Characteristics of a Seasonal Front in the Southern Bay of Bengal: Dynamics, Mixing, and Water Mass Transport


a National Research Council, Naval Research Laboratory, Stennis Space Center, Mississippi
b Naval Research Laboratory, Stennis Space Center, Mississippi
c Ocean University, Colombo, Sri Lanka
d Department of Civil Engineering, Environmental and Earth Sciences, University of Notre Dame, Notre Dame, Indiana
e SRR International, Inc., Riviera Beach, Florida

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ABSTRACT: The formation of a sharp oceanic front located south-southeast of Sri Lanka during the southwest monsoon is examined through in situ and remote observations and high-resolution model output. Remote sensing and model output reveal that the front extends approximately 200 km eastward from the southeast coast of Sri Lanka toward the southern Bay of Bengal (BoB). This annually occurring front is associated with the boundary between the southwest monsoon current with high-salinity water to the south, and a weak flow field comprised of relatively fresh BoB water to the north. The front contains a line of high chlorophyll extending from the coastal upwelling zone, often for several hundred kilometers. Elevated turbulent diffusivities $10^{-2}$ m$^2$s$^{-1}$ along with large diapycnal fluxes of heat and salt were found within the front. The formation of the front and vertical transports are linked to local wind stress curl. Large vertical velocities ($\sim 50$ m day$^{-1}$) indicate the importance of ageostrophic, submesoscale processes. To examine these processes, the Ertel potential vorticity (PV) was computed using the observations and numerical model output. The model output shows a ribbon of negative PV along the front between the coastal upwelling zone and two eddies (Sri Lanka Dome and an anticyclonic eddy) typically found in the southern BoB. PV estimates support the view that the flow is susceptible to submesoscale instabilities, which in turn generate high vertical velocities within the front. Frontal upwelling and heightened mixing show that the seasonal front is regionally important to linking the fresh surface water of the BoB with the Arabian Sea.

SIGNIFICANCE STATEMENT: Within the ocean, motions span extraordinarily wide ranges of sizes and time scales. In this study we focus on a narrow, intensified feature called a front. This front occurs in the southern Bay of Bengal during the summer monsoon and forms a boundary between fresher water to the north and saltier water to the south. Features such as this are difficult to study, however, by combining observations made from ships and satellites with output from numerical models of the ocean, we are able to better understand the front. This is important because fronts like the one studied here play a role in determining the pathways of heat within the ocean, which, in turn, may feedback into the atmosphere and weather patterns.

KEYWORDS: Fronts; Instability; Atmosphere-ocean interaction; Diapycnal mixing; Ekman pumping/transport; Mesoscale processes

1. Introduction

The southern Bay of Bengal (BoB), south and east of the coast of Sri Lanka, is unique because it is located at the confluence of multiple seasonally reversing currents. Freshwater input from rivers and rainfall to the north is exchanged with high-salinity waters of the Arabian Sea through export pathways to the south and west in order to maintain the basinwide balance. During the winter monsoon (December–March), fresh surface water from the BoB is transported into the Arabian Sea via the northeast monsoon current, while during the summer monsoon (June–September), the strong southwest-erly winds of the southwest monsoon act as basin-scale forcing, spinning up the southwest monsoon current (SMC; Praveen Kumar et al. 2012; Kripalani et al. 2004). While the basin-scale forcing provided by the Indian Ocean monsoon (IOM) generates large-scale flows such as the SMC, it also
provides the background for a variety of mesoscale and submesoscale motions, fronts, and instabilities that eventually provide a pathway for turbulent mixing and vertical transports. Many studies indicate that the southern BoB and the SMC act as an important linkage between the water masses of the Arabian Sea and the BoB (Murtu et al. 1992; Jensen 2001, 2003; Vinayachandran et al. 2013; Mahadevan et al. 2016; Lee et al. 2016; Jensen et al. 2016; Pirro et al. 2020a); however, the detailed dynamics of how these two basins exchange water are not fully understood.

Observational and modeling studies have highlighted the importance of oceanic–atmospheric coupling within the Bay of Bengal in determining the time of onset, duration, and strength of the IOM (e.g., Schott and McCreary 2001; Bhat et al. 2001; Webster et al. 2002; Rao et al. 2011; Wijesekera et al. 2016a; Vinayachandran et al. 2018; Shroyer et al. 2021). Understanding of water mass exchanges and associated heat and salinity fluxes that are driving factors for upper-ocean heat content, mixed-layer structure, and SST of the BoB is critical for improving our understanding of the BoB circulation as well as the predictability of the IOM. Mixing forced by submesoscale features and fronts is likely to play an important role for the vertical transfer of heat and salinity between subsurface and surface waters and, as a consequence, drives water mass exchange. Submesoscale cross-frontal transport has been shown to enhance transport within regional models (Jensen et al. 2018). Forcing from the monsoon drives the large-scale circulation in the BoB; however, energetic mesoscale and submesoscale features interact with and complicate the regional circulation. Hence, the mechanisms that drive turbulent mixing and water mass exchange between the BoB and the Arabian Sea, and the details of how these processes influence air–sea interactions in the southern BoB, are not well understood.

During the summer monsoon, a sharp frontal feature is formed south and east of Sri Lanka. The front is characterized by a narrowing and sharpening of the SMC between the Sri Lanka Dome (SLD) to the north and an anticyclonic eddy (AE) to the south. A number of studies have broadly noted this feature (Schott et al. 1994; Vinayachandran et al. 1999, 2004; Pirro et al. 2020a; Wijesekera et al. 2016b). For instance, Vinayachandran et al. (1999) describes the path of the SMC into the BoB as an intrusion, where the flow of the SMC is forced by both Ekman pumping as it passes south and southeast of Sri Lanka as well as by Rossby wave radiation connected to the Wyrtki jet (Wyrtki 1973). Some studies have reported that the SMC in this location undergoes interactions with eddies to the north and south; for example, Wijesekera et al. (2016b) noted that a cyclonic and anticyclonic eddy pair influenced the location and strength of the SMC during the summer of 2014, and Pirro et al. (2020b) reported the presence of an anticyclonic eddy southeast of Sri Lanka in July 2018 with surface velocities up to 1 m s⁻¹ and a signature extending approximately to the depth of the thermocline (~100 m).

South of Sri Lanka, the frontal region exhibits coastal upwelling driven in part by monsoon winds (Vinayachandran 2004; Pirro et al. 2020a). The cold upwelled water is then advected offshore to the north and east, forming a pronounced filament in sea surface temperature (de Vos et al. 2014). The offshore advection appears to be associated with an eastward oceanic jet at the boundary of the SLD. Several hypotheses have been suggested for SLD formation, the first being that the cyclonic wind stress curl over the southwestern BoB provides the forcing (McCreary et al. 1996; Vinayachandran and Yamagata 1998; Schott and McCreary 2001). A second mechanism, proposed by de Vos et al. (2014), suggests that flow separation of SMC from the southeastern coastal boundary of Sri Lanka may lead to the generation of the SLD. As the SMC passes Sri Lanka, frictional drag imparts cyclonic (anticyclonic) relative vorticity ωc, causing the flow to turn northward, leading to the development of the SLD. Vinayachandran and Yamagata (1998) propose that Rossby waves, propagating from the coast of Sumatra, interact with the SMC, leading to the generation of the AE, while Pirro et al. (2020a) hypothesize that the AE is generated by the arrested Rossby wave response of the SMC to perturbations by the Sri Lankan coast.

The focus of the work presented here will be on the formation and evolution of the sharp frontal feature south and east of Sri Lanka (Fig. 1a). The front formation appears to be linked with wind stress forcing, coastal upwelling to the west, and both the SLD to the north and the AE to the south.

