The Shallow Overturning Circulation in the Indian Ocean

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ABSTRACT

The number of in situ observations in the Indian Ocean has dramatically increased over the past 15 years thanks to the implementation of the Argo profiling float program. This study estimates the mean circulation in the Indian Ocean using hydrographic observations obtained from both Argo and conductivity–temperature–depth (CTD) observations. Absolute velocity at the Argo float parking depth is used so there is no need to assume a level of no motion. Results reveal previously unknown features in addition to well-known currents and water masses. Some newly identified features include the lack of an interior pathway to the equator from the southern Indian Ocean in the pycnocline, indicating that water parcels must transit through the western boundary to reach the equator. High potential vorticity (PV) intrudes from the western coast of Australia in the depth range of the Subantarctic Mode Water, which leads to a structure similar to a PV barrier. The subtropical anticyclonic gyre retreats poleward with depth, as happens in the subtropical Atlantic and Pacific. An eastward flow was found in the eastern basin along 15°S at the depth of the Antarctic Intermediate Water—a feature expected from property distributions but never before detected in velocity estimates. Meridional mass transport indicates about 10 Sv (1 Sv = 10⁶ m³ s⁻¹) southward flow at 6°S and 18 Sv northward flow at 20°S, which results in meridional convergence of currents and thermocline depression at about 16°–20°S. These estimated absolute velocities agree well with those of an ocean reanalysis, which lends credibility to the strictly databased analysis.

1. Introduction

The shallow overturning circulation refers to wind-driven flow in approximately the upper 1000 m of the ocean, which significantly contributes to meridional heat transport (Klinger and Marotzke 2000). In this circulation, water subducts into the thermocline at midlatitudes and flows equatorward. A part of this subducted water flows into the western boundary and turns poleward in the subtropical gyre. The remaining water flows equatorward, either via the western boundary or in the interior, feeding the equatorial undercurrent and upwelling in the eastern equatorial ocean. This overturning circulation, consisting of midlatitude subduction and equatorial upwelling, is called the subtropical cell. It has been described theoretically (e.g., Liu 1994) and numerically (e.g., McCreary and Lu 1994; Lu and McCreary 1995) and is observed in the Pacific and Atlantic Oceans (Johnson and McPhaden 1999; Zhang et al. 2003). The subtropical cells in the Pacific and Atlantic Oceans are roughly symmetric about the equator, being driven by wind systems that are approximately meridionally symmetric, with easterlies at low latitudes and westerlies in the midlatitudes.

In contrast, the shallow overturning circulation in the Indian Ocean is asymmetric about the equator, being driven by easterly winds in the tropical Southern Hemisphere and monsoonal winds to the north (Miyama et al. 2003; Schott et al. 2002, 2004, 2009). Owing to the absence of mean easterly winds along the equator there is no equatorial upwelling in the Indian Ocean as in the Pacific and Atlantic. Water subducts in the southern midlatitudes, flows equatorward, and partly upwells in the region of the Seychelles Chagos thermocline ridge, which is located along 10°S (Schott et al. 2002). The remaining water crosses the equator in the lower branch of
the cross-equatorial cell (CEC) and upwells off Somali and off Oman (Schott et al. 2002). Water is also injected into the thermocline in the northern Arabian Sea and marginal seas (Prasad et al. 2001; Prasad and Ikeda 2002a; Beal et al. 2000), which spreads to the interior and possibly feeds upwelling off Somali (Schott et al. 2002). There are also interbasin exchanges with the Pacific through Indonesian Throughflow (ITF; Sprintall et al. 2009) and with the Atlantic through the Agulhas Current (Beal et al. 2011).

Johnson and McPhaden (1999) and Zhang et al. (2003) have described the subtropical cells in the Pacific and Atlantic Oceans, respectively, from observations. They used temperature $T$ and salinity $S$ data obtained from conductivity–temperature–depth (CTD) and other hydrographic observations to estimate geostrophic velocity with an assumption of a level of no motion at a certain depth. An equivalent observational study for the Indian Ocean circulation has not been systematically undertaken, owing to the paucity of hydrographic observations. Reid (2003) described the basinwide circulation in the Indian Ocean using 2187 hydrocast stations, far less than were available to Johnson and McPhaden in the Pacific (15 693 stations) or Zhang et al. in the Atlantic (86 131 stations). Qu and Meyers (2005) estimated currents based on in situ data, but they restricted their analysis to the narrow region between Indonesia and Australia, where observations are relatively dense. Also, Qu and Meyers had to calculate density from temperature and a mean $T/S$ relationship because the number of salinity observations was an order smaller than for temperature (47 420 profiles for temperature vs 6473 for salinity). Schott and McCreary (2001) and Schott et al. (2009) provided schematics for the basin-scale circulation in the Indian Ocean, but these were based on model simulations, ocean reanalyses, and relatively few observations.

Since Reid’s (2003) and Qu and Meyers’s (2005) studies were published, there has been a dramatic increase in the availability of ocean observations. Argo floats have been deployed globally since 2004 (Roemmich et al. 2009), with an array of more than 3000 floats sustained in the World Ocean after 2007. Argo floats nominally measure temperature and salinity in the upper 2000 m with the vertical intervals of 10 m every 10 days. Since the initiation of the Argo program, the number of $T/S$ observations has significantly increased in the Indian Ocean (Fig. 1a).

In addition to $T$ and $S$, absolute velocity at a deep reference level can be estimated from Argo float observations (Lebedev et al. 2007; Park et al. 2005). The standard mission of Argo floats is to park at 1000-m depth for nominally 9 days, then descend to 2000-m depth and ascend to the surface while measuring temperature and salinity. After drifting near the surface to transmit data to satellites, floats sink to the parking depth again. The positions and times when floats descend and ascend are known from the satellite positioning system and the float clocks, respectively. The distance between ascending and descending positions is approximately equal to float displacement at the parking depth, from which velocity can be calculated. The resulting absolute velocity data increases in number in the Indian Ocean in tandem with the $T/S$ observations (Fig. 1b). When parking depth velocity data are plentiful, it is possible to calculate a mean absolute velocity field, which obviates the need to arbitrarily assume a level of no motion.

The purpose of this study is to estimate the mean Indian Ocean circulation using recently accumulated $T/S$ and absolute velocity observations, mainly from the Argo float program. Gray and Riser (2014) similarly estimated velocity but for the global domain, focusing on vertically integrated transport. Here, we focus on the Indian Ocean region, where implementation of the Argo array has had a relatively high impact on data availability to describe currents.

