Physical oceanography of the present day Indonesian Throughflow

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Abstract: The Indonesian Throughflow (ITF) transfers c. 15 Sv (1 Sv = 10^6 m^3 s^-1) of relatively cool, fresh water from the tropical Pacific Ocean to the tropical Indian Ocean. Additionally, the ITF is a key interocean component of the global ocean warm water route, which returns water from the Pacific Ocean to the Atlantic Ocean to close the loop of the thermohaline overturning circulation associated with North Atlantic Deep Water. That flow consequently freshens the Indian Ocean and transports heat between basins. The ITF can also be described by the island rule, which relates the winds over the entire South Pacific Ocean to the magnitude of the ITF. El Niño-Southern Oscillation (ENSO) dominates the regional variability in the Pacific Ocean and exerts a strong control over the variability of ITF transport. The Indian Ocean responds to the ENSO signal as well, but is also influenced by the Indian Ocean Dipole, a climate phenomenon that may act independently of ENSO to affect the ITF.

On a local scale, the surface layer of the ITF is controlled by local winds, which are primarily monsoonal, but at depth the ITF responds to the pressure gradient between the Pacific Ocean and the Indian Ocean. Newly available observational data from within the major straits of the Indonesian Seas allow for an improved resolution of the total ITF and its variability. Makassar Strait, which is the main inflow channel of the ITF, transported an average of 11.8 Sv of North Pacific water. Lifamatola passage, a route for deep South Pacific water, transported 2.7 Sv below 1250 m and returned 0.9–1.3 Sv to the north. The Karimata Strait has not been monitored observationally, but models estimate a contribution of 1.4 Sv, which is likely important for its role in the transport of freshwater. Transport through the outflow passages was 2.6 Sv through Lombok Strait, 4.9 Sv through Ombai Strait, and 7.5 Sv through Timor Passage. Temporal variability within the straits is seen on timescales from the interannual ENSO signal to semi-diurnal tidal signals.

The Indonesian Throughflow (ITF) transports c. 15 Sv (1 Sv = 10^6 m^3 s^-1) from the Pacific to the Indian Ocean. The structure and magnitude of the ITF varies on temporal scales from days to decades and even over geological time. As the water is transported, its hydrological characteristics are altered by heat and freshwater inputs from the Indonesian Seas and by strong vertical mixing. The physical oceanography of the present day ITF therefore cannot be characterized by a mean transport alone and requires an understanding of the surrounding current systems and climatic conditions. The ITF can be understood as both a local phenomenon, affecting the Indonesian Seas and the edges of the western Pacific and eastern Indian basins, and as a global phenomenon, providing a path for water to circumnavigate the world ocean.

The objective of this paper is to understand the significance of the ITF and its variability in both a global and a local sense. The first section places the ITF in the context of global ocean circulation, showing its importance to thermohaline circulation, estimates of volume, heat, and salt transport, and the framework of large-scale wind-driven circulation. The next section narrows in focus from a global to a regional scale, providing the physical oceanography and prevailing climate phenomena that interact with the ITF.

The ITF is considered on a local scale in the following two sections. The third section develops the theoretical frameworks for understanding the existence of the ITF, while the fourth provides observational data for understanding the mean flow and temporal variability within individual straits. The study concludes with a discussion of the implications of the ITF and directions for future work.

Global ocean circulation

Thermohaline circulation

The circulation of the world ocean is often compared to a conveyor belt. In the North Atlantic, cold, salty surface water sinks and spreads southward through the Atlantic Ocean as North Atlantic Deep Water (NADW). That NADW is then carried by the Antarctic Circumpolar Current and deep western boundary currents to the Pacific and Indian Oceans. In order for this transport to occur, there must be a return flow of comparatively warm water balancing the mass transport. Gordon (1986) proposed that this exchange must happen in the

oceans’ thermocline, the layer of seawater between the surface mixed layer and the cooler deepwater in which the water temperature changes rapidly with depth. This occurs via the ITF and the subsequent export of those waters from the Indian Ocean into the South Atlantic through the Agulhas Leakage. This concept, now well accepted in principle, requires that the ITF balance the transport of mass, heat, and salt from the Pacific to the Indian Ocean.

The metaphor of the ‘conveyor belt’ originated with Broecker & Peng (1982) and referred to a conceptual model of global circulation (for a discussion of the history of the ‘conveyor belt,’ see Richardson 2008). The concept was popularized through a magazine article (Broecker 1987) that showed a simplified version of the circulation described in Gordon 1986 (Fig. 1). This version has been reproduced many times, although it fails to show the true path of the ITF through the Indonesian Straits and neglects the formation of Antarctic Bottom Water. A more sophisticated conveyor belt, developed by Gordon (1991) and updated by Schmitz (1996) and Lumpkin & Speer (2007) attempts to show a more three-dimensional view of global ocean circulation (Fig. 2). The structure of these later diagrams emphasize the fact the ITF is the only low latitude connection between ocean basins, and the largest interocean flow outside of the Southern Ocean.