There is a large body of literature examining the formation and evolution of mixed-layer fronts in the open ocean. Early work focused on the conditions that lead to frontogenesis. Hoskins and Bretherton (1972) propose a mechanism in which convergent surface flow imposed upon an initially small gradient in horizontal density intensifies in the presence of lateral strain and convergence. Studies by De Szoeke (1980) and De Szoeke and Richman (1981, 1984) highlight the importance of horizontal variations in the wind stress field in driving frontogenesis, while Cushman-Roisin (1981) examines how wind stress drives both horizontal advection and vertical mixing, leading to both convergent and confluent frontogenesis. Expanding on this earlier work, Rudnick and Davis (1988) examine analytic solutions for three conditions that lead to frontogenesis: two processes that rely on mixed-layer convergences and divergences and a third process that occurs during the shoaling of the barotropic tide in shallow water. The former two are of relevance to this manuscript. Spall (1995) explores the structure of ageostrophic circulation that leads from baroclinic instability and the assumptions of conservation of potential density and potential vorticity.

Later work focused on the time evolution of upper-ocean front as well as submesoscale processes occurring on scales of O(1–10) km. There have been several mechanisms proposed to explain the development of submesoscale structures in the presence of horizontal density gradients. Boccaletti et al. (2007) model mixed-layer instabilities arise as density fronts, which slump and restratify the upper ocean. Thomas and Lee (2005) suggest that alongfront winds can generate cross-front Ekman transport, which can drive dense water over buoyant, resulting in secondary ageostrophic circulation that further sharpens the front. Thomas and Ferrari (2008) examine the importance of friction versus frontogenesis for the restratification of the mixed
layer and find that wind-induced friction often dominates frontogenesis, implying that interactions between a front and the wind forcing must be carefully considered. For further review of frontal processes, McWilliams (2021) gives an excellent overview of the present understanding of frontogenesis in the ocean. While in situ observations of fronts have been made in other regions (Pollard and Regier 1992; Jinadasa et al. 2016; Ramachandran et al. 2018), to our knowledge there are relatively few observations of the frontal feature described in this manuscript.

Motivated by this, we present a detailed analysis of the front, using both remote-sensed and in situ data as well as high-resolution regional model output to create a complete description of the front from the mesoscale to small-scale turbulent mixing properties. Additionally, the Ertel potential vorticity (PV) is used as a diagnostic tool to shed light on possible mechanisms that drive submesoscale dynamics within the front.

The paper is organized as follows: in section 2, we describe the ship surveys and in situ observations, the satellite data, and the model output, along with briefly outlining the methodology used for estimates of turbulent kinetic energy (TKE) dissipation rates. In section 3a, the large-scale characteristics of the front, seasonal variability, and the relationship between the front and local wind forcing are described. The frontal cross sections are described in section 3b. In section 3c, estimates of vertical velocities due to local wind forcing through the process of Ekman pumping as well as frontal velocities using horizontal convergence are presented. In section 3d, temperature and salinity measurements in conjunction with the microstructure data are presented in order to estimate turbulent heat and salinity fluxes. PV estimates from both in situ measurements and model output are presented in section 3e. Last, in section 3f, model output of the front is examined. A discussion of the results is presented in section 4, and conclusions are given in section 5.

2. Instrumentation, observations, and methods

We utilize in situ measurements, remote observations, and model output to aid in the characterization of the frontal feature as part of an international effort sponsored by the U.S. Naval Research Laboratory and U.S. Office of Naval Research (Shroyer et al. 2021; Wijesekera et al. 2022). These include ship-based cross-front conductivity–temperature–depth (CTD) profiles from two different shipboard campaigns. Additionally, vertical microstructure profiler (VMP) data were collected and analyzed concurrently for the second of these ship-based frontal surveys. To relate the cross-frontal scales observed through ship surveys with larger-scale dynamics in the region, we
employ high-resolution ocean model output, high-resolution atmospheric-forcing fields from atmospheric models, and remotely sensed data such as satellite altimetry and chlorophyll-a (Chl-a) concentration. In the following subsections, we will briefly describe each individual dataset and model output used.

a. Frontal survey in 2018

During the 2018 research cruise, a ScanFish survey of the front was undertaken from the R/V Thomas G. Thompson. A ScanFish is a towed, undulating vehicle system (Jarosz et al. 2014). Flaps control the up and down movement of the towed vehicle. During the survey, the ScanFish was equipped with both a CTD and an optical instrument package to measure chlorophyll-a concentration. Chl-a concentrations reported in here are converted from instrument counts to concentration by the following procedure: Chl-a (µg L⁻¹) = scale factor × (total counts – dark counts), where dark counts for this particular instrument were 60, and a scale factor of 0.0112 µg L⁻¹ count⁻¹ was predetermined using a monoculture of phytoplankton (Thalassiosira weissflogii). Because the relationship between fluorescence and in situ Chl-a concentrations is highly variable, the absolute magnitudes of Chl-a concentration should be regarded with caution, as the primary focus within this manuscript is on the relative magnitudes and spatial patterns rather than absolute concentrations.

Between 18 and 19 July 2018, five cross-frontal transects were performed between 82.2° and 82.5°E, beginning north of the front at the easternmost section and ending south of the front at the westernmost section. The ScanFish was “flown” between depths of 5 and 100 m and was towed at 3.5 m s⁻¹, covering approximately 0.02° of latitude during each 10-min up and down profiling sequence. The ScanFish data were interpolated to a 2-m vertical grid and a 0.02° horizontal grid, which corresponds to a resolution of about 2 km. Figure 1c shows the track of the 2018 ScanFish survey along with the potential density anomaly σθ and velocity at 22 m depth. During the frontal survey, ship-mounted acoustic Doppler current profiler (ADCP) data were collected with a 75-kHz ADCP; therefore, the shallowest velocity measurement was at the first data bin of 22 m for the 2018 frontal survey. The 75-kHz ADCP was sampled with 5-min ensembles yielding a horizontal resolution of about 1 km.

b. Frontal survey in 2019

The 2019 frontal survey was conducted in June from the R/V Sally Ride. Three north–south cross-frontal sectional lines were made using the ship-based CTD, with stations separated by approximately 0.05° of latitude. The CTD sampled at 24 Hz and downcast profiles were binned at 1-m vertical resolution. During the CTD survey, shipboard ADCP data were collected with 150- and 300-kHz ADCPs with a 2- and 5-min ensemble averaging, respectively. Velocity profiles from 300- and 150-kHz ADCPs were binned horizontally from their original resolution onto a 3-km grid, which yields velocity coverage between 100- and 10-m depths, which is concurrent with the CTD surveys. The locations of three CTD frontal surveys are displayed in Fig. 1b, where the western cross section along 81.80°E was taken on 18 June, and the eastern cross sections along 82.05°E were taken on 19 and 21 June, respectively.

After the first CTD frontal survey section was completed on 19 June, a VMP survey line was performed at the location of the frontal cross section. The duration of the VMP survey was approximately 12 h, and stations were spaced between 3 and 6 km apart, with an average spacing of 4.7 km. The VMP records data internally and carries two airfoil shear probes, an FP07 fast-response thermistor, a three-axis accelerometer, and a pressure sensor. The VMP was manufactured by Rockland Oceanographic Services, Inc. (RSI), Canada, and is similar to the instrument described in Wolk et al. (2002). The specific instrument configuration for the VMP survey described here is similar to the configuration used in Lozovatsky et al. (2019) and Jarosz et al. (2014). Reference temperature and salinity were also recorded from a Sea-Bird SBE3 and four other sensors attached to the VMP. Owing to calibration problems and malfunctions of the SBE3 sensor, we instead used CTD data collected on 18 June to provide information on background temperature, salinity, potential density, and buoyancy frequency. The VMP was deployed from the starboard side of the ship with a free-falling rate of about 0.7 m s⁻¹. VMP casts were limited to depths of ~100–150 m. At each individual VMP station, 2–3 individual casts were taken, and the results were then bin averaged by depth.

