The role of precipitation and salinity effect in multi-decadal changes and long-term trends of the Indonesian Throughflow

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Abstract

Multi-decadal variability of the Indonesian Throughflow (ITF) is crucial for the Indo-Pacific and global climate due to significant inter-basin exchanges of heat and freshwater. Previous studies suggest that both wind and buoyancy forcing may drive ITF variability, but the role of precipitation and salinity effect in the variability of ITF on multi-decadal time scales remains largely unexplored. Here, we investigate the multi-decadal changes and long-term trend in the ITF transport during the past six decades, with a focus on the role of precipitation and halosteric contribution. The diverse datasets consistently indicate a substantial upward trend in the halosteric component of geostrophic transport of ITF in the outflow region at 114°E during the six decades. We find that the meridional differences of salinity trend in the outflow region explain the increase of the halosteric component of ITF transport. On a larger scale, the tropical western Pacific Ocean and Indonesian seas have experienced significant freshening, which has strengthened the Indo-Pacific pressure gradient and thus enhanced the ITF. In contrast, the equatorial trade wind in the western Pacific Ocean has weakened over recent decades, implying that changes in wind forcing have contributed to weakening the ITF. The combined effect of strengthened halosteric and weakened thermosteric components has resulted in a weak strengthening for the total ITF with large uncertainties. Although both the thermosteric and halosteric components are associated with natural climate modes, our results suggest that the importance of salinity effect is likely increasing given the enhanced water cycle under global warming.

Keywords: Salinity effect, multi-decadal variability, precipitation variability, Indonesian Throughflow

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1. Introduction

The Indonesian Throughflow (ITF) represents a distinctive tropical conduit within the global circulation system, facilitating the interconnection between the Pacific Ocean and the Indian Ocean (Sprintall et al., 2014; Hu et al., 2015; Feng et al., 2018; Ummenhofer et al., 2021). It traverses the Maritime Continent and transports heat and freshwater from the Pacific Ocean to the Indian Ocean (Hu and Sprintall, 2017; Fan et al., 2018; Phillips et al., 2021). Therefore, the ITF plays a crucial role in the Indo-Pacific regional air-sea exchange and the global climate system (Wijffels et al., 2008; Feng et al., 2010; Lee et al., 2015; Liu et al., 2016; Hu et al., 2019; Sprintall et al., 2019; Gunn et al., 2022; McMonigal et al., 2022; Santoso et al., 2022).

The large-scale Indo-Pacific pressure gradient governs the ITF and is primarily influenced by wind forcing on multiple time scales (Wyrtki, 1987; Wijffels and Meyers, 2003; Meng et al., 2004; Feng et al., 2011; Li et al., 2018; Shilimkar et al., 2022; Shi et al., 2023). The role of wind forcing in the ITF has been extensively studied through various methods, including observations (Gordon and Fine, 1996; Sprintall et al., 2009, 2014; Gordon et al., 2019; Li et al., 2020), numerical models (Lebedev and Yaremchuk, 2000; Talley and Sprintall, 2005; van Sebille et al., 2014; Feng et al., 2016, 2017; Wei et al., 2016; Ivanova et al., 2021; Eabry et al., 2022), as well as a deep-learning approach (Xin et al., 2023).

Godfrey (1989) proposed the “Island Rule”, which estimates the ITF transport by integrating wind stress around Australia and the South Pacific. The ITF transport, calculated using the Island Rule, is approximately $-16 \pm 4$ Sv (Wajsowicz, 1993; Meng et al., 2004; Liu et al., 2009), which is consistent with the observed value of $-15$ Sv during the International Nusantara Stratification and Transport (INSTANT) program from 2004 to 2006 (Sprintall et al., 2009) and that determined using the Indo-Pacific pressure difference on a decadal time scale (Tillinger and Gordon, 2009). Recently, Shi et al. (2023) proposed a multiple-island rule that considers the intricate geometry within the Indo-Pacific Maritime Continent, obtaining an estimated ITF strength of approximately $14.5$ Sv.

