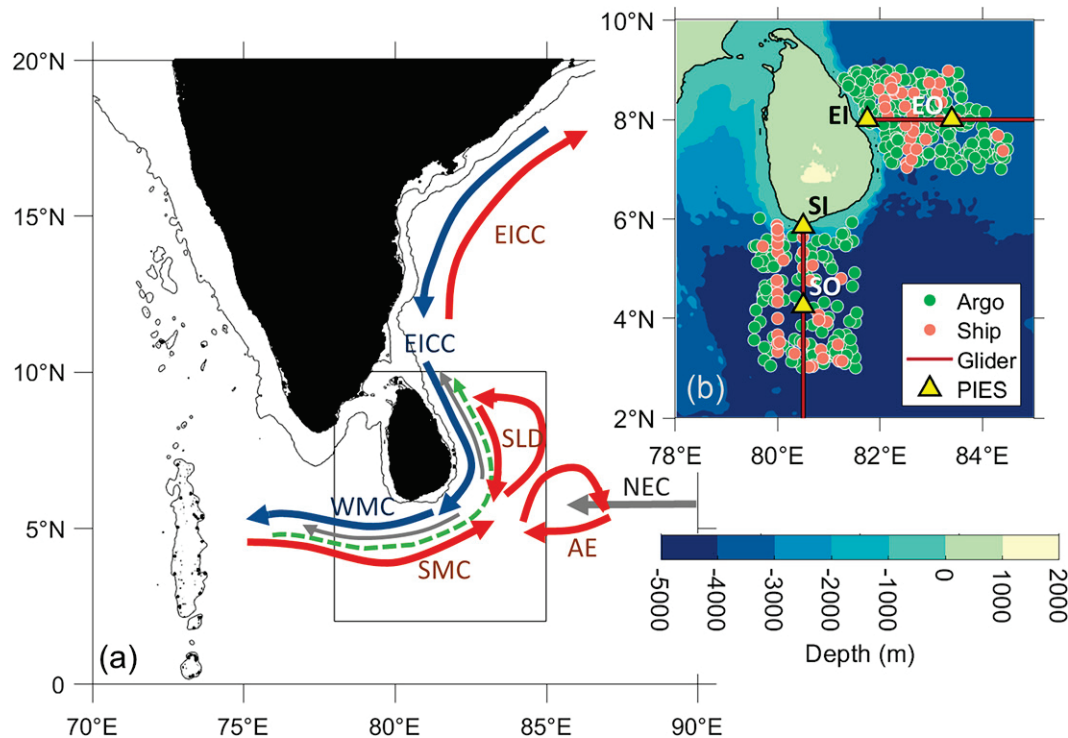


FIG. 1. (a) Map of the observing area and the dominant currents during the southwest monsoon (red), northeast monsoon (blue), spring monsoon transition (gray), and fall monsoon transition during some years (e.g., 2015 and 2018; green dashed). Black contour lines are the 1000-m isobaths. EICC is the East Indian Coastal Current, WMC is the Winter Monsoon Current, SMC is the Summer Monsoon Current, SLD is the Sri Lanka Dome, AE is the seasonal anticyclonic eddy, and NEC is the North Equatorial Current. (b) The inset shows the bathymetry of the Indian Ocean around Sri Lanka with the location of the historical shipboard (orange) and Argo (green) hydrographic profiles, nominal glider tracks (maroon lines), and PIES instruments (yellow triangles) where EI, EO, SI, SO denote different PIES sites: EI eastern inshore; EO eastern offshore; SI southern inshore; SO southern offshore.

Weller et al. 2016) as the warmer SST encourages deep atmospheric convection, particularly during the summer monsoon (Shenoi et al. 2002). In addition, using the wind–evaporation–SST feedback (Xie and Philander 1994) the subsequent rainfall location over the BoB can be predicted when the location of warm SST is identified. Thus, understanding the intensity and fluctuation of the surface salinity layer in the BoB could improve our understanding of the ocean–atmosphere coupling that promotes convection associated with monsoon rainfalls in the northern IO.

One pathway of freshwater export from the BoB is via the seasonal East India Coastal Current (EICC) that occurs during the northeast monsoon (Fig. 1). The EICC extends southward from the eastern coast of India to the eastern coast of Sri Lanka (Shetye et al. 1996). The current then turns westward to merge with the Winter Monsoon Current (WMC; approximately November–February) (Schott et al. 1994; Hacker et al. 1998; Reppin et al. 1999; Rao et al. 2008) that flows along the southern coast of Sri Lanka transporting low-salinity water out of the BoB (Han and McCreary 2001; Wijesekera et al. 2015). The WMC exhibits a similar structure to the EICC in that it intensifies near the coast and in the upper 100–200 m of the water column, although it can extend to 1000-m depth (Schott et al. 1994; Schott and McCreary 2001).

During the southwest monsoon, the eastward Summer Monsoon Current (SMC) transports high-salinity water originating in the Arabian Sea (AS) into the BoB at $\sim 85.5^\circ\text{E}$ (Han and McCreary 2001; Vinayachandran et al. 2013; Wijesekera et al. 2016a; Sanchez-Franks et al. 2019). A seasonal anticyclonic eddy trapping upwelled water off the southern coast of India and Sri Lanka is present to the east of the SMC (Pirro et al. 2020). Along the Sri Lankan eastern coast, the Sri Lanka Dome (SLD), a seasonal cyclonic eddy, dominates the regional circulation (Vinayachandran and Yamagata 1997; Vinayachandran et al. 1999; Shankar et al. 2002; de Vos et al. 2014). Numerical simulations by Jensen (2001) show the intrusion of the AS-originated water into the BoB at the end of the southwest monsoon; Jensen (2003) attributed this to northward Ekman transport arising from the monsoon wind reversal. Analyses using multiple numerical simulations and reanalysis products show that the depth-integrated salt transport across 6°N into the BoB reaches its maximum in July at $\sim 84^\circ\text{E}$, whereas the maximum salt transport out of the BoB across the same transect reaches its maximum in January at $\sim 82^\circ\text{E}$ (D’Addezio et al. 2015).

Although the WMC and SMC are still found along the southern coast of Sri Lanka during the spring and fall monsoon transitions, respectively (Schott et al. 1994; Schott and McCreary 2001),

the altimetry-derived surface geostrophic flow is northward during both monsoon transitions (Eigenheer and Quadfasel 2000; Durand et al. 2009; Lee et al. 2016). In addition, a seasonal opposing undercurrent below the surface layer is present in boreal spring along the eastern coast (Anutaliya et al. 2017). To date, there have been few continuous or simultaneous measurements of flows off the Sri Lankan eastern and southern coasts to demonstrate the regional circulation and its contribution to BoB freshwater export.

In this study, we use continuous measurements from nearly simultaneous 4-yr deployments of seafloor pressure-sensing inverted echo sounders (PIES) from 2014 to 2019 off the Sri Lankan eastern and southern coasts and surface velocity from satellite altimetry to estimate the volume transport within these boundary currents and understand the regional circulation. Together with remotely sensed sea surface salinity (SSS), we use the PIES measurements to estimate eddy salt fluxes along the Sri Lankan eastern and southern coasts. In addition to the seasonal monsoon, we also examine the impact on this circulation from interannual climate modes, such as the Indian Ocean dipole (IOD) and El Niño–Southern Oscillation (ENSO). The datasets and method are described in sections 2 and 3, respectively. Section 4 presents the observed velocity, volume transport, and eddy salt flux across the Sri Lankan eastern and southern sections. Section 5 contains the summary.

2. Datasets

a. *In situ measurements*

The PIES were deployed on the seafloor to measure bottom pressure and acoustic travel time. The bottom pressure fluctuation captures the barotropic flow fluctuation while the acoustic travel time reflects baroclinic flow, mainly contributed by temperature variations in the water column plus the effect of sea surface height changes (e.g., Meinen and Watts 2000; Lee et al. 2016). As suggested by altimeter data and also Seaglider measurements (Rainville et al. 2022), the boundary currents extend from the coasts to approximately 170 km offshore, although the location and width of the flow along the southern coast varies seasonally (Schott et al. 1994). A pair of PIES were deployed off the eastern coast [8.00°N, 81.75°E: site EI (eastern inshore) and 8.00°N, 83.40°E: site EO (eastern offshore)] and off the southern coast of Sri Lanka [5.85°N, 80.50°E: site SI (southern inshore) and 4.25°N, 80.50°E: site SO (southern offshore)], spaced ~180 km on each section to bracket the expected position of the boundary currents along each of the Sri Lankan coasts (Fig. 1). The eastern pair were deployed from November 2014 to March 2019, and the southern pair were deployed from December 2015 to November 2019. Both inshore PIES were at ~600-m depth (24 km off the eastern coast and 8 km off the southern coast), and both offshore PIES were at ~4000-m depth. PIES bottom pressure was processed to eliminate the known exponential-linear instrument drift and tidal signals (Kennelly et al. 2007). The resulting pressure fluctuations resolve variability shorter than the deployment period (~4 years). Acoustic travel time data

off the southern coast at the inshore location (SI) are missing because of instrument issues.

A priori *in situ* measurements of the water column structure are necessary to estimate water transport from the PIES travel time and bottom pressure. Local conductivity–temperature–depth (CTD) profiles from historical shipboard hydrography, autonomous gliders, and Argo floats (Anutaliya et al. 2017) were used to establish a relationship [see section 3a(1)] between the integrals of the baroclinic flow relative to a bottom reference (i.e., the PIES bottom pressure) and a linear combination of PIES acoustic travel time and satellite altimetry. There were 281 profiles (30 profiles from shipboard hydrography and 251 profiles from Argo floats) sampled within the eastern box and 185 profiles (39 profiles from shipboard hydrography and 146 profiles from Argo floats) sampled within the southern box (see inset of Fig. 1). The CTD profiles sampled to depths between 600 and 2000 m. All temperature–salinity profiles were used to simulate the acoustic travel time signal for the inshore PIES (~600-m depth), but only those that sampled to 2000 m (214 profiles off the eastern coast and 136 profiles off the southern coast) were used to simulate the acoustic travel time signal for the offshore PIES (~4000-m depth). The available CTD profiles only sample to 2000-m depth, but the potential temperature and salinity between 2000- and 4000-m depth over the region do not vary much (Shetye et al. 1993; Garternicht and Schott 1997; Beal et al. 2003). Therefore, temperature and salinity were extrapolated from 2000 to 4000 m by assuming that potential temperature and salinity remained constant below 2000 m.

Seagliders were deployed nominally along these same sections as the PIES because they were part of larger observational programs: Air–Sea Interactions Regional Initiative (ASIRI; Wijesekera et al. 2016b) and Monsoon Intra-Seasonal Oscillations in Bay of Bengal (MISO-BoB; Shroyer et al. 2021). The Seagliders repeatedly sampled temperature and salinity from the surface to ~1000-m depth from December 2013 to January 2016 off the Sri Lankan eastern coast and from April 2016 to December 2019 off the southern coast (Lee et al. 2016; Rainville et al. 2022; Fig. 1). Although the glider sections extend farther to the east or south than the offshore PIES locations considered here, there is significant overlap between the PIES and glider sections (by design). Gliders take 1–2 weeks to travel the distance between the PIES. There are 14 crossings, with 6 of the glider crossings overlapping with the PIES observing period, at the eastern section and 36 crossings, with 32 of the glider crossings overlapping with the PIES observing period, at the southern section. In this study, the glider-derived geostrophic velocity and salinity measurements are primarily used to calibrate and validate the time series of volume transport and eddy salt flux as estimated by the PIES and satellite measurements. In addition, the typical vertical structure (modes) calculated from the glider velocity profiles were used to retrieve details of the velocity structure once total transports in the surface and subsurface layers were determined from PIES and satellite altimetry data. In addition, errors associated with the eddy salt flux estimation method based on PIES, satellite altimetry, and SSS at the sections are estimated using simulations of the method with salinity and

velocity measurements from the Seagliders. Water properties and transports estimated from gliders are discussed in detail elsewhere (Rainville et al. 2022).

b. Remotely sensed data

The gridded SLA is available over the 1993–2019 period on a $1/4^\circ$ grid at daily resolution (Ducet et al. 2000). The satellite SLA measures expansion of the water column due to both temperature and salinity variations plus mass changes. The SLA was linearly interpolated onto the 8.00°N eastern and 80.50°E southern sections (Fig. 1) to combine with the PIES acoustic travel time and estimate baroclinic flow across the sections.

