1. Introduction

The Indian Ocean has warmed rapidly since 2000, accounting for one-half of the global ocean warming above 700 m (Desbruyères et al. 2017; Zhang et al. 2018). This warming has impacted Indian Ocean rim countries by raising sea levels and changing winds, rainfall, and storm tracks (Han et al. 2014; Beal et al. 2020). Globally, Indian Ocean warming has potential impacts on the Atlantic Ocean meridional overturning circulation and on North Atlantic sea surface temperatures through rainfall and surface wind teleconnections (Hu and Fedorov 2019, 2020). The Estimating the Circulation and Climate of the Ocean (ECCO) reanalysis suggests that changes in oceanic fluxes more than air–sea fluxes have driven the rapid Indian Ocean warming over the past 20 years (Zhang et al. 2018). Models and observations have linked the warming to an increase in the volume and temperature transport of the Indonesian Throughflow (ITF) (Lee et al. 2015; Hu and Sprintall 2017). However, the role of circulation and heat transport within the Indian Ocean has yet to be investigated.

A better understanding of the relative roles of the ITF, the Agulhas Current, and the basin interior on the rapid warming will improve future climate projections. Although the ITF has strengthened over 2002–10 (Hu and Sprintall 2017), it is expected to weaken in the long term due to climate change (Ma et al. 2020; Stellema et al. 2019; Sen Gupta et al. 2016; Feng et al. 2017). If the Indian Ocean heat budget is controlled primarily by the ITF, we might expect the rapid warming to abate as the ITF weakens. However, other long-term changes may affect the heat budget. The interior gyre circulation strengthened between 1987 and 2002 (Palmer et al. 2004; McDonagh et al. 2005; McMonigal et al. 2018) while a proxy based on observations shows no trend in Agulhas Current volume transport over 1993–2015. Instead, the current is broadening, which could reduce southward heat transport even without a reduction in volume transport (Beal and Elipot 2016). A broadening current is related to more meandering (Beal and Elipot 2016), and the transport-weighted temperature of the Agulhas Current is cooler during meanders (McMonigal et al. 2020), even while the volume transport is maintained (Leber and Beal 2014). Thus, a broader current may transport less heat. In the long term, models project a future decrease in southward heat transport due to climate change, linked to a projected decrease in the Agulhas Current volume transport (Ma et al. 2020).

The deep and shallow overturning circulations could also play a role in changes to the Indian Ocean heat budget. The shallow overturning circulation has increased in strength by about 1 Sv (1 Sv ≡ 10^6 m^3 s^-1) since 1957 (Meng et al. 2020). The structure and strength of the deep overturning circulation varies widely across previous estimates, even those that are based on the same data (Table 1), and temporal variability in the deep overturning circulation is largely unknown. To assess how these changes combine to alter the Indian
Ocean heat budget, we need robust estimates of the southern Indian Ocean heat transport, Indian Ocean heat convergence, and the dominant drivers of variability to each.

Table 1. Previously estimated Indian Ocean heat fluxes from inverse models. Positive heat transport values indicate northward heat transport (i.e., into the Indian Ocean). Positive heat convergence values indicate a warming tendency over the Indian Ocean.

<table>
<thead>
<tr>
<th>Authors</th>
<th>Method (Indian Ocean data)</th>
<th>Deep overturning (Sv)</th>
<th>Agulhas (Sv)</th>
<th>ITF (Sv)</th>
<th>Heat transport across the southern boundary (PW)</th>
<th>Heat convergence (PW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Macdonald (1998)</td>
<td>Global inverse (1987)</td>
<td>17 ± 5</td>
<td>−93</td>
<td>10 ± 10</td>
<td>−1.30</td>
<td>−0.1 ± 0.2</td>
</tr>
<tr>
<td>Ganachaud et al. (2000)</td>
<td>Global inverse (1987)</td>
<td>10.6 ± 4</td>
<td>−74 ± 7</td>
<td>15 ± 5</td>
<td>−1.5 ± 0.2</td>
<td>−0.1 ± 0.2</td>
</tr>
<tr>
<td>Lumpkin and Speer (2007)</td>
<td>Global inverse (1987)</td>
<td>12.4 ± 2.6</td>
<td>−67 ± 5</td>
<td>13.2 ± 1.8</td>
<td>−1.55 ± 0.12</td>
<td>−0.1 ± 0.2</td>
</tr>
<tr>
<td>Robbins and Toole (1997)</td>
<td>Regional inverse w/silica  constraint (1987)</td>
<td>12 ± 3</td>
<td>−87</td>
<td>5.3 ± 3.5</td>
<td>−0.42</td>
<td></td>
</tr>
<tr>
<td>Sloyan and Rintoul (2001)</td>
<td>Regional inverse w/air-sea fluxes (1987)</td>
<td>23 ± 2</td>
<td>−48.3</td>
<td>10 ± 3</td>
<td>−0.87 ± 0.06</td>
<td></td>
</tr>
<tr>
<td>Bryden and Beal (2001)</td>
<td>Barotropic adjustment (1987)</td>
<td>10.1</td>
<td>−66.5</td>
<td>12.3</td>
<td>−0.66</td>
<td></td>
</tr>
<tr>
<td>Talley (2003)</td>
<td>Regional inverse (1987)</td>
<td>13.8</td>
<td>−48.5</td>
<td>8</td>
<td>−0.99</td>
<td>−0.59</td>
</tr>
<tr>
<td>Talley (2008)</td>
<td>Regional inverse (1987)</td>
<td>16.4 ± 6.1</td>
<td>−73</td>
<td>10</td>
<td>−1.33</td>
<td>−0.83</td>
</tr>
<tr>
<td>McDonagh et al. (2008)</td>
<td>Regional inverse with LADCP (2002)</td>
<td>10.3</td>
<td>−69.7 ± 4.3</td>
<td>12</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hernández-Guerra and Talley (2016)</td>
<td>Indo-Pacific inverse (2002)</td>
<td>12 ± 9</td>
<td>−75</td>
<td>11 ± 11</td>
<td>−1.1 ± 0.2</td>
<td>−0.47</td>
</tr>
<tr>
<td>Hernández-Guerra and Talley (2016)</td>
<td>Indo-Pacific inverse (2009)</td>
<td>8 ± 7</td>
<td>−92</td>
<td>11.9 ± 8.3</td>
<td>−1.5 ± 0.2</td>
<td>−0.82</td>
</tr>
<tr>
<td>This work</td>
<td>Argo mapping with moorings (2010–19)</td>
<td>9.6</td>
<td>−76</td>
<td>15</td>
<td>−1.09 ± 0.11</td>
<td>−0.01 ± 0.15</td>
</tr>
</tbody>
</table>