The seasonal cycle of the ITF is closely linked to the regional monsoon winds and is stronger during the southeast monsoon and weaker during the northwest monsoon (Wijffels and Meyers, 2003; Sprintall et al., 2009; Gordon et al., 2010; Susanto et al., 2012; Peña-Molino et al., 2022).
Lu et al. (2023) recently found that the freshwater input and salinity effect related to monsoon is an import mechanism that controls the seasonality of Makassar Strait throughflow. On the interannual time scale, both the El Niño-Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) affect the surface wind stress that subsequently impacts the ITF transport (Potemra and Schneider, 2007; Sprintall and Révelard, 2014; Liu et al., 2015; Hu and Sprintall, 2016; Li et al., 2020; Susanto et al., 2022). On the decadal time scale, when the Pacific Decadal Oscillation (PDO) turned to a positive phase in 1976/77, the tropical Pacific trade winds weakened and the ITF tended to diminish (Wainwright et al., 2008; Liu et al., 2009; Feng et al., 2010). The weakened easterly trades turned to be intensified in 1993, and the ITF subsequently increased due to the stronger Indo-Pacific pressure gradient when the Interdecadal Pacific Oscillation (IPO) entered a negative phase (Feng et al., 2011, 2016; Zhuang et al., 2013).

Although often closely associated with wind forcing, salinity also plays a crucial role in the variability of ITF. The Indonesian archipelago is a region characterized by substantial precipitation and vigorous tidal mixing (Lebedev and Yaremchuk, 2000; Atmadipoera et al., 2009; Susanto et al., 2022). Low salinity water from the Pacific Ocean becomes fresher here and feeds the tropical eastern Indian Ocean, piling up in the ITF outflow region and forming a downstream buoyancy pool (DBP), where buoyancy forcing modulates the pressure gradient (Andersson and Stigebrandt, 2005; Hu and Sprintall, 2016). On seasonal time scales, Gordon (2003) introduced the concept of the “freshwater plug” to elucidate the influence of freshwater input from the Karimata Strait during monsoon seasons, which generates a northward pressure gradient in the Makassar Strait and subsequently weakens the ITF during boreal winter (Gordon et al., 2003). Winter-spring freshening due to precipitation in the Maritime Continent also weakens the pressure gradient and the ITF (Lee et al., 2019), and the freshwater input through the Karimata Strait is equivalent to 42% of the rainfall in the Maritime Continent in boreal winter (Xu et al., 2021). On interannual time scales, during El Niño events, low salinity water transport through the Karimata Strait is reduced due to weakened South China Sea circulation (Zu et al., 2018, 2019), while buoyant water through the Sulu Sea blocks western Pacific water from the northern Makassar Strait (Gordon et al., 2012). Hu and Sprintall (2016) found that the halosteric component explains (36±7)% of the total interannual variability of the ITF. The enhanced rainfall in the Maritime Continent and the salinity effect led to an intensified ITF over 2004–2014 (Hu...
and Sprintall, 2017), and a recent study also confirmed that the strengthening of the ITF over 1993-2018 is primarily due to salinity effects (Guo et al., 2023).

In the context of global warming and natural decadal variability, the global mean ocean circulation has been accelerated during the past decades (Hu et al., 2020), and the Indo-Pacific salinity exhibits significant multi-decadal fluctuations and trends, which can exert a crucial influence on the ITF (e.g., Hu et al., 2019). Previous studies suggest that, during the first decade of the 2000s, sea surface salinity in the southeastern Indian Ocean decreased (Du et al., 2015), and approximately 40% of sea level rise in the south Indian Ocean was attributed to the halosteric component (Jyoti et al., 2019). Climate models predict a weakening trend of the ITF by 2100 under global warming (e.g., Sen Gupta et al., 2016; Ma et al., 2020; Sun et al., 2020), but previous studies have also indicated that the climate models overestimate the effect of global warming scenarios (Hausfather et al., 2022; Voosen, 2022). Insufficient observations and a lack of rigorous inter-dataset error comparison pose challenges for investigating the multi-decadal variability and long-term trends of the ITF transport. Given the context of global warming and increased precipitation in the Maritime Continent (Iskandar et al., 2020), uncertainties persist regarding the influence of temperature and salinity variability on the ITF.