Remotely sensed SSS is derived from Soil Moisture Active Passive V4.0 (SMAP; Meissner et al. 2018). The product is an 8-day running average with 40-km spatial resolution interpolated onto a $1/4^\circ$ grid and is available from April 2015 to December 2019. The SSS is linearly interpolated onto the 8.00°N eastern and 80.50°E southern sections (Fig. 1) and is used to estimate the eddy salt flux carried by the boundary currents.

3. Method

a. Volume transport calculation

1) VOLUME TRANSPORT FLUCTUATION FROM PIES AND SATELLITE ALTIMETRY

PIES provide acoustic travel time and bottom pressure measurements. Traditionally, acoustic travel times can be related to the existing historical CTD temperature and salinity profiles of the measured water column to create lookup tables for each variable; the technique is referred to as the gravest empirical mode (GEM) method (Meinen and Watts 2000; Watts et al. 2001). From the lookup tables, the temperature and salinity structure of the water column can be inferred from the measured acoustic travel time. Thus, the baroclinic volume transport can be calculated over the depth layer being investigated. However, the water column structure is highly variable. As a result, the GEM method does not work well in the region.

Instead of parameterizing temperature and salinity profiles, we developed a technique that directly relates the quantity of interest (integral of the geopotential, referenced to the bottom, over the target layer; G') to known quantities, which are the PIES acoustic travel time and the satellite SLA. The quantity G' is combined with the geopotential fluctuation at the reference depth, that is, the bottom, Φ derived from the PIES bottom pressure (i.e., $\Phi = p'/\rho$, where ρ represents mean density at the PIES depth) to obtain the total (absolute) geopotential integral G over the surface or subsurface layer with a thickness of Δz :

$$G = G' + \Phi\Delta z, \quad (1)$$

where Δz is 200 m for the surface layer and 400 m for the subsurface layer. This new technique gives a more robust relationship of the measured quantities with the quantities of

interest since G' is a double integral of specific volume anomaly over the desired pressure range. The relationship between G' and a combination of acoustic travel time and satellite SLA is found empirically through multivariable linear regression and it is determined separately for each PIES site. At each PIES site, the PIES acoustic travel time is simulated from the nearby historical hydrography (see inset Fig. 1) adjusted for the height of the water column using collocated (in both space and time) satellite SLA. The height adjustment is required to accurately calculate the travel time integral from the PIES depth to the actual sea surface. Because historical hydrography shows that the boundary current is concentrated in the upper 200 m of the water column (Anutaliya et al. 2017), the integral of geopotential was estimated over the upper 200 m of the water column (surface layer) separately from that over the 200–600-m-depth (subsurface) layer.

To obtain the accuracy of the G' estimates, a scatterplot between the surface-layer G' that is directly computed from historical hydrography (CTD G') and G' that is determined from the empirical relationship based on the travel time and satellite altimetry (model G') at each PIES location is examined (Fig. 2). For the surface layer (sfc), the correlation between CTD G'_{sfc} and model G'_{sfc} is very high (>0.7) at all four locations and is significant at the 95% confidence level. Since the travel time is missing at the SI site, only SLA is used to linearly regress against G'_{sfc} , and this yields the lowest correlation coefficient of 0.74 between CTD G'_{sfc} and model G'_{sfc} . In the 200–600 m subsurface layer (sub), the correlation between CTD G'_{sub} and model G'_{sub} is significant at the 95% confidence level at the EI, EO, and SO sites, with correlation coefficients ranging from 0.75 to 0.83. The correlation at the SI site is only 0.13 and is not significant at the 95% confidence level because the acoustic travel time at this site is missing and altimetry SLA apparently does not constraint the 200–600-m dynamic height. To examine whether bottom pressure alone can constrain the subsurface layer dynamic height, the barotropic component [$\Phi(t)\Delta z$; Eq. (1)] is estimated by calculating bottom pressure from historical CTD measurements adjusted for the surface height using collocated satellite SLA. The result shows that the barotropic contribution from the PIES bottom pressure dominates the subsurface absolute geopotential fluctuation at this site; calculating the absolute geopotential integral fluctuation (both baroclinic and barotropic contributions) using $G'_{\text{sub}}^{\text{In}}$ estimated without the acoustic travel time results in only $\sim 8\%$ root-mean-square (rms) difference from that computed using $G'_{\text{sub}}^{\text{In}}$ estimated using both satellite SLA and acoustic travel time.

From the geopotential estimated separately at the inshore and offshore locations, the temporal fluctuations of the volume transport Q' at each time step t in the surface layer can be calculated as

$$Q'_{\text{sfc}}(t) = \frac{1}{f} [G'_{\text{sfc}}^{\text{Off}}(t) - G'_{\text{sfc}}^{\text{In}}(t)], \quad (2)$$

where f is the Coriolis parameter at the midsection location (8.00°N for the eastern section and 5.05°N for the southern section). The superscript denotes the location of the inshore

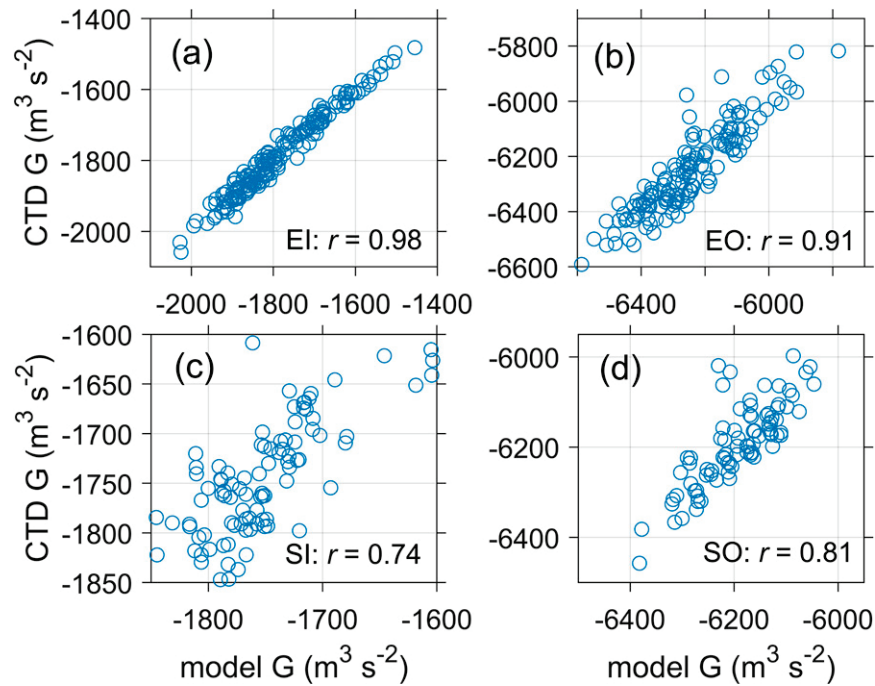


FIG. 2. Scatterplots of the vertical integral of the geopotential G_{sfc} over the 0–200-m layer referenced to the PIES depths calculated from historical CTD profiles (y axis) vs that obtained through a linear regression model against the CTD-simulated acoustic travel time and satellite SLA (x axis) at the (a) EI, (b) EO, and (d) SO sites. Only satellite SLA was used for the model G_{sfc} at the (c) SI site. The corresponding correlation coefficients are shown in the lower-right corners of each subplot. The G_{sfc} is calculated with reference levels at the PIES depths, i.e., 600-m depth for the inshore location and 4000-m depth for the offshore location.

(In) and offshore (Off) PIES. Volume transport fluctuations in the 200–600-m subsurface layer are calculated correspondingly, but by using G_{sub} instead of G_{sfc} .

2) ABSOLUTE VOLUME TRANSPORT AND ACROSS-SECTION MEAN VELOCITY PROFILE CONSTRUCTION

The geostrophic transports obtained with the above approach have an unknown time-mean offset that corresponds to the unknown absolute levelling of the two PIES bracketing each section. To estimate the mean offset, we use volume transport calculated from the Seaglider measurements over the section spanned by the PIES and over the surface (0–200-m depth) and subsurface (200–600-m depth) layers. There are 6 and 32 glider crossings that overlap with the PIES/altimetry volume transport off the eastern and southern coasts, respectively. For each Seaglider transect crossing, the PIES/altimetry-derived volume transport fluctuation Q' is averaged over the time period of the glider crossing, and Q' is computed separately in the surface and subsurface layers. The time-mean adjustment to the PIES transports is the average of the differences between Seaglider absolute volume transport and the PIES/altimetry-derived transport fluctuation over all of the Seaglider crossings. This time-mean offset of -0.3 Sv was added to the PIES/altimetry surface-layer volume transport fluctuation, while that of -2.8 Sv was added to the subsurface-layer fluctuation across the eastern section. In addition, this

time-mean offset of -2.2 Sv was added to the PIES/altimetry surface-layer volume transport fluctuation, while that of 0.5 Sv was added to the subsurface-layer fluctuation across the southern section. The adjusted volume transport will be referred to as PIES/altimetry absolute volume transport. Although there is some uncertainty in these offsets because of the small number of overlapping glider crossings, the main focus of this study is on the fluctuating component of the flow, and thus our analysis is not sensitive to the offset choice.

For the salt transports (see section 3b), we need to construct the vertical structure of the across-section mean velocity in the upper water column from our continuous PIES/altimetry volume transport estimates in the surface and subsurface layers. Vertical empirical orthogonal functions (EOFs) from the Seaglider velocity profiles sampled along the PIES section were calculated and used to serve as a priori information about the vertical structure. At each time step, the PIES/altimetry-derived volume transports in the surface and subsurface layer were projected onto the two most dominant EOFs calculated from Seaglider measurements to obtain the velocity profile (Fig. 3; Anutaliya et al. 2019).