To address the gaps in knowledge of the variability in southward heat export from the Indian Ocean, we estimate the decadal mean heat transport over 2010–19, as well as the scale of subseasonal to seasonal variability in heat transport for the first time. These estimates are made using unique measurements of the Agulhas Current temperature and velocity from moorings. We then add these estimates of the southern boundary heat transport to estimates of the heat transport into the basin via the ITF to estimate the Indian Ocean heat convergence. We find a double-celled overturning, similar to other recent studies (McDonagh et al. 2008; Hernández-Guerra and...
Table 2. Data sources used in heat transport estimates.

<table>
<thead>
<tr>
<th>Temperature</th>
<th>Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>2010–19 mean</strong></td>
<td><strong>2016–18 time series</strong></td>
</tr>
<tr>
<td>Agulhas</td>
<td>ASCA</td>
</tr>
<tr>
<td>Upper interior Argo and satellite</td>
<td>ACT; ASCA</td>
</tr>
<tr>
<td>Ekman</td>
<td>Argo and satellite</td>
</tr>
<tr>
<td>Leeuwin</td>
<td>ERA5</td>
</tr>
<tr>
<td>ITF</td>
<td>Feng et al. (2003)</td>
</tr>
<tr>
<td></td>
<td>Sprintall et al. (2009); Gordon et al. (2019)</td>
</tr>
</tbody>
</table>

Talley 2016). We find that heat transport out of the Indian Ocean over 2010–19 is weaker than previous estimates from the 2000s and that this change has contributed about one-third of the recent rapid warming of the Indian Ocean.

2. Materials and methods

a. Heat transport calculation

Ocean heat transport is defined as

$$Q = \rho C_p \int \int \int \theta \, dx \, dz,$$  \hspace{1cm} (1)

where \(\rho\) is density, \(C_p\) is the specific heat capacity of seawater, \(v\) is meridional velocity, and \(\theta\) is potential temperature. The product \(\rho C_p\) is taken as the constant 4.093 J K\(^{-1}\) cm\(^{-3}\) (Hall and Bryden 1982). Heat transport can be summarized as the volume transport of each flow,

$$V = \int \int \int v \, dx \, dz,$$  \hspace{1cm} (2)

multiplied by the transport-weighted temperature (TWT) of that flow,

$$\text{TWT} = \frac{\int \int \int \theta \, dx \, dz}{\int \int \int v \, dx \, dz},$$  \hspace{1cm} (3)

multiplied by \(\rho C_p\).

The integration of Eq. (1) is typically taken over full depth and from coast to coast, because heat transport is only independent of a reference temperature when mass is conserved. We follow previous authors and use the term “temperature transport” to refer to \(Q\) when mass is not conserved. These temperature transports are relative to 0°C.

Historically, decadal repeat hydrographic sections have been combined with Ekman transport calculated from wind products to calculate the heat transport of the southern Indian Ocean near 32°S [references in Table 1; hydrography taken as parts of World Ocean Experiment (WOCE) and U.S. Global Ocean Ship Based Hydrographic Investigations Program (U.S. GO-SHIP)]. These hydrographic sections occurred in 1987, 2002, and 2009, and an upcoming section is planned for 2022. Then, a convergence or divergence of heat over the Indian Ocean can be calculated by adding the westward ITF temperature transport into the basin. However, we cannot evaluate how much short-term variability may be aliased in these decadal repeats. Short-term variability is likely large, for instance seasonal variability in AMOC strength at 26.5°N has an amplitude that is 2.2 times the multidecadal trend (Kanzow et al. 2010; Caesar et al. 2018). Moreover, previous studies based upon hydrographic data rely on inverse models to solve for the unknown geostrophic reference velocity while conserving mass, salt, and sometimes silica, and meeting a number of other oceanographic and/or air–sea flux constraints. The specific details of an inverse model matter, and adding constraints or changing the initial reference velocity, can substantially change the inferred circulation and heat transport (Hernández-Guerra and Talley 2016; Robbins and Toole 1997; Toole and Warren 1993). While repeat hydrography and inverse models are powerful tools to deduce full-depth large-scale circulation and property fluxes, their sensitivity to constraints and their sparse temporal resolution are drawbacks when it comes to determining the variability and possible trend in ocean heat transport.

We take an alternative approach by taking advantage of the large number of hydrographic profiles collected by Argo floats over the period 2010–19. In combination with data from a moored array across the Agulhas Current and satellite sea surface height and wind data, these measurements give a picture of the seasonal variability of heat transport, as well as decadal time scale changes since previous estimates.

b. Decadal mean heat transport

To estimate a decadal mean meridional heat transport near 36°S, we divide the Indian Ocean into subregions according to how each is observed:

$$Q_{36S} = Q_{\text{ASCA}} + Q_{>2000m} + Q_{>2000m} + Q_{\text{Eddy}}$$

+ $$Q_{\text{Leeuwin}} + Q_{\text{Eddy}},$$  \hspace{1cm} (4)

where \(Q_{36S}\) is the total meridional heat transport, composed of the Agulhas Current component \(Q_{\text{ASCA}}\), the upper-2000-m interior geostrophic component \(Q_{>2000m}\), the below-2000-m interior component \(Q_{>2000m}\), the Eddy component \(Q_{\text{Eddy}}\), the Leeuwin Current component \(Q_{\text{Leeuwin}}\), and the eddy heat transport.
transport $Q_{\text{eddy}}$. Data sources are described below and are summarized in Table 2.

Our estimate is nominally located at 35.75°S, about 4° of latitude southward of the previous estimates (Fig. 1). We expect the heat transport at 32° and 35.75°S to be similar, as integrated time-mean net air–sea fluxes between the latitudes from an ensemble mean of six data products implies that the heat transport at 35.75°S is smaller by 0.03 PW, much smaller than our estimated errors [see section 2d; air–sea flux data products are described by Beal et al. (2019)]. In addition, time series of heat transport at these two latitudes calculated from the global ocean state estimate ECCO4v4 (Forget et al. 2015; Fukumori et al. 2020) shows a decrease in heat transport of 0.08 PW between 32° and 35.75°S and a Pearson’s correlation coefficient of 0.9 (Fig. 2).