This study aims to investigate the impact of multi-decadal changes of precipitation in the Maritime Continent and salinity in the Indonesian seas on the multi-decadal changes of the ITF. The remaining sections of the manuscript are organized as follows. Data and methods used to estimate the ITF transport will be described in Section 2. In Section 3, we will investigate the multi-decadal variability and long-term trend of ITF transport, and examine the role of the halosteric component, precipitation, and the effect on salinity. Finally, Section 4 summarizes the main conclusions and discusses the remaining issues.

2. Data and methods

a. Data

Temperature and salinity datasets spanning from 1958 to 2017 were used to calculate the ITF transport, including an observation-based climatological mean dataset as well as five monthly products.
The World Ocean Atlas 2018 (WOA18, Garcia et al., 2019) includes six climatological mean temperature and salinity fields analyzed objectively every ten years for the period 1955–2017 (1955–1964, 1965–1974, 1975–1984, 1985–1994, 1995–2004, 2005–2017), with a horizontal resolution of 0.25°×0.25° and 31 layers in the upper 1100 m. The five monthly products spanning 1958–2017 consist of four observation-based products: (a) Ocean salinity and temperature gridded product (version 1.0) from Institute of Atmospheric Physics (IAP) with a horizontal resolution of 1°×1° and 33 layers in the upper 1100 m (Cheng et al., 2020); (b) an objective analysis product provided by Ishii (version 7.3.1) with a horizontal resolution of 1°×1° and 21 layers in the upper 1100 m (Ishii et al., 2017); (c) European Centre for Medium-Range Weather Forecasts Ocean Reanalysis System 5 (ECMWF-ORAS5) with a horizontal resolution of 0.25°×0.25° and 48 layers in the upper 1100 m (Zuo et al., 2019); (d) the Simple Ocean Data Assimilation (SODA, version 3.15.2) with a horizontal resolution of 0.5°×0.5° and 29 layers in the upper 1100 m (Carton et al., 2018); and a high-resolution numerical model output: (e) the OGCM (oceanic general circulation models) of Earth Simulator (OFES) model output provided by the Applications Laboratory (APL) with a horizontal resolution of 0.1°×0.1° and 36 layers in the upper 1100 m (Sasaki et al., 2008).

The 0.25°×0.25° mapped evaporation (E) and precipitation (P) over the period 1958–2017 from the ECMWF Reanalysis v5 (ERA5) (Hersbach et al., 2020) are utilized to estimate the “evaporation minus precipitation (E-P)” flux, and the 0.25°×0.25° mapped ERA5 sea surface wind over the period 1958–2017 is used to investigate the influence of the large-scale wind forcing.

b. Method

As defined in previous studies (Wainwright et al., 2008; Feng et al., 2010; Liu et al., 2015; Hu et al., 2021), the ITF volume transport in this paper is the zonal geostrophic transport across the 114°E (8.5°S–22.5°S) section located in the ITF outflow region and bounded by Java and Australia (Fig. 1). The zonal geostrophic velocity \( V \) perpendicular to the 114°E section between stations \( A \) and \( B \) from isobaric surfaces \( P_1 \) and \( P_2 \) is given by:

\[
V = \frac{1}{fL} \left( \int_{P_1B}^{P_2B} \alpha_B dp - \int_{P_1A}^{P_2A} \alpha_A dp \right)
\]  

(01)
where $f$ is the Coriolis parameter, $L$ is the horizontal distance between station $A$ and $B$, and $\alpha$ is the specific volume anomaly, which is a function of ocean salinity ($S$, psu), temperature ($T$, °C) and pressure ($p$, dbar):

$$\alpha = \alpha(S, T, p)$$  \hspace{1cm} (02)

Depth across the 114°E section is mostly less than 3000 m, and velocity beneath 1100 m is near zero (Wijffels et al., 2008; Sprintall et al., 2009). The 1100 m surface is thus designated as the reference depth of no flow, while stations in water depths shallower than 1100 m are referenced to the bottom levels, and hence the velocity of each layer can be obtained through Eq. (1).