3) UNCERTAINTY ASSOCIATED WITH THE CALCULATED VOLUME TRANSPORT

Uncertainties associated with the PIES/altimetry-based volume transport are mainly contributed by two sources:

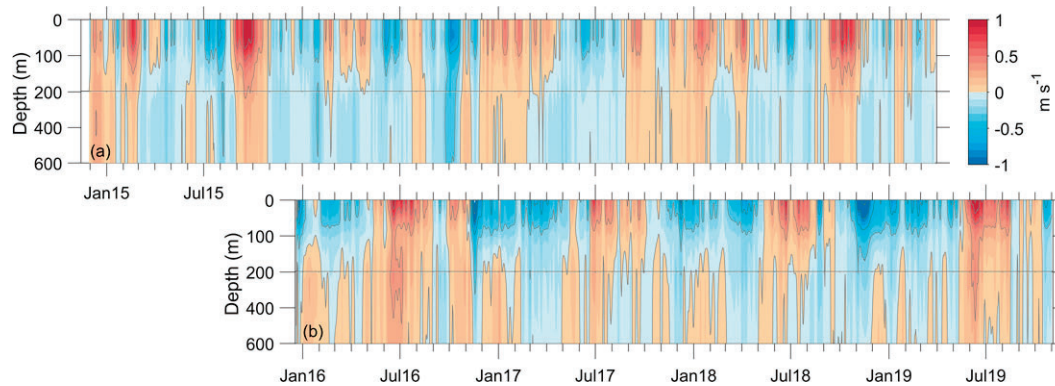


FIG. 3. Across-section mean geostrophic velocity (m s^{-1}) estimated from the PIES/altimetry surface and subsurface volume transport for the (a) eastern and (b) southern sections. Color contour intervals are 0.05 m s^{-1} , and gray contours are plotted every 0.3 m s^{-1} . Note the expanded y axis over the 0–200-m-depth layer.

instrument accuracy and the calculation technique [Eqs. (1) and (2)], together with the regression relationships derived from Fig. 2]. At the eastern and southern sections, uncertainty in the surface-layer volume transport of 0.6 and 0.8 Sv, respectively, is associated with the instrument accuracy as the instrument error is $\sim 0.01\%$ of the pressure range. The error associated with the acoustic travel time measurement is included in uncertainty of the calculation technique (Meinen and Watts 1998), which is shown in the scatterplot (Fig. 2); the uncertainty is 3.9 and 6.4 Sv at the eastern and southern sections, respectively. The total uncertainties, calculated as the sum of variances produced by both components, are 4.0 and 6.4 Sv for the surface volume transport across the eastern and southern sections, respectively. The PIES/altimetry surface-layer absolute volume transport at the southern section agrees with the glider transport within the uncertainty (rms difference is 4.5 Sv). At the eastern section, however, the rms difference between the PIES/altimetry and glider surface-layer volume transport is 7.9 Sv, larger than the estimated uncertainty, which is likely a consequence of fewer overlapping measurements by the PIES and the glider [section 3a(2) above]. For the subsurface layer, the uncertainty for the flow along the eastern coast is 7.1 Sv, of which 7.0 Sv is from the calculation technique and 1.2 Sv is from the instrument accuracy, and the uncertainty for the flow along the southern coast is 10.0 Sv, of which 9.9 Sv is from the calculation technique and 1.6 Sv is from the instrument accuracy. The uncertainty due to the calculation technique is very high for the subsurface flow along the southern coast because the acoustic travel time measurement at the SI site is missing; hence, the subsurface G' cannot be accurately estimated, as noted above.

b. Eddy salt transport estimation

The eddy salt flux component is then determined as the product of the temporal fluctuations of salinity [$\tilde{S}(s, z, t)$] and velocity [$\tilde{v}(s, z, t)$] at each section, where a tilde denotes the temporal fluctuation relative to the mean vertical profile $\bar{x}(z)$ [the overbar denotes averages over the section (section mean) and over the observing period (time mean) of a quantity x]; that is, $\tilde{x}(s, z, t) = x(s, z, t) - \bar{x}(z)$. The time series will be

hereinafter referred to as the eddy salt transport $F_s(t)$, and the time-mean of the eddy salt transport is the eddy salt flux $\bar{F}_s$. The eddy salt transport is calculated as follows:

$$F_s(t) = \int_{z=0}^{200} \int_{s^{\text{In}}}^{s^{\text{Off}}} \bar{\rho}(z) \tilde{S}(s, z, t) \tilde{v}(s, z, t) ds dz, \quad (3)$$

where s^{In} and s^{Off} represent the location of PIES at the inshore and offshore sites, respectively, and $\bar{\rho}(z)$ is the time-mean density that is determined from the historical hydrography averaged over each section. The $\tilde{v}(z)$ is derived from the reconstruction based on the PIES/altimetry volume transport and Seaglider EOFs [section 3a(2); Fig. 3], and $\tilde{S}(z)$ is from the local hydrography. At the surface, the SMAP mean SSS is 33.4 and the mean SSS from local hydrography is 33.8 at the eastern section whereas the SMAP mean SSS is 34.1 and the mean SSS from local hydrography is 34.5 at the southern section. The different time-mean SSS from the hydrography and from the satellite SSS is likely a result of the combination of the different sampling periods, the extrapolation of the hydrographic salinity profile to the surface, and potential biases in the satellite product (Tang et al. 2017; Qin et al. 2020). Therefore, the hydrography-derived $\tilde{S}(z)$ used for the eddy salt flux estimation is scaled and shifted (linearly transformed); constants for shifting and scaling are determined so that $\tilde{S}(z)$ passes through the satellite time-mean SSS while salinity at 600-m depth remains the same. From Eq. (3), $\tilde{S}(s, z, t)$ and $\tilde{v}(s, z, t)$ are needed but cannot be directly obtained from either the PIES measurements or satellite SMAP SSS. Thus, these quantities have to be inferred from available information as described in the following sections.

1) SALINITY PROFILE CONSTRUCTION

To calculate eddy salt transport, a salinity profile at locations along the section at each time step [$S(s, z, t)$] is required; however, satellite measurements can only provide the time series of salinity at the surface along the section [$S(s, 0, t)$]. Therefore, we used the local historical CTD measurements to estimate the salinity structure below the surface at each time along each section based on either altimetry-derived surface

velocity or SMAP SSS. The details of the salinity profile construction can be found in the [appendix](#).

Because there is no continuous measurement of 200–600 m salinity available for the salinity profile construction, we are not confident of the estimated salinity over the deeper subsurface layer. Therefore, eddy salt transport will only be calculated for the upper 200 m of the water column. Also, both observations and numerical simulations show that the eddy salt flux below 200 m is likely small ([Anutaliya et al. 2017](#); [Rainville et al. 2022](#)).

2) VELOCITY PROFILE CONSTRUCTION

While the time series of velocity profiles (PIES/altimetry volume transport with Seaglider EOFs) were constructed as along-section means [[section 3a\(2\)](#); [Fig. 3](#)], the velocity profile at locations *along* each section [$v(s, z, t)$] is needed for the eddy salt transport estimation [[Eq. \(3\)](#)]. This quantity is estimated by projecting the altimetry-derived surface velocity at each section onto the two leading modes of the Seaglider EOFs, constrained by the total across-section transports in the surface (0–200 m) and subsurface (200–600 m) layers from the PIES. Among the possible solutions, the one with the smallest rms velocity in the subsurface layer is chosen, which we found to give results resembling the actual flow distributions.

3) SECTION-MEAN AND SECTION-VARYING COMPONENTS OF EDDY SALT TRANSPORT

The $F_s(t)$ can be decomposed into two components: the first component results from the section-mean salinity fluctuation, $S^*(z, t)$ defined as $\langle \tilde{S}(s, z, t) \rangle$, and velocity fluctuation, $v^*(z, t)$ defined as $\langle \tilde{v}(s, z, t) \rangle$, where the angle brackets indicate the section-mean, and the second component is the departure from the section-mean (spatially fluctuating) parts [e.g., $S'(s, z, t)$]. By decomposing $\tilde{S}(s, z, t)$ and $\tilde{v}(s, z, t)$ into the section-mean and spatially fluctuating part, [Eq. \(3\)](#) can be rewritten as follows:

$$F_s(t) = \int_{z=0}^{200} \int_{s^{\text{In}}}^{s^{\text{Off}}} \bar{\rho}(z) S^*(z, t) v^*(z, t) + \bar{\rho}(z) S'(s, z, t) v'(s, z, t) ds dz. \quad (4)$$

We can compute the first component of [Eq. \(4\)](#) from the time-varying section-mean velocity (i.e., [Fig. 3](#)) and the section-mean salinity profile constructed from the section-mean SMAP SSS and the local hydrography (see the [appendix](#)). In situ measurements of salinity and velocity profiles from the Seaglider can be used to evaluate the percentage of variance in eddy salt transport captured by the calculation using only the section means, that is, the first term in [Eq. \(4\)](#). In addition, Seaglider measurements allow the estimation of uncertainties associated with the eddy salt transport calculation using solely the first (section mean) component as compared with the total eddy salt transport (section-mean and spatially fluctuating components). Thus, we can determine the appropriate technique for the eddy salt transport calculation at each section. Simulations with the Seaglider measurements show that the section-mean eddy salt transport alone [first component; calculated on the basis of $S^*(z, t)$ and $v^*(z, t)$] accounts for 62% of the total eddy salt transport variability at the eastern section. At the

TABLE 1. Uncertainties associated with time series mean and annual mean (in parentheses) salt flux estimation ($\times 10^6 \text{ kg s}^{-1}$) using section-mean and section-depth velocity v and salinity S at the eastern and southern sections. The estimated mean salt flux values ($\times 10^6 \text{ kg s}^{-1}$) at the eastern section are based on the section-depth v and S , whereas the mean salt flux values ($\times 10^6 \text{ kg s}^{-1}$) at the southern section are based on the section-mean v and S [$\langle v(s, z, t) \rangle$ and $\langle S(s, z, t) \rangle$].

	Uncertainty ($\times 10^6 \text{ kg s}^{-1}$)		Estimated time series mean salt flux ($\times 10^6 \text{ kg s}^{-1}$)
	$\langle v(s, z, t) \rangle$ and $\langle S(s, z, t) \rangle$	Section-depth v and S	
East	0.20 (0.39)	0.08 (0.17)	0.42
South	0.16 (0.31)	0.21 (0.42)	1.29

southern section, the section-mean eddy salt transport accounts for 83% of the total eddy salt transport variance. The uncertainties in eddy salt flux are 0.20×10^6 and $0.16 \times 10^6 \text{ kg s}^{-1}$ at the eastern and southern section, respectively ([Table 1](#)). Thus, the eddy salt flux component derived from $S^*(z, t)$ and $v^*(z, t)$ alone approximately represents the total eddy salt flux for the sections, particularly off the southern coast.

For a more accurate estimate of the total eddy salt transport [[Eq. \(4\)](#)], both section-mean [$S^*(z, t)$ and $v^*(z, t)$] and spatially varying salinity and velocity profiles [$S'(s, z, t)$ and $v'(s, z, t)$] are required. Each PIES section is divided into four equally spaced ($\sim 45 \text{ km}$) grid cells beginning at the coast (81.75°E for the eastern and 5.75°N for the southern section). To obtain the along-section/depth distribution of salinity, SMAP SSS is linearly interpolated to the center of each grid cell where subsurface salinity structure is separately determined based on the historical hydrography (see the [appendix](#)). The along-section/depth distribution of velocity is estimated according to [section 3b\(2\)](#), where surface velocity calculated from satellite altimetry is also linearly interpolated to the center of each grid cell. Again, the Seaglider measurements were used to simulate the additional variance of the eddy salt transport captured by including both section-mean and spatially fluctuating components in the calculation. Then, we compare the resulting eddy salt transport from the simulated sections with that calculated from the full Seaglider salinity and velocity profiles. At the eastern section, the eddy salt transport calculated from the section–depth salinity/velocity [both $S^*(z, t)$, $v^*(z, t)$ and $S'(s, z, t)$, $v'(s, z, t)$] explains a higher percentage of variance (88%) of the eddy salt transport. The uncertainty in eddy salt flux also decreases to $0.08 \times 10^6 \text{ kg s}^{-1}$ ([Table 1](#)). However, on the southern section the Seaglider measurements show that both $S^*(z, t)$, $v^*(z, t)$ and $S'(s, z, t)$, $v'(s, z, t)$ together only explain 53% of the total eddy salt transport variance, decreasing the skill in estimating the eddy salt transport in comparison with the section-mean salinity/velocity. Also, the uncertainty in eddy salt flux estimation increases to $0.21 \times 10^6 \text{ kg s}^{-1}$ ([Table 1](#)). The higher uncertainty at the southern section reflects the inability of the constructed along-section/depth distributions of salinity/velocity to adequately replicate the vertical structure of the salinity and velocity along the section. Indeed, the glider