Velocity shear and temperature measurements were processed using software provided by RSI. We provide a brief overview of this processing; however, a more comprehensive explanation of the data-processing procedures is available in RSI’s Technical Note 028 (Lueck 2016), and further discussion can be found in Lozovatsky et al. (2019) and Jarosz et al. (2014). The small-scale shear spectra were fitted to the Nasmyth shear spectrum (e.g., Nasmyth 1970; Wolk et al. 2002) in order to calculate the TKE dissipation rate ε within consecutive vertical segments at 2-s intervals, which corresponds to a vertical resolution of ~1.5 m. Coherent acceleration signals were removed from the shear spectra using the Goodman et al. (2006) algorithm. The VMP data could be contaminated by ship motion between the sea surface and a depth of about 15 m, and consequently our analysis focuses on data below 15 m. Within each vertical section, frequency spectra of the 512-Hz shear signal are taken and transformed into wavenumber k space using velocity derived from the VMP pressure sensor. The shear spectra are then integrated, providing an estimate of ε:

\[ \varepsilon = \frac{15}{2} \nu \int_{0}^{\infty} \Psi(k)dk, \]  

where \( \Psi(k) \) is the wavenumber spectrum of horizontal shear (Lueck 2016; Osborn 1974; Gregg 1991), and ν is the kinematic viscosity of seawater. A third-order polynomial was fit to the shear spectra in order to determine the location of the spectral minimum, an indicator for the onset of noise domination (Lueck 2016).

c. Satellite observations and model output

To help contextualize our frontal surveys, we utilize a variety of remotely sensed observations and model output. Sea surface height (SSH) anomalies and geostrophic velocities are derived from AVISO products and provided on a
GlobColour-merged Chl-a products are based on satellite observation [SeaWiFS, Medium-Resolution Imaging Spectrometer (MERIS), MODIS, VIIRS, and Ocean and Land Color Instrument (OLCI)] and are retrieved from the Copernicus Marine Environmental Monitoring Service (CMEMS; https://hermes.acri.fr/). The spatial resolution is 4 km for Chl-a. Wind stress is retrieved directly from ECMWF ERA5 reanalysis output (https://cds.climate.copernicus.eu/). ERA5 output is originally provided in 1-h increments, however, in this manuscript, we examine daily averages of ERA5 wind stress.

To further examine spatial features of the front and to qualitatively compare observations with the model output, we utilize the output from a regional ocean model, the Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS). The COAMPS regional run used in this study has been shown to resolve a variety of submesoscale fronts and features (Jensen et al. 2018). COAMPS includes an atmospheric, ocean, and wave models that are fully coupled, covering the Indian Ocean and the BoB to just south of the equator (Jensen et al. 2016). The COAMPS model uses 60 vertical levels for both the atmosphere and the ocean. The ocean contains 45 sigma levels with 15 z levels below depths of 330 m, and in the upper 10 m of the ocean, the vertical resolution is 0.5 m or less, depending on water depth. Vertical mixing is computed using the 2.5-level turbulence closure scheme by Mellor and Yamada (1982), modified to include contributions from Stokes drift and Langmuir circulations (Kantha and Anne Clayson 2004). COAMPS has horizontal resolutions of 2 km in the ocean and 6 km in the atmosphere and covers the BoB from 70° to 100°E and from 1.5°S to 23°N. The ocean model is covered by a spherical grid with a resolution of 1/54° on a 1600 × 1320 grid with a time step of 30 s. The ocean model is initialized, and its boundaries are forced every 6 h by output from the 1/12° global Hybrid Coordinate Ocean Model (HYCOM; Metzger et al. 2014). The ocean model includes eight semidiurnal and diurnal tidal-forcing components. Monthly climatological discharges of river outputs are included in the ocean model. For the analysis presented here, we have chosen model output from when the modeled front is qualitatively similar to the frontal development seen in the 2019 frontal survey. The daily averaged output from 26 June 2019 is used for a calculation of PV in section 3.

3. Results

a. Large-scale frontal features

We begin by describing the large-scale spatial features and temporal variability of the front. Figure 2 shows the along-
track velocities from the 150-kHz ship-based ADCP during the 2019 cruise at 40-m vertical resolution starting at a depth of 16 m and ending at a depth of 216 m (Fig. 2). The lack of continuous, stationary data during both frontal surveys makes it difficult to remove the effects of tides aliased into our ADCP measurements; however, by using moored estimates of tidal amplitudes in the southern BoB, we can estimate some of the impacts of tides. Wijesekera et al. (2019) report maximum diurnal amplitudes less than 1 cm s$^{-1}$, while M2 and S2 components are less than 5 cm s$^{-1}$. Because the magnitudes of the currents described in this survey are much larger than those of tides, we do not expect the aliasing of tides to cause major errors in broad structures of the velocities reported here. Furthermore, the fastest moving mode-1 internal tide has a phase speed and a wavelength of about 2.2 m s$^{-1}$ and 100 km, respectively, and the inertial period is about 5–6 days. These long waves have relatively less impact on the short space–time record examined here. However, high-frequency and short wavelength motions can contaminate ADCP signal.

Within the upper 100 m, the intensified eastward flow of the SMC can be seen around 6°N and between 80.5° and 82.5°E (Figs. 2a,c). Beyond this, the SMC can be seen as it hooks to the northeast at 7°N, between 83° and 84°E. In these areas, currents peak at about 1.5 m s$^{-1}$. Immediately north of the SMC, there is a sharp transition to generally southward weaker flow. This abrupt transition between the eastward flow of the SMC and southward flow on the northern track marks the position of the seasonal front. At depths deeper than 96 m, the SMC drastically decays, being replaced by a south-southwest current flowing parallel to the coast of Sri Lanka (Figs. 2d–f).

Because the duration of ship surveys was limited to several days, both satellite remote sensing and atmospheric model reanalysis products were used to better understand temporal characteristics and evolution of the front. Figure 3 shows Ekman drift (derived from ECMWF reanalysis fields), geostrophic currents (derived from SSH), and Chl-a (from GlobColour) for a number of snapshots throughout 2018. To examine low-frequency background flow, both the geostrophic currents (red arrows in Fig. 3) and combined surface currents, i.e., the sum of the geostrophic currents and the Ekman drift derived from ERA5 wind stress (black arrows in Fig. 3), were estimated by following Vinayachandran (2004). During periods of time where Ekman transport does not contribute significantly to the combined velocity, the combined velocities (black arrows) overlay the geostrophic velocities.

FIG. 3. Large-scale surface currents and Chl-a (mg m$^{-3}$) for a number of snapshots throughout 2018. Geostrophic currents (calculated from SSH) are shown by the red arrows. The sum of both Ekman drift currents derived from ERA5 wind stress and geostrophic currents are shown by the black arrows. The 1 m s$^{-1}$ scale is given by the blue bar.
red arrows) and indicate that the total flow is primarily geostrophic. When the arrows differ, however, a higher percentage of the overall flow in that area is due to Ekman drift, indicating that estimated Ekman transport is nonnegligible relative to geostrophic flow in the surface layer. Spatially, the front varies between 200 and 500 km in length, beginning at approximately 81°E and extending to approximately 83°–86°E depending on its relative strength and stage of development, which is concomitant with the monsoonal cycle. The frontal axis runs approximately east–west, with the cross-front direction being north–south during June and July, while in August, the front initially extends east-southeast and then hooks northeast following the SMC.

The surface currents south of Sri Lanka are relatively weak, just prior to the onset of the southwest monsoon (28 May; Fig. 3). At that time, elevated Chl-a can be seen off the southern coast of Sri Lanka, but a line of Chl-a had not yet developed toward the east along the SMC. By 21 June 2018 (Fig. 3), the flow shifted east, indicating that the SMC has begun to spin up. The early stages of a front can be seen as an east–west filament in Chl-a between 81° and 82.5°E. As the monsoon progressed in July–August 2018, the SMC was fully developed, and the Chl-a signature of the front is clearly visible in Fig. 3. During June–August 2018, the geostrophic and combined velocities differed considerably, indicating that Ekman drift may be important to the formation of the front, although this does not conclusively show that the convergence of Ekman flow is the primary cause of front formation (Cushman-Roisin 1981). During most of the duration of the front, the prevailing wind was in alignment with the front. Throughout July–August 2018, the cyclonic circulation of the SLD appeared north of the SMC and east of Sri Lanka. By early September (Fig. 3), the front deteriorated, and the SMC weakened as the southwest monsoon weakened. The surface currents at this time were primarily geostrophic (Fig. 3). In February 2019, the flow south of Sri Lanka reversed to southwest flows, representing the typical pattern of the winter monsoon (Fig. 3).