The respective contributions of temperature and salinity to the total geostrophic transport are separated by isolating the thermosteric and halosteric components (Feng et al., 2015; Hu and Sprintall, 2016; Hu et al., 2021). The thermosteric ITF($\bar{S}, T$) component is estimated using the climatological mean salinity ($\bar{S}$) and time-dependent temperature time series ($T$), and the halosteric ITF($S, \bar{T}$) component is estimated using climatological mean temperature ($\bar{T}$) and time-dependent salinity time series ($S$). The ITF variability is thus divided into the thermosteric component and the halosteric component, with a negligible residual portion corresponding to the coupled effect (see the supplement materials and Appendix A of Hu and Sprintall, 2016). To remove the seasonal cycle and ENSO-associated cycle of 2–7 years, the ITF transport time series derived from the monthly products are filtered using a 7-year fourth-order Butterworth filter.

Contributions ($P_i$) of the thermosteric and halosteric ITF components to the multi-decadal variations of the total geostrophic ITF transport are calculated following Hu and Sprintall (2016):

$$P_i = \frac{|ITF'_i|}{|ITF'_T| + |ITF'_S| + |ITF'_R|}$$  \hspace{1cm} (03)

where $i$ is $T$ or $S$, $ITF'_T$ is the thermosteric ITF transport anomaly, $ITF'_S$ is the halosteric ITF transport anomaly, and $ITF'_R$ is the residual term ($ITF' - ITF'_T - ITF'_S$).

The net heat transport ($HT_{ITF}$) to the Indian Ocean across the 114°E section induced by the ITF is calculated by:

$$HT_{ITF} = \int \int \rho C_p U_{114} (T_{114} - T_{IO}) dy dz$$  \hspace{1cm} (04)

where $\rho$, $C_p$, $U_{114}$, and $T_{114}$ are density, the specific heat capacity of seawater, zonal geostrophic velocity, and temperature of each grid throughout the 114°E section, respectively, and $T_{IO}$ is the
spatially mean temperature of the Indian Ocean. The $\rho$, $C_p$, $T_{114}$, $U_{114}$, and $T_{IO}$ are all time-dependent to investigate the contribution of ITF to the heat content of the Indian Ocean. The $\rho$ and $C_p$ are calculated using the UNESCO polynomial (Fofonoff, et al., 1983).

The net freshwater transport ($FWT_{ITF}$) to the Indian Ocean across the 114°E section induced by the ITF is calculated by:

$$FWT_{ITF} = \iint U_{114} \frac{S_{10} - S_{114}}{S_{IO}} dydz$$  \hspace{1cm} (05)

where $S_{114}$ is salinity of each grid throughout the 114°E section, and $S_{IO}$ is the spatially mean salinity of the Indian Ocean. The $U_{114}$, $S_{114}$, and $S_{IO}$ are all time-dependent to indicate the net freshwater transport via ITF.

**Fig. 1.** Topography (shaded, unit in km), climatological mean geostrophic currents in the upper 100 m from ORAS5 (black arrows, unit in m/s), and location of the 114°E section (8.5–22.5°S, blue dashed line). Streamlines of the climatological mean geostrophic currents in the upper 100 m from ORAS5 are shown with red lines.

### 3. Results

#### a. Decadal ITF variability

The climatological mean geostrophic and the direct zonal velocities across the 114°E section from ORAS5, SODA, and OFES show similar structures (Figs. 2 a–f), which indicates that the geostrophic current is the major component of ITF and is consistent with previous studies.
(Wijffels et al., 2008; Wainwright et al., 2008). Although the geostrophic transport across the 114°E section is a bit weaker than that of the total zonal transport, the transport time series show a good agreement at annual and decadal time scales in all three products (Figs. 2 g–i). The geostrophic transport is therefore capable of accounting for the ITF low-frequency characteristics and thus can be used to investigate the multi-decadal variability.