As the conceptual framework of a conveyor belt expands to include more details about ocean circulation, it builds in complexity and becomes more difficult to show schematically. However, a more crucial concern is that it gives the impression of an ocean governed by a strong mean flow, with little, if any, variability. Such a deterministic view of ocean circulation is contradicted by observational evidence, including evidence that the ITF is highly variable in its transport not only of volume, but of heat and salt as well.

Transport of volume, heat and salt

Whether ocean circulation is best described by the mean flow or by temporally variant eddies and currents, transport of mass, heat and salt must balance over long timescales. The World Ocean Circulation Experiment (WOCE) project attempted to quantify these fluxes using observational data. These data, along with satellite and other data, are then incorporated into general circulation models which can provide an estimate for both the mean state and the variability of these fluxes (see Wunsch et al. 2009 for a full discussion of these methods). Volume transport is the rate of volume flow across a unit area, calculated as the product of the integrated velocity through the area and the cross section of that area. Given that the mass of seawater is not variable, conversions between volume and mass transport are simply a matter of units and the two terms are often used interchangeably. The

Fig. 1. The ocean conveyor belt as proposed by Broecker (1987).
volume transport of the ITF has been estimated from to 5 to 15 Sv (Gordon 1986; Godfrey 1996; Gordon et al. 2009).

An accurate estimation of volume transport is required for any calculation of heat or salt transport. However, the calculation of heat transport, more precisely referred to as internal energy transport (see Warren 1999 for discussion), also requires the use of a reference temperature. This constraint is inherent to the calculation and arises from the need to consider not only the flow through a given section but the eventual export (or previous import) of those waters as well. The heat transport is the difference in heat capacities of the water being imported and the water being exported. Thus, all heat transport calculations are somewhat dependent on the choice of reference temperature, although volume transport has a far larger effect.

In the case of the ITF, heat transport estimates range from 0.24 PW (Vranes et al. 2002) (1 PW = 10^{15} watts) to 1.15 PW (Schiller et al. 1998), and reference temperature estimates range from 0 °C to 3.4 °C, the presumed temperature of Pacific Ocean inflow or Indian Ocean outflow.

Because salt comprises only a small fraction of seawater, it is computationally simpler to think of the flux of freshwater, rather than the flux of salt. The ITF transports water from the fresher Pacific Ocean to the saltier Indian Ocean. Like heat flux, freshwater flux considers the difference in salinity between the inflow and outflow waters. In this case, the water entering the Indian Ocean via the ITF is less saline than the water exiting the Indian Ocean. The difference results in a freshwater transport of 0.23 Sv. Since the Indian Ocean is not becoming increasingly fresh, freshwater must be removed via net evaporation over the Indian Ocean (Talley 2008).

**Wind-driven circulation**

Although much of the large-scale circulation in the ocean is ascribed to thermohaline forcing, it can also be understood in terms of global patterns of wind forcing. Wind stress, the drag exerted by air moving over the ocean, can be used to calculate the volume of water integrated from the sea floor to the sea surface being transported north or south, assuming an ocean in which local velocity is governed by density differences on a rotating Earth. This relationship, the Sverdrup balance (Sverdrup 1947), was extended to calculate flow around islands (Godfrey 1989). According to Godfrey’s island rule, the transport around an island can be calculated using the integral of the wind stress along a closed contour encompassing the island and the entire width of the ocean basin located eastward of that island. In the case of the ITF, the relevant island is Australia-New Guinea and the path can be seen in Figure 3. Godfrey’s original

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**Fig. 2.** A more accurate schematic of global ocean circulation (Gordon 1991).
calculation led to a long-term mean ITF of 16 ± 4 Sv. The island rule was later modified to include the effects of friction and bottom topography (Wajsowicz 1993), showing that the Indonesian Seas modify the ITF and reduce its magnitude. That study further suggested that the island rule could be extended to include interannual variability of ITF in addition to the long-term mean. Wajsowicz (1996) used observational wind data to calculate the interannual ITF variability.

Using an ocean circulation model, Humphries & Webb (2008) compared interannual variability of the ITF to that produced by the island rule. They found that the island rule agreed well with the model when run using annually repeating monthly climatological winds, but less so when the model was run with more realistic winds. This may be due to the time required for signals generated within the Pacific Ocean to propagate to the ITF. However, given that the same wind stress data were used in both the model and the island rule calculation, it is not possible to conclusively quantify the relationship between the in situ wind stress and the in situ transport.

**Regional circulation and climate**

**Western equatorial Pacific Ocean**

The western equatorial Pacific Ocean contains a complex set of currents and several eddy systems (Fig. 4). The dominant currents are the westward-flowing North and South Equatorial Currents and the eastward-flowing North and South Equatorial Counter Currents and Equatorial Undercurrent. The inflow area for the ITF lies at approximately the same latitude as the North Equatorial Counter Current, in an area with two substantial eddies: the Mindanao Eddy to the north and the Halmahera Eddy to the south.

North Pacific Intermediate Water (NPIW) is defined as the salinity minimum in the subtropical North Pacific (Talley 1993). That water mass is formed by the mixing of cold, fresh subpolar waters with saltier subtropical waters and occupies the intermediate depths (300–700 m) of the northern equatorial Pacific Ocean. In the south, the thermocline contains Antarctic Intermediate Water (AIW), which is colder and saltier than NPIW and is formed by overturning in the Southern Ocean. Overlying AIW is South Pacific Equatorial Water, which is also colder and saltier than its northern counterpart, North Pacific Equatorial Water.