### b. Frontal cross sections

Frontal cross sections taken during 2018 and 2019 allow local vertical and small-scale horizontal processes to be examined. Figures 4–8 show depth–latitude cross sections of the front taken from the 2018 ScanFish survey. Temperature $T$, salinity $S$, Chl-a, the potential density anomaly $\sigma_\theta$, as well as the east–west or zonal $U$ and north–south or meridional $V$ components of velocity are shown. In all figures, the frontal...
axis is clearly visible, indicated by the tightly grouped, vertically outcropping isopycnals between 6.0° and 6.3°N. In each transect, the frontal axis is not always at the same latitude, perhaps due to spatial–temporal variability including tides. The front is relatively shallow and confined to the upper 75–100 m of the water column, with a well-mixed layer on the southern side of the front extending to ~50 m. On the northern side of the front, however, the mixed layer is often confined to the upper 10 m. Because of the shallow nature of the 2018 front, part of the frontal outcropping is not totally resolved. In Figs. 4a, 5a, 6a, and 7a, the water temperatures in the vicinity of the isopycnal outcropping are 1°–2°C cooler than temperatures on either side of the front. Cross-sectional plots of T, S, U, and V from the 2019 ship-based CTD survey are shown in Figs. 9–11. In 2019, the frontal axis varies between 6.2° and 6.5°N. Due to variations in the position of the front throughout the CTD sections, Fig. 9 has the best coverage of the southern side of the front, while Fig. 11 primarily shows the northern side of the front. In all three transects, the steeply slanted isopycnals of the frontal axis can be seen. The MLD appears to be slightly deeper in 2019 than was observed in 2018, extending to ~60 m on the southern side of the front and ~20 m on the northern side where freshwater stratification is strong. Although there are some differences between the observed front during both cruises, there are many features that remain consistent. The front has a strong cross-frontal density gradient, and relative contributions of salinity and temperature were examined by comparing the haline contraction and thermal expansion terms of the linearized equation of state. Density is dominated by salinity within the upper 20–25 m. Below 25 m both salinity and temperature contribute to density, although the salinity contribution diminishes with depth. Density varies by as much as 1 kg m⁻³ over the course of 20 km in the cross-frontal direction. The water mass to the north is warm and fresh with a salinity around 34 psu, while for the water on the southern side, salinity exceeds 35 psu. During both years, an increased salinity “tongue” or intrusion can be seen at depths of 50–100 m, flowing under the fresher water to the north (Figs. 4–11). During both ship surveys, the SMC appears as a surface-intensified zonal jet flowing eastward on the southern side of the front, extending to a depth of ~100 m (Figs. 2, 4–11). The meridional velocity was found to be convergent at the front, and in the case of ScanFish (section 1) in 2018 and CTD (sections 2 and 3) in 2019, a reversal in the meridional velocity was observed, with currents on the north side of the front flowing southward and currents on the south side of the front flowing northward. This is especially evident during the 2019 cruise and can also be seen in Fig. 1.
c. Estimates of vertical velocities

Prompted by the observations of convergent flow in the region of the front, we employ two different methods to estimate vertical velocities within the region of the front. We begin by calculating Ekman pumping from ERA5 wind stress. Ekman pumping may give a representative estimate of open-ocean upwelling due to local wind forcing. Figure 12 shows the monthly averaged Ekman pumping velocities calculated from ERA5 output over the course of 2018. During the summer monsoon, there is a dominant pattern of alternating positive and negative wind stress curl, and during the period between June and August, there is a pronounced area of alternating upwelling and downwelling extending northeastward from the eastern coast of Sri Lanka. The area of upwelling contained in this pattern appears to be close to the position of the front. Both the positive and negative areas of wind stress curl that occur on the east coast of Sri Lanka appear to be caused by the topographic wind shadow of the island. Close to the coast, predicted vertical velocities exceed 10 m day\(^{-1}\), which tapers off to about 1.5 m day\(^{-1}\).

Although Ekman pumping provides a rough estimate of vertical velocities in an area, it does not directly address upwelling/downwelling that is forced by convergent motions near the front due to the resolution of ECMWF output. Thus, we utilize mass continuity to make estimates of vertical velocities within the front, and the vertical velocity \(w\), at a given depth \(h\), can be expressed as

\[
w(h) = -\int_{h}^{z} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dz,
\]

where \(u\), \(v\), and \(w\) are the zonal, meridional, and vertical components of velocity, respectively; \(dz\) is the differential of the vertical coordinate \(z\); and \(w = 0\) at \(z = 0\). The terms \(\partial u/\partial x\) and \(\partial v/\partial y\) are zonal and meridional gradients of \(u\) and \(v\), respectively. Observations indicate that the alongfront velocity gradients are smaller than cross-front velocity gradients. Therefore, we assume that the zonal flow (in the alongfront direction) is steady, and the alongfront velocity gradient is negligible, i.e., \((\partial u/\partial x) = 0\). To observationally test these assumptions, we compare gradients of velocity and density averaged over the upper 50 m in both the alongfront and across-front directions. Alongfront gradients were calculated as a centered difference between 2019 CTD survey line “1” and line “2.” The across-front zonal and meridional velocity gradients were found to be 12.2 and 6.2 times larger than in the alongfront direction, while the across-front density gradient was found to be 8.4 times larger than the gradient in the alongfront direction. Following these assumptions, \(w\) is estimated in the cross-front direction as a function of depth by integrating Eq. (2) with a surface boundary condition of \(w = 0\).
at z = 0 and also by assuming that velocities are vertically uniform between the shallowest ADCP velocity record and the surface. We integrate cross-frontal velocity gradients between the surface and 50 m to obtained vertical velocity estimates at 50-m depth. Because of the surface intensification nature of the front, especially in 2018, we expect that the lack of shallow velocity measurements will introduce some errors and may underestimate w within the front. For both 2018 and 2019, we smoothed the meridional velocities to 8 km following the approach of Ramachandran et al. (2018) to eliminate the effect of high-frequency internal-wave noise with wavelengths of several kilometers or less. Owing to the nature of our sampling, we were unable to remove the effects of larger-scale horizontal motions such as inertial oscillations and tides, which we expect are the main source of errors in this calculation. We use an instrument error of 1 cm s$^{-1}$, and by utilizing the maximum reported amplitude for the M2 tide of \( \sim 3 \) cm s$^{-1}$ from Wijesekera et al. (2019), we infer that the total error in the cross-frontal velocity measurements is \( \sim 5 \) cm s$^{-1}$. This error is then propagated through our calculations of vertical velocities from continuity. Figure 13 presents estimates of frontal vertical velocities at 50 m for all 2019 CTD sections (Figs. 13a–c) and 2018 ScanFish sections (Figs. 13d–h). For each section, the 95% confidence intervals of propagated error estimates are shown with black dashed lines. Figures 13a–h also show contours of \( \sigma_z \) for each cross section. To further estimate possible errors in our estimates of vertical velocity, all five of the ScanFish sections are averaged. Each section is aligned with respect to the frontal axis using the location of maximal cross-frontal density gradient averaged over the upper 25 m to define the frontal axis. Vertical velocity estimates from each section are then averaged using 5-km bins. Figure 13i shows the vertical velocities from each individually aligned front as well as the averaged cross-frontal vertical velocity. Although in both cases the total error in vertical velocities is large, there remains a clear pattern of up/downwelling across the front. During the 2019 frontal surveys, upwelling velocities as large as 40 m day$^{-1}$ seem to be linked to the frontal axis in the vicinity of outcropping isopycnals in Figs. 13a–d and 13g, although the exact location of the upwelling varies between sections. Downwelling of \( \sim 10 \) m day$^{-1}$ is also observed between 6.1$^\circ$ and 6.2$^\circ$N along the southern edge of the front in Figs. 13a and 13c; however, the frontal axis is positioned in such a way that this area of the front was positioned on the southernmost end of the third CTD cross section taken in 2019 (Fig. 13c), and it is, therefore, difficult to determine if the full region of downwelling is resolved in this section. The 2019 results appear to be somewhat consistent with the ScanFish results in 2018. Upwelling velocities of \( \sim 50 \) m day$^{-1}$ in the ScanFish survey appear to be linked to the frontal axis in the immediate area of...
outcropping, although estimates may have large uncertainties due to lack of ADCP velocities in the upper 20 m. This may be due to the lack of shallow ADCP velocities during the ScanFish surveys or simply a result of noise contamination due to motions less than the averaging length of 8 km. Vertical velocities will be further examined in section 3f through the use of the COAMPS model output. In Figs. 13b and 13h, the pattern of vertical velocities appears to be consistent with single-celled frontal circulation as is depicted in Spall (1995); however, in Figs. 13c and 13d-g, frontal downwelling is accompanied by velocity maxima on either side of the front, which resembles the circulation pattern for cold filamentary intensification proposed by McWilliams et al. (2009).