**Fig. 2.** Geostrophic velocity, zonal ocean current velocity, and volume transport across the 114°E section of (a, d, g) ORAS5, (b, e, h) SODA and (c, f, i) OFES over 1958–2017. The 7-year low-pass (thick lines) and 13-month running mean (thin lines) geostrophic ITF transports (black) and direct calculated ITF transports (gray) are exhibited in panels g, h, and i. The correlation coefficients between the 7-year low-passed geostrophic and directly calculated ITF transports (shown in panels g, h, and i) are significant at the 95% confidence level. Positive values represent eastward currents and transports. Note different scales in the depth axis below 300 m in panels a–f.

The ITF geostrophic velocity and volume transport of the 114°E section are shown in Fig. 3. The velocity structures of the WOA18 and the ensemble mean of the five monthly datasets are quite similar (Figs. 3a, c) and consistent with previous studies (Wijffels et al., 2008; Wainwright et al., 2008), and large standard deviations of velocity are concentrated in the upper 300 m layers. Zonal geostrophic currents across the 114°E section are mostly westward, with a westward core at 10–12°S corresponding to the northern part of the South Equatorial Current (SEC), reaching 1000 m depth with a maximum of ~ -0.3 m/s. The eastward current in the upper 200 m at ~14–21°S is the
East Gyral Current (EGC) with a geostrophic velocity maximum of $\sim 0.1 \text{ m/s}$. The westward current beneath EGC is the southern part of the SEC, and the westward current close to Australia is the Leeuwin Current (LC).

**Fig. 3.** Geostrophic velocity and ITF transports across the 114°E section over the period 1958–2017: (a, b) WOA18 and (c, d) the ensemble mean of IAP, Ishii, ORAS5, SODA, and OFES. Black contours in panels a and c indicate standard deviations of velocity, and shaded colors in d indicate standard deviations calculated using the five products. Solid lines in panels b and d indicate the total ITF transport (black, unit in Sv), thermosteric ITF transport (red), halosteric ITF transport (blue), and dashed lines indicate the long-term linear trend of the total ITF transport (black), halosteric ITF transport (blue) and thermosteric ITF transport (red). Positive values represent eastward currents and transports. Note the change in depth scale below 300 m in panels a and c.

The climatological mean geostrophic transport is approximately -15 Sv from the WOA18 (Fig. 3b), and (-9.1 ± 1.5) Sv from the ensemble mean in the upper 1100 m over the period 1958–2017 (Fig. 3d). Geostrophic volume transport across the IX1 expendable bathythermograph (XBT) section over the depth range 0–700 m is $\sim 8.6 \text{ Sv}$ during the period 1984–2013 (Liu et al., 2015) and (-8.2 ± 0.2) Sv during the period 1993–2018 (Guo et al., 2023), slightly weaker than the geostrophic ITF transport of the ensemble mean over the depth range 0–1100 m. The ITF transport of the WOA18 and the 7-year low-pass ITF transport of the ensemble mean show similar decadal
variability consistent with that from previous studies, with a reversal to a weakened ITF in 1976/77, and an intensified trend since the mid-1990s (Zhuang et al., 2013; Liu et al., 2015; Feng et al., 2016, 2017). All trends mentioned in this paper pass the Mann-Kendall (M-K) test at the 95% confidence level unless otherwise specified.

The ITF transport from the WOA18 shows a long-term trend of -0.8 Sv/decade (negative anomalies indicate intensified ITF) over 1955–2017 (Fig. 3b). However, the ensemble mean ITF shows a weak but significant trend of -0.11 Sv/decade over 1958–2017 (Fig. 3d). The difference of ITF trends between the WOA18 and the ensemble mean indicates that there may large uncertainties in the estimating of long-term trend of ITF transport. The thermosteric ITF transport of the WOA18 shows no significant trend (Fig. 3b), while the ensemble mean shows a weakening trend of (1.1 ± 0.21) Sv/decade (Fig. 3d). The halosteric ITF transport of the WOA18 shows a strengthening trend of -0.77 Sv/decade (Fig. 3b), while the strengthening trend of the ensemble mean is stronger at (-1.2 ± 0.18) Sv/decade (Fig. 3d). It is notable that the thermosteric trend of the ensemble mean rises sharply from 0.9 Sv/decade over 1958–2010 to 6.9 Sv/decade over 2011–2017. Similarly, the halosteric trend of the ensemble mean increases from -1.1 Sv/decade over 1958–2010 to -2.7 Sv/decade over 2011–2017. Together, these trends cause an increased ITF of -0.2 Sv/decade over 1958–2010 reversing to a decreased ITF of 4.2 Sv/decade over 2011–2017 in the ensemble mean. In addition, the ITF time series of the ensemble mean (Fig. 3d) spanning 1958–2017 begin with a sharp increase and end with a sharp decrease, which might lead to an underestimation of the long-term intensified trend.