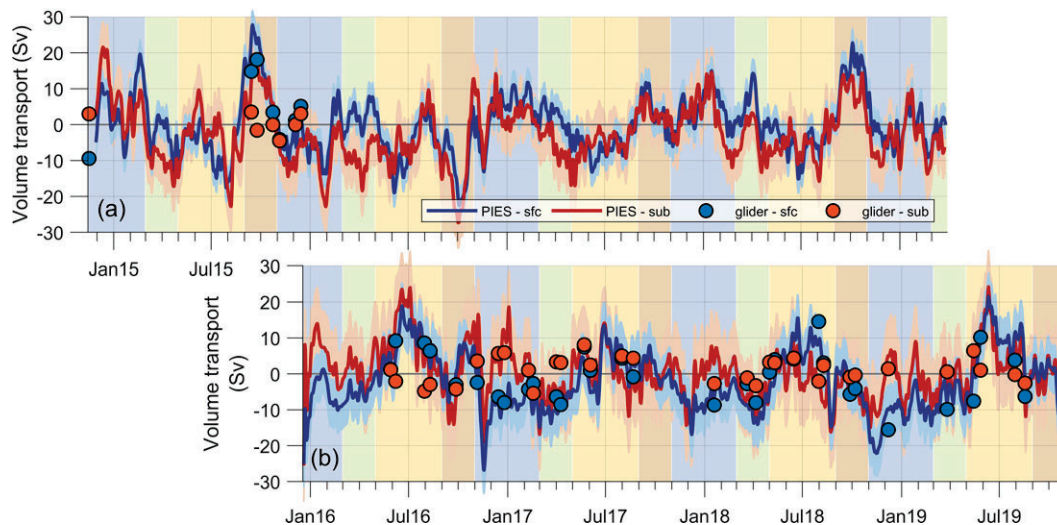


FIG. 4. Volume transport (Sv) across the section off the Sri Lankan (a) eastern and (b) southern coasts over the 0–200-m-depth (surface; blue) and 200–600-m-depth (subsurface; red) layers. Results from the PIES measurements and satellite SLA are shown as lines, with the associated uncertainties shaded. Results from the Seaglider measurements are shown as dots. Background shading denotes the seasons of northeast monsoon (blue), spring monsoon transition (green), southwest monsoon (yellow), and fall monsoon transition (brown).

measurements along the southern section show that the salinity and velocity in the upper 200 m of the water column are complex; a subsurface salinity maximum is sometimes present between 40- and 120-m depth, while the velocity section often shows subsurface reversing flow (Rainville et al. 2022). Therefore, $F_s(t)$ across the eastern section is determined from both components in Eq. (4) [$S^*(z, t)$, $v^*(z, t)$ and $S'(s, z, t)$, $v'(s, z, t)$], whereas that across the southern section is determined from the section [only the first component in Eq. (4): $S^*(z, t)$ and $v^*(z, t)$].

4. Results and discussion

a. Overall description of the circulation along Sri Lankan eastern and southern coasts

The volume transports along the eastern and southern coasts of Sri Lanka have similar mean values and ranges in both the surface (0–200-m depth) and subsurface (200–600-m depth) layers. In the surface layer, volume transport along the eastern coast ranges from -22.6 to 27.8 Sv with a mean of -0.1 Sv over the whole observing period of November 2014–March 2019 and volume transport along the southern coast ranges from -26.8 to 24.2 Sv with a mean of -0.2 Sv over the whole observing period of December 2015–November 2019 (Fig. 4). A positive sign designates northward flow along the eastern coast and eastward flow along the southern coast. The mean surface-layer volume transport along the eastern coast over the full 4-yr period from the beginning of 2015 to the end of 2018 is -0.2 Sv. Along the southern coast, the PIES observations do not cover the whole 4-yr period; data from most of November and the first half of December, when the flow is seasonally westward, are missing. Thus, we expect the full 4-yr mean to be lower than -0.2 Sv (i.e., stronger westward). The estimated mean volume

transport along the southern coast is roughly consistent with previous mooring transport observations where the volume transport is not significantly different from 0 over a 3.75° – 5.87° N section in the 0–300-m-depth layer (Schott et al. 1994). In the subsurface layer, the mean volume transport along the eastern coast is -2.7 Sv, ranging from -27.3 to 21.5 Sv, and the mean volume transport along the southern coast is 0.5 Sv, ranging from -25.6 to 24.2 Sv. At the eastern section, the 2015–18 mean subsurface volume transport is -3.0 Sv (Fig. 4). Along both coasts, the volume transports in the surface and subsurface layers are correlated with correlation coefficient of 0.74 and 0.68 at the eastern and southern sections, respectively; the correlations are significant at the 95% confidence level.

Along the Sri Lankan eastern coast, the geostrophic velocity time series confirm the surface circulation variability found by previous observational studies (e.g., Hacker et al. 1998; Shankar et al. 2002; de Vos et al. 2014; Wijesekera et al. 2015; Lee et al. 2016); the surface current reverses its direction twice a year flowing southward during the monsoon seasons and northward during transition periods (Figs. 3a and 4a). The PIES-altimetry volume transport also shows a subsurface current that occasionally flows in the opposite direction to the surface current (Fig. 3a). The timing of these appearances generally corresponds to previously identified periods during the spring monsoon transition and southwest monsoon when a subsurface countercurrent is present (Anutaliya et al. 2017). Along the Sri Lankan southern coast, the velocity derived from the PIES-altimetry observations exhibits a strong seasonal cycle in the surface layer consistent with observations of the surface current based from satellite altimetry and moorings (Fig. 5; Eigenheer and Quadfasel 2000; Durand et al. 2009; Schott et al. 1994). The dominant surface current along

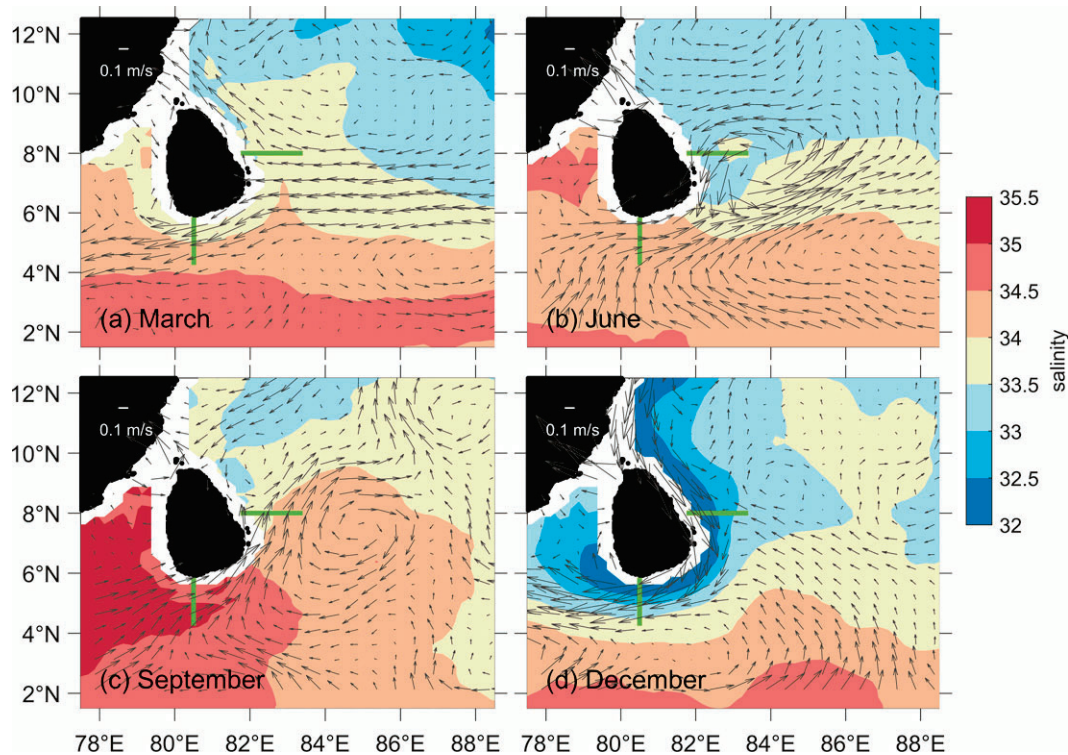


FIG. 5. Monthly mean SMAP sea surface salinity for April 2015–December 2019 (color contours) and geostrophic sea surface velocity derived from satellite altimetry (gray arrows) for the months representing (a) the spring monsoon transition corresponding to green shading in Figs. 4 and 6 (March), (b) the southwest monsoon corresponding to yellow shading in Figs. 4 and 6 (June), (c) the fall monsoon transition corresponding to brown shading in Figs. 4 and 6 (September), and (d) the northeast monsoon corresponding to blue shading in Figs. 4 and 6 (December).

the southern coast is eastward in June–October and westward in November–May; however, the subsurface flow does not show a distinct seasonal cycle. The PIES and altimetry provide the first set of nearly simultaneously observations along the Sri Lankan eastern and southern coasts that continuously capture the regional circulation as well as the eddy salt transport, which will be described in the next sections (sections 4b–d).

b. Overall eddy salt transport by the boundary currents

The estimated salt transport [$F_s(t)$] across the eastern section in the upper 200 m of the water column has a mean of $0.42 \times 10^6 \text{ kg s}^{-1}$ over the April 2015–December 2018 period (when the SMAP SSS is available). At the Sri Lankan southern coast, the salt transport is mainly positive over the observing period (Fig. 6b) yielding the mean eddy salt flux of $1.29 \times 10^6 \text{ kg s}^{-1}$ over the period of December 2015–November 2019 (Table 1). A positive time-mean eddy salt flux indicates that either fresher water is being transported southward or more saline water is being transported northward along the Sri Lankan eastern coast and either fresher water is being transported westward or more saline water is being transported eastward along the Sri Lankan southern coast; both conditions result in an increase of the salinity in the BoB. Note that the estimated eddy salt transport is missing from 19 June 2019 to 23 July 2019

because of an interruption of the SMAP instrumentation sampling. The estimated salt transport [$F_s(t)$] based on PIES, satellite altimetry, and SMAP SSS exhibits high seasonal and year-to-year variability and agrees with that estimated from Seaglider measurements at both sections when they overlap (Fig. 6). Any inconsistency is likely a consequence of the subsurface salinity profile that shows large variability and so cannot always be successfully replicated through our simple construction (see the appendix).

Previous estimates, from climatology, reanalysis products, and numerical simulations, give an overall long-term freshwater input into the BoB of $\sim 0.13 \text{ Sv}$ annually (Rao and Sivakumar 2003; Sengupta et al. 2006; Wilson and Riser 2016), which translates into $\sim 4.5 \times 10^6 \text{ kg s}^{-1}$ of salt input to maintain the salinity balance in the BoB. Our measurements show that the mean eddy salt flux along the eastern coast of Sri Lanka contributes approximately 9% while that along the southern coast contributes about 27% to the estimated BoB annual salt budget (Table 1). The year-to-year changes in the boundary current eddy salt flux can be 12%–28% of the BoB total salt budget (Table 2). Also, the eddy salt transport is highly variable on the seasonal time scale (Fig. 6).