1) AGULHAS CURRENT

The temperature transport of the Agulhas Current $Q_{\text{ASCA}}$ was measured with the Agulhas System Climate Array (ASCA). This mooring array was a collaboration between the United States, South Africa, and the Netherlands, and consisted of seven full-depth moorings and five current and pressure recording echosounders deployed from May 2016 to April 2018 (McMonigal et al. 2020; Gunn et al. 2020). We integrate over the full 300 km of the mooring array and assume that the ASCA volume transport is representative of two years. We combine this with the 2010–13 Agulhas Current Time series (ACT) volume transport estimate (Beal et al. 2015), which we assume is representative of three years. Taking a weighted mean yields a decadal mean Agulhas Current volume transport estimate of $-76.2$ Sv. Although this is based on data from only five years of the decade, it agrees to within error with an extension of the satellite proxy produced by Beal and Elipot (2016). The satellite proxy shows a 2010–19 mean volume transport of $-84$ Sv, from the coast to 220 km offshore. The transport from 220 km offshore to 300 km offshore is 4.7 Sv (weighted mean of ACT and ASCA). Thus, the satellite proxy decadal mean Agulhas Current volume transport...
transport estimate from the coast to 300 km offshore is −79.3 Sv. Considering the 4-Sv error on the mean (McMonigal et al. 2020), these two estimates are the same. Time-varying temperature was not measured during ACT, and so we assume the 2016–18 temperature from ASCA is representative of the full decade. Therefore, we use 5 years of velocity and 2 years of temperature data from the Agulhas Current and assume that this is representative of the mean over 2010–19.

2) UPPER-2000-M GEOSTROPHIC INTERIOR

The geostrophic temperature transport of the upper-2000-m interior \( Q_{0–2000m} \) was calculated by mapping Argo profile density and temperature and 1000-m drift velocities over the period 2010–19. The regression between higher-temporal-resolution satellite sea surface height and lower-temporal-resolution Argo profile measurements is used to suppress aliased mesoscale (less than 90 day) variability in the mapping method (Willis and Fu 2008). This method was used in the North Atlantic to determine the time-mean and time-varying volume and heat transports associated with the Atlantic meridional overturning circulation strength (Willis 2010; Hobbs and Willis 2012), and in the southern Indian Ocean to estimate the seasonal cycle of the gyre (McMonigal et al. 2018).

3) EKMAN TRANSPORT

The Ekman volume transport was calculated using ERA5 monthly winds over 2010–19 and assumed to advect water with a temperature corresponding to the near surface monthly mapped Argo surface temperature, averaged across the interior of the basin. This method follows previous studies that assumed that the Ekman volume transport advects waters at the sea surface temperature (Johns et al. 2011). The monthly Ekman volume transport times the monthly near surface temperature yields monthly Ekman temperature transport. The Ekman volume transport is relatively small, 4.2 Sv in the mean, and thus the exact depth (and therefore temperature) of the Ekman layer is not a significant source of error [section 2d(4)]. The monthly Ekman volume transport ranges from −0.1 to 8.7 Sv.

4) LEEUWIN CURRENT

The Leeuwin Current is estimated to transport −3.4 Sv at a TWT of 20°C, as determined from 50 years of sea level and temperature data near 32°S (Feng et al. 2003). Although the volume transport of the Leeuwin Current is relatively small, its warm temperatures make it an important contributor to the basinwide heat transport. A more recent estimate of the Leeuwin Current volume transport using gridded hydrographic data finds a smaller southward transport, however, this difference is primarily due to differences in the reference level and the zonal extent of the integration (Furue et al. 2017). The Argo mapping of the interior section does not show any evidence of a southward eastern boundary current (not shown), and so we use the wider integration from Feng et al. (2003). There is uncertainty in the Leeuwin Current transport, which we address in section 2d.

5) INDONESIAN THROUGHFLOW

To conserve mass, we need an estimate of the ITF volume transport. The mean ITF transport is 15 Sv, estimated from 13.3 yr of observations in Makassar Strait between 1996 and 2017 (Sprintall et al. 2009; Gordon et al. 2019). Data from Makassar Strait, which represents 77% of the total ITF transport, are available for 6 years over the 2010s. The Makassar Strait transport was relatively small in 2014–16 and was large from 2016 to 2017 such that the mean during the 6 years of observations between 2010 and 2019 is similar to the mean from the full period of observations (Gordon et al. 2019). Therefore, we assume that the 2010–19 mean ITF transport is equal to the longer-term mean value of 15 Sv.

6) BELOW-2000-M INTERIOR

The below-2000-m interior volume transport was estimated as the difference between the other terms and the ITF transport of 15 Sv (Sprintall et al. 2009; Gordon et al. 2019). This is equivalent to assuming that volume is conserved in the Indian Ocean north of 36°S; that is,

\[
0 = V_{\text{ASCA}} + V_{0–2000m} + V_{>2000m} + V_{\text{Ekman}} + V_{\text{Leeuwin}} + V_{\text{ITF}},
\]

(5)

where \( V \) is volume transport.

The temperature of the below-2000-m interior was taken to be 1.6°C, the areal-average below-2000-m potential temperature from the 2009 hydrographic crossing. The few deep Argo temperature profiles in the region show temperature profiles within the range of the 2009 hydrographic section temperature profiles (not shown).