The five monthly products indicate different multi-decadal variability, especially over 2000–2010 when SODA and IAP are significantly out of phase with the other products, and the long-term trends are different: they are intensified in the IAP, ORAS5, and SODA, but weakened in the Ishii and OFES data sets (Fig. 4a). Although the range of transport anomalies are close (Fig. 4d), the inter-dataset difference makes it difficult to confirm a significant trend in the ensemble mean like that found in the WOA18. However, all five monthly datasets indicate a weakened thermosteric ITF (Fig. 4b) and an intensified halosteric ITF (Fig. 4c) during the period 1958–2017.
Fig. 4. The (a) total, (b) thermosteric, (c) halosteric, and (d) product distribution of the geostrophic ITF transport from IAP, Ishii, ORAS5, SODA, and OFES across the 114°E section over the period 1958–2017. Shaded colors in panels a–c indicate standard deviations computed over the five products. Transports are filtered using a 7-year fourth-order Butterworth filter. In panel d, the central mark (red) indicates the median, the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively, and the whiskers extend to the most extreme data points. Positive values represent eastward currents and transports.

The halosteric component contributes 50% of the total ITF multi-decadal variability in the WOA18, and (44.1±15.6) % in the ensemble mean (Fig. 5a). Overall, although the thermosteric contributions are stronger than the halosteric contributions of the ensemble mean across the 114°E section, the halosteric contribution of the ensemble mean is comparable with the thermosteric contributions within error bars, and exhibit a similar pattern (Figs. 5b, c). The halosteric contribution is less than 40% in the upper 100 m, but increases with depth and shows two high-value regions roughly corresponding to the northern and southern parts of the SEC: one broad strong region of ~50% over the depth below 200 m at the coast of Java and another wider region of more than 40% at 14–18°S below 300 m depth.
Fig. 5. (a) Contributions to the total ITF variability of the ensemble mean averaged over the 114°E section and for (b) the thermosteric and (c) the halosteric contributions of the ensemble mean over the period 1958–2017. Note the change in depth scale below 300 m in panels b and c.

b. Intensified halosteric component and the enhanced salinity effect

Zonal geostrophic currents across the 114°E section can be modulated by the meridional dynamic height (DH) difference associated with both the temperature and the salinity. Fig. 6 shows the temperature and salinity trends across the 114°E section over 1958–2017. Temperature and salinity trends of the WOA18 rarely pass the 95% confidence level with low degrees of freedom (less than 6), however, their patterns are somewhat similar to the ensemble mean (Fig 6). In WOA18, the temperature trend is mostly a two-mode vertical pattern, with surface warming in the upper 50 m and subsurface cooling below that to 800 m (Fig. 6b). The surface warming and subsurface cooling in the upper 200 m are significant at 10–13°S, where the northern part of the SEC is located (Fig. 6b). The temperature trend of the ensemble mean shows a similar two-mode
vertical structure, but with stronger surface warming reaching deeper to 100 m, and stronger subsurface cooling extending shallower to ~500 m (Fig. 6d). Overall, the long-term trends of the ensemble mean are qualitatively consistent with that in the WOA18 over the upper 500 m layers. Notably, the WOA18 is saltier at 18°–20°S within the depth range 200–300 m (Fig. 6a), where the ensemble mean shows freshening (Fig. 6c). The surface warming of the ensemble mean is deeper than in the WOA18, with a significant cooling of WOA18 at 9°S and at 150 m which is in stark contrast to the warming shown in the ensemble mean.