d. Microstructure

A VMP survey taken across the front during the 2019 cruise allows for examination of turbulent mixing in the immediate vicinity of the front. Figure 14a shows $\epsilon$ for the entire cross-frontal VMP section. Data from depths shallower than 10 m have been discarded due to possible contamination from ship motion. Vertical dashed lines in Fig. 14 indicate the locations of each VMP station along the section. The frontal axis at the time of the VMP survey can be identified by the vertically outcropping isopycnals located at 6.3°N. TKE dissipation rates in the area of the front vary by several orders of magnitude, between $\epsilon = 10^{-4}$ and $10^{-9}$ W kg$^{-1}$. On the southern side of the front, the dissipation rate is slightly elevated (between $\epsilon = 10^{-5}$ and $10^{-7}$ W kg$^{-1}$) to the base of the mixed layer at a depth of about 50 m and then drops below $10^{-8}$ W kg$^{-1}$ for 20-50 m before becoming slightly elevated at depths below 50 m. Peak TKE dissipation exceeding $10^{-5}$ W kg$^{-1}$ occurs between 6.1° and 6.25°N at a depth between 60 and 80 m, which is collocated with the base of the mixed layer on the south side of the front as well as the top of the salty intrusion seen in Fig. 9b. Dissipation is also elevated and above $\epsilon = 10^{-5}$ W kg$^{-1}$ on both the north and south sides of the location where the 22 kg m$^{-3}$ isopycnal outcrops at the surface around 6.3°N. Last, at a depth of around 100 m to the north of the front, there is a patch of active turbulence where $\epsilon = 10^{-6}$ W kg$^{-1}$. The largest values of turbulent dissipation within the front are near the base of the mixed layer and on the southern side of the frontal axis. Using the microstructure profiles, we estimate eddy diffusivity using

$$K_p = \frac{\Gamma_\epsilon}{N^2}$$  

(Osborn 1980), with a mixing efficiency of $\Gamma = 0.2$ (Gregg et al. 2018) and buoyancy frequency $N$ calculated with a background stratification from the CTD survey (section 1). The spatial patterns of $K_p$ (Fig. 14b) mirror those of TKE dissipation (Fig. 14a).
with the highest values exceeding $\sim 10^{-2} \text{ m}^2 \text{s}^{-1}$ between 60 and 80 m close to the frontal axis. Equation (3) assumes a downgradient property transport; our estimates from within the mixed layer tend to become unrealistically large in the mixed layer where $N$ becomes small, in which case the diffusivity is determined by dynamics of homogeneous turbulence (Large et al. 1994; Conry et al. 2020).

Last, Figs. 14c and 14d present estimates of the vertical heat flux $J'_q$ and salt flux $J'_s$, where heat and salt fluxes are defined as

$$J'_q = -\rho_0 c_p K_T T_z,$$

$$J'_s = -\rho_0 K_S S_z,$$

where $c_p$ is the heat capacity of seawater, $K_T$ and $K_S$ are the eddy diffusivities for temperature and salinity, $T_z$ is the local vertical temperature gradient, $S_z$ is the local vertical salinity gradient, and $\rho_0$ is the density of seawater. Here, we take $K_T = K_S$. Because $T_z$ is negative and monotonic throughout the water column, all heat flux is downward, representing heat from surface water being mixed downward. Maximum downward heat flux of over 100 W m$^{-2}$ occurs in the location of the high-salinity intrusion, following the 22 kg m$^{-3}$ isopycnal. Patterns of salinity flux, however, take both signs due to subsurface salinity maxima associated with the intruding Arabian Sea water. Within the mixed layer on the southern side of the front, the salinity flux is generally positive, owing to a slightly negative vertical gradient of salinity. At the base of the mixed layer and top of the pycnocline, the salinity flux is negative and greater than $10^{-3}$ psu m s$^{-1}$, suggesting intense mixing of low-salinity BoB and high-salinity Arabian Sea water masses. When averaged over depth and across the front, salinity flux downward is of the order $10^{-4}$ psu m s$^{-1}$.

e. Ertel potential vorticity

Ertel PV is useful as a diagnostic to better understand where and how the front may be susceptible to various mechanisms of instability. Submesoscale instabilities can arise when the PV is of the opposite sign to the local planetary vorticity (Hoskins 1974; Haine and Marshall 1998; Thomas and Shakespeare 2015; Yu et al. 2021). Neglecting terms with vertical velocity, the PV $q$ is given by

$$q = (f + \zeta) \nabla^2 + \left( \frac{\partial u}{\partial x} \frac{\partial B}{\partial y} - \frac{\partial u}{\partial y} \frac{\partial B}{\partial x} \right) k, \tag{6}$$
where \( B = - g(\rho/\rho_0) \) is the buoyancy, \( g \) is the gravitational acceleration, \( \rho \) is the potential density, \( \rho_0 \) is the reference density taken to be 1027 kg m\(^{-3}\), \( N^2 = \partial B/\partial z \), \( f \) is the Coriolis parameter, and \( \xi = (\partial u/\partial x) - (\partial u/\partial y) \) is the vertical component of relative vorticity. For consistency, all fields used in the calculation of PV were smoothed to 8 km to eliminate high-frequency internal wave noise, with the exception of the 2019 CTD-derived density, which had a station spacing close to 8 km. For our calculations of the frontal potential vorticity, we use the one-ship method (Shcherbina et al. 2013; Ramachandran et al. 2018). This approximation assumes that the ship track is in the cross-frontal direction and neglects gradients of both density and velocity in the alongfront direction. Following this simplification, the PV is

\[
q \sim \left( f - \frac{\partial u}{\partial y} \right) N^2 + \left( \frac{\partial u}{\partial z} \frac{\partial B}{\partial y} \right).
\]

(7)

Negative or anomalously low values of PV can occur during a variety of flow instabilities: 1) gravitational instability occurs when the fluid column is unstably stratified, 2) sufficiently large horizontally sheared flows can cause centrifugal instability (Haine and Marshall 1998), and 3) anticyclonic along-isopycnal shear can result in symmetric instability (Thomas et al. 2013). To identify the causes of negative PV, we separate \( q \) into both vertical and baroclinic components, where the vertical component of \( q \) is given by \(( f - (\partial u/\partial y)) N^2 \), and the baroclinic component consists of \((\partial u/\partial z) \partial B/\partial y\), as was done in Yu et al. (2021). Figure 15 shows PV calculated for the 2019 CTD frontal cross sections plotted as function of depth and latitude. For each of the three sections, total PV (top), vertical PV (middle), and baroclinic PV (bottom) are shown. In all three cross sections, some isolated areas of negative total PV can be seen. In both CTD (sections 1 and 2), the largest most coherent “bolus” of negative total PV is located within the south side of the front, at the base of the mixed layer, immediately adjacent to the frontal outcropping of isopycnals. In some parts of these areas, the total negative PV reaches \(-1 \times 10^{-3} \text{ s}^{-3}\). The negative PV anomaly appears to be mainly coming from the baroclinic term, a product of the vertical alongfront shear and the horizontal, cross-frontal buoyancy gradient. Vertical shear of the zonal current \( \partial u/\partial z \) is generally positive, while the lateral cross-frontal buoyancy gradient is sufficiently large and negative resulting in a region of negative baroclinic PV. The positive vertical shear of the zonal current is associated with the SMC. In the vertical PV term, \( f \) is positive, and \( \partial u/\partial y \) is generally negative, so the entire term is generally positive. The only factor that allows the vertical term to become anomalously low (but still positive) is the vertical stratification, which becomes small within the mixed layer. Below the mixed layer, the vertical PV is strong enough
to overcome the slightly negative baroclinic PV, as it is typically several times larger ($5 \times 10^{-8} \text{ s}^{-3}$); however, within the mixed layer, vertical PV vanishes, and as a result, the total PV becomes negative. This effect is particularly pronounced on the southern side of the front, where the mixed layer is much deeper than on the northern side.