The seasonal cycle in $F_s(t)$, constructed as the sum of the annual and semiannual harmonics, explains 21% and 23% of the total variance at that the eastern and southern sections,

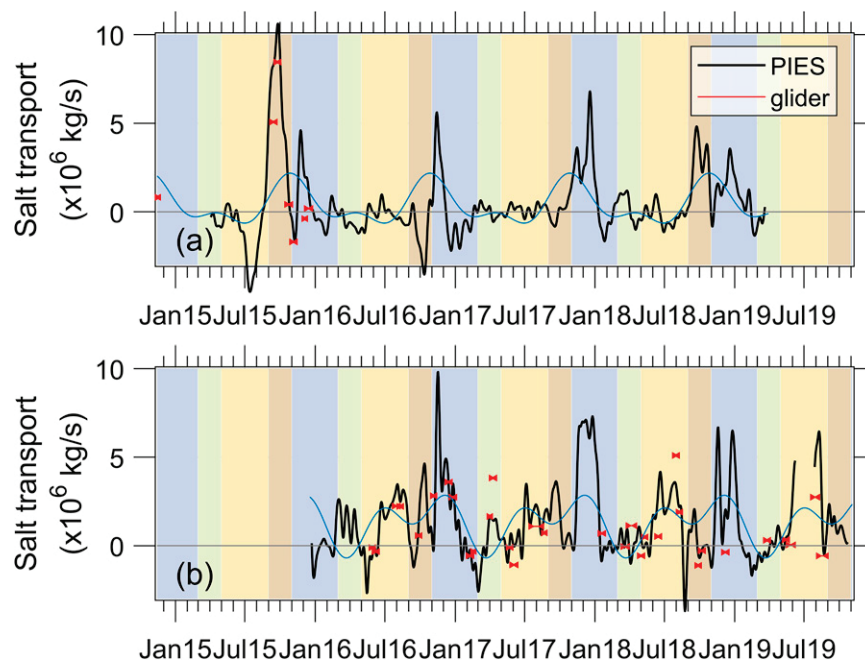


FIG. 6. Estimated eddy salt flux across the Sri Lankan (a) eastern and (b) southern sections. Values in (a) are computed from section-depth velocity and salinity, whereas values in (b) are computed from section-mean velocity and salinity. Glider results are shown in red, and results from PIES, satellite altimetry, and SMAP SSS are shown in black. The lengths of the red symbols represent the time it took the glider to complete the crossing bracketed by the PIES. The seasonal cycles, constructed as the sums of the annual and semiannual harmonics, are plotted in blue. Background shading denotes monsoon seasons as in Fig. 4.

respectively (Fig. 6). The seasonal cycle of eddy salt flux off the Sri Lankan eastern coast exhibits a maximum in October and minimum negative salt transport at the end of June. Off the southern coast, the seasonal cycle is roughly equally influenced by the annual and semiannual harmonics and consists of a maximum seasonal eddy salt flux at the beginning of December with a secondary lower maximum at the beginning of July and a minimum in March. The low-frequency (>120 days) component of $F_s(t)$ explains 15% of the total eddy salt transport variability at the eastern section, whereas it explains 4% of the total variability at the southern section.

TABLE 2. Annual-mean eddy salt flux ($\times 10^6 \text{ kg s}^{-1}$) computed from the section-depth velocity and salinity at the eastern section and from the section-mean velocity [$\langle v(s, z, t) \rangle$] and salinity [$\langle S(s, z, t) \rangle$] at the southern section. Note that the average value along the eastern coast in 2015 only includes the eddy salt transport from April to December and that the average value along the southern coast in 2019 only includes the transport from January to November.

	East	South
2015	1.05	—
2016	−0.23	1.51
2017	0.47	1.48
2018	0.62	1.17
2019	—	0.98

c. Seasonal variability of the regional circulation

1) SOUTHWEST MONSOON (MAY–AUGUST)

Off the Sri Lankan southern coast, the eastward SMC dominates during the southwest monsoon (yellow shading in Fig. 4) with a maximum surface transport of 13.4–21.5 Sv in June (Fig. 4b). The eastward flow is surface intensified but can extend to the deepest analyzed depth of 600 m (Fig. 3b). The PIES-altimetry and SMAP SSS show that the SMC transports saline water originating in the AS along the Sri Lankan southern coast during the southwest monsoon in agreement with findings from prior studies based on hydrographic surveys, climatological salinity, and model simulations (Vinayachandran et al. 2013, 1999; Han and McCreary 2001; Sanchez-Franks et al. 2019). However, this eastward transport of the AS water is not necessarily contiguous around the Sri Lankan eastern coast due to the presence of the SLD that arises from westward-propagating Rossby waves and Ekman pumping over the BoB (Vinayachandran et al. 1999; Wijesekera et al. 2016a; Fig. 5b).

Our PIES measurements show that the southward-flowing flank of the SLD is very strong in this season and usually extends to the deepest analyzed depth of 600 m; it transports 14–20 Sv in the upper 200 m of the water column (Fig. 4a). The occasional opposing subsurface flow is only briefly apparent at the eastern section in June of each year (Fig. 3a). As Anutaliya et al. (2017) suggested, the core of the undercurrent in June is very deep with the maximum speed typically

occurring at 900 m, and so is below the depth resolved by our observations. Satellite altimetry shows that the saline water during the southwest monsoon is often wrapped around the eastern flank of the SLD (around $\sim 85.5^\circ\text{E}$) and thus flows northward farther from Sri Lanka (Fig. 5b). The imported saline water is sometimes recirculated southward along the eastern coast by the SLD resulting in convergence at the southeastern coast of Sri Lanka (Fig. 5b). The recirculation can also produce negative eddy salt transport along the Sri Lankan eastern coast, for example like that observed in 2015 (Fig. 6a).

2) FALL MONSOON TRANSITION (SEPTEMBER–OCTOBER)

Along the Sri Lankan southern coast, the SMC often continues into the fall monsoon transition (brown shading in Fig. 4). With the exception of 2016 [when a strong negative IOD (nIOD) occurs], surface northward flow is present along the Sri Lankan eastern coast in this season, suggesting a continuous pathway of volume transport along the Sri Lankan eastern and southern coasts during these years (Figs. 3 and 4). Also, satellite altimetry patterns confirm the connection of these boundary currents at the surface during the fall monsoon transition of 2017 and 2018 similar to the monthly mean (Fig. 5c). The reconstructed velocity profile shows that the northward flow off the Sri Lankan eastern coast is not only present at the surface but can also extend over the 0–600-m-depth layer (Fig. 3a).

In 2017, the altimetry-derived circulation maps show the northward flow of the SMC at the eastern section more offshore, centered around $\sim 83.5^\circ\text{E}$ (farther offshore than presented in the monthly map; Fig. 5c). This may be the reason that the northward current through the eastern PIES section in 2017 transports much fresher water with a salinity of 33.1–33.2, as compared with that along the southern coast where the salinity is 34.6–34.8. During the fall monsoon transition of 2018, satellite measurements show a strong eastward surface current associated with the saline water from the low-latitude region (to the south of 5°N) of the central or western AS (to the west of 70°E) occurring off the southern coast. The eastward current turns sharply around the southeastern coast of Sri Lanka centered around $\sim 82^\circ\text{E}$ (closer to the coast than shown in the monthly mean map; Fig. 5c) to transport saline water into the BoB producing a positive eddy salt flux along both the southern and eastern Sri Lankan coasts (Fig. 6). Similarly, during the 2015 fall monsoon transition, the PIES at the eastern section, satellite altimetry, and satellite SSS also suggest the import of saline water into the BoB along the southern and eastern coasts of Sri Lanka. This BoB salt import around Sri Lanka originating in the western equatorial IO during the fall monsoon transition is consistent with previous numerical studies (Jensen 2001, 2003; Sanchez-Franks et al. 2019). Thus, the PIES observations highlight the role of saline water imported into the BoB via the boundary currents during the fall transition of some years.

3) NORTHEAST MONSOON (NOVEMBER–DECEMBER)

During the northeast monsoon (blue shading in Fig. 4), the southward EICC off the Sri Lankan eastern coast observed in previous studies (Shetye et al. 1996; Wijesekera et al. 2015) is

apparent for ~ 1 –2 months between October and January (roughly the northeast monsoon season) in every year of the PIES record (Fig. 4a). The EICC is strongest near the surface and extends to the deepest observed depth of 600 m (Fig. 3a), which is comparable to that derived from a single hydrographic survey at 11°N (Shetye et al. 1996). Off the southern coast, the WMC occupies the region; thus, the flow is predominantly westward particularly in the surface layer (0–200-m depth; Figs. 3b and 4b). Southward flow along the Sri Lankan eastern coast occurs at the same time as westward flow along the southern coast every year, strongly suggesting a connection between the EICC and the WMC (Figs. 1, 3, 4, 5d). The satellite surface geostrophic velocity also confirms the westward pathway of the surface EICC around the Sri Lankan southern coast during these periods (Fig. 5d). The PIES-satellite observations provide the first observation-based estimate of simultaneous eddy salt flux showing the export of freshwater by the EICC and the WMC along the Sri Lankan eastern and southern coasts over multiple years. Although the continuous flow along the two coasts persists over only a relatively short period (1–2 months), 4×10^6 – $6 \times 10^6 \text{ kg s}^{-1}$ of eddy salt transport is present (Figs. 5d and 6).

The relatively short duration of the continuous flow along the Sri Lankan eastern and southern coasts is likely a result of a westward-propagating Rossby wave radiated from the eastern boundary of the BoB during the northwest monsoon that often affects the circulation off the Sri Lankan southeastern coast. The westward propagation of the Rossby wave from the eastern boundary of the BoB is also apparent in satellite altimetry (Sreenivas et al. 2012). The Rossby waves arriving offshore of Sri Lanka cause bifurcation of the southward EICC with one branch flowing southward or southeastward leaving a weaker westward flow along the southern coast.

The surface velocity derived from satellite altimetry indicates that the EICC and the WMC usually do not fully extend over the width of the PIES section (Fig. 5d; Wijesekera et al. 2015; Rainville et al. 2022). Thus, with only two PIES deployed at the section end-points, the true EICC width and the associated eddy salt transport may not always be accurately represented as the section may also include recirculation in the interior of the BoB. Also, the SMAP SSS product within $\sim 50 \text{ km}$ off the eastern coast and $\sim 20 \text{ km}$ off the southern coast is unavailable because the satellite-measured SSS is highly contaminated in the land–sea transition region (Grodsky et al. 2018; Fig. 5). Since the low-salinity water is often transported out of the BoB along the Sri Lankan eastern and southern coasts in a narrow alongshore current during the northeast monsoon, the calculated eddy salt transport might underrepresent the role of the EICC and the WMC in the BoB freshwater export. Thus, the estimated values should serve as a lower bound of the volume transport and actual eddy salt transport, particularly during the northeast monsoon season.

4) SPRING MONSOON TRANSITION (MARCH–APRIL)

During the spring transition (green shading in Fig. 4), the westward-flowing North Equatorial Current (NEC) is present in the southern BoB at 5° – 8°N (Cutler and Swallow 1984;

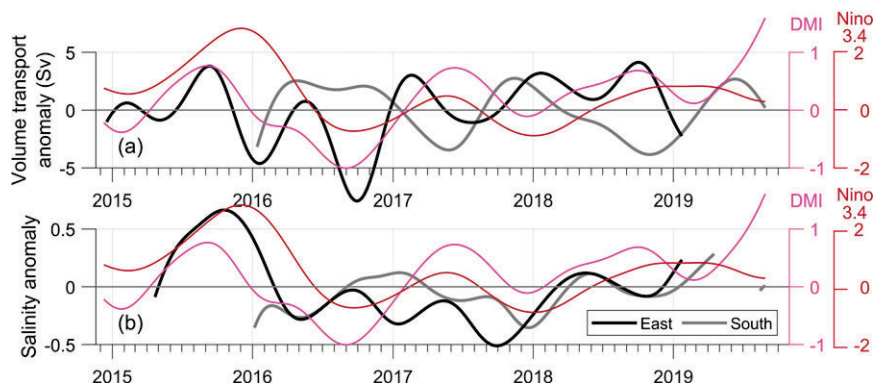


FIG. 7. (a) Volume transport anomalies (Sv) in the surface layer (0–200-m depth) along the Sri Lankan eastern (black) and southern (gray) coasts, and (b) section-mean SMAP SSS anomalies at the eastern (black) and southern (gray) coasts, with the same 120-day low-pass-filtered DMI (magenta) and Niño-3.4 index (maroon) being depicted in both panels. The anomaly values are calculated by removing the seasonal cycles and frequencies higher than 120 days.