Godfrey et al. (1993) argued that although the relatively fresh waters of the ITF suggested a northern source for the throughflow, the source must actually be to the south to account for the island rule’s dependence on wind stress to the south of the ITF. Although most of the South Equatorial Current retrofects to join the North Equatorial Counter Current, some could conceivably leak from the Halmahera eddy to form all or part of the ITF. South Pacific water is distinctly saltier than both North Pacific and ITF water, so they argued that this South Pacific source water was freshened by rainfall so that it appeared to be North Pacific water by the time it reached the ITF.

The distinction here between North and South Pacific water is partly semantic, given that all North Pacific water was, at some time in the past, South Pacific water. North Pacific water forms when South Pacific water, following the South Equatorial Current, turns back to the east in the Halmahera Eddy (Fig. 4). That water flows eastward along the North Equatorial Countercurrent, then turns again to flow westward along the North Equatorial Current. Only when it reaches the western boundary of the Pacific a second time does it join the Mindanao Current and enter the Indonesian Straits.

Gordon (1995) argued that the ITF source water was from the North Pacific and that its characteristic low salinity was the result of the excess precipitation throughout that region. He did, however, mention...
that a small amount of South Pacific water might enter the ITF below the thermocline. Nof (1996) used a theoretical approach to show that the complex geography of the Indonesian archipelago and the existence of retroreflecting currents controls the source water of the ITF and leads to an ITF that is primarily from the North Pacific with only a small contribution from the South Pacific. Later studies of water mass analysis have confirmed the primacy of the North Pacific as the source of the ITF (Ilahude & Gordon 1996).

The surface conditions are dominated by the trade winds, which blow from east to west and towards the equator in both hemispheres. They meet at the intertropical convergence zone, which has an average location of 5°N. The trade winds establish a sea level gradient, with higher sea level in the western side of the basin, and an associated gradient in the thermocline, with a deeper, thicker layer of warm water in the west forming the western Pacific warm pool.

**El Niño-Southern Oscillation**

El Niño-Southern Oscillation (ENSO) is centred in the Pacific Ocean but can influence global climate...
through oceanic and atmospheric teleconnections. Its effect on the ITF is far more localized, as it changes the ITF source waters directly. When the trade winds slacken and the zonal sea surface height and temperature gradients weaken or reverse, creating an El Niño event, the magnitude of the ITF decreases. The opposite effect is seen during a La Niña event, with increased trade winds, an increased sea surface height and temperature gradient, and an anomalously large ITF. Meyers (1996) used a twelve-year record of observational hydrological data from the outflow of the ITF to construct a record of ITF variability. He found that the strongest signal within the data corresponded to the ENSO signal. The same result has been seen in modelling studies (Clark & Liu 1994; England & Huang 2005). The strength of ENSO can be recorded by various indices, among them the NINO3.4 index, which measures the sea surface temperature anomaly in the eastern-central Pacific. Although major ENSO events can be very clearly seen in all indices and measures of ENSO, smaller events can elude classification and remain a subject of scientific debate (Meyers et al. 2007).

**Equatorial Indian Ocean**

The Indian Ocean is notable for its seasonal variability. The SE Asian monsoon dominates the equatorial Indian Ocean and leads to current reversals. During the winter (all seasons are defined relative to the North Hemisphere), the monsoon brings north-easterly winds from the high pressure centre over Asian landmass. During this time, the current system resembles that of other ocean basins, with a North Equatorial Current, a South Equatorial Current, and an Equatorial Counter Current (Fig. 4, top panel). In the west, the Somali Current brings water south from the North Equatorial Current to the Equatorial Counter Current. In the east, the South Java Current flows toward the Indonesian archipelago and curves to the south, bringing water from the Equatorial Counter Current to the South Equatorial Current. The situation reverses during the summer, when there is an atmospheric low over the Asian landmass and corresponding southwesterly winds (Fig. 4, bottom panel). The Somali Current reverses to flow to the north and South Java Current reverses to flow to the west. In place of the North Equatorial Current and Equatorial Countercurrent, the SW Monsoon Current flows from west to east. During both seasons, the Leeuwin Current flows southward along the western coast of Australia. The transitional periods between summer and winter monsoons, typically during May to June and October to November, are accompanied by Wyrtki jets, which are strong equatorial currents travelling from west to east. Over the most of the Indian Ocean and the Indonesian Seas, the monsoonal winds are directed from the NW in winter and from the SE in summer.

In the Indian Ocean, the warm pool is on the eastern side of the basin and can be seen as an extension of the western Pacific warm pool. Accordingly, the water in the eastern Indian Ocean is warmer and fresher than the water in the western Indian Ocean.