Figure 16 shows PV calculated for three of five 2018 ScanFish frontal cross sections. As in Fig. 15, we plot total PV (top), vertical PV (middle), and baroclinic PV (bottom). The trend is similar between the 2018 and 2019 frontal surveys; however, the total magnitude of the PV measured in 2018 is lower than in 2019. Furthermore, in the 2019 sections, the generally negative baroclinic term is large enough to overcome the positive vertical PV within the mixed layer. In 2018, there are very few regions where this is the case. Differences in PV estimates between the two ship surveys may be attributed to variability within the front. A number of frontal attributes such as mixed-layer depth and cross-frontal density structure do change between the two surveys. Background conditions associated with 2018 and 2019 monsoons are not identical, which in turn influence the strength of SMC and coastal upwelling off Sri Lanka (Wijesekera et al. 2022). There are some differences in sampling between 2018 and 2019, although care has been made to make our PV estimates as consistent as possible between the two years. In the 2018 estimates of PV, some noise appears to be introduced via the $\partial u/\partial y$ term, which appears as faint vertical banding. Because this signal appears at scales near the filtering length scale of 8 km, we suspect this may be due to high-frequency motions not completely eliminated by the filter. Although this noise does not appear to change the broad patterns seen in the PV estimates, it illustrates the difficulties of estimating PV with relatively sparse observations. Within the front, the horizontal cross-frontal density gradients and vertical shear contained in the baroclinic PV term appear to be the largest factor in determining areas of negative PV, while the cross-frontal horizontal velocity gradient and vertical stratification within the vertical component tend to increase the local PV. For this reason, instances of negative PV are observed mainly in areas with 1) low stratification and 2) sharp cross-frontal density gradients.

f. COAMPS front

The use of the COAMPS model output provides a better spatial understanding of the front as well as a qualitative assessment of how well the front is represented within a high-resolution coupled ocean model. Figure 17 depicts 24-h averages of model temperature, salinity, vertical velocity $w$, and PV from 26 June 2019. Model $T$, $S$, and PV are shown at a
depth of 15 m, while the vertical velocity is an estimate integrated from the surface to a depth of 50 m. To remain consistent with the observational estimates of \( w \), vertical velocities were calculated via integrated horizontal convergence, as in Eq. (2). Because the frontal axis in the model appears to be oriented in a more east-northeasterly direction than was observed, we did not neglect terms with horizontal gradients in the east–west direction. The same applies to our calculations of model PV. Spatially, the front exhibits similar horizontal scales and sharp gradients of salinity and temperature. A low-temperature anomaly extends along the front exhibiting temperatures of \( 1^\circ \text{C} \), cooler than water to the north or south of the front. Likewise, the cross-frontal salinity gradient is sharp and coherent along the length of the front, extending from 81.5° to 85°E. These structures are more consistent with our ship-based and remotely sensed observations presented earlier. In temperature, salinity, and velocity, the characteristic signatures of both the SLD to the north of the front and AE to the south can also be seen. Calculations of the vertical velocity resulting from horizontal convergence (Fig. 17c) reveal that the front had a broad upwelling that is \( \sim 50 \text{ km in width} \) with vertical velocities ranging from 50 to 80 m day\(^{-1}\). This upwelling is present along the entire extent of the front. Additionally, a much narrower \( \sim 25 \text{ m day}^{-1} \) downwelling along the southern edge of the front can be seen. Although both the observational estimates and model output both show strong vertical velocities in the region of the front, the upwelling associated with the front appears much more coherent within the model output than in the estimates derived from the cross-frontal ship surveys. The model exhibits a ribbon of elevated PV along the front between the coastal upwelling zone and the two eddies (SLD and AE). On the northern side of the front, the PV is anomalously high, while on the southern extent of the front, the PV is negative. The negative PV anomaly begins south of Sri Lanka and extends to 85°E, at which point only a positive PV anomaly can be seen. Just south of Sri Lanka, the negative and positive PVs are located in an area with a submesoscale cyclonic eddy with strong upwelling and cold-water anomaly.

Because the COAMPS model is coupled to atmospheric forcing, local forcing terms within the model can be examined. This not only provides valuable insight into the nature of the ocean–atmosphere coupling in the region of the front but also allows for comparisons to the earlier estimates of vertical velocities based on ECMWF wind stress, which is spatially much coarser. Figure 18 shows 24-h averages of model wind stress, net heat flux, wind stress curl, and resultant Ekman pumping from 26 June 2019. The wind stress in COAMPS qualitatively resembles the ECMWF ERA5 output, where the dominant feature is a northeastward wind stress resulting from the seasonal southwesterly monsoon winds. A weakening of these winds can be seen east of Sri Lanka, which results in an alternating pattern of wind stress curl and resultant Ekman pumping. Figure 18 shows 24-h averages of model wind stress, net heat flux, wind stress curl, and resultant Ekman pumping from 26 June 2019. The wind stress in COAMPS qualitatively resembles the ECMWF ERA5 output, where the dominant feature is a northeastward wind stress resulting from the seasonal southwesterly monsoon winds. A weakening of these winds can be seen east of Sri Lanka, which results in an alternating pattern of wind stress curl and resultant Ekman pumping.
important to note, however, that the wind stress curl shown in Fig. 12 depicts monthly averages, while the COAMPS wind stress is a daily averaged product. Additionally, COAMPS is output at higher spatial resolution than ECMWF. For these reasons, we expect the COAMPS wind stress to contain high-frequency intrinsic variability, whereas the ECMWF product is much smoother. Despite model resolution, the large-scale patterns in both can be broadly compared. In both the COAMPS wind stress curl and Ekman pumping, the front is clearly visible as a positive anomaly between 81.5° and 84°E. This anomaly is closely linked to the upwelling/cold surface signature seen in Fig. 17a. To further examine the structure of the front utilizing the model, a cross section of the front is scrutinized. Figures 19a–d show representative model cross sections of temperature, salinity, $U$, and $V$. The position of the cross section is 82.4°E and was selected to be consistent with the locations of the ship-based surveys. The frontal axis is positioned just north of 6°N, marked by the steeply slanted isopycnal outcropping that was noted in the ship surveys. In COAMPS, the MLD appears to be slightly shallower than was observed on the southern side of the front, extending to ~40 m, while on the northern side of the front, the MLD is slightly deeper than was seen in observations at ~30 m. Similarly, the SMC in the model appears to be confined to the upper 30 m, whereas in the frontal observations, the SMC often extended past 50 m. Although there are some differences between the modeled and the observed front, it should be noted that the variation between the model and both observational surveys is of similar magnitude to the differences noted between the 2018 and 2019 surveys. As was the case in the observations, the water mass to the north is warm and fresh, with the salinity around 34 psu, while on the southern side the
salinity exceeds 35 psu. The SMC appears as a surface-intensiﬁed zonal jet ﬂowing eastward on the southern side of the front, which is also consistent with the observations.

Figure 19f shows modeled cross-frontal density overlaid with the vertical velocity calculated in a similar manner to those calculated from the 2018 and 2019 in situ data. As is seen in Fig. 17c, the front appears to be associated with sharp upwelling that peaks in the area of the front with the largest lateral buoyancy gradient. The upwelling occurs for approximately 25 km across the front, with the strongest upwelling occurring in a band about 10 km wide. The vertical velocity across the front is of a similar magnitude to the observational estimates; however, the upwelling in the model is more coherent with the frontal axis. Possible reasons for this will be discussed in the following section. The cross-sectional modeled PV presented in Fig. 19e shows areas of low PV occurring in alignment with the frontal axis. This is somewhat consistent with the results presented in section 3f in showing general background PV is positive; however, in the presence of the strong lateral buoyancy gradients associated with the front, there are regions of negative PV.