The rest of this paper is organized as follows. Sections 2 and 3 describe the data and methods to process them. Section 4 describes the wind fields, velocity, and volume transport across the Indian Ocean basin. We also
compare tracer patterns with velocity fields in section 4. We mainly focus on the south Indian Ocean, where the mean velocity field is better defined. Section 5 summarizes the main results.

2. Data and processing

a. Hydrographic observations

We obtained T/S from CTD and profile float data (PFL) from the World Ocean Database archive (WOD; Boyer et al. 2013) for the region 50°S–30°N and 30°–120°E. For PFL data, 96% of T and 99% of S data are contributed from Argo floats (Boyer et al. 2013). We applied three quality checks to these data. First, data were discarded if the WOD quality flag was other than 0, which is the “accepted value.” Second, data were discarded if they deviated from the mean by more than three standard deviations. For this check, we obtained the mean and standard deviation on a 5° × 5° grid from the World Ocean Atlas 13 (Locarnini et al. 2013; Zweng et al. 2013). Third, we manually eliminated obviously erroneous data. These three checks eliminated 16%, 0.5%, and 1.2% of the data, respectively. Density inversion checks were also done, but no profile failed this test. The total number of profiles available to us after quality control is 221,614, of which 92% are from PFL data and 8% from CTD. This total is two orders of magnitude larger than the number of hydrographic observations available to Reid (2003). In most of the interior of the Indian Ocean, more than 30 observations are available in each 1° × 1° box since 1961 (Figs. 2a–c). Data are mostly confined to the upper 2000 m (Fig. 2d), which is the standard depth range for Argo float observations.
We calculated potential density $\sigma_t$ using Fofonoff and Millard’s (1983) equation of state and interpolated temperature, salinity, and pressure onto $\sigma_t$ levels. Also, we calculated mixed layer depth (MLD) from potential density as the depth where density is higher than the 10-m value by 0.03 kg m$^{-3}$ following de Boyer Montégut et al. (2004).

b. Absolute velocity at the Argo parking depth

We obtained Argo float parking depth velocity data from YoMaHa’07 (Lebedev et al. 2007). Data for the global domain were used to make a gridded field. We applied three quality checks, following Katsumata and Yoshinari (2010). First, data were discarded if the time period between a float’s ascent and descent was shorter than 3 days. This test eliminated 0.7% of the data. Second, we discarded data if floats likely hit the ocean bottom. We judged likelihood of this happening if parking depth was deeper than 0.8 times the depth of the ocean bottom and if a float’s drift speed was smaller than 50% of its average. The bottom depth is obtained from the 5-minute gridded elevations/bathymetry for the world (ETOPO5). We also discarded data if parking depth was shallower than 400 m or deeper than the ocean bottom. In total, 4.1% of the data were eliminated by these criteria. Third, we discarded data if a float’s drift speed was larger than its average by more than five standard deviations. This process eliminated 0.07% of the data. The total number of parking depth data is 1170206 for the globe and 179663 for the Indian Ocean region (50°S–30°N, 30°–120°E).

Velocity data were adjusted to the common reference pressure (1000 db) by adding geostrophic shear calculated from a hydrographic dataset based on the thermal wind relationship. This adjustment was done only for the data outside of the equatorial region (0.5°S–0.5°N). We use the WOCE Global Hydrographic Climatology (Gouretski and Koltermann 2004) for this purpose, following Katsumata and Yoshinari (2010). This dataset provides the mean $T$ and $S$ for the global domain with spatial intervals of 0.5° in longitude and latitude. We use this dataset rather than our own hydrographic data, because we fit velocity observations to global harmonic functions [see section 3a(2)], which requires use of data in the global domain, whereas our $T/S$ analysis is restricted to the Indian Ocean. We confirmed that the WOCE Global Hydrographic Climatology and ours describe similar horizontal density gradients in the Indian Ocean domain. Errors for this depth adjustment are calculated from standard deviations of $T$ and $S$ obtained from the WOCE Global Hydrographic Climatology, which is used in the gridding procedure [section 3a(2)].

Note that parking depth velocity is an estimate obtained from float’s position and time. This estimate leads to at least three kinds of errors. First, Argo floats, especially those deployed in the early period, used Service Argos for positioning, which is less accurate than the more recently used global positioning system. Second, the position and time of the float are not measured at the exact time of its ascent or descent, but measured when it communicates with the satellite. After the float surfaces, it needs to wait typically for an hour for the arrival of the satellite (Ichikawa et al. 2001), during which the float has drifted under the influence of surface currents. This leads to a discrepancy between the position of float when it reaches the surface and the position when it communicates with the satellite. A similar discrepancy happens after the float communicates with the satellite and waits to descend. Third, it takes about 6 h for the Argo float to ascend from the parking depth to the surface (Ichikawa et al. 2001), during which the float is advected by the currents. Ichikawa et al. (2001) tracked four Argo floats and estimated the above three errors: the positioning error owing to the Service Argos was 0.1 cm s$^{-1}$, the surface drift error was 0–1.2 cm s$^{-1}$, and the drift error while ascending was 0.1–1.3 cm s$^{-1}$. It is difficult to estimate other errors, such as that resulting from the drift of the float’s clock. Hence, we assume a measurement error of 0.5 cm s$^{-1}$ for all parking depth velocity data.

The number of parking depth velocity observations in each 1° × 1° grid is shown in Fig. 2e. Parking depth velocity observations are sparser in the Northern Hemisphere compared to $T/S$ observations because some of the Argo floats deployed there lack information necessary for the calculation of parking depth velocity. Nevertheless, more than 30 observations are available in most 1° × 1° boxes, except near the east coast of the African continent and the Arabian Peninsula.

c. Other datasets

We use two different ocean reanalysis datasets for the following reasons. First, a reanalysis dataset is necessary to remove seasonal variability from velocity observations. We gridded velocity observations following the method of Davis (2005) and Katsumata and Yoshinari (2010) [details described in section 3a(2)]. The approach focuses on estimating the mean field, treating any temporal variability in input data as the sampling error. To reduce temporal variability, the seasonal cycle in velocity was identified and removed using an ocean reanalysis product. For this purpose, we used the reanalysis provided by Masuda et al. (2003, 2009), which is described below. Second, we needed an ocean reanalysis to compare with observational results. As Masuda et al.’s reanalysis was already used in processing velocity data, we used a different dataset for this purpose, which is
Ocean Reanalysis System, version 4 (ORAS4; Balmaseda et al. 2013).