7) EDDY HEAT TRANSPORT

We do not have sufficient data to quantify the eddy heat transport directly because the Argo array does not resolve mesoscale eddies. The time-mean eddy heat transport is substantial in some parts of the global ocean, such as across the Agulhas Return Current poleward of 40°S (Ansorge et al. 2006; Volkov et al. 2008). We expect the eddy heat transport at 36°S to be small, because it is near the middle of the subtropical gyre, where eddy kinetic energy is low and the meridional temperature gradient is small (Roulet et al. 2014). We can estimate the magnitude of the time-mean eddy heat transport by assuming that the eddy heat flux over the top 2000 m follows the form of Fickian diffusion:

\[
\nabla' T' = -\kappa \frac{\partial T}{\partial y},
\]

(6)

where \( \kappa \) is the eddy diffusivity, primes represent anomalies from the time mean, and the overbar represents the time mean. The meridional temperature gradient between 35° and 37°S taken from the mapped Argo product over May 2016–April 2018 (ASCA mooring period) is used to calculate
\(dT/\partial y\). Taking 500 m² s⁻¹ as a generous value for horizontal eddy diffusivity over the upper 2000 m (Stammer 1998; Abernathey and Marshall 2013; Bachman et al. 2020) yields an eddy heat transport of \(-0.12\) PW. This estimate agrees well with an early modeling estimate in the region (Jayne and Marotzke 2002) and with an Argo based estimate at a similar latitude in the North Pacific (Roemmich and Gilson 2001).

We will show later that our estimated error (section 2d) is the same magnitude as this eddy heat flux. Therefore, we conclude that the eddy heat transport is not significant in the time mean.

c. 2-yr time series of heat transport

To construct a 24-month time series for May 2016–April 2018, we follow the same method that we use for the decadal mean, with two modifications. First, the above 2000-m interior, Agulhas Current, and Ekman components now vary monthly. Argo profile coverage over 2016–18 is adequate (Fig. 3). The Leeuwin Current component remains constant in time because there are not sufficient data to quantify its monthly variability. Second, we infer the below-2000-m flow by making a first guess that it is the decadal mean value of 9.6 Sv. Then, to conserve volume, we make a monthly-varying, spatially uniform barotropic adjustment to the full-depth interior so that the total flow across 36°S is equal to the monthly-varying ITF transport (Sprintall et al. 2009), similar to Bryden and Beal (2001). This barotropic adjustment is very small: 0.0012 cm s⁻¹ in the mean over the 24-month period. We apply this barotropic adjustment based on mass conservation rather than using Argo drift velocities because the monthly-mapped drift velocities imply a mean southward transport of deep waters, which is implausible given there is no deep-water formation in the Indian Ocean. This is likely due to an insufficient number of Argo float displacements per month to map the small and highly variable 1000-m absolute velocity. Conclusions on variability inferred from the 2-yr time series do not qualitatively change if the barotropic adjustment is not applied (not shown).

d. Errors

For each component of the heat transport, we estimate errors associated with temperature and volume transport, following McDonagh et al. (2015). The temperature-derived error is defined as \(\sigma_T = V\delta\theta\), where \(V\) is the spatial and temporal mean volume transport and \(\delta\theta\) is the temperature error. The volume transport-derived error is defined as \(\sigma_V = \theta\delta V\), where \(\theta\) is the spatial and temporal mean temperature and \(\delta V\) is the volume transport error. These two error sources are combined while assuming they are independent and yield a total uncertainty for each component. The total error is estimated by combining the uncertainty from each component, assuming that they are independent. This yields an error on each monthly estimate. For the error of the 2010–19 mean, we divide the monthly error by the square root of the degrees of freedom, assuming a 3-month decorrelation time scale on the 24 monthly estimates (based on the 90-day smoothing used in the Argo mapping). A summary of the errors is given in Table 3.

1) AGULHAS CURRENT

The ASCA array temperature error is a combination of the temperature-derived error of 0.07 PW with the transport-derived error of 0.28 PW, yielding a total error of 0.28 PW (McMonigal et al. 2020).
2) UPPER-2000-M GEOSTROPHIC INTERIOR

To estimate an error in the upper-2000-m geostrophic interior, we compare the Argo/SSH (sea surface height) mapping during 2009 with the 2009 hydrographic crossing, which we assume is the true ocean state. This yields a volume transport error of 3.1 Sv (McMonigal et al. 2018). Multiplying by the spatial mean temperature of 7.6°C yields a transport-derived error of 0.10 PW. To estimate a temperature-derived error, we consider the Argo/SSH data during April 2009, as compared with the 2009 hydrographic crossing, which we assume is the true ocean state. The difference is small in the spatial mean, −0.06°C, with an RMS difference of 0.24°C (Fig. 4c). Multiplying 0.24°C by the time-mean volume transport of 50.8 Sv gives a temperature-derived error of 0.05 PW. Combining these, the upper-2000-m geostrophic interior error is 0.11 PW.

3) EKMAN TRANSPORT

For the transport-derived error, we use the 1.0-Sv RMS difference between the NCEP- and ERA5-derived Ekman transports. Multiplying by the mean temperature of 17.7°C gives a transport-derived error of 0.07 PW. For the temperature-derived error, we consider the impact of using the 50-m temperature rather than the near-surface temperature, which effectively assumes a deeper and cooler Ekman layer. This leads to a difference of 0.02 PW. Combining these in quadrature, the total Ekman transport error is 0.07 PW.

<table>
<thead>
<tr>
<th></th>
<th>Mean $V$ (Sv)</th>
<th>Mean $\theta$ (°C)</th>
<th>$\delta V$ (Sv)</th>
<th>$\delta \theta$ (°C)</th>
<th>$V$-derived error (PW)</th>
<th>$\theta$-derived error (PW)</th>
<th>Total error (PW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agulhas</td>
<td>−76</td>
<td>12</td>
<td>14.6</td>
<td>0.23</td>
<td>0.28</td>
<td>0.07</td>
<td>0.28</td>
</tr>
<tr>
<td>Upper interior</td>
<td>51</td>
<td>7.6</td>
<td>3.1</td>
<td>0.23</td>
<td>0.10</td>
<td>0.05</td>
<td>0.11</td>
</tr>
<tr>
<td>Ekman</td>
<td>4</td>
<td>17.7</td>
<td>1.0</td>
<td>0.4</td>
<td>0.07</td>
<td>0.02</td>
<td>0.07</td>
</tr>
<tr>
<td>Leeuwin</td>
<td>−3</td>
<td>20.0</td>
<td>1.0</td>
<td>0.5</td>
<td>0.08</td>
<td>0.01</td>
<td>0.08</td>
</tr>
<tr>
<td>ITF</td>
<td>15</td>
<td>17.6</td>
<td>4.0</td>
<td>1.0</td>
<td>0.29</td>
<td>0.06</td>
<td>0.29</td>
</tr>
<tr>
<td>Deep interior</td>
<td>10</td>
<td>1.6</td>
<td>0</td>
<td>0.04</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>Total transport (convergence)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.32 (0.43)</td>
</tr>
</tbody>
</table>