Wijesekera et al. 2016a; Fig. 1) and bifurcates at the south-eastern coast of Sri Lanka (Fig. 5a; Shetye et al. 1993; Hacker et al. 1998). The bifurcation produces the observed northward surface flow along the eastern coast (Eigenheer and Quadfasel 2000; Durand et al. 2009; Anutaliya et al. 2017) and westward flow along the Sri Lankan southern coast where the westward-flowing WMC is still present (Figs. 3, 4, 5a). The reconstructed velocity profile shows that the northward current at the eastern section is usually confined to the upper 150–250 m of the water column (Fig. 3a). Below the northward surface layer, a southward-flowing spring undercurrent is apparent along the Sri Lankan eastern coast in March and April throughout the PIES observing period, consistent with the previous study of Anutaliya et al. (2017) that is based on sporadic observations and numerical simulations. As the water being transported by the NEC originates in the southern BoB, the salinity does not deviate much from the mean salinity; thus, eddy salt transport during the spring monsoon transition is often small. Still, the northward flow along the Sri Lanka eastern coast sometimes yields weak positive eddy salt flux, e.g., in 2018 (Fig. 6).

5) SUMMARY ON SEASONAL VARIABILITY

The PIES observations highlight periods of high positive $F_s(t)$ along the eastern coast of Sri Lanka that occur not only during the northeast monsoon season, as found in previous studies (Han and McCreary 2001; Wijesekera et al. 2015), but positive $F_s(t)$ also occurs during the fall monsoon transitions in some years, e.g., 2015 and 2018 (Fig. 6). Despite the continuous eddy salt transport between the south and east sections that mainly occurs during the fall monsoon transition and the northeast monsoon, the correlation between salt fluxes along the eastern and southern coasts is only 0.34 (significant at the 95% confidence level) over the overlapping observed period (Figs. 5a and 6).

d. Year-to-year variability of the regional circulation

1) YEAR-TO-YEAR VARIABILITY OF EDDY SALT FLUX

At the eastern section, the mean eddy salt fluxes for the individual years over the observing period vary from

-0.23×10^6 to $1.05 \times 10^6 \text{ kg s}^{-1}$ with the 2015 annual-mean eddy salt flux calculated only from April to December due to because of the limited availability of the SMAP SSS during that year (Table 2). Still, the time-mean eddy salt flux over the missing months (January–March) is usually small (Figs. 5a and 7a). The uncertainty associated with the annual-mean salt flux estimation is $0.17 \times 10^6 \text{ kg s}^{-1}$ (Table 1). The eddy salt flux across the southern section averaged over each individual year ranges from 0.98×10^6 to $1.51 \times 10^6 \text{ kg s}^{-1}$ with an uncertainty of $0.31 \times 10^6 \text{ kg s}^{-1}$ (Tables 1 and 2), generally higher than that at the eastern section. Note that the low value of mean annual eddy salt flux in 2019 is simply due to the observing period that ends at the beginning of November, and thus does not fully capture the seasonally high eddy salt transport during the northeast monsoon (Fig. 6b).

2) ROLE OF THE 2016 NIOD EVENT ON THE BOUNDARY CURRENTS

To examine year-to-year variability of the boundary currents and salt being transported, the low-frequency components (anomaly) of the volume transport and SSS are calculated by removing the 4-yr seasonal cycle and signals with period shorter than 120 days (Fig. 7a). The anomalies are compared with the low-frequency component of the dipole mode index (DMI) that indicates IOD activity (Saji et al. 1999) and the mean sea surface temperature in the Niño-3.4 region (Niño-3.4 index). The correlation between the transport anomaly off the eastern coast and the DMI is 0.65, suggesting an anomalous southward flow during an IOD event, and likewise the correlation between the transport anomaly off the southern coast and the DMI is -0.52 , suggesting an anomalous eastward flow during an IOD event (Table 3); the correlation is significant at the 95% confidence level. Negative IOD events are often associated with weakening of the easterlies over the equatorial IO in August–September (Saji et al. 1999; Schott et al. 2009), which results in the development of strong downwelling Kelvin waves at the eastern boundary of the IO and around the coastal rim of the BoB, that is, equatorward flow anomaly. The Kelvin waves

TABLE 3. Correlation coefficients r between the low-frequency components of the climate mode indices [dipole mode index (label DMI) and mean sea surface temperature in the Niño-3.4 region (label Niño-3.4)] and the low-frequency components of volume transport, section-mean SSS, and eddy salt transport at the eastern and southern sections. Asterisks denote values that are not significant at the 95% confidence level.

	East		South	
	DMI	Niño-3.4	DMI	Niño-3.4
Volume transport	0.65	0.02*	−0.52	−0.38
Section-mean SSS (entire period)	0.31	0.76	0.13*	−0.06*
Section-mean SSS (July 2016 onward)	0.07*	0.32	—	—
Eddy salt transport	0.67	0.44	−0.40	−0.52

have seemingly little impact off the southern coast (Sreenivas et al. 2012). Also, the seasonal Rossby waves radiated from the eastern BoB are modified by changes in the equatorial winds during IOD events. This usually affects the circulation off the eastern coast of Sri Lanka more than that off the southern coast (Sreenivas et al. 2012). Thus, the volume transport along the Sri Lankan eastern coast is anomalously southward during negative IOD in agreement with findings from previous numerical studies (Thompson et al. 2006; Dandapat et al. 2018), while that along the southern coast exhibits little deviation from its seasonal cycle during the 2016 nIOD event (Figs. 3 and 4). As the nIOD event in 2016 reached its peak in October 2016, the anomalous southward flow along the eastern coast opposed the seasonal northward boundary current during the fall monsoon transition and obscured it.

Negative eddy salt transport along the Sri Lankan eastern coast, corresponding to BoB freshwater import or saline water export, is likely during an nIOD event as suggested by the positive correlation (Table 3). To determine how SSS at the section fluctuates during IOD events, correlation between section-mean SSS and DMI is calculated. Although the correlation between the DMI and the mean SSS off the eastern coast is positive, it is mainly dominated by the anomalously high SSS during the second half of 2015 when a weak positive IOD event and the 2015/16 El Niño simultaneously occur (Fig. 7a). The SSS deviation is in fact small and the SSS increases as the 2016 nIOD develops, hence this is not consistent with the calculated positive correlation for the entire period. Correlation between the low-frequency DMI and low-frequency SSS from July 2016 onward, when the 2015/16 El Niño no longer dominates, is near zero and is not significant at the 95% confidence level (Table 3). The evidence suggests no connection between the IOD variability and the surface salinity pattern at the eastern section even though a previous study indicates low BoB freshwater input during nIOD events (Durand et al. 2011). Therefore, the positive correlation between the low-frequency DMI and eddy salt transport at the eastern section (Table 3) is likely caused by the anomalous volume transport alone (Figs. 4a and 7). The deviation of the SSS at the eastern section does not significantly contribute to

the variation of the eddy salt transport under the influence of the IOD events. Similarly, the DMI only significantly correlates with the volume transport anomaly and does not correlate with the section-mean SSS at the southern section (Fig. 7; Table 3). Thus, the significant anticorrelation there between the low-frequency eddy salt transport and the DMI is likely caused by the anomalous volume transport along the southern coast of Sri Lanka.

During the 2016 nIOD event, the anomalously southward flow along the Sri Lankan eastern coast and anomalously eastward flow along the southern coast cause an anomalous confluence of flow at the southeastern coast slightly after the event is fully developed, during the fall monsoon transition. The confluence is associated with southward flow of saline water yielding a negative eddy salt transport along the Sri Lanka eastern coast and anomalously eastward flow along the southern coast yielding a positive eddy salt transport (Figs. 3, 4, 6; Table 3). Thus, the lowest annual-mean eddy salt transport at the eastern section and highest annual-mean eddy salt transport at the southern section are found in 2016 (Table 2). However, the anomalous confluence of saline water tends to continue northeastward into the BoB at $\sim 85.5^\circ\text{E}$. The circulation pattern is similar to that observed during the southwest monsoon of a neutral year (Fig. 5b). Thus, the overall eddy salt transport into the BoB is expected to increase during the 2016 nIOD event.

3) ROLE OF THE 2015/16 EL NIÑO EVENT ON THE BOUNDARY CURRENTS

Along the Sri Lankan eastern coast, El Niño conditions potentially strengthen the northward flow during the fall monsoon transition and weaken the southward-flowing EICC during the northeast monsoon (Pant et al. 2015) due to the suppression of seasonal downwelling Kelvin waves (Sreenivas et al. 2012). However, the eastern-section transport anomaly does not exhibit such strong deviation from the seasonal cycle during the 2015/16 El Niño (Fig. 7a). The correlation of the transport anomaly with the Niño-3.4 index is near zero and not significant at the 95% confidence level (Table 3). Off the southern coast, weak negative correlation is found between the volume transport anomaly and the low-frequency component of the Niño-3.4 index ($r = -0.38$), suggesting anomalously westward volume transport during El Niño conditions, and vice versa. A previous study based on satellite measurements found that the eastern equatorial wind is anomalously westward during an El Niño causing suppression of the seasonal downwelling Kelvin waves during the northeast monsoon (Sreenivas et al. 2012). As a result, the boundary current along the Sri Lankan southern coast is anomalously westward.

A correlation between the SSS anomaly at the Sri Lankan eastern section and the Niño-3.4 index suggests anomalously saline water during an El Niño event ($r = 0.76$; Fig. 7b; Table 3). However, interannual SSS variability in other parts of the BoB, e.g., the northeastern and central BoB, has shown to be unrelated to ENSO variability (Chaitanya et al. 2015; Pant et al. 2015). As the correlation between volume transport along the eastern coast and the Niño-3.4 index is not significant at the 95% confidence

level at low-frequency time scales, the increase in eddy salt transport along the eastern coast during an El Niño event is mainly dominated by the presence of anomalously saline water (Table 2; Fig. 7b). In contrast, the significant anticorrelation between the low-frequency components of eddy salt transport off the southern coast and Niño-3.4 index likely contributes from the presence of anomalously westward flow during an El Niño event as correlation between the Niño-3.4 and the section-mean SSS is not significant (Table 3).

The PIES observations during the 2015/16 El Niño event show the largest annual eddy salt flux along the Sri Lankan eastern coast in 2015 when highly saline water, up to $10.6 \times 10^6 \text{ kg s}^{-1}$ of salt originating from the AS, is transported northward by the seasonal current during the fall monsoon transition (Figs. 3a, 4a, 5c). The anomalously saline water occurs shortly before the 2015/16 El Niño is fully developed (Fig. 7b). The satellite-derived surface geostrophic current and SSS show that the northward flow of saline water along the eastern coast is a part of an anticyclonic eddy and does not effectively result in a net transport of saline water into the BoB. Off the Sri Lankan southern coast, the PIES were unfortunately not yet deployed during the 2015/16 El Niño. The altimetry-derived surface current in fall of 2015, however, does not deviate from the seasonal mean (not shown), despite the significant anticorrelation between the volume transport anomaly and the low-frequency Niño-3.4 (Table 3). Therefore, the impact of an El Niño on the BoB eddy salt transport is still unclear.