**Indian Ocean Dipole**

Just as an El Niño event is associated with a slackening or reversal of the trade winds, the Indian Ocean can experience a similar phenomenon when the summer southwesterly monsoon winds blow from east instead. This leads to anomalous warming and freshening of the western Indian Ocean and the upwelling of cooler water in the eastern Indian Ocean. This is considered the positive phase of the Indian Ocean Dipole (IOD). In its negative mode, the IOD strengthens the average conditions of the Indian Ocean, further warming the east and cooling the west (Saji et al. 1999).

The IOD and ENSO can co-vary, prompting a debate over the existence of the IOD as an independent climate phenomenon or merely as an extension of ENSO dynamics (see Meyers et al. 2007, for a review). Positive IOD events generally occur during El Niño years, and negative IOD events generally occur during La Niña years. They can, however, occur independently (Yamagata et al. 2004). Like ENSO, the IOD is tracked by an index of sea surface temperatures, called the Dipole Mode Index (DMI). In this case the difference in sea surface temperature anomaly between the western and southeastern Indian Ocean. The IOD can also be tracked by considering just the eastern sea surface temperature anomalies (Meyers et al. 2007). As in the case of ENSO, there remains uncertainty in the classification of many years as IOD events.

**The Indonesian Seas**

The complex geography of the Indonesian Seas can be seen in Figure 5. The area shows complex bathymetry as well, with deep basins and many sills (Gordon et al. 2003a). The main path of the ITF consists of water entering from the North Pacific between the Philippines and New Guinea into the Sulawesi (Celebes) Sea and continuing through Makassar Strait. From there, the water can exit via Lombok Strait or circulate through the Banda and Flores Seas and can enter the Indian Ocean via Ombai Strait or Timor Passage. In addition, deeper South Pacific waters enter Lifamatola Passage and provide cooler, saltier water to ventilate the Banda
Sea (van Aken et al. 2009). Surface water also enters the Flores Sea from the South China Sea via Karimata Strait (Qu et al. 2005). The Indonesian Seas experience significant freshwater input from rainfall and river runoff and are warmed by surface heat fluxes south of Makassar Strait (Wijffels et al. 2008). Internal tides are trapped within the various semi-enclosed seas, resulting in strong vertical diffusivity that mixes the buoyant surface water downwards. Although the presence of these strong tides does not change the magnitude of the ITF, it changes the temperature and salinity of the outflow water (Koch-Larrouy et al. 2007).

Local forcing of the ITF

Pressure driven flow. Wyrtki (1987) observed that a pressure gradient between the western Pacific Ocean and the eastern Indian Ocean must exist and govern the ITF. This is true whether that gradient exists due to the island rule, thermohaline circulation, or any other cause. Although pressure differences alone cannot give a numerical value for the ITF, they can quantify its variability relative to an unknown mean. To investigate this, Wyrtki compared sea level from Davao in the Philippines and Darwin in Australia to develop a time series of ITF variability relative to an unknown mean. He hypothesized that the trade winds in the Pacific built up a pressure head from the Pacific Ocean to the Indian Ocean. However, the sea surface signal at Darwin is more representative of the Pacific Ocean than the Indian Ocean, which may explain why the resultant time series showed no correlation with ENSO. Since modern observations show a strong ENSO signal, this approach must be modified by using data that are more representative of the Indian Ocean. Furthermore, differences in the sea surface do not fully describe differences in pressure below the sea surface. The ITF has been shown to be primarily baroclinic (Waworuntu et al. 2001), meaning that pressure surfaces, such as the sea surface, are inclined relative to density surfaces.

To address the problem of accurately representing sea surface height in both the Indian and Pacific Oceans, Potemra et al. (1997) and Potemra (2005) used sea level derived from satellite altimetry at multiple locations on the Pacific and Indian sides of the ITF to calculate the pressure difference between the two ocean basins. In addition, they expanded the concept by allowing the sea level changes to lead or lag the ITF signal to account for the time needed for signals to propagate through the basins. The resultant ITF series does not show a correlation with ENSO, which may...
again be explained by the baroclinic nature of the flow. Potemra et al. (2003) noted that the ENSO signal propagates through deeper layers of the ITF. To address the issue of subsurface forcing on the ITF, Burnett et al. (2000a, b) explored the possibility that drag from the complex bathymetry of the Indonesian Seas could cancel the pressure head from the Pacific Ocean, meaning that the pressure head would not control the ITF. The model used in that study is barotropic, neglecting the relative tilt of pressure surfaces to density surfaces, meaning that the pressure head was calculated solely as the difference in sea surface heights. The results showed that transport through individual channels would change in response to increased or decreased sea surface height differences, but that the total ITF would not.

Burnett et al. (2003) and Kamenkovich et al. (2003) further explored the relationship between the interocean pressure difference and the total ITF transport in a two-part study. Their results emphasized the importance of bathymetry and indicated the pressure head, as calculated by a variety of different methods, was correlated with the seasonal variation of the ITF but could not uniquely determine its value. That is, a particular pressure head could exist for multiple values of ITF transport. They found that the transports from the Pacific currents into and away from the Indonesian Seas also influenced the total transport of the ITF.