FIG. 4. (a) Temperature from the April 2009 hydrographic crossing located at nominally 32°S. (b) April 2009 monthly mean of Argo temperature mapped onto the 2009 hydrographic station locations. (c) Difference between (a) and (b).
4) LEEUWIN CURRENT

The Leeuwin Current error in volume transport is taken as 1.0 Sv, due to the interannual variability in current strength (Feng et al. 2003). Surface temperatures vary by about 0.5°C over the seasonal cycle (Feng et al. 2003), and so we take 0.5°C as an estimate of the error in temperature. Neither of these errors were directly assessed by Feng et al. (2003), however, we take these values as conservative estimates, with the volume transport error 29% of the mean value. This leads to a transport-derived error of 0.08 PW and a temperature-derived error of 0.01 PW, which combine to a total error of 0.08 PW. This is substantially smaller than the errors on the Agulhas and ITF components, and so our assumptions here have only a small impact on the total estimated error.

5) INDONESIAN THROUGHFLOW

For the Indonesian Throughflow we take monthly-varying values from the INSTANT array (Sprintall et al. 2009). We consider a plausible error of 4 Sv and 1°C on each monthly estimate, informed by the large variability in ITF volume transport and TWT (Sprintall et al. 2009). ENSO can cause approximately 4-Sv fluctuations in ITF volume transport (Sprintall and Revelard 2014). The possible variability in TWT is complicated to assess, due to the different response of each ITF channel to the seasonal cycle and to El Niño–Southern Oscillation (ENSO) (Sprintall et al. 2009). However, it is expected that the velocity makes the largest contribution to the error, since it dominates variability in ITF temperature transport (Gruenburg and Gordon 2018). These error values yield a transport-derived error of 0.29 PW, and a temperature-derived error of 0.06 PW, combining to 0.29 PW.

6) BELOW-2000-M INTERIOR

We do not consider a transport-derived error for the below-2000-m component, because we infer this volume transport from the volume transport of the other components through the barotropic adjustment. We take 0.04°C as a maximum temperature error, because that is the spread of the below-2000-m mean temperature from the three hydrographic crossings. Multiplied by the 9.6-Sv volume transport yields a negligible below-2000-m interior error of 0.002 PW.

7) TOTAL ERROR

For an error in the heat transport at 36°S, we combine errors from all its components, giving an error of 0.32 PW on each monthly estimate. Dividing by the square root of the degrees of freedom reduces the error on the mean to 0.11 PW. For an error in the basinwide heat convergence, we combine errors from all components including the ITF. This yields a 0.43-PW error on each monthly estimate, reducing to 0.15 PW in the time mean.

e. Air–sea fluxes

To assess the role of the ocean heat convergence as compared with air–sea fluxes on the Indian Ocean warming, we integrate four air–sea flux products over the Indian Ocean, north of 36°S, over the period 2000–19. The Red Sea and Persian Gulf are included in the integration. Air–sea fluxes come from three reanalysis products (NCEP, ERA5, and MERRA-2) and one blended data reanalysis product (TropFlux). TropFlux is only integrated south to 30°S because that is the geographical limit of the data product.

3. Results

a. Inferred circulation

The Agulhas Current participates in the gyre, overturning, and throughflow circulations. The transport per unit depth profile (Fig. 6b) shows the subtle balance at all depths between the southward flowing Agulhas Current and the northward flowing interior that make up the gyre circulation. In the mean, the gyre circulation is 50.8 Sv (defined as the upper-2000-m northward flow), in line with previous estimates that range from 41 to 58 Sv (Palmer et al. 2004; McMonigal et al. 2018). The Agulhas Undercurrent (Beal and Bryden 1997) is not visible in the zonal mean across the boundary layer, as its weak northward transport over the continental slope is swamped by the larger southward transport of the Agulhas Current offshore. The interior flow is weakly southward below 1600 m, which is evidence of the upper arm of the deep overturning circulation. Added together, the interior and Agulhas transports are southward at all depths, reflecting the export of waters to match the import via the ITF.

Our analysis implies that there is a net 9.6-Sv flow into the Indian Ocean below 2000 m over the period 2010–19 (Figs. 5 and 7a). This agrees well with the deep transport found using the 2002 and 2009 hydrographic crossings (McDonagh et al. 2008; Hernández-Guerra and Talley 2016) and is 5–11 Sv less than the deep transport reported by previous authors who analyzed the 1987 data (Robbins and Toole 1997; Toole and Warren 1993; Sloyan and Rintoul 2001; Bryden and Beal 2001).

The overturning circulation is best illustrated in neutral density space (Fig. 6b). The Ekman transport is included in the lightest density layer, and contributes 4.2 Sv of northward flow, leading to an overall smaller southward transport of surface waters. There is a net southward transport in all layers except the Subantarctic Mode Water (SAMW) layer, compensating for water added to the basin via the ITF. SAMW is transported northward in all months, with a standard deviation of 4.7 Sv. Our overturning circulation has a structure similar to the inverse model circulation from the 2009 hydrographic crossing (Fig. 6b; Hernández-Guerra and Talley 2016). However, southward flow appears at higher density classes in our derived circulation and we estimate a larger influx of SAMW than that of Hernández-Guerra and Talley (2016). Our analysis adds novel information about the seasonal variability of the circulation, which we find is largest in the SAMW density class (Fig. 6b). This makes sense in terms of the known seasonal subduction of SAMW into the eastern arm of the gyre (Hanawa and Talley 2001).