The different patterns of the temperature and salinity trends influence the pattern of the DH trend. The DH trend of the WOA18 is mainly concentrated in the upper 600 m, with a prominent meridional difference associated with both the salinity and the temperature trends (Figs. 6a, b). The meridional DH pattern of the WOA18 is characterized by two weakening cores on either side of the significant freshening core at 13°S (Fig. 6a), while the increasing DH at 20°S is consistent with the warming in the upper 200 m (Fig. 6b). Subsequently, the meridional pressure gradient and the associated zonal currents of the WOA18 would be intensified. The DH trend of the ensemble mean shows a similar structure with two weakening cores, which are a result of both salinity and temperature trends at 12°S and 16°–19°S in the upper 200 m (Figs 6c, d). The difference in the meridional DH trends across the section would enhance the meridional pressure gradient which leads to intensified zonal currents.
**Fig. 6.** Salinity and temperature trends across the 114°E section over the period 1958–2017: (a, b) WOA18; (c, d) the ensemble mean (the five datasets show similar trends). Shaded colors indicate the S trend (a, c, unit in 10⁻² psu/10 yrs) and T trend (b, d, unit in 10⁻¹ °C/10 yrs), contour lines (black) indicate the dynamic height trend (unit in cm/10yrs). Trends significant at the 95% confidence level are hatched. Note the change in depth scale below 300 m.

c. Basin-scale salinity trend

Both temperature and salinity variability along 114°E are modulated by the Indo-Pacific heat and freshwater exchange (Du et al., 2023), and have great influences on the zonal currents (Ummenhofer et al., 2021). The basin-scale patterns of the salinity and temperature trends of the WOA18 are quite similar to that of the ensemble mean in the upper 400 m over the period 1958–2017. The basin-scale temperature trend in the upper 400 m could be roughly characterized as warming in both the Pacific Ocean and the Indian Ocean, except for the cooling observed in equatorial regions (10°S–4°N) in the Pacific Ocean and south of the equator (about 17°S–3°S) in the Indian Ocean (Figs. 7a, c). The cooling regions in the upper 400 m are coincident with the ITF inflow and outflow regions, indicative of the cross-basin transport from the western Pacific Ocean to the Indian Ocean via the ITF.
Fig. 7. Linear trends of temperature and salinity. Upper 400 m mean temperature (T) and salinity (S) trends over the period 1958–2017 of (a, b) WOA18 and (c, d) the ensemble mean, and (e) temperature trend and (f) salinity trend of the Pacific Ocean (0°-20°N, 120°E–180°E) and the Indian Ocean (20°S-0°, 60°E–120°E) of the ensemble mean over the period 1958–2017. Shaded colors represent the temperature trend (a, c, e, unit in 10⁻¹°C/decade) and salinity trend (b, d, f, unit in 10⁻² psu/decade). Trends significant at the 95% confidence level are hatched in panels a–d. Boxes in panels b and d indicate the Pacific Ocean and Indian Ocean regions.

The sea surface temperature (SST) indicates a basin-scale warming trend in both the Pacific and the Indian Oceans (Fig. 7e), consistent with the warming SST over the period 1982–2012 (Ummenhofer et al., 2021). The cooling of the ensemble mean appears at 100 m in the Pacific Ocean, consistent with Hu and Sprintall (2016). However, in the ITF outflow region in the Indian Ocean, the cooling appears at 200 m with a deepening to the west (Fig. 7e). In the ITF outflow region, the surface warming deepens from the upper ~100 m at 120°E to ~300 m at 100°E, and to ~500 m at 80°E (Fig. 7e), which might lead to a weakened ITF due to higher sea level in the south-east Indian Ocean related to this warm water.
In contrast, the salinity trend in the upper 400 m illustrates significant basin-scale differences in both the WOA18 and the ensemble mean. Strong freshening occurs in the western Pacific Ocean, the Indonesian seas, and the western coast of Australia, while the central and western Indian Ocean is getting saltier (Figs. 7b, d). Consequently, the DH would increase in the freshening regions because of buoyancy forcing, and decrease in the tropical Indian Ocean due to the water column getting saltier and denser. Subsequently, the opposite basin-scale salinity trends lead to an intensified Indo-Pacific pressure gradient that would act to intensify the ITF. Thus, both the thermosteric and the halosteric components play important roles in the ITF long-term trend, and the contribution of the salinity effect appears to be increasingly important with time.