The interocean pressure gradient can be used to provide a numerical value for velocity in addition to its variability. This is done by calculating a geostrophic balance, which assumes that velocity must balance a pressure gradient, the effect of the Earth’s rotation and friction. Unlike the pressure difference alone, which can at best calculate the variability of a flow relative to an unknown mean, geostrophy provides a velocity relative to a deeper layer, ideally one presumed to have no motion. This method was used by Meyers et al. (1995) and Meyers (1996) with observational temperature data collected in the Indonesian Seas and Indian Ocean by volunteer merchant ships using expendable bathythermographs (XBTs). The XBT temperature data and mean salinity were used to calculate velocity in the top 400 m of the water column. They found a strong relationship between monsoon winds, which can change the pressure gradient from the Indian side, and the ITF. Potemra et al. (2002) also used this method to calculate velocities in the outflow channels and compare them to in-situ observational data and model output. They found the flow to be in geostrophic balance. Wijffels et al. (2008) used an expanded XBT dataset and found an average ITF of 8.9 Sv. However, they suggest that this estimate is too low because it does not account for the deep portion of the ITF.

Geostrophic balance calculations are useful for the individual outflow straits, but are difficult to apply to the inflow straits and thus to the ITF as a whole. This is because geostrophic conditions do not hold along the equator and in narrow straits, such as Makassar (Burnett et al. 2000a). This is particularly problematic because most of the ITF is transported through Makassar Strait. In addition, the Banda Sea acts as a capacitor, holding water for a varying amount of time before it is released into the Indian Ocean (Gordon & Susanto 2001). Fieux et al. (1996) noted that geostrophic currents could also be affected by friction resulting from the proximity of strong currents to coastlines.

Song (2006) used both sea surface height and ocean bottom pressure to estimate the ITF, combining the techniques of geostrophy and hydraulic control of a strait. The study found the combination of both factors gave a better approximation to the total ITF than did either one alone. Tillinger & Gordon (2009) used interocean pressure differences calculated on layers of constant density to construct a timeseries of ITF variability. The resultant variability was found to match observational data on an interannual timescale and island rule calculations on a decadal timescale.

**Local wind driven flow.** In addition to the geostrophic pressure-driven flow, a portion of the ITF is ageostrophic and driven directly by local winds. The wind drives Ekman transport, which relates local wind stress to transport, modulated by the rotation of the Earth. The ITF can be divided into several vertical layers, with the top layer forced by surface winds and the deeper layers responding to remote forcing (Potemra et al. 2002). The Ekman and geostrophic transports may interact, with the southward Ekman flow raising sea level and consequently deepening the thermocline and changing the associated pressure gradient.

Although the existence of Ekman transport in the ITF is clear, its magnitude is uncertain. Murray & Arief (1988) noted that the surface flow could reverse to the north and attributed those events to cyclonic wind events that are part of the monsoonal cycle. Potemra et al. (1997) found that the ageostrophic Ekman transport was much smaller than the geostrophic transport but could account for as much as 1.8 Sv or, 25% of the total ITF, and was highly variable. Sprintall & Liu (2005) used scatterometer and SST data to more precisely calculate the Ekman contribution to the ITF, and found that Ekman transport can be the same order of magnitude as the geostrophic flow through the outflow straits, but reverses with the phase of the monsoon.
The summer monsoon is associated with sustained winds leading to southward transport. The winter monsoon leads to strong but sporadic wind bursts over the Indonesian Seas, which lead to northward transport that balances the previous southward transport. So despite the significant contribution of Ekman transport to the total ITF at the outflow straits on sub-annual timescales, it does not significantly impact the ITF and timescales longer than one year. These reversals were confirmed by observational data from within the straits (Sprintall et al. 2009). Tillinger & Gordon (2009) found that the surface layer flow within Makassar Strait correlated with local wind variability. Between Australia and Java, further from the outflow straits, calculations from wind data suggest that Ekman transport accounts for 40% of the total transport over a twenty-year period (Wijffels et al. 2008).

**Individual straits**

The International Nusantara Stratification and Transport (INSTANT) program deployed 11 moorings over a three-year period to obtain simultaneous observations of the major straits of the ITF: Makassar Strait, Lifamatola Passage, Lombok Strait, Ombai Strait, and Timor Passage (Fig. 5). Earlier studies of the ITF exist, but they did not observe all of the passages simultaneously. By convention, transport values are described here as positive when they are directed from the Pacific to the Indian Ocean. The most recent transport values from INSTANT (shown in Table 1) give a total transport of 15 Sv.

**Makassar Strait.** Makassar Strait was monitored by two moorings placed at its narrowest constriction, the 45 km wide Labani Channel, analysed by Gordon et al. (2008). The resultant velocity profile (Fig. 6, top panel) shows a peak within the thermocline, at 110 to 140 m, with maximum velocities of nearly 1 m/s. The surface flow shows average velocities of c. 0.5 m/s, dropping to nearly zero below the 680 m depth of the Dewakan sill. However, velocities remain slightly positive on average down to a depth of 1500 m. A similar profile was seen during the Arus Lintas Indonen (ARLINDO) observational period of 1997, when moorings were placed in the Labani Channel (Gordon et al. 1999; Susanto & Gordon 2005). The summer monsoon is associated with greater maximum velocities, but lower velocities in the deeper flow below c. 250 m.