From our results, we interpret the Indian Ocean overturning circulation as characterized by two overturning cells. The deep cell consists of 9.6 Sv of waters below 2000 m (structure
not resolved here), which flow northward (Hernández-Guerra and Talley 2016), upwell in the Indian Ocean, and then flow out of the basin as Upper Deep Water (UDW) and lower Antarctic Intermediate Water (AAIW). The shallow overturning cell consists of northward flowing SAMW that upwells in the Indian Ocean and exits as surface waters (Fig. 6b). A portion of this shallow cell crosses the equator to upwell in the Arabian Sea during the summer monsoon (Schott and McCreary 2001).
addition to these upwelling cells, which have been documented previously, our study suggests that a portion of the SAMW inflow is transformed into denser waters that exit the basin below the thermocline (Fig. 6c). We interpret this net downwelling of SAMW to reflect the formation of Red Sea Water (RSW), which plunges over the sill at Bab El Mendeb into the northwestern reaches of the Indian Ocean, entraining thermocline waters as it sinks (Peters et al. 2005). It is these high-salinity waters that set the warm and salty property of Indian Ocean intermediate waters—a mixture of RSW and AAIW—that exit the basin via the Agulhas Current (Beal et al. 2000; Nagura and McPhaden 2018).

b. Heat transport

The 2010–19 decadal mean heat transport is $-1.09 \pm 0.11$ PW, where the error bar is the standard error of the mean as described in section 2d. This is a smaller southward heat transport than previous studies (Fig. 7b; Table 1). A weak seasonal cycle is seen in the heat transport time series, with maximum southward heat transport in austral summer, January–March, which is 0.3 PW larger than the minimum in austral winter, July–October (Fig. 8a). The same phasing is found in heat transport from 1992 to 2017 in ECCO and in the seasonal cycle of the Agulhas Current volume transport (Beal et al. 2015). However, a seasonal cycle was not seen in the temperature transport of the Agulhas Current (McMonigal et al. 2020). The seasonal cycle does not account for the difference between our new estimate of southward heat export from the Indian Ocean and previous estimates from hydrographic cross-ings in March–April 2002 and March–May 2009. Therefore, our decadal mean of $-1.09 \pm 0.11$ PW as compared with the mean of $-1.3$ PW from Hernández-Guerra and Talley (2016) suggests southward heat transport has weakened by 0.21 PW between the 2000s and the 2010s (Fig. 7; Table 4).

Over our 2-yr monthly time series, peak-to-peak variability in heat transport is from $-1.17$ to $-1.76$ PW, with a standard deviation of 0.16 PW. Monthly meridional heat transport estimates at 35°, 30°, 25°, and 20°S in the Atlantic, constructed from Argo and satellite SSH data, have similar standard deviations of 0.22–0.16 PW (Majumder et al. 2016). The seasonal cycle in southward heat transport out of the Indian Ocean (Fig. 8a) suggests that measurements from June–July to February–March will be biased by seasonal aliasing. This should be taken into account in planning of future hydrographic crossings.

c. Components of heat transport

To investigate the dynamics responsible for variability in the southward heat transport, we calculate the gyre and overturning components in depth space following McDonagh et al. (2015) and Bryden et al. (2011) and decompose the heat transport across 36°S into throughflow, overturning, and gyre components. First, velocity and potential temperature are divided into components:

$$\mathbf{v} = \overline{\mathbf{v}} + \langle \mathbf{v} \rangle (z) + \mathbf{v}'(x, z)$$

(7)

$$\theta = \overline{\theta} + \langle \theta \rangle (z) + \theta'(x, z).$$

(8)

where the overbar represents a basinwide average, the angle brackets indicate zonally averaged deviations from the basinwide average, and the prime indicates residual deviations from the zonal average.
The throughflow component is the temperature transport due to the difference in temperature between the ITF inflow and the southern boundary outflow:

\[ Q_{\text{Throughflow}} = \rho C_p V_{\text{ITF}} (T_{\text{ITF}} - T_{\text{36S}}), \]  

where \( V_{\text{ITF}} \) is volume transport of the ITF. The overturning component is due to vertical overturning in depth space and is calculated as

\[ Q_{\text{Overturning}} = \rho C_p \int V_{\text{Pud}}(z) \langle \theta(z) \rangle \, dz, \]

where \( V_{\text{Pud}} \), the volume transport per unit depth, is

\[ V_{\text{Pud}} = \int \langle y(z) \rangle \, dx. \]

The gyre component is due to horizontal flows:

\[ Q_{\text{Gyre}} = \rho C_p \int \mathbf{v}(x,z) \mathbf{\theta}(x,z) \, dx \, dz. \]

The Ekman component is treated separately (Fig. 9). There is a small residual in the calculation, because the Agulhas Current temperature transport is 90-day smoothed in our heat transport estimate, while in this calculation, the velocity and potential temperature data from the Agulhas Current are smoothed independently, so that the separate heat transport components can be calculated.

At all times, the throughflow temperature transport and the overturning and gyre heat transports are southward (Fig. 9). The Ekman temperature transport is northward at all times. The overturning component contributes the most to the total heat transport in both the mean (−1.14 PW) and variability (0.18 PW standard deviation). The gyre is the second largest contributor to the mean (−0.48 PW) while the Ekman transport is the second largest contributor to the variability (0.17 PW standard deviation). The throughflow component has low variability, although this is likely due to our assumption of a constant TWT of the ITF. The overturning heat transport has the highest correlation with the total heat transport, with an \( R^2 \)-squared value of 0.6. If we consider that our 90-day smoothed estimates imply 8 degrees of freedom, this is significant at the 95% level. The Ekman and gyre transports are also correlated with the total, with \( R^2 \)-squared values of 0.4. Although these correlations are not significant for our short

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<th>Table 4. Heat transport and convergence estimates.</th>
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<td>ITF heat transport (PW)</td>
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<td>(i) 1987 (mean of previous authors, listed in Table 1)</td>
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<td>(ii) 2000s (Hernández-Guerra and Talley 2016)</td>
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<td>(iii) 2010s</td>
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<td>2010s − 2000s (i.e., iii − ii)</td>
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FIG. 8. (a) Seasonal cycle of heat transport across 36°S (gray dashed line; left axis), heat transport across 35.75°S from ECCO (gray line; left axis), and Agulhas Current volume transport (red line; right axis). Black closed circles show monthly means, and open circles show each of the 24 monthly estimates. (b) Seasonal cycle of heat convergence in the Indian Ocean (gray dashed line; left axis) and ITF volume transport (red line; right axis). Black closed circles are monthly means, and open circles are each of the 24 monthly realizations.
time series, they suggest that the Ekman and gyre components may play equally important roles in heat transport variability on monthly time scales.