4. Discussion

In this section, we will discuss the results of this study in a broader context. Here, we attempt to explain the observed sharpness of the front based on theories of frontogenesis. Additionally, we will show evidence that the front is susceptible to submesoscale instabilities due to the presence of negative PV. We then will discuss the impact of the front on regional mixing within the BoB. Next, we will utilize COAMPS output to offer a plausible explanation for the observed negative PV anomaly. Using both observations and model output, we will provide a qualitative interpretation of mechanisms that may be important to the formation and maintenance of the front. Although a detailed quantitative analysis of model output is out of the scope of this paper, we will contextualize our observations in the framework of dynamical frontal theories. Finally, we will discuss the consequences of turbulent mixing within the front and how the model output compares with the observations.

a. Frontal formation

The results presented in section 3a show that on the first order, a formation of the front is linked to local wind stress curl and spinup of the SMC during the monsoon. The magnitude of the Ekman currents during the formation and evolution of the front is roughly on the same order as the geostrophic flow, which suggests that wind forcing is important to the front. However, questions remain: what are the important factors that support the sharpening of the front? Is the local
interaction between the SMC and the local wind important for explaining the sharpness of the front? Both the observations and model suggest that the frontal circulation is consistent with ageostrophic frontal secondary circulation, with upwelling occurring near the frontal axis, within the less-dense water mass, and local downwelling occurring to the south; however, in the model output, the upwelling appears to be shifted toward the frontal axis, which may be attributed to a broader-scale upwelling generated by Ekman pumping. The upwelling/downwelling patterns seen along the light/dense side of the front are consistent with theories of frontal formation and secondary circulation (e.g., Hoskins and Bretherton 1972; Spall 1995). Thomas and Lee (2005) propose a mechanism in which down-front winds lead to vertical instability within a front. Mixing by this mechanism redistributes the buoyancy across the front and leads to the ageostrophic secondary circulation. As mentioned in section 3a, the dominant wind during the times of peak frontal intensity is oriented down front, meaning that the wind is blowing in the direction of the frontal jet; however, the density gradient is aligned in the opposite direction (with dense water to the south and the alongfront wind blowing eastward) to that observed in Thomas and Lee (2005), and as a result, the down-front wind does not have the same destabilizing effect. In the front described here, the resulting southward Ekman transport should advect the lighter, low-salinity water on the north side of the front over the salty water to the south, resulting in strengthening of the vertical stratification.

Although the down-front wind may not account for the necessary forcing to sharpen the front, there may be other mechanisms acting to positively reinforce the frontal structure. In both observations and model output, the frontal axis is linked to the intense upwelling. This upwelling is likely to be a combination of the vertical Ekman pumping from the local wind stress curl and ageostrophic secondary circulation associated with the front. Another possible mechanism to explain the sharpness of the front could be cold filamentary intensification (McWilliams et al. 2009). As noted in section 3c, in some survey sections, downwelling in the region of the front is accompanied by upwelling on either side of the front. This double-cell circulation resembles the circulation pattern for cold filamentary intensification. This process is analogous to frontogenesis but is adapted for flow conditions in which a narrow filament of the cool, dense water is bordered on both sides by the less dense water. Horizontal deformation acts to sharpen the filament, accelerated by characteristic double-celled ageostrophic circulation accompanied by convergent surface flow and central downwelling. The front discussed in this paper certainly shares many of the features, such as the heavy cool water near the surface and the light warm water on the northern and southern sides of it. This is particularly evident in the cross sections presented in Figs. 10a and 11a and can also be seen in model output (Fig. 17a), where the front appears as a narrow band of the cool water near the surface extending to the east and northeast. Surface salinity is low to the north and high to the south, forming a sharp front.
and 17b). It is likely that some of the processes described by cold filamentary intensification occurs; however, it is also important to note that the front has both a cold filament in temperature and a north–south gradient in salinity, which results in a feature that has some analog to a cold filament, while also retaining some features of an oceanic front.

Geostrophic balance explains the surface-intensified current, and the observed background isopycnal slopes from both the 2018 ScanFish and 2019 CTD surveys predict eastward geostrophic flows on the order of 0.5–0.7 m s$^{-1}$. However, this does not provide a mechanism for maintaining a sharp front. In both the ScanFish and CTD cross sections, as well as in the COAMPS output, the northward flow is associated with the south side of the front. We can consider that the front would be in the partially semigeostrophic momentum balance, where $f\nu = Du_y/Dt$ and $u_y$ is the cross-front ageostrophic velocity, $u_0$ is the alongfront geostrophic velocity, and $D/Dt$ is the total derivative (Hoskins 1975). If the front accelerates in the alongfront direction, this would imply a northward ageostrophic flow, which is consistent with the observed northward and eastward currents south of the front and the weak currents to the north of the front in both the ScanFish and CTD cross sections as well as in the model output. Observations in the alongfront direction are too sparse to demonstrate the existence of an alongfront acceleration; however, if such acceleration is present, the resulting northward flow would push the upwelled high-density water into the low-density water along the northern edge of the front, intensifying the cross-front density gradient and sharpening the front. Frontal upwelling decreases the SST along the front, which, together with topographic effects from Sri Lanka, results in an increase in the local wind stress curl. The impact of the front can be seen in Figs. 18c and 18d, and the impact of SST gradients on the local atmospheric boundary layer and wind stress has been previously noted in mesoscale to large-scale fronts (Chelton et al. 2001; Small et al. 2008) as well as on smaller scales (Sullivan et al. 2020, 2021). The local increase in positive wind stress curl results in Ekman pumping (or upwelling) along the front, thereby providing a positive feedback for frontal maintenance and could be an explanation for why the front remains sharp for such a long distance and extended time.

b. Evidence of instabilities

In both the 2019 CTD survey and the model output, substantial negative PV anomalies exist within the frontal region, which in turn provide the conditions for frontal instabilities. In the 2018 ScanFish survey, there is less evidence for the existence of large regions with negative PV, which may be due to the depth of the available velocity data combined with the relatively shallow nature of the front during that year. However, even the low PV may still be enough to provide conditions for frontal instability. The low PV may be susceptible to forced symmetric instability if “subjected to subtractive fluxes” (Thomas et al. 2013; Ramachandran et al. 2018). This would include surface buoyancy and wind stress, and given how coupled both surface heat flux and wind stress curl appear to be with the front in COAMPS (Fig. 18), it may not be
unreasonable to assume that these fluxes can play an active role in the generation of instabilities.

c. **Consequences of mixing on regional heat and salt transport**

During the cross-frontal VMP survey, elevated turbulence was found to result in the estimated vertical turbulent heat flux exceeding $-100$ W m$^{-2}$, which was of the same order as the daily averaged model surface heat flux, and the salinity flux exceeded $10^{-3}$ psu m s$^{-1}$, which is equivalent to a freshwater exchange of 100 mm h$^{-1}$ using a reference salinity of 35 psu. Assuming that our estimated average turbulent salinity flux of $10^{-4}$ psu m s$^{-1}$ is representative, we estimate that the front is responsible for 10 mm h$^{-1}$ of freshwater exchange. The basinwide estimate for freshwater input reported in Sengupta et al. (2006) is about 0.2 mm h$^{-1}$, implying that freshwater exchange in the area of the front may be occurring at 50 times the basin-averaged rate. These are averaged quantities; there are individual points where the estimates are an order of magnitude higher than these, making net fluxes difficult to quantify. This is expected based on how these quantities are statistically distributed (e.g., Lozovatsky et al. 2017, 2023), and it should be noted that we may underestimate the fluxes in some areas of locally high turbulent mixing. Even so, the downward fluxes of these magnitudes represent a significant source of freshwater transformation and heat transport. Coupling these observations with the length of the front suggests that the front has an important influence on mixing in the BoB.