Masuda et al.’s (2003, 2009) ocean reanalysis is based on version 3 of the Modular Ocean Model (MOM3; Pacanowski and Griffies 1999), which uses a four-dimensional variational method to assimilate Argo float and other T/S observations such as CTDs, expendable bathythermographs (XBTs), mooring buoy observations, satellite altimetry, and satellite sea surface temperature observations. In their analysis, surface boundary conditions and the initial condition were the control variables of assimilation. The horizontal grid spacing is 1° and the model has 46 vertical levels. We use monthly averages from 1990 to 2010.

ORAS4 uses a three-dimensional variational assimilation method to assimilate Argo float observations and other T/S observations such as CTDs, XBTs, mooring buoy data, and satellite altimetry observations into the Nucleus for European Modelling of the Ocean (NEMO) ocean general circulation model (OGCM; Madec 2008). ORAS4 does not assimilate velocity data. The model is forced by surface fluxes from the ECMWF atmospheric reanalysis and the control variables are temperature, salinity, and sea level. We use monthly averages of ORAS4 for the period from 2005 to 2015. The grid system for NEMO is a tripolar grid called ORCA1, and we use data interpolated onto a 1° × 1° grid. Potential density is calculated on a monthly basis, and zonal and meridional velocity is interpolated onto the same σθ levels as those used for observations. MLD is calculated using the same definition as that for the observational MLD (i.e., the depth where potential density is higher than the 10-m value by 0.03 kg m⁻³). Note that we calculate MLD from monthly averages for ORAS4, which can lead to 10–20-m underestimates (de Boyer Montégut et al. 2004; Toyoda et al. 2017).

We use surface wind stress data (specific output labeled “instantaneous eastward/northward turbulent surface stress”) from the ECMWF interim reanalysis (Dee et al. 2011). This output is on a 0.75° × 0.75° grid. We calculated a climatology for the period from 2005 to 2015 for consistency with in situ observations.

3. Methods

a. Mapping

1) IN SITU HYDROGRAPHIC OBSERVATIONS

We objectively mapped observed MLD and temperature, salinity, and pressure on isopycnals on a 1° × 1° grid for each calendar month using the method of Bretherton et al. (1976) assuming a Gaussian covariance function. We used decorrelation scales of 6° in longitude, 3° in latitude, and 45 days in time. The spatial decorrelation scales are chosen so that they are larger than the radius of deformation at all latitudes [the largest radius of deformation is the one near the equator for the lowest baroclinic mode, which is about 3° according to the parameters shown by Nagura and McPhaden (2010)]. A wider decorrelation scale in the zonal direction than in the meridional direction is chosen consistent with the anisotropy of tropical circulation away from the boundaries. The 1000 profiles with the highest weights were used in the gridding.

Objective analysis yields nondimensional analysis errors, which range from 0 to 1, depending on the density of observations. The error is large on shallow isopycnals, which is due to periodic outcropping and thus a reduction of analyzed values. We discarded estimates if the analysis error was larger than 0.3, which roughly limits our analysis to depths below the surface mixed layer. Note that this process eliminated virtually none of the interior estimates, as the error is smaller than 0.1 at almost all grid points in the pycnocline.

We define dimensional errors for pressure as standard deviations of pressure at each grid point multiplied by nondimensional analysis errors, following Bretherton et al. (1976). Standard deviation was calculated using the 300 hydrographic observations closest to the grid point. The resulting error was used to calculate uncertainties for velocity estimates.

2) PARKING DEPTH VELOCITY

Parking depth velocity was gridded exactly following the method described in Davis (2005) and Katsumata and Yoshinari (2010), which we briefly summarize here. This method calculates the mean velocity field, with any temporal variability in input data treated as the sampling error. To reduce the sampling error, seasonal variability was subtracted first. Empirical orthogonal functions were calculated using velocity at 1000-m depth obtained from Masuda et al.’s (2003, 2009) ocean reanalysis and fit to parking depth velocity observations for each calendar month. The resulting seasonal variability was subtracted from parking depth velocity observations. Deseasonalizing the velocity data using ORAS4 instead of Masuda et al’s reanalysis made very little difference in the results, indicating that the results are not sensitive to which reanalysis product is used.

Second, area averages were calculated at 1° grid intervals, using deseasonalized parking depth velocity data within 300 km around each grid point. Using the resulting area average, the sampling error for velocity observations was calculated as diffusivity, which was defined in terms of the difference in velocity from the
area average (Davis 1991). The effective degrees of freedom for input data were calculated from the temporal and spatial separation between each pair of observations, with a length scale of 100 km and a time scale of 10 days.

Third, the deseasonalized parking depth velocity data were locally fit to nondivergent, linear functions at 1° grid intervals. We again used observations within 300 km of the grid point. The uncertainty of input data was considered in this fitting, which was calculated from the depth adjustment error, the measurement error described in section 2b and the sampling error obtained in the area averaging. The fitting was done by minimizing a cost function described by Davis (2005) and Katsumata and Yoshinari (2010). The sampling error was recalculated in terms of diffusivity using the residual of the fitting.

Finally, the mean velocity obtained from local function fitting was fit to global functions. The global functions were obtained as the solution of the Poisson equation, \( \nabla^2 \phi = \exp(ik \cdot x) \), where \( k \) is wavenumber vector, \( x = (x, y) \) is zonal and meridional coordinates, and \( \phi \) is geostrophic pressure. The Poisson equation was solved with the boundary condition that \( \phi = \) constant along the coast of landmasses obtained from ETOP05. Here we considered functions corresponding to zonal and meridional wavenumbers from 1 to 48. In addition, a series of functions that have nonzero value along the coastline of each landmass was considered, in order to allow transport between two different landmasses. The uncertainty of input data was calculated from the depth adjustment error, the measurement error described in section 2b and the sampling error obtained from the local function fitting. The fitting was done by minimizing the cost function described by Davis (2005) and Katsumata and Yoshinari (2010). We discarded results if the error of the global function fit was larger than 0.08 m s\(^{-1}\).