Our results broadly match similar decompositions from the 1987 hydrographic crossing (Bryden and Beal 2001; Talley 2003). Bryden and Beal (2001) find that the overturning and gyre components contribute similarly to the heat transport (20.62 PW overturning; 20.75 PW gyre). Talley (2003) decomposes the transport in density space, and also finds a roughly 50:50 split (20.24 PW overturning; 20.27 PW gyre). Our calculation has limitations that are due to our treatment of the deep flow as a uniform slab, but this agreement with previous studies provides support for our method. Overall, this decomposition suggests that the Ekman transport and the overturning cell dominate heat transport variability on annual time scales, with a smaller contribution from the gyre component. On longer time scales, other components may dominate.

d. Heat convergence

Next, we estimate the Indian Ocean heat convergence by adding our estimate of southward heat transport out of the Indian Ocean to the estimate of heat transport into the basin via the ITF. Adding the ITF temperature transport of 1.08 PW to the 21.09-PW southward heat transport, we estimate a 2010–19 decadal mean heat convergence of 20.0160.15 PW. This estimate is significantly different from previous studies, which have found a divergence of heat within the Indian Ocean, implying a net gain of heat from the atmosphere (Fig. 7c).

We find that the convergence of heat in the Indian Ocean over the past decade has been 0.6 PW larger than previous estimates made using data from 1987, 2002, and 2009 (Table 4; Fig. 7c). This increase is robust above a semiannual cycle in heat convergence of 0.01 ± 0.15 PW. An important finding is that convergence is driven, in part, by a weakened southward heat transport from the Indian Ocean. A 0.21 PW reduction in heat exported from the basin across its southern boundary explains 33% of the change in convergence, while a 0.42 PW increase in the heat imported via the ITF explains the other 67%.

We deduce an estimated 0.42 PW increase in ITF heat transport per decade by comparing our basinwide budget analysis using Agulhas array and Argo data from the 2010s with the budget analyses conducted by Hernández-Guerra and Talley (2016) using hydrographic sections from the 2000s. This inferred ITF increase agrees well with an estimate based on mooring and satellite data of 5.9 Sv decade⁻¹ (Hu and Sprintall 2017), which is equivalent to 0.43 PW decade⁻¹ assuming a constant TWT of 17.6°C. Recent strengthening of the ITF has been linked to a shift toward La Niña–like conditions (Lee et al. 2015) and enhanced rainfall over the Maritime Continent (Hu and Sprintall 2017).

The reduced heat transport at the southern boundary has not been reported previously, and so we can use our budget analysis to investigate what dynamics may be responsible. We find no implied change in the deep overturning cell between the 2000s and 2010s (Fig. 7a). Thus, the Ekman-driven shallow overturning (Fig. 6b) and/or the gyre must have played a role in the decreased heat transport. In a recent analysis, Meng et al. (2020) find that the shallow overturning circulation has increased over 2000–17, largely due to decadal variability (see their Fig. 5). Yet, a stronger shallow overturning would transport more heat poleward, opposite to the sign of the changes we observe. Therefore, changes in the gyre circulation must dominate the decrease in meridional heat transport. This interpretation makes sense in terms of the observed Agulhas Current broadening since 1993 (Beal and Eliot 2016). The TWT of the Agulhas Current is cooler during meanders (McMonigal et al. 2020) and a broader current is consistent with more meandering (Beal and Eliot 2016).

We estimate the seasonal variability in ocean heat convergence for the first time, finding a semiannual cycle with largest heat divergence in October, February, and March, and smallest heat divergence in June, July, August, and December. The phase agreement between the ITF seasonal cycle (red line in Fig. 8a) and the heat convergence seasonal cycle suggests the

FIG. 9. Time series of components of the monthly heat transport, as labeled in the legend. Values for each component are shown as the mean plus standard deviation.
ITF is the primary driver of monthly variability in basinwide heat convergence. Monthly estimates range from $-0.69$ to 0.05 PW with a standard deviation of 0.19 PW. Hence, the month-to-month variability is the same order as the observed decadal change.

### e. Link to warming

Last, we consider our $-0.01 \pm 0.15$ PW ocean heat convergence in the context of the observed 20-yr rapid warming of the Indian Ocean (Desbruyères et al. 2017; Zhang et al. 2018). Integrating monthly mapped temperature north of 36°S from the Roemmich–Gilion Argo climatology (Roemmich and Gilson 2009) gives a monthly estimate of upper-2000-m ocean heat content. Over 2005–15, Indian Ocean heat content has increased by $1.97 \times 10^{22}$ J or 0.06 PW. This balances, within error bars, the heat convergence, implying that net air–sea fluxes over the Indian Ocean were near zero over the 2010s. In contrast, previous studies all found divergences of Indian Ocean heat of order 0.5 PW (Table 1), implying warming of the Indian Ocean through the air–sea interface and net buoyancy gain during the 2000s (Bryden and Beal 2001). If the net air–sea heat flux has reduced since the 2000s, even while the Indian Ocean has rapidly warmed, this is further evidence that the warming has been driven by significant changes in oceanic fluxes and not by surface forcing.

For an independent assessment of this change in air–sea fluxes, we integrate four air–sea flux products over the Indian Ocean, north of our southern boundary at 36°S (Table 5). Positive signs indicate fluxes into the ocean. The products do not agree on the sign of the decadal mean integrated fluxes over the Indian Ocean over the 2000s or 2010s, nor do they agree on the sign of the change between the 2000s and the 2010s. NCEP and ERA5 reanalyses agree reasonably well, however, TropFlux, which is more consistent with independent observations (Kumar et al. 2012), shows a different sign for the decadal means, and a much smaller change between the 2000s and 2010s, even when accounting for the discrepancy in geographical coverage of TropFlux (which extends only to 30°S (not shown)). MERRA-2 yields a drastically different magnitude of integrated air–sea fluxes than the other products, likely due to the large global heat imbalance ($+21 \text{ W m}^{-2}$; Josey et al. 2013) and different regional pattern of fluxes (Beal et al. 2019), and it shows the opposite signed change from the 2000s to 2010s. Putting MERRA-2 aside, the ensemble standard deviation of NCEP, ERA5, and TropFlux over the 2016–18 ASCA time period is still 0.8 PW. This is more than 5 times the estimated error on our ocean heat transport as well as convergence (section 2d). Because no consistent change is found from the air–sea fluxes, and the spread between products is much larger than the estimated error on ocean heat convergence, these independent products cannot substantiate a decadal change in air–sea heat flux. Still, based on our analysis, we conclude that the ocean has been driving a rapid increase in Indian Ocean heat content, in agreement with previous studies (Zhang et al. 2018; Lee et al. 2015; Hu and Sprintall 2017) and that about one-third of the warming is due to a weakening in the heat exported by the horizontal gyre circulation via the Agulhas Current.