Transport (Fig. 7, top panel) was calculated by multiplying the velocity by the depth-dependent cross section of the Labani channel. The minimum transport, 8.8 Sv, is seen from October to December, when the monsoon is in a transitional phase. The April to June transition also shows low transport (11.7 Sv), while transport is higher from July to September (12.6 Sv) and reaches a maximum of 13.1 Sv in January through March. The mean transport during INSTANT was 11.6 Sv, which is an increase of 27% from the ARLINDO value of 9.2 Sv. The difference can be attributed primarily to the El Niño event of 1997–1998, which would have substantially changed conditions in the region. The INSTANT period was primarily neutral or in a weak El Niño, and ended with a La Niña event. Both observational periods occurred during positive IOD events. Tillinger & Gordon (2009) suggest that the surface flow, normally due to Ekman transport and therefore independent of the pressure-driven flow, would have been dominated by ENSO signal during the ARLINDO period, lowering the total transport. These observational data confirmed the strong effect of ENSO on the volume of ITF transport.

The transport-weighted temperature (TWT) of the flow is the average temperature in the channel, weighted by the transport at different depths. Gordon et al. (2008) found an average TWT of 15.6 °C. The seasonal pattern of TWT follows the seasonal transport pattern, with a minimum in the October to December transitional period between monsoons and a peak in the January to March winter monsoon. During ARLINDO, Vranes et al. (2002) found a TWT of 15.2 °C. Since the TWT is a function of the vertical distributions of temperature and velocity, it can be considered the heat transport per unit volume transport. A change in the total transport but not in the velocity distribution will not change the TWT, but it will change the heat transport (see section 1.2 for a discussion of heat transport). In periods of greater transport, the location of the maximum current speed deepens and therefore contains cooler water. Therefore when volume transport increases, the TWT decreases, but the heat transport increases. This disconnect between the TWT and the heat transport is a result

### Table 1. Transport and TWT through the straits of the ITF

<table>
<thead>
<tr>
<th>Strait</th>
<th>Transport (Sv)</th>
<th>TWT (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Makassar Strait</td>
<td>11.6</td>
<td>15.6</td>
</tr>
<tr>
<td>Karimata Strait</td>
<td>1.4*</td>
<td>–</td>
</tr>
<tr>
<td>Lifamatola Passage</td>
<td>2.7</td>
<td>3.2</td>
</tr>
<tr>
<td>Lombok Strait</td>
<td>2.6</td>
<td>21.5</td>
</tr>
<tr>
<td>Ombai Strait</td>
<td>4.9</td>
<td>15.2</td>
</tr>
<tr>
<td>Timor Passage</td>
<td>7.5</td>
<td>17.8</td>
</tr>
</tbody>
</table>

*There are no observational data available for Karimata Strait. This value is from the modelling results of Tozuka et al. (2007). No TWT was available.*
of both the intensified flow within the thermocline through Makassar and the tendency of the velocity maximum to deepen when total transport increases (Tillinger & Gordon 2010). Gordon et al. (2003b) attribute the subsurface transport maximum and the corresponding low TWT as the result of fresh Java Sea water pushed in to the southern end of Makassar Strait during the winter monsoon. This thin sheet of freshwater in the top 50 to 100 m prevents the flow of warm Pacific surface water into the Indian Ocean.

Karimata Strait. An alternative explanation for the profile of transport through Makassar Strait is the circulation of water from the South China Sea through the Karimata Strait, known as the South China Sea Throughflow (SCST). Qu et al. (2005) suggest that Pacific water travels between Taiwan and the Philippines through the Luzon Strait to the South China Sea, through Karimata Strait into the Java Sea, then northward through Makassar Strait to return to the Pacific Ocean. This process is a direct response to the winds in the western Pacific Ocean and therefore contains a strong ENSO signal. With the inclusion of the SCST, the ITF through Makassar Strait is a combination of two circulations rotating in the same direction. Both are forced by the Pacific wind regime: Makassar contains a thermocline flow around Australia while the SCST contains a surface flow around the Philippines. Modelling results suggest that without this flow, the vertical profile of the ITF through Makassar Strait would be constant through the thermocline and surface layers, leading to a higher TWT and heat transport (Qu et al. 2005).

By modelling the ITF with and without the SCST, Tozuka et al. (2007) find that the SCST lowers the volume transport of ITF through Makassar Strait by 1.5 Sv and decreases its TWT by 2.1 °C. Other studies (Gordon 2005) suggest that the main impact of the SCST on the ITF is the addition of freshwater to the Indonesian Seas, which changes the hydrographic characteristics of the outflow water without necessarily modulating the inflow water. Koch-Larrouy et al. (2008) demonstrated the importance of the SCST to the salinity budget of the Indonesian Seas and the Indian Ocean despite its small mass contribution.