Results compare well with absolute velocity at 1000-m depth from ORAS4 (Fig. 3). Both observations and the reanalysis show westward and northward velocity south of 20°S, except for eastward flow at about 40°S and in the region between 30° and 60°E related to the Antarctic Circumpolar Current. Both analyses show westward currents at about 10°S and weak currents in the Northern Hemisphere. Differences are generally minor or limited geographically, such as in the magnitude of eastward flow between 0° and 10°S, which is strong in the observations but weak in the reanalysis.

b. Velocity estimates

We calculated the pressure anomaly streamfunction (Zhang and Hogg 1992) with a reference level of 1000 m from gridded temperature, salinity, and pressure data. We then estimated geostrophic velocity shear except near the equator (5°S–5°N). We estimated absolute velocity as a function of depth from the combination of velocity at 1000-m depth plus vertically integrated...
geostrophic velocity shear. The corresponding streamline for absolute velocity is obtained from the pressure anomaly streamfunction and geostrophic pressure for parking depth velocity, the latter of which is obtained from our gridding procedure [8 in section 3a(2)]. Errors for geostrophic velocity shear were calculated from errors for pressure on isopycnals obtained from objective mapping. Errors for parking depth velocity were obtained from the global function fitting.

c. Mean fields

The above procedures give us monthly estimates for geostrophic velocity shear, isopycnal depth, temperature, and salinity from in situ hydrographic observations. For these fields, we calculated the mean simply by averaging monthly estimates. To check the validity of the arithmetic mean, we fitted monthly values to a five-parameter regression model for the mean and the one- and two-cycle-per-year harmonics using a least squares fit. The mean obtained from least squares fit was almost the same as the arithmetic mean. Note that in some regions isopycnals outcrop in winter. For those regions, we calculate the mean from all available estimates, with the results representing summertime conditions. To identify the outcropping regions, we hatch the areas where the annual mean depth of an isopycnal is shallower than wintertime MLD in the following figures.

4. Results

Here we describe basinwide patterns of wind forcing in section 4a. We describe velocity and transport in section 4b. We compare velocity fields with tracer patterns in section 4c.

a. Wind forcing

Wind stress curl at midlatitudes (20°–30°N and S) is negative in the Northern Hemisphere and positive in the Southern Hemisphere, which drives the equatorward interior branch of the subtropical circulation (Fig. 4a). The region with large positive wind curl (>1.2 × 10^{-7} N m^{-3} marked by dark red) in the Southern Hemisphere midlatitudes is larger in the Indian Ocean compared to that in the Pacific and Atlantic. Narrow bands of positive wind curl related to the intertropical convergence zone (ITCZ) are seen at 10°–20°N in the North Pacific and Atlantic. Wind curl tends to be negative from 10°N to 15°S except in the western North Pacific and the central South Pacific, which is related to the South Pacific convergence zone (SPCZ).

Ekman pumping velocity was calculated from wind stress as \( \nabla \times [\tau/(\rho_0 f)] \), where \( \tau \) denotes the wind stress vector, \( \rho_0 \) is the mean seawater density, and \( f \) is the Coriolis parameter. It tends to be negative at latitudes of 10°–30°N and S except for regions near the coasts (Fig. 4b). Ekman pumping is positive in some regions near the equator related to the ITCZ and SPCZ, such as in the western and eastern parts of the tropical Pacific, the eastern parts of the tropical Atlantic, the region near Sri Lanka, and between 0° and 10°S east of 40°E in the tropical south Indian Ocean. Positive Ekman pumping near the equator in the south Indian Ocean is larger in space and stronger in magnitude.
compared to the low-latitude positive Ekman pumping velocity in other basins.

b. Velocity and transport

In this section we describe velocity and transport. Section 4b(1) introduces four isopycnals on which we describe velocity. Sections 4b(2) and 4b(3) describe zonal and meridional currents, respectively. Section 4b(4) compares results from observations with those from ORAS4. Section 4b(5) presents transport estimates.

1) CIRCULATION ON ISOPYCNAL SURFACES

In this and the following subsections we illustrate currents on four isopycnal surfaces. The shallowest is 24σθ, which is located in the middle pycnocline and varies in depth between 60 and 150 m (Fig. 5a). This isopycnal outcrops at southern latitudes and the northern Arabian Sea, and is shallow in the Seychelles–Chagos thermocline ridge region between 10°S and the equator, 40°–80°E (Hermes and Reason 2008; Yokoi et al. 2008). The second shallowest isopycnal is 25.5σθ, which is in the lower pycnocline and varies in depth between 100 and 250 m deep (Fig. 5b). This isopycnal outcrops in the south and exhibits a thermocline ridge as well. Also this isopycnal corresponds to the density of the South Indian Subtropical Water (STW; Wijffels et al. 2002). The third isopycnal is 26.8σθ, which is found at about 400-m depth north of 10°S but deepens to 800 m in the south Indian Ocean (Fig. 5c). Two well-known water masses are located on this isopycnal surface, which are the Persian Gulf Water (PGW) in the Arabian Sea (Rochford 1964; Morrison 1997; Prasad et al. 2001) and Subantarctic Mode Water (SAMW) in the south Indian Ocean (McCartney 1982; McCarthy and Talley 1999; Karstensen and Quadfasel 2002). This isopycnal does not outcrop in the analysis domain but does at about 45°S where SAMW is generated (Karstensen and Quadfasel 2002). The deepest isopycnal we consider is 27.1σθ, which is located at about 600-m depth north of 15°S and deepens to the south up to 1000 m or more (Fig. 5d). The Antarctic Intermediate Water (AAIW; Fine 1993; Toole and Warren 1993; McCarthy and Talley 1999) and the Red Sea Water (Rochford 1964; Quadfasel and Schott 1982; Beal et al. 2000) are found on this isopycnal.

Fig. 5. Mean depth of isopycnals (colors), absolute velocity (vectors), and streamlines for absolute velocity (contours) obtained from in situ observations. (a) 24σθ, (b) 25.5σθ, (c) 26.8σθ, and (d) 27.1σθ. Contour intervals for streamlines are (a) 0.5, (b) 0.5, (c) 0.3, and (d) 0.3 × 10^4 m^2 s^-2. Hatching shows outcropping regions, where the annual mean depth of the isopycnal is shallower than the wintertime MLD. The MLD is defined from potential density as the depth where density is higher than the 10-m value by 0.03 kg m^-3 following de Boyer Montégut et al. (2004).
Streamlines (Fig. 5) give a rough idea about the circulation. The most notable feature is the anticyclonic circulation south of 10°S, which is consistent with the wind stress curl forcing (Fig. 4a; Stramma and Lutjeharms 1997) and the related westward South Equatorial Current (SEC). The center of the subtropical circulation retreats poleward with depth (Figs. 5b and 5c) as is seen in other basins (e.g., Reid 1981). This poleward retreat is consistent with the concept of the $\beta$ spiral (Pedlosky 1996, 201–204), which indicates that in the downwelling region of the Southern Hemisphere, the velocity vector turns anticlockwise with depth.