5. Summary

This study provides the first observations of high-frequency continuous volume transport and eddy salt transport measured simultaneously off both the eastern and southern coasts of Sri Lanka. The results provide a better understanding of the regional circulation as well as its contribution to the salt budget in the BoB and the variability on seasonal and interannual time scales. In addition, the PIES measurements together with the vertical structure derived from the Seaglider velocity measurements resolve the temporally varying section-mean velocity of the boundary currents providing the vertical structure of the regional currents, such as the EICC, SLD, WMC, and SMC (e.g., McCreary et al. 1996; Han and McCreary 2001; Wijesekera et al. 2015; Schott et al. 1994). High-frequency variability of the vertical structure of currents occurring during monsoon transitions, that has previously only been examined via numerical simulations (e.g., Jensen 2001; Anutaliya et al. 2017), is also revealed.

Together with the satellite SSS, the PIES measurements and satellite altimetry allows a continuous estimation of the eddy salt transport with reasonable accuracy at these sections. Glider measurements indicate that both velocity and salinity structures are more complex at the southern section and our integrated technique produces an estimated eddy salt transport that has higher uncertainty at that section. Despite this higher uncertainty, the eddy salt transports across both sections compare well to values independently computed from the Seaglider measurements while permitting a view of the higher-frequency component of the transports. The observations

reveal that large sudden changes in eddy salt transport often occur, thus high-frequency and continuous monitoring of the eddy salt transport is required to achieve accurate estimates. The high-frequency observations are also valuable for obtaining a better understanding of intraseasonal events, such as those associated with rainfall events during the monsoon season. The eddy salt transports also show high seasonal and interannual variability. The interannual changes in the boundary current eddy salt flux can be as large as 12%–28% of the entire salt budget of the BoB, while the mean eddy salt flux contributes 9%–27% to the BoB salt budget. As mentioned above, this should be viewed as a lower bound.

The estimates of volume transport and eddy salt transport presented here can be further improved with additional PIES and mooring measurements. The deployment of only two end-point PIES does not always capture the lateral extent of the boundary currents as the width of the boundary current can vary within the section covered. When the boundary current is narrower than the region between the PIES, the estimated volume transport within the section length could include opposing currents resulting in an underestimation of the boundary current transport. Simulations using glider velocity measurements along each PIES section show that an addition of one or two more PIES between the two existing PIES would significantly reduce the transport uncertainty associated with the varying width of the boundary current. With an additional two PIES, the transport uncertainty would be reduced from 3.7 to 0.7 Sv. In addition, neither SMAP nor the Seaglider measures the salinity very close to the coast where a strong salinity signal is expected, particularly during the northeast monsoon. The average innermost salinity measurement taken at the eastern section by the gliders is ~50 km offshore, and the innermost measurement taken by the SMAP satellite is ~50 km offshore. Likewise, the average innermost salinity measurement taken at the southern section by the gliders is ~40 km offshore, and the innermost measurement taken by the SMAP satellite is ~20 km offshore. Additional in situ salinity measurements nearer to the coast, such as might be obtained from a simple shelf mooring, would provide the missing salinity signal and likely improve the accuracy of the eddy salt transport estimation.

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Data availability statement. PIES measurements are freely available on OceanSITES (https://dods.ndbc.noaa.gov/thredds/catalog/oceansites/DATA_GRIDDED/SLBC/catalog.html).

Data collected by Seaglider are freely available online (<http://hdl.handle.net/1773/48130>). Hydrographic profiles are from the National Centers for Environmental Information (Boyer et al. 2013; <https://www.nodc.noaa.gov/access/index.html>). Argo data were collected and made freely available by the International Argo Program and the national programs that contribute to it (<http://www.argo.ucsd.edu>, <http://argo.jcommops.org>). The Argo Program is part of the Global Ocean Observing System. This study has been conducted using altimeter data from the EU Copernicus Marine Service Information (http://marine.copernicus.eu/services-portfolio/access-to-products/?option=com_csw&view=details&product_id=SEALEVEL_GLO_PHY_L4_REP_OBSERVATIONS_008_047). SMAP salinity data are produced by Remote Sensing Systems and sponsored by the NASA Ocean Salinity Science Team and are available online (<https://www.remss.com>).

APPENDIX

Salinity Profile Construction

At each section, local historical hydrography was used to create two salinity lookup tables; the first table has the CTD salinity profiles arranged as a function of surface velocity, while the second table has the profiles arranged as a function of SSS. Satellite-derived surface velocity or salinity at a time and location of interest is used to determine subsurface salinity structure on the basis of the first or second lookup table, respectively. The vertical structure obtained from the lookup table is then scaled such that it passes through the SMAP SSS at the corresponding time and location and through 35.0 at 600-m depth (which is the historical mean salinity at 600-m depth at both sections). The two lookup tables provide us with two options for the section-depth salinity estimation and were used to assess the skill of this method.

To test the lookup tables for the subsurface structure of salinity at the eastern and southern sections, Seaglider salinity measurements were compared with those derived from each lookup table. In most cases, the salinity profiles derived from the two lookup tables are similar. However, the second lookup table arranged as a function of SSS generally provides results that agree better with the Seaglider salinity profiles at the eastern section. At the southern section, the first lookup table with salinity profiles sorted as a function of surface velocity generally gives profiles that agree better with Seaglider salinity profiles. Therefore, the following calculation is based on these “best” salinity lookup tables determined accordingly for each section.

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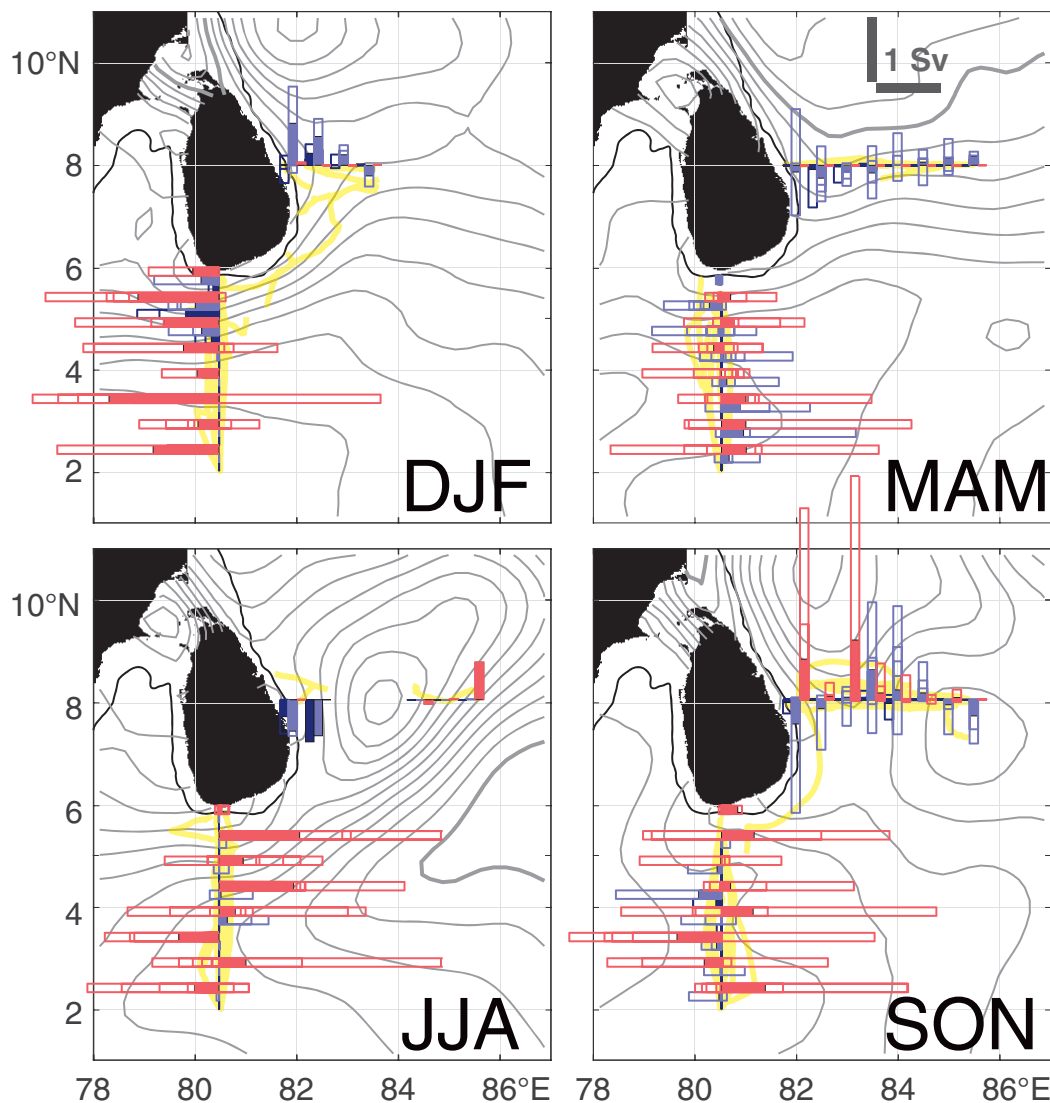


FIG. 14. Seasonal circulation of water masses [salty waters from the Arabian Sea (A1), in red; freshwaters from the Bay of Bengal (B1 and B2), in blue] captured by the gliders over several years, overlaid on the mean SSH contours (2.5-cm intervals) for each season. 1000-m isobath is shown in black. Seasonal mean transports are shown by the solid bars (over all years), and open bars represent estimates from individual sections, in the direction they are transported. Yellow lines show the glider tracks for each season.

important. While the interannual variability of the EUC is likely very important in controlling how and in what volume the high-salinity waters from the Arabian Sea enter the Bay of Bengal, it appears likely that filaments, instabilities, and other processes happening near the Maldives and the southern tip of India play a dominant role for salt transport into the Bay of Bengal. Even in models, the relative importance of these pathways is unclear. More work, including but not limited to the data presented here, will provide insight in this broader circulation.

Interannual monsoon variability

Across the well-sampled 80°30'E line south of Sri Lanka, the transport across each section sampled by the gliders as

a function of time, for each of the three water masses (Fig. 16), shows interannual, seasonal, and intraseasonal variability are well captured. As seen above and in Fig. 16, the general monsoon circulation can be deciphered from all the sections: there is enhanced transport of high-salinity waters by SMC in the SW monsoon (season marked with the orange background), and westward transport of low-salinity waters by NMC during the NE monsoon (periods with green background).

The probability density functions (PDFs) of salinity in the top 150 m (Fig. 17) show the presence of low-salinity water in all seasons. Different colored lines in the figure show different years. The freshwater transports (B1, B2) emphasized above

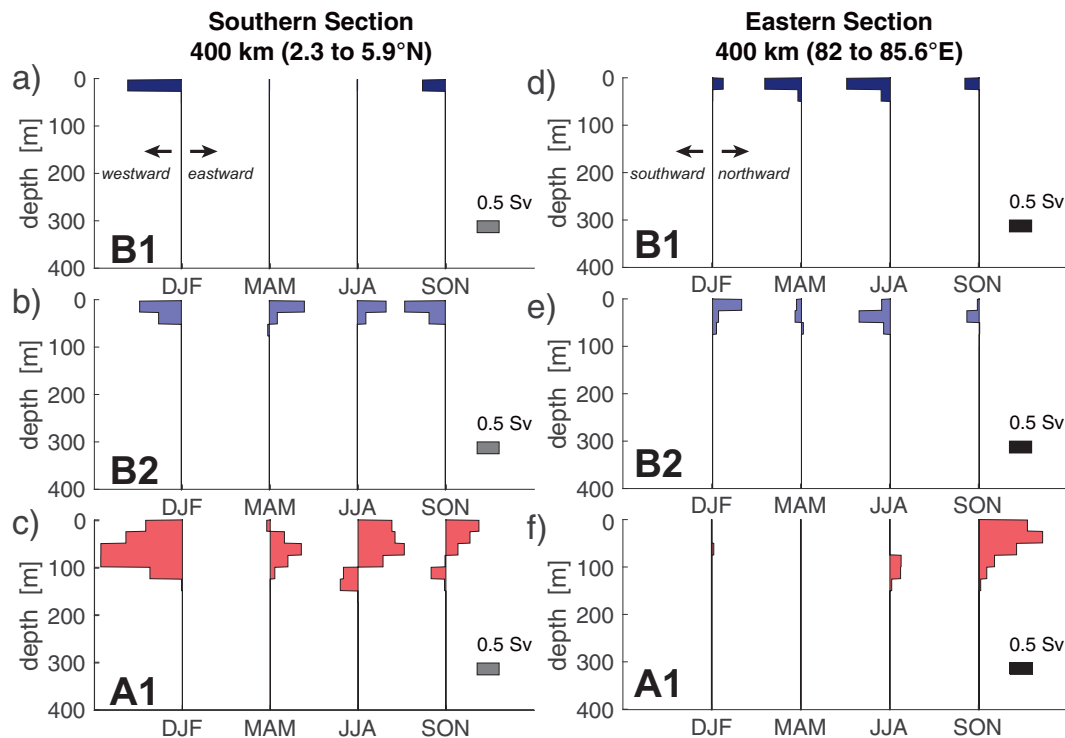


FIG. 15. Transport of water with the properties of water masses (a),(d) B1, (b),(e) B2, and (c),(f) A1 as a function of depth in 25-m bins integrated over the whole section (400 km) for each season for the (left) southern section and (right) eastern section.

are associated with the very low-salinity waters (found near the coast of Sri Lanka), which are quickly advected westward, primarily during the NE monsoon (DJF). The PDFs show that there is proportionally more freshwater in the transition seasons (MAM in particular), but it is more distributed and advected in both directions (see Fig. 14).