Fig. 6. Mean velocity profiles during the INSTANT observational period of (a) Makassar Strait (Gordon et al. 2008), (b) Lifamatola Passage (van Aken et al. 2009), (c) Lombok Strait, (d) Ombai Strait and (e) Timor Passage (Sprintall et al. 2009).
Lifamatola passage. While Makassar Strait provides most of the volume transport of the ITF, it cannot ventilate the deep Banda Sea. The deep ITF transport travels an eastern route through the Maluku Sea to Lifamatola Passage, which has a sill depth near 2000 m (van Aken et al. 1988). This cold water sinks to depths of 5000 m, and then mixes and upwells to near 1000 m (van Aken et al. 1991). Water mass analysis from WOCE data has shown that the water in the deep Banda Sea is derived from saltier South Pacific water. Those data also suggest an intermediate ITF of 3 to 7 Sv (Talley & Sprintall 2005).

Observational data from INSTANT were gathered using moorings located slightly downstream of the Lifamatola sill for 34 months and analysed by van Aken et al. (2009). Due to instrument error and unexpectedly strong tides, there were difficulties with data collection, particularly during the first half of the observational period. From January 2004 through July 2005, uninterrupted current data are available from 1000 to 1500 m and from July 2005 through April 2006, they are available from 1000 to 2000 m. Limited velocity data are available from 1000 to 300 m during the second half of the observational period.

Most of the variability in the currents comes from tidal forcing, showing fortnightly variability. The average flow (Fig. 6, second panel) is near zero or towards the Pacific Ocean at c. 0.1 m/s in the upper part of the water column, from 250 m to 1250 m (data shallower than 250 m are not available). Below 1250 m, the flow is directed towards the Indian Ocean at velocities increasing to 0.7 m/s near 1950 m. Velocities decrease below that depth to c. 0.5 m/s at 2000 m, probably due...
to frictional interaction with the sea floor. As with Makassar Strait, transport was calculated by multiplying the velocity by the depth-dependent cross section of the channel, which decreases from 36 km at 1000 m depth width to only 10 km wide at 1750 m. The resulting transport ranges from 0.8 Sv towards the Pacific Ocean to 5.2 Sv towards the Indian Ocean, with an average of 2.7 Sv towards the Indian Ocean (Fig. 7, second panel). This is comprised of a 2.5 Sv flow towards the Indian Ocean below 1250 m, and a return flow of c. 0.9–1.3 Sv towards the Pacific Ocean above 1250 m.

The TWT of the flow through the Lifamatola Passage below 1250 m is 3.2 °C. It is expected to be much lower than the TWT of the flow through Makassar Strait due to the much greater depth, and therefore the colder temperature, of the flow. Combining the TWTs of Makassar Strait and Lifamatola Passage during non-contemporaneous measurements suggests an overall TWT of 12.2 °C.

**Lombok Strait.** Lombok Strait is the westernmost path through which ITF water can enter the Indian Ocean. It is 35 km wide and 300 m deep, in contrast with the numerous nearby straits which are shallower than 50 m and thus do not appreciably contribute to the ITF. Previous observational data suggested a total transport of 1.7 Sv (Murray & Arief 1988), with brief periods of northward transport. This was confirmed in a 1996–1997 study using shallow pressure gauges, which found an upper layer (0 to 100 m) that was primarily directed towards the Indian Ocean but periodically reversed and flowed towards the Pacific Ocean. A deeper layer (100 m to 200 m) consistently transported water to the Indian Ocean, with no flow beneath that. The average transport from that study was 2.6 Sv (Hautala et al. 2001).

During INSTANT, two moorings were placed in the strait. The eastern mooring was at a depth of 1144 m and remained in place from January 2004 through December 2006, and the western mooring was at a depth of 921 m and remained in place from January 2004 through June 2005. The results are analysed by Sprintall et al. (2009). The western side of Lombok Strait showed slightly higher velocities than did the eastern side, but the two were highly positively correlated. The flow through Lombok Strait (Fig. 6, third panel) is primarily surface intensified with a maximum at 50 to 80 m with velocities of c. 0.6 m/s. Velocity decreases quickly to the western sidewall and more slowly to the eastern sidewall and weakens considerably below 150 m. The flow was towards the Indian Ocean, except for weak surface reversals during the winter monsoon. The maximum surface velocity occurs in June or July, with a subsurface velocity maximum in August. The average transport (Fig. 7, third panel) is 2.6 Sv, with a TWT of 21.5 °C. This very warm outflow temperature is due to the shallow sill in Lombok Strait and the surface intensified flow. The TWT is at a maximum during the summer monsoon and at a minimum during the winter monsoon, when the surface flow is reversed.

**Ombai Strait.** Ombai Strait, 35 km wide and 3250 m deep, is located to the east of Lombok Strait. An average transport of 5 Sv was calculated from a one-year current mooring during JADE (Molcard et al. 2001). That study found an upper layer flow in the top 200 m with mean velocities of 0.6 m/s and some as high as 1.5 m/s. There is a deeper layer flow from 200 to 400 m of 0.2 m/s, and the flow decreases to nearly zero by 1200 m. The transport maximum was found during the summer monsoon and minimum during the winter monsoon, which included a reversal in the surface flow.