The mean currents are weak in the Northern Hemisphere on all the isopycnals. Surface winds north of the ITF isopycnals feed into a westward current of about 10°S between the African continent and Madagascar (Swallow et al. 1988; Schott et al. 1988, 2009). The northward NEMC feeds into a westward current at about 10°S between the African continent and Madagascar (Swallow et al. 1988; Schott et al. 1988), which is visible at about 10°S in the maps on all the four isopycnals (Figs. 6a–d) and velocity sections along 43°E (Figs. 7a–d). The southward SEMC feeds into a westward current south of Madagascar at about 26°S, which is also observed at all the four isopycnals.

Near Indonesia, a westward current related to the outflow of the ITF is clearly seen near 10°S in the upper two levels (Figs. 6a, 6b, 7i, and 7j). Westward outflow from the west coast of Australia is seen on the 26.8$\sigma_\theta$ and 27.1$\sigma_\theta$, which is fed by the Leeuwin Undercurrent (LUC; Smith et al. 1991; Domingues et al. 2007; Furue et al. 2017). Westward velocity shows the peak of magnitude southwest of Australia at about 35°S between 110° and 120°E on the lower two isopycnals (Figs. 6c and 6d), which is a manifestation of the Flinders Current (Bye 1972; Hufford et al. 1997).

Zonal velocity shows various eastward countercurrents. Currents on the upper two levels are cyclonic between the equator and 10°S, 40°–80°E in the vicinity of the thermocline ridge (Figs. 5a and 5b). In relation to this thermocline ridge, clear eastward currents are observed at about 5°S, which is referred to as the South Equatorial Countercurrent (SECC; Figs. 6a, 6b, 7a, 7b, 7e, and 7f; Schott et al. 2009). The eastward current between 25° and 30°S on the upper two levels is the South Indian Countercurrent (SICC; Menezes et al. 2014). Divakaran and Brassington (2011) showed that, at higher horizontal resolution, the SICC consists of a series of eastward currents interspersed by meridionally narrow westward currents. Our estimates aimed at broadscale structures and are missing these small-scale features. Eastward flow at about 15°S east of 105°E is the Eastern Gyral Current (EGC; Figs. 6a–c; Bray et al. 1997; Menezes et al. 2013). The EGC was not distinguished from the SICC in many past studies (e.g., Qu and Meyers 2005), but our study shows that they are separate currents in the pycnocline. Note that the two currents merge at the surface according to Menezes et al. (2013, 2014), who estimated geostrophic currents using Argo observations mapped on pressure surfaces.

An eastward current is observed at about 15°S east of 70°E on 27.1$\sigma_\theta$ (Figs. 6d and 7l). This flow can be interpreted as the deep extension of the EGC, which is trapped near the surface but extends to 1000-m depth or more (Fig. 8a). This deep eastward flow between 500 and
Wijffels et al. (2002) showed eastward spreading of high salinity water at about 900 m between 15°S and 20°S east of 90°E (their Fig. 4) and speculated the presence of an eastward flow. Our study shows that this mean eastward current actually exists. North of 10°S, currents tend to be eastward at the lower two levels between 50° and 90°E (Figs. 6c, 6d, 7g, 7h, 7k, and 7l). Reid (2003) described this feature, which is confirmed by our analysis.

3) MERIDIONAL VELOCITY

Meridional velocity is noisier and the signals are weaker than for zonal velocity, so it is difficult to distinguish detailed features. Here, we discuss overall patterns for flow.

Meridional velocity on the upper two isopycnals tends to be poleward north of about 15°S and equatorward south of 15°S (Figs. 9a, 9b, 10a, and 10b). The southward flow near the equator is likely a part of the cross-equatorial cell (Miyama et al. 2003; Horii et al. 2013; Wang and McPhaden 2017) and/or southward flow in the cyclonic gyre of the Seychelles–Chagos thermocline ridge. There is no interior path to the equator east of 60°E, which can be expected from strong Ekman upwelling between 0° and 10°S (Fig. 4b). Near-surface meridional convergence around 15°S feeds the westward
SEC and downwelling, which is evident in the depression of the shallow pycnocline in this latitude range (Figs. 5, 16, and 17).

On the deeper two isopycnals (Figs. 9c, 9d, 10c, and 10d), equatorward currents dominate south of 15°S and are largest in magnitude in the southeastern part of the gyre (25°–35°S, 90°–110°E). This feature is expected from ventilated thermocline theory (e.g., Luyten et al. 1983). In contrast to the upper two levels, poleward currents do not dominate north of 15°S.

4) COMPARISON WITH THE REANALYSIS

ORAS4 velocity compares well with that from in situ observations in terms of both spatial patterns and

**Fig. 7.** Absolute zonal velocity as a function of latitude along (a)–(d) 43°E, (e)–(h) 60°E, and (i)–(l) 95°E. The first, second, third, and fourth row from the top is for 24σθ, 25.5σθ, 26.8σθ, and 27.1σθ, respectively. Solid lines are for observational estimates, and dashed lines are results from ORAS4.
In particular, reanalysis velocities and observational estimates for zonal velocity agree well in latitude and magnitude. For example, there is good correspondence between the two westward currents near the western boundary (Figs. 7a–d), the northern and southern parts of SEC and the eastward SECC (Figs. 7e–h), and the westward current related to the ITF and the eastward SICC (Figs. 7i and 7j). Both the observational results and the reanalysis exhibit the eastward EGC, although the EGC in the reanalysis is weaker and less extensive compared to observations (Figs. 6a, 6b, 6e, and 6f). The reanalysis shows the deep extension of the eastward EGC along about 15°S (Figs. 6h and 8b) and eastward flow north of 10°S on the lower two isopycnals (Figs. 6g, 6h, 7g, 7h, 7k, and 7l), currents that are also seen in observational estimates. For meridional velocity, both the reanalysis and observational estimates show the convergence at 15°S on the upper two isopycnals and the dominance of equatorward flow south of 15°S on the lower two levels (Figs. 9e–h and 10).