There is significant interannual variability. The distributions of both low-salinity waters (blue box in Fig. 17) and high-salinity waters (red box in Fig. 17) changes from year to year, suggesting that the water observed south of Sri Lanka has taken different paths and been modified by different processes. For example, focusing on the last two SW monsoon periods (Fig. 16), there are larger transport of high-salinity waters in 2018 relative to 2019, despite the PDF showing a bit more of salty waters in 2019 (Fig. 17c). This interannual variability is not surprising, particularly given that the ocean conditions around Sri Lanka respond to both local winds (e.g., de Vos et al. 2014) and the timing of annual Rossby waves, linked to Kelvin wave propagating along the equatorial ocean waveguide (Shankar et al. 2002; Rao et al. 2010), both of which are highly variable from year to year.

While a full interpretation of the interannual variability of the circulation presented here requires a much broader perspective, it is worth noting that in 2019 the Sri Lanka Dome was much smaller relative to other years, as seen by the time series of SSH in middle of the mean location of the SLD (Fig. 16e). The SLD is a large cyclonic circulation that develops east of Sri

Lanka during the summer months. Strength and evolution have been linked to the local wind stress, regional forcing and Rossby wave propagation (Vinayachandran and Yamagata 1998; Cullen and Shroyer 2019; Pirro et al. 2020a). The seasonal SSH minimum centered in July (Fig. 16e) is the signature of the SLD. SSH anomalies are negative in 2018 (strong SLD, and large transport of high-salinity water) but positive in 2019 (weak SLD). The data presented here, along with measurements from other sources (e.g., Anutaliya et al. 2017; Hormann et al. 2019; Anutaliya et al. 2022) can be used to explore potential connections between SMC transport south of Sri Lanka and Sri Lanka Dome strength.

5. Summary

Six years of repeated observations, with autonomous gliders collecting over 12 000 profiles along two sections extending south and east from the Sri Lankan coast quantify seasonal variability in the boundary circulation and exchange between the Bay of Bengal and Arabian Sea. The boundary currents within a few hundred kilometers of Sri Lanka carry significant fresh and high-salinity waters between the Arabian Sea and the Bay of Bengal. In addition to reversing with the monsoon winds, this circulation is highly variable at seasonal and interannual time scales, with significant vertical structure (Jensen et al. 2016)

Currents are surface intensified, as observed previously (e.g., Schott et al. 1994), but also show significant vertical

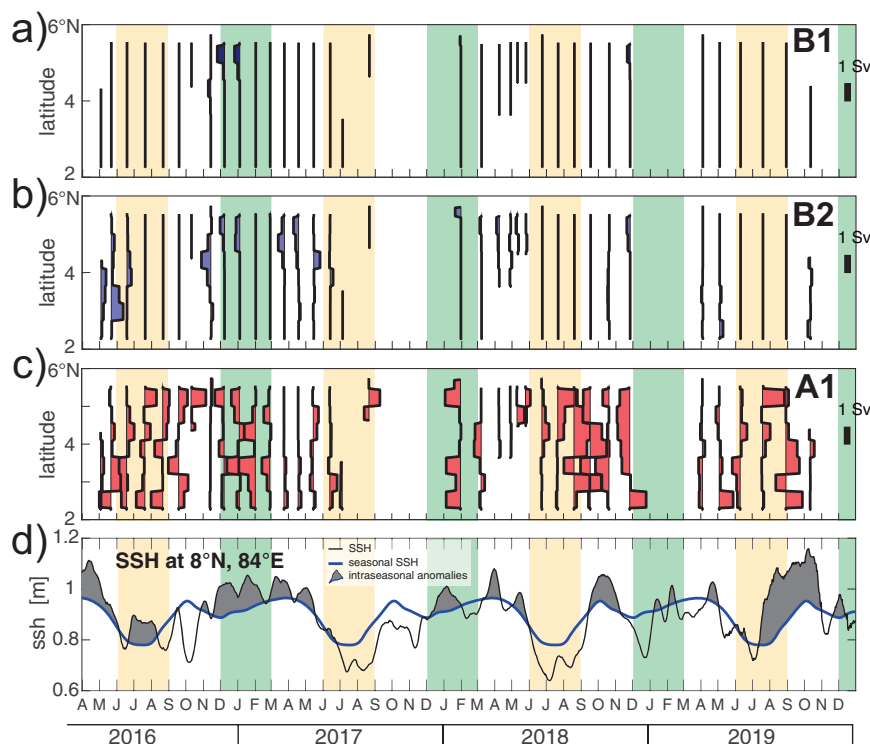


FIG. 16. Transport of water with the properties of water masses (a)–(c) B2, B1, and A1 for all sections across 80.5°E as a function of yearday and latitude (in 50-km bins). (d) Sea surface height time series in the middle of the Sri Lanka Dome, with the 10-yr seasonal average (blue) and positive intra-annual anomalies highlighted.

structure. Flows often reverse with depth, such that subsurface transports of high-salinity waters oppose those in the surface layer. This introduces uncertainty to transports inferred from surface currents and water mass properties. The region south of Sri Lanka, but north of the Equatorial Undercurrent (roughly 2°N) appears to be a dominant pathway for high-salinity Arabian Sea waters entering the Bay of Bengal. While this differs from the pathways recently proposed by Sanchez-Franks et al. (2019), it suggests that circulation in the southeast Arabian Sea (e.g., Lakshadweep eddy, upwelling near the Maldives; de Vos et al. 2014; Pirro et al. 2020b) likely plays an important role in modulating the import of high-salinity water into the Bay of Bengal. The glider estimates of transport are a good step toward quantifying the relative importance of the various pathways of water exchange in the northern Indian Ocean.

As noted previously, the freshwater export during the NE monsoon and spring (MAM) from the Bay of Bengal plays an important role in setting up the upper-ocean stratification in the southeast Arabian Sea (mini warm pool between the Maldives and India; Fig. 7), which is linked to SST and the onset of the SW monsoon (e.g., Roman-Stork et al. 2020a). Irregular sampling and the relatively short 3-yr period covered by gliders make it difficult to draw conclusions from glider data alone, but the results presented should be used to better understand the connections between upper-ocean variability in the southeast

Arabian Sea and the circulation pattern near Sri Lanka. A comprehensive analysis can be conducted by combing with model outputs in future studies, but it is out of the scope of the present paper.

The extensive time series of glider-based observations presented here were one component of a sequence of broad, collaborative, international programs supported by the U.S. Office of Naval Research. These programs provide an unprecedented wealth of sustained observations and focused process studies (Wijesekera et al. 2016b; Centurioni et al. 2017; Anutaliya et al. 2017; Hormann et al. 2019) which, when combined with remotely sensed sea surface height, temperature, and salinity, offer new opportunities to advance understanding of the variability and underlying dynamics of circulation around Sri Lanka.

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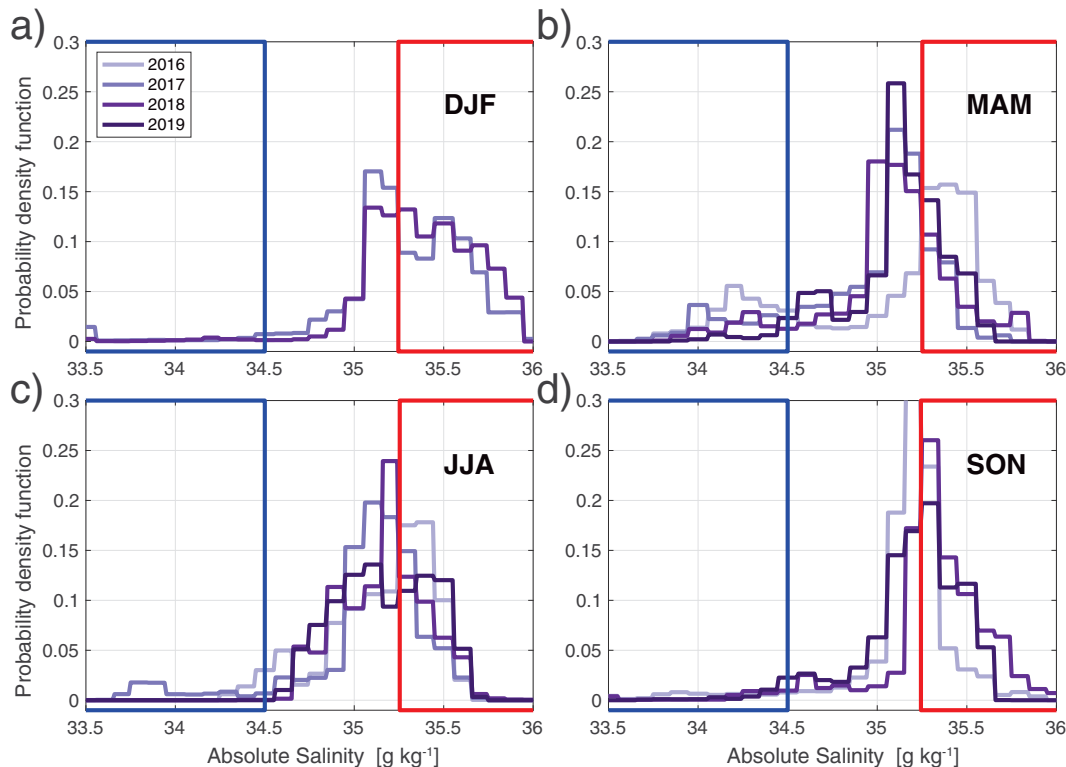


FIG. 17. Probability density function of the salinity along the southern section in the top 150 m from 2° to 5°30'N, for each year and season. Blue and red boxes show the limits of the B2 and A1 water masses, respectively.

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Data availability statement. The gridded sections and seasonal averages obtained from the Seaglider data presented in this paper are available at the University of Washington Digital Library (<http://hdl.handle.net/1773/48130>). The glider data can be downloaded from the IOP group (Lee and Rainville) website (<https://iop.apl.washington.edu>). Remote sensing SST and SSS are available at www.remss.com. Maps of sea surface height and currents were produced and distributed by the Copernicus Marine and Environment Monitoring Service (CMEMS) (<https://doi.org/10.48670/moi-00148>).

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