Two moorings were placed in Ombai Strait for the INSTANT observational period and were analysed by Sprintall et al. (2009). The northern mooring was located at a depth of 1329 m and was in place from January 2004 through December 2006, while the southern mooring was at a depth of 3224 m from August 2003 through December 2006 (the same location as the Molcard et al. (2001) mooring described in the previous paragraph). Although these sites are less than 15 km apart, the southern mooring recorded stronger flow. Despite its sill depth of 1450 m, the outflow of Ombai Strait is limited by the c. 900 m Sumba Strait and the c. 1150 Savu/Dao Strait.

Ombai Strait (Fig. 6, fourth panel) shows two distinct velocity maxima of c. 0.45 m/s, one near the surface and another near 150 m, in the thermocline, both located slightly towards the western side of the channel. In the subsurface flow is directed away from the Indian Ocean with flows of 0.1 to nearly 0.2 m/s which extend down to 80 m. The flow in the top 300 m of Ombai Strait displays variability similar to that of Lombok Strait, with weak surface reversals during the winter monsoon and velocity maxima during the summer monsoon. Below 300 m, the flow reverses towards the Indian Ocean semi-annually, during the transitions between monsoons. Transport through Ombai Strait (Fig. 7, fourth panel) has a total average transport of 4.9 Sv, with maxima in August and February and minima in May and November. The average TWT is 15.2 °C, but this value increases during the semi-annual flow reversal and decreases when all of the transport is directed towards the Indian Ocean.

**Timor passage.** The widest of the outflow channels, Timor Passage is located the furthest south and transports the largest amount of ITF outflow
water. It is 160 km wide, with depths of 1250 m in the east to 1890 m in the west. Two current meters were deployed there during JADE and a mean transport of 4.3 Sv was found, more than half of it in the upper 200 m (Molcard et al. 1996). Variability was found on annual, semi-annual, and monthly timescales, with a maximum in spring and summer and minimum in winter.

Timor Passage was monitored by four moorings during INSTANT, again analysed by Sprintall et al. (2009). From NW to SE, the moorings were located at: 741 m depth (from January 2004 to December 2006), 1890 m depth (from December 2003 to December 2006), 1386 m depth (from December 2003 to December 2006), and 902 m depth (from December 2003 to December 2006). Flow through Timor passage (Fig. 6, bottom panel) is nearly always towards the Indian Ocean, with reversals at the northernmost mooring suggesting the presence of rotating eddies. The maximum velocity is located at the surface with flows of 0.5 m/s. Although most of the flow is found in the upper 200 m, there is a weak secondary core of increased velocity near 1200 m, with speeds up to 0.1 m/s.

Transport within Timor Passage (Fig. 7, bottom panel) does not co-vary with Lombok and Ombai Straits. Reversals are seen during the monsoon transitional periods in the flow between c. 800 to c. 1800 m, with the strongest reversals below 1400 m. The mean transport is 7.5 Sv, or half of the total outflow, with variability dominated by the annual cycle. The large difference in transport between JADE and INSTANT is due to a difference in the width of the strait used to calculate transport. The earlier study assumed that the ITF encompassed 85 km of channel width, but the INSTANT measurements, using four moorings instead of two, showed that the ITF has significant velocity through the full 160 km width of the channel. The associated TWT is 17.8 °C, which is warmer than Ombai Strait but colder than Lombok Strait. The total outflow is dominated by Timor Passage transport. During INSTANT, the total outflow value was 15 Sv, which is larger than previous estimates. Most of that transport was confined to the upper 300 m.

Conclusions and future work

The ITF has now been considered on global, local, and regional scales. It sits at the intersection of ENSO, IOD, and the SE Asian monsoon and provides a conduit from the Pacific Ocean to the Indian Ocean. Broadly speaking, the ITF is controlled by a pressure gradient in the thermocline and wind forcing in the surface layer, with a deep transport presumably driven by pressure as well. But the existence of the pressure gradient between the Pacific and Indian Oceans can be seen to be the result of a geographical configuration of landmasses in which the Indian Ocean lacks a northern basin to export heat. Luyten et al. (1983) presented a theoretical model of circulation proposing that the ocean can be divided into multiple layers of constant density. The uppermost layer, located in the tropics is driven directly by Ekman pumping, while the deeper layers interact with the atmosphere only where they outcrop at higher latitudes. The Indian Ocean is therefore ventilated by the northern Pacific Ocean since it has no northern basin of its own. The Indian Ocean remains at a steady temperature because its high evaporation rate leads to cooling through latent heat loss.

Questions remain about the ITF on all of these scales. The INSTANT program and the growing availability of high quality observational data will allow models with realistic values for the mass and heat transport of the ITF to answer many of these questions. They will also allow for a comprehensive understanding of the flow of heat, freshwater, and mass between ocean basins. Within the observational data, however, there remains significant uncertainty about the shallow flow through Lifamatola Strait and the contribution and importance of the SCST to the ITF. Although it is understood how the monsoon and ENSO affect the ITF, it is still not known if there are feedbacks by which the ITF can influence those phenomena. A growing appreciation of the IOD as an independent climate phenomenon offers another avenue of research, as the relationship between the IOD and the ITF is not well documented. A deeper understanding of the ITF can be expected to provide the basis for better regional forecasts and climate prediction. It will also contribute to our understanding of the ITF in previous geological and climatic conditions.

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References