This good agreement is likely due to the density of Argo float observations, on which our analysis is based and which ORAS4 assimilates. However, our analysis method and that of ORAS4 are very different. Our method is empirical, based only on observations and statistical gridding procedures, while ORAS4 uses a numerical model and sophisticated data assimilation procedures. The good agreement described above indicates that results are not sensitive to the method being used and lends credibility to the results presented here. In addition, the velocities from ORAS4 are the total velocities whereas our analysis is for the geostrophic component of velocity. The two agree well in general, which indicates that most of the flow is geostrophic in the interior of the ocean.

The largest differences between the observations and the reanalysis are found mainly in coastal regions. Our observational estimates are obtained with a spatial decorrelation scale of 6° in longitude and 3° in latitude, which is likely to heavily smooth out coastal boundary currents, such as the Leeuwin Current (Smith et al. 1991), the LUC, SEMC, NEMC, the East African Coastal Current, and the Agulhas Current (Schott et al. 2009). These currents are represented in ORAS4 along the west coast of Australia and the eastern coasts of Madagascar and the African continent. In addition, noticeable discrepancies are found in meridional velocity in outcropping regions. We base our analysis on geostrophy, and our estimates do not include Ekman currents, whereas absolute velocity obtained from ORAS4 does. Winds are easterly (or westerly), and wind-driven Ekman currents are southward (northward) north (south) of the center of the Mascarene high at about 35°S. As a consequence, compared to observational estimates, reanalysis meridional velocity is smaller between 20° and 35°S on 24σθ (Figs. 9a, 9e, and 10a) and larger south of 35°S on 25.5σθ (Figs. 9b, 9f, and 10b). Otherwise, in the interior ocean, the two analyses agree well.

5) VOLUME TRANSPORT

Figure 11 shows meridional transport accumulated from the eastern boundary between the 24.5σθ and 27.1σθ surfaces. Errors for transport were estimated from those for vertically integrated geostrophic velocity shear and parking depth velocity. Meridional transport is about −10 Sv (1 Sv = 10⁶ m³ s⁻¹) southward along 6°S and +18 Sv northward along 20°S. The observational estimate tends to be larger in magnitude at 6°S compared to transport obtained from the reanalysis, but the observational and reanalysis results generally agree well within the range of error. Discrepancies near the western end of the domain at 6°S are due to the western boundary current, which the observational analysis misses. At 20°S, the observational estimates agree well with the reanalysis results east of 70°E, but are smaller than the reanalysis by about 3 Sv between 50° and 65°E. Note that the two reanalysis results differ, one of which is obtained from reanalysis’s velocity output, and the other from velocity shear calculated from reanalysis temperature and salinity added to reanalysis absolute...
velocity at 1000 m. The discrepancy is noticeable in particular along 6°S. It is not clear why this discrepancy occurs, but it may be due to computational errors, caused for example by assumptions in the derivation of the pressure anomaly streamfunction (Zhang and Hogg 1992) and/or contributions from any ageostrophic currents.

Vertically integrated geostrophic transport per unit width is shown in Figs. 12a–f. We compared results between transports estimated from wind, in situ hydrographic observations, and absolute velocity from the reanalysis. The estimate from wind was obtained from ECMWF wind stress, subtracting Ekman transport from Sverdrup transport in a similar manner to Landsteiner et al. (1990) and Gray and Riser (2014). We applied Godfrey’s (1989) island rule to the Australian and Madagascar landmasses and added the resulting transport to the estimates. Transport from observational velocity estimates was calculated for the range of 24.5–27.1σθ. Results were essentially the same if we included deeper levels, but errors for transport became larger than the transport itself at deeper levels, and thus we limit our analysis to isopycnals shallower than 27.1σθ. In addition to horizontal transport, we also compared vertical velocity at the base of the Ekman layer. Vertical velocity was computed from the ECMWF wind stress derived as Ekman pumping velocity (Fig. 13a). For the in situ observational estimates and reanalysis, vertical velocity at the base of the Ekman layer was calculated from meridional transport using the relationship, 

\[ w = \frac{\beta}{f} \frac{\partial P}{\partial y} dz, \]

where \( \beta \) is the meridional gradient of \( f \) (Figs. 13b and 13c). This relationship is derived from the Sverdrup relation, \( \beta v = f_0 \omega \), ignoring vertical velocity at the bottom depth of the vertical integration.

Fig. 9. As in Fig. 6, but color shadings are for meridional velocity.
Zonal transport (Figs. 12a,c,e) tends to be westward south of 8°S and eastward north of it, which roughly agrees in the three estimates. The estimate obtained from wind shows a sharp peak of westward and eastward flow at about 10°S, which is due to the island rule calculation and would be smoothed out by diffusion and instabilities if the full dynamics of the flow system were considered. The observational estimates and the reanalysis do not show such sharp peaks.

Meridional transport (Figs. 12b,d,f) is poleward north of 15°S and equatorward south of it in all the three estimates. The magnitude of transport compares well, although results from in situ hydrographic observations are patchy near 5°S. Vertical velocity at the base of the Ekman layer (Fig. 13) tends to be upward north of 15°S and downward south of it, which illustrates the circulation of the subtropical cell consisting of midlatitude subduction and low-latitude upwelling. The pattern and magnitude show a rough agreement between the three

**Fig. 10.** Absolute meridional velocity averaged over 60° to 100°E on (a) 24°S, (b) 25.5°S, (c) 26.8°S, and (d) 27.1°S for observational estimates (solid line) and ORAS4 (dotted line).

**Fig. 11.** Meridional mass transport zonally accumulated from the eastern boundary to various longitudes along (a) 6°S and (b) 20°S between 24.5°S and 27.1°S. Solid line is for estimates from in situ observations. Dashed line is for the equivalent estimate obtained from ORAS4 (i.e., geostrophic velocity shear calculated from reanalysis temperature and salinity and added to reanalysis absolute velocity at 1000-m depth). Dotted line is for results obtained from velocity output of ORAS4. Vertical line denotes typical error for observational estimates of transport.
estimates, although the estimates from the in situ hydrographic observations and the reanalysis are patchy.

c. Tracers on isopycnal surfaces

In this section we compare tracer fields with velocity estimates and discuss their consistency. Sections 4c(1) and 4c(2) describe planetary potential vorticity (PV) and salinity, respectively. We present vertical sections in section 4c(3) and describe the circulation patterns on these sections.