### 4. Discussion

Using one-time ship-based sections and an inverse model, Hernández-Guerra and Talley (2016) found that the heat transported out of the Indian Ocean was greater in 2009 than in 2002, implying an opposing change to that we observe between the 2000s and the 2010s. The 2002 and 2009 hydrographic crossings were conducted in the same season, so that aliasing of the seasonal variability we estimated cannot explain this difference. Interannual variability may play a role, as both ENSO and the southern annular mode influence the strength of the Agulhas Current (Elipot and Beal 2018; Putrasahan et al. 2016), but we cannot quantify this with our 2-yr time series. Reanalyses show that the Indian Ocean heat transport is reduced during El Niño (Trenberth and Zhang 2019), however, neither 2002 nor 2009 were El Niño or La Niña years and our decadal mean using Agulhas array and Argo data is an average over several ENSO states. More data are needed to fully understand the variability in Indian Ocean heat transport and convergence. This dataset needs to include high-resolution measurements where Argo data cannot resolve the flow, namely, in the Agulhas Current, the Indonesian Throughflow, and the Leeuwin Current.

Our treatment of the below-2000-m interior section is crude due to lack of data in the deep ocean. The deep section contains an overturning cell that is not fully resolved here. The sensitivity of the Indian Ocean heat transport to this deep cell is within our estimated errors, because most of the mean heat transport is carried by the shallow overturning cell and the gyre (Ferrari and Ferreira 2011; Sloyan and Rintoul 2001; McDonagh et al. 2008). Nonetheless, it is possible that the deep overturning cell contributes to the heat transport variability. The crude treatment of the below-2000-m section is likely to lead to errors in the decomposition of the heat transport into gyre and overturning components, more than it contributes to errors in the total, time-mean heat transport. To monitor changes in the heat budget more robustly would take a dedicated observing system at the southern boundary of the Indian Ocean able to resolve the boundary fluxes directly and capture the top-to-bottom basinwide density gradient for the

| 2000–09 | −0.59 | −0.72 | 0.79 | 12.09 |
| 2010–19 | −0.44 | −0.59 | 0.81 | 11.95 |
| ASCA time period | −0.55 | −0.92 | 0.78 | 12.05 |

Table 5. Integrated air-sea fluxes over Indian Ocean.
interior geostrophic fluxes. Such a system is a priority for the Indian Ocean community (Beal et al. 2020).

Using Argo data and boundary current arrays to estimate heat transport and convergence can supplement the more common method of using hydrographic crossings in conjunction with inverse models. Because Argo does not resolve the mesoscale, this method is best suited to latitudes where eddy heat transport is expected to be small, such as the centers of the subtropical gyres. With appropriate boundary flow monitoring, the estimated errors are similar in magnitude as those from inverse models (Ganachaud et al. 2000). XBT data may also be useful in making heat transport estimates in some regions (Goes et al. 2020). Estimating the seasonal cycle in heat transport and convergence at key latitudes in each ocean basin would be one way to leverage combined Argo/boundary current monitoring methods. This would provide a baseline to avoid aliasing of the seasonal cycle when comparing temporally sparse hydrographic data. Because the Atlantic Ocean is already well monitored by RAPID-MOCHA (Johns et al. 2011) and other programs, the North and South Pacific Oceans could be important areas to target. Errors in the implied air–sea fluxes from these ocean observation-based estimates of ocean heat transport and ocean warming are one-fifth of the spread between air–sea flux products. Thus, these estimates of ocean heat convergence could serve to assess which air–sea flux products are in best agreement with the ocean data.

5. Conclusions

We find net zero heat convergence within the Indian Ocean over the last decade, a substantial change from estimates of divergence during the 2000s. This change is consistent with the observed rapid warming of the Indian Ocean over the past two decades. About two-thirds of the increase in ocean heat convergence since the 2000s is related to a strengthening of the ITF (Lee et al. 2015; Hu and Sprintall 2017). The other one-third is caused by a decrease in heat export across the southern boundary at 36°S. Because the below-2000-m flow has not changed since the 2000s, and the shallow overturning circulation has increased in strength (Meng et al. 2020), the change must be due to the gyre circulation. We conclude that this change in the gyre component of the heat transport is due to the broader, and therefore cooler, Agulhas Current (Beal and Eliot 2016; McMonigal et al. 2020). On annual time scales, the overturning and Ekman components dominate the variability in heat transport, but on longer time scales the gyre appears to become dominant. The observed change in Indian Ocean heat convergence implies net air–sea fluxes over the basin were near zero over 2010–19. Because of the substantial role of the southern Indian Ocean gyre circulation in the recent rapid warming of the Indian Ocean, as found here, we cannot overlook its potential influence on future climate change scenarios. The expected reduction in ITF transport due to climate change (Ma et al. 2020; Stellemma et al. 2019; Sen Gupta et al. 2016; Feng et al. 2017) may be partially offset by a further reduction in the heat transport at the southern boundary and the Indian Ocean could continue to warm at a faster rate than other subtropical ocean basins.

Acknowledgments. This work was funded by U.S. National Science Foundation Award 1459543, the Agulhas System Climate Array. We acknowledge our South African partners, led by principal investigator Juliet Hermes and project manager Tamaryn Morris of the South African Environmental Observation Network (SAEON), without whom this science would not have been possible. We thank the South African Department of Science and Innovation and the National Research Foundation of South Africa for supporting this project through SAEON and the Department of Environment, Forestry and Fisheries (DEFF, Branch: Oceans and Coasts). We thank Bradley Blows and Nauti-Buoys for building the ASCA moorings and leading deck operations on deployment and recovery, the Royal Netherlands Institute for Sea Research for their instrumentation contributions, and the crews and science parties of the RS Algoa and SA Agulhas. Author Willis acknowledges support of the Jet Propulsion Laboratory, California Institute of Technology, under a contract with the National Aeronautics and Space Administration (80NM0018D0004).


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