d. **Origin of low-PV water along front**

PV is a semiconserved quantity, meaning that following the flow, PV is only modified via mixing, diabatic, or frictional processes such as surface heating/cooling or wind stress. The model output shows positive PV on the north side of the front and a negative PV on the south side. As shown in Figs. 12 and 18, the wind stress curl is positive. D’Asaro (1988) suggests a mechanism for the generation of submesoscale vortices and the formation of PV through boundary layer effects and flow separation. Following D’Asaro (1988), boundary friction south of Sri Lanka can lead to the generation of positive PV. The combination of the boundary mixing and wind stress
forcing may be the source of the coherent, positive PV seen on the northern side of the front in the model output (Fig. 17d). The spatial extent of the negative PV anomaly along the modeled front suggests that some of the low-PV water may originate near the coast via coastal upwelling (Fig. 17c). As the water upwells from the coast and is advected along the SMC, the local meridional buoyancy gradient becomes negative enough to dominate the overall potential vorticity. This effect would be reinforced by the frontogenetic mechanism proposed in section 4a and is likely the result of both the advection of coastally upwelled water and reinforcement through frontal upwelling/circulation. This explanation fits the patterns of negative PV coming from the baroclinic and vertical components shown in Figs. 15 and 16. An examination of the model output reveals that the frontal buoyancy gradient in the baroclinic PV term is responsible for regions of the negative PV along the front; however, there is also a component of the negative PV arising from the negative absolute vorticity in the vertical component. The fact that these two terms are comparable in magnitude in many areas suggests that other factors such as a north–south gradient in velocity along the southern edge of the front may be important but was not well resolved in the ship surveys.

e. Model comparison and vertical frontal velocities

From the examination of the COAMPS output presented in section 3f, it is clear that the representation of the front in the model is qualitatively consistent with observations. Modeled vertical velocities show upwelling as large as 50 m day$^{-1}$ at the front, which brings cold and high-salinity water to the mixed layer/surface with the outcropping isopycnals at the front. The model also produces strong northward and eastward currents south of the front and weak currents north of the front. The upwelled high-density water, just south of the frontal edge, can be pushed northward and thus intensify the frontal density gradients. It is likely that the wind stress curl is increased due to air–sea interactions with the cooler upwelled water along the front, resulting in a positive reinforcement of the local Ekman pumping.

The properties of the involved water masses, presence of both the SMC and intensified front, and lateral gradients match between the observations and model. However, the estimates of the vertical velocities show that the model upwelling is more consistently aligned with the frontal axis and, as a consequence, is more consistent with ageostrophic secondary frontal circulation in which upwelling is associated with the light side of the front and downwelling occurs on the heavy side of the front, as in Spall (1995). To some extent, evidence of the cold filamentary intensification as described in McWilliams et al. (2009) can also be seen because downwelling occurs on either side of the front (filament). A plausible explanation for the differences in modeled and observed frontal vertical velocities is the introduction of unresolved velocity variability from inertial waves and internal tides being aliased into the observational estimates (Jithin et al. 2019; Jensen et al. 2020). The short duration of the ship-based observations makes it
dif
difficult or impossible to remove these effects. Because the
front most likely contains energetic meanders and eddies,
two-dimensional analysis cannot fully characterize these
more complex dynamics. Nonetheless, the estimates devel-
oped in this study can provide plausible explanations for the
formation of the front and its regional impact on mixing and
water mass transport.

5. Conclusions

This study presents analyses of satellite-based and in situ ob-
servations along with high-resolution numerical model outputs
in order to examine the formation, evolution, and consequen-
ces of a sharp oceanic front located south and southeast of Sri
Lanka during the IOM. Key findings are as follows:

1) Using ECMWF data, we show that the magnitude of the
Ekman drift in the region of the front is of a comparable
magnitude to that of the surface geostrophic currents.
This suggest that local wind forcing and wind stress curl
are important forcing factors to the front. On the first or-
der, frontal formation and vertical transports seem linked
to the local wind stress curl. Estimated Ekman transport
and large vertical velocities on the order of 50 m day\(^{-1}\) indi-
cate the importance of submesoscale processes.

2) Diffusivity within the front is relatively high: \(K_e \sim 10^{-4} \text{ m}^2 \text{s}^{-1}\) or above over most of the region covered by
the VMP survey, with values near the frontal axis being
above \(10^{-2} \text{ m}^2 \text{s}^{-1}\). During peak intensity, the observa-
tions and model output reveal the front to be laterally
sharp and over 200 km in length, consistent with fronto-
genesis and possible baroclinic instabilities based on the
PV anomaly calculations summarized below. Elevated
mixing and turbulent diffusivities result in high vertical
turbulent fluxes of heat and salt, suggesting that the front
is an important factor for water mass transformation
within the BoB. The observed heat and salinity fluxes, in
conjunction with our local estimates of vertical velocities,
suggest that the front is an important region for the ex-
change of the BoB and Arabian Sea water, i.e., the fresh
warm water is readily transported to the interior and
mixed with the saltier water to the south.

3) To examine these ageostrophic processes, the PV was
computed using the observations and numerical model

\[ \text{Model Transect at 82.3518° E} \]

\[ \text{Fig. 19. Modeled cross sections of (a) temperature (°C), (b) salinity (psu), (c) zonal and (d) meridional velocities}
\[ \text{(m s}^{-1}\text{), and (e) Ertel PV (s}^{-1}\text{) taken on 26 Jun 2019. (f) The potential density (kg m}^{-3}\text{) with vertical velocity at}
\[ \text{50 m (m day}^{-1}\text{; scale on right y axis) overlaid with a black line. The vertical velocity is similar to those displayed in}
\[ \text{Fig. 13. In each plot, contours of potential density anomaly } \sigma_b \text{ (kg m}^{-3}\text{) are overlaid.} \]
output. Cross-frontal estimates show low and negative PV patches in the frontal vicinity. The model output shows a ribbon of negative PV along the front. The PV estimates support the view that the flow is susceptible to a variety of submesoscale instabilities, which in turn likely generate high vertical velocities within the front. PV within the front can be anomalously low and even negative, which creates favorable conditions for a variety of frontal instabilities. Possibly these instabilities play a role in sharpening the front. As the front intensifies, eventually processes such as shear and gravitationally generated instabilities will lead to vertical mixing.

The regional model output from COAMPS shows many similarities with the observations. Key properties of the front as represented in the model such as current velocities and vertical and lateral gradients of temperature and salinity are shown to qualitatively match our observations, making the model a useful tool in the understanding of features such as the front described in this paper. Model vertical velocities show upwelling as large as 50 m day$^{-1}$ at the front, which brings cold and high-salinity water to the mixed layer/surface with outcropping isopycnals at the front. The model also produces strong northward and eastward currents south of the front and weak currents to the north of the front. The upwelled high-density water just south of the frontal edge can be pushed northward, thus intensifying the frontal density gradients, leading to upwelling and a cool-surface expression. The cooler temperature along the front results in air-sea interactions and increase in the wind stress, which further increase the local upwelling, providing a mechanism for maintenance of the front.

Small-scale frontal dynamics are difficult to study through observations; however, combining the in situ data from two ship-based surveys with the remotely sensed data and the regional model output yields several key insights into the nature of a seasonal front in the southern BoB. The formation of the front is linked to local wind stress provided by the seasonal monsoon. Vertical velocities inferred in the area of the front suggest that frontal processes produce ageostrophic secondary circulation. The presence of the low and negative PV both in the observations and model output suggest that submesoscale instabilities within the front may sharpen the front and eventually lead to elevated turbulent mixing. The examination of the turbulent salt and heat fluxes showed that the front has significant regional consequences to the overall transport of these quantities.

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Data availability statement. Community supported satellite data are available from https://www.avisio.altimetry.fr/en/data/products/sea-surface-height-products/global.html (sea surface height and associated geostrophic velocities); https://psl.noaa.gov/ (daily maps of SST); and https://cds.climate.copernicus.eu/ (ECMWF-ERA5 reanalysis products). Project-supported datasets are subjected to the signed agreement (memorandum of understanding) between the United States and Sri Lanka, and restrictions will apply until 2025. This time frame is intended to allow for students, postdoctoral fellows, and investigators supported under the project to have sufficient time to publish their results. After the restriction period, data may be requested from the corresponding author.

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