1) PLANETARY PV

PV acts as a tracer, which is conserved along particle trajectories, if the fluid is inviscid and adiabatic (Pedlosky 1987, 38–42). However, it is an active tracer, the structure of which partly determines the circulation. For example, a band of high PV collocated with the ITCZ exists in the North Pacific and Atlantic Oceans, which acts as a barrier to inhibit midlatitude water from flowing into the equatorial region directly via the interior (Lu and McCreary 1995; Johnson and McPhaden 1999). Here we present the absolute value of planetary PV, calculated as $|f(\rho_0\psi, z)|$ with $\rho_0 = 1025$ kg m$^{-3}$. We calculated $\partial_\psi z$ as $\Delta z/\Delta \rho$, where $\Delta z$ and $\Delta \rho$ are the vertical distance and difference in density, respectively, between two isopycnals. The resulting planetary PV is defined at midpoints of the two isopycnals. For this reason, we use slightly different density levels for PV from those for velocity. Note that planetary PV is the dominant component of PV in the ocean interior away from the equator. We do not discuss results near the equator in this section, where other components dominate total PV.

At 23.8$\sigma_\theta$ (Fig. 14a), planetary PV between 5$^\circ$ and 15$^\circ$S is high in the thermocline ridge region (45$^\circ$–90$^\circ$E). PV related to the ITF is lower, but increases to the west. This is consistent with Ekman upwelling and the resulting squeezing of isopycnals. PV at 25.2$\sigma_\theta$ (Fig. 14b) is relatively low in 15$^\circ$–25$^\circ$S, 70$^\circ$–90$^\circ$E compared to that at other longitudes, while PV of the ITF and the LUC is relatively high. At 26.7$\sigma_\theta$ (Fig. 14c), PV is low south of 20$^\circ$S and 50$^\circ$–100$^\circ$E, indicative of SAMW, which flows to the north from the outcropping region being advected by the subtropical circulation. PV from the eastern boundary, possibly related to the lower part of the ITF and the outflow from the LUC, is higher and extends to the west. This indicates that high PV coming from the eastern boundary gives rise to a PV barrier between 15$^\circ$ and 20$^\circ$S in the Indian Ocean. Consistent with the dynamics of an inviscid, adiabatic fluid, streamlines do not cross the high PV barrier and bend toward the west at 20$^\circ$S. At 27.0$\sigma_\theta$ (Fig. 14d), PV is high in the southeastern region related to AAIW (McCarthy and Talley 1999).
2) SALINITY

Salinity is usually considered as a passive tracer and has been used by many studies to mark distinct water masses. Here we show maps in both the Northern and Southern Hemispheres for a basin-scale perspective. Four well-known saline water masses can be identified: Arabian Sea High Salinity Water (ASHSW) in the northern Arabian Sea (Figs. 15a and 15b; Morrison 1997; Kumar and Prasad 1999; Prasad and Ikeda 2002b), STW south of 15°S (Fig. 15b), PGW in the northern Arabian Sea (Fig. 15c), and RSW at the exit of the Red Sea (Fig. 15d). Also, two low-salinity water masses are found: the Indonesian Water that originates from the Indonesian Seas (Figs. 15a and 15b; Gordon 2005) and AAIW in the south Indian Ocean (Fig. 15d). Salinity is low on 24°S in the Bay of Bengal, which is likely related to the massive supply of freshwater by rain and river runoff (Rao and Sivakumar 2003).

The fresh Indonesian Water spreads to the west between 10° and 20°S, which is consistent with the westward SEC. A high salinity tongue on 24°S and 25.5°S along the equator (light blue colors between 5°S and 5°N in Figs. 15a and 15b) is likely an extension of ASHSW in the Arabian Sea, and its eastward spreading can be accounted for by eastward currents such as the SECC and/or the seasonal equatorial undercurrents. The eastward spreading of the high salinity tongue along the equator was also reported by Prasad and Ikeda (2002a) and Masson et al. (2002). The pattern of saline STW on 25.5°S suggests that this water mass is advected by the subtropical gyre, and the boundary of STW and the Indonesian freshwater coincides with that of the subtropical gyre and the extension of the ITF.

PGW and RSW in diluted form spread all the way to the equator (Figs. 15c and 15d; Rochford 1964; You and Tomczak 1993). The spreading of these water masses is not supported by the mean circulation, which suggests contributions from variable currents and/or eddies. The northward spreading of AAIW is consistent with the mean northward currents, but eddy transport also contributes (Katsumata 2016). Because of the different water masses at different depths, salinity in the Southern Hemisphere decreases to the north on the upper two levels, but increases to the north along the lower two levels.

3) VERTICAL SECTIONS

Here we show vertical sections in the upper 1200 m, which includes the intermediate water and the ventilated thermocline. Three water masses are identified (Figs. 16a and 16b): the saline STW (14°–35°S centered at 25–26°S), SAMW characterized by low PV (15°–40°S centered at 26.5–27°S), and AAIW characterized by high PV and low salinity (15°–40°S centered at about 27–27.2°S). Note that high PV related to AAIW is apparently shallower than the low salinity associated with this water mass, in particular near the southern end of the domain. AAIW is generated in the Drake Passage as low PV water (Sallée et al. 2010), but it loses this signature when it reaches the southern Indian Ocean (McCarthy and Talley 1999). Isopleths of PV at 600–1200-m depths are not parallel to isopycnals south of about 34°S, which possibly indicates diabatic modification of AAIW properties while being advected to the north.

Zonal velocity is mostly westward in this section, which shows the SEC and the ITF extension (Fig. 16c). The westward current retreats poleward between 15° and 25°S as the depth increases, which reflects the poleward retreat of the anticyclonic circulation in the subtropical gyre. Currents in the upper 200 m are eastward between 20° and 30°S in the SICC and north of 6°S in the SECC. Currents are also eastward north of 8°S between 200- and 1200-m depths. Meridional velocity in the upper 1000 m tends to be equatorward south of about 34°S, which possibly indicates diabatic modification of AAIW properties while being advected to the north.
with the depressed pycnocline in the upper few hundred meters between 16° and 20°S. Vertical sections along other longitudes show similar patterns of zonal and meridional currents, except that the SECC is not found east of 90°E (figure not shown).

Finally, we show sections along 60°E (Fig. 17), which includes the region with major upwelling in the southern Indian Ocean in the region of the Seychelles–Chagos thermocline ridge. We omit currents here, as their patterns are similar to those along 75°E. IsopyCNals and isothermals near the surface are shallowest at about 6°S, which defines the thermocline ridge (Fig. 17a). The shoaling can be seen down to 360-m depth, 10.4°C, and 26.8σθ. A high salinity tongue between 15° and 35°S marks STW, whereas a high salinity tongue north of 6°S is likely an extension of ASHSW (Fig. 17b). Lower salinity is found near the surface within the depth of the winter-time mixed layer and below 140 m from the equator to 18°S. At about 100-m depth, salinity is lowest at 8°S between the two saline water masses. This local minimum in salinity may be related to the Indonesian Water advected by the westward SEC and/or freshwater that upwells.

5. Summary

The mean shallow overturning circulation in the Indian Ocean, which connects the midlatitude subduction zone with upwelling zones north and south of the equator, was poorly described in previous studies owing to the paucity of in situ hydrographic observations. The recent success of the Argo program has dramatically increased the number of hydrographic observations by two orders compared to just 15 years ago, when Reid (2003) described the basinwide circulation. In addition, absolute velocity at the Argo float parking depth obviates the need to arbitrarily assume a level of no motion. Using these data, we are able to describe the Indian Ocean circulation in greater detail and with more reliability than in earlier work.

We obtained hydrographic data and parking depth velocity data from the WOD (Boyer et al. 2013) and YoMaHa’07 (Lebedev et al. 2007), respectively. We objectively mapped temperature, salinity, and depth on isopycnals and used Davis’s (2005) and Katsumata and Yoshinari’s (2010) method to grid parking depth velocity. Geostrophic shear was calculated from the gridded hydrographic data, and gridded parking depth velocity was added to vertically integrated geostrophic velocity shear. The resulting velocity field shows known features of the circulation, such as the anticlockwise subtropical circulation in the south Indian Ocean, the westward SEC and ITF, the cyclonic circulation related to the Seychelles–Chagos thermocline ridge, and the eastward SICC. Gridded tracer fields (salinity and PV)
show known water masses, such as high salinity water in the Arabian Sea (ASHSW, PGW, and RSW), mid-latitude waters in the south Indian Ocean (STMW, SAMW, and AAIW), and the low-salinity Indonesian Water mass.

In addition, our results reveal circulation features that have not been described in previous studies. Specifically, we find that

- The subtropical circulation in the south Indian Ocean retreats poleward as occurs in the Atlantic and Pacific. This retreat is qualitatively consistent with midlatitude downwelling and the $\beta$ spiral relationship.
- PV in the south Indian Ocean tends to be high at $5^\circ$–$15^\circ$S near the surface ($23.8 \sigma_\theta$) and in the lower pycnocline ($26.7 \sigma_\theta$). High PV near the surface is due to Ekman upwelling in the Seychelles–Chagos thermocline ridge region, which thins the upper layer. In the lower pycnocline, high PV intrudes from the west coast of Australia and is located north of low PV related to the mode water (SAMW), which leads to a structure similar to a PV barrier in the North Pacific and Atlantic Oceans. PV is largest near the western coast of Australia, which suggests that high PV is generated in the eastern boundary region and injected westward into the interior.
- There is no interior pathway in the Indian Ocean that connects the Southern Hemisphere interior to the equatorial region, which is consistent with a band of high PV between $5^\circ$ and $15^\circ$S. Water parcels must transit through the western boundary to reach the equator. This is in contrast to the South Pacific Ocean, where the absence of an atmospheric convergence zone in the eastern basin allows the interior pathway; it is similar, however, to the North Pacific and Atlantic Oceans, where a PV barrier blocks interior pathways.
- The westward SEC splits into two at about $65^\circ$E near Mascarene Plateau in the lower part of the pycnocline ($25.5$–$27.1 \sigma_\theta$). This split has not been documented in observations before to our knowledge.
- An eastward flow is observed on the $27.1 \sigma_\theta$ surface along $15^\circ$S from the eastern boundary (120$^\circ$E) to the midbasin (about 70$^\circ$E), which can be interpreted as a

![FIG. 15. As in Fig. 5, but colors are for the mean salinity obtained from in situ observations.](image-url)
deep extension of the EGC. The existence of this flow has been deduced from property distributions (e.g., Wijffels et al. 2002), but its velocity has not been described previously to our knowledge.

- The EGC and the SICC are separate currents in the pycnocline. Previous studies (e.g., Qu and Meyers 2005) did not distinguish these two. The EGC is centered around 15°S east of 105°E on the 24σθ and 25.5σθ surfaces, whereas the SICC is located around 25°–30°S between Madagascar and Australia. They merge at the surface, however, according to Menezes et al. (2013, 2014). Also note that the SICC itself consists of a series of narrow eastward currents, as is revealed by estimates with higher horizontal resolution (Divakaran and Brassington 2011). Our basin-wide analysis smooths these features out.

Our analysis based on observations compares well with the ORAS4 ocean reanalysis, which assimilates Argo float observations. The good agreement indicates that our empirical statistical approach and ORAS4’s numerical approach give essentially the same results, probably owing to very dense Argo float observations that we use and that are assimilated into the reanalysis. This agreement lends credibility to our results and represents a cross validation of the reanalysis.

We also estimated geostrophic mass transport between 24.5σθ and 27.1σθ and vertical velocity at the base of the Ekman layer. The spatial patterns of estimated transport per unit width and vertical velocity compare well to estimates obtained from surface wind and from the reanalysis. Zonally accumulated meridional mass transport is 10 Sv southward at 6°S and 18 Sv northward at 20°S, respectively, which results in meridional convergence of currents and thermocline depression at about 16°–20°S.

Our current study has focused on the mean field and primarily the circulation in the Southern Hemisphere of the Indian Ocean. The circulation in the north Indian Ocean undergoes significant seasonal variations with reversing ocean currents due to strong monsoon wind forcing (Schott et al. 2009). Hence, a different strategy is needed for detailed examination of the circulation in the northern Indian Ocean. We have undertaken a study of this seasonal variability and will report on that analysis at a later date.
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