Freshening in the western Pacific Ocean reaches 50 m, deepening to 200 m within the Indonesian seas where strong tidal mixing occurs (Koch-Larrouy et al., 2015, Susanto et al., 2022), and fills the surface DBP (8°S–15°S, 100°E–120°E) at the ITF outflow region, where the 114°E section is located (Fig 7f). Freshwater from the Indonesian seas is then propagated westward by the SEC and southward by the LC, leading to freshening in the eastern Indian Ocean and along the western coast of Australia (Fig. 7f). The pressure gradient between the DBP and the adjacent Indian Ocean would drive an anticlockwise circulation in the ITF outflow region (Fig. 1) that acts to sustain the DBP (Andersson et al., 2005; Hu and Sprintall, 2016).

The net heat and freshwater transports across the 114°E section via the geostrophic current show similar multi-decadal variability (Fig. 8). The ITF maintains a continuous positive net heat transport of 0.34 PW (1PW = 10^{15}W) into the Indian Ocean with a long-term trend of 1.4 TW/decade (1TW = 10^{12}W). The freshwater transport of the ITF is approximately 102.0 mSv (1 mSv = 10^3 m^3/s) with a long-term trend of 3.6 mSv/decade under the background of global warming over 1958–2017. Errors subject to the calculation of geostrophic transport and transport-weighted temperature and salinity in estimations of heat and freshwater transport are calculated (McCarthy et al., 2015; McMonigal et al., 2022), and the error is 0.07 PW in the heat transport and 36.2 mSv in the freshwater transport.
Fig. 8. Net transports of heat and freshwater across the 114°E section to the whole Indian Ocean over 1958–2017. The 7-year low-pass transports (thick lines), 13-month running mean transports (thin lines), and long-term trends in transport (dashed lines) are shown.

d. The role of increasing precipitation in the Indonesian seas

Enhanced precipitation in the Maritime Continent was observed over the period 1982-2013 (Iskandar et al., 2020) and the strong freshening observed over the period 2002–2010 intensified the ITF (Hu and Sprintall, 2017). The basin-scale opposing salinity trend (Fig. 7f) might be associated with the increasing precipitation in the Maritime Continent, and so we investigate the relationship of salinity to the E-P flux over the period 1958–2017.
Fig. 9. Climatological mean and linear trends of (a, c) the E-P flux and (b, d) zonal wind stress over the period 1958–2017. Shaded colors indicate the climatological mean E-P flux (a, unit in cm/yr.), linear E-P flux trend (c, unit in cm/10 yrs\(^2\)), climatological mean zonal wind stress (b, unit in 10\(^{-2}\) N·m\(^{-2}\)) and linear zonal wind stress trend (d, unit in 10\(^{-2}\) N·m\(^{-2}\)/10 yrs). Arrows indicate the climatological mean wind stress in panel b and the linear trend in panel d. Trends significant at the 95% confidence level are hatched.

The climatological mean E-P flux represents a meridional pattern bounded roughly by 10°S in the Indo-Pacific region, with negative values (P>E) to the north including the western Pacific Ocean and the Indonesian seas, and positive values (P<E) to the south centering on the eastern Indian Ocean (Fig. 9a). The strong tidal mixing in the Indonesian seas propagates the surface freshening to the deep (Koch-Larrouy et al., 2015), and thus leads to a strong buoyancy forcing contributing to the Indo-Pacific pressure gradient. The long-term trend of the E-P flux (Fig. 9c) represents an enhancement of the climatological mean, with strongly negative values in the Indonesian seas and positive values in the southern Indian Ocean. That is, the wet and fresh Indonesian seas where strong mixing occurs are getting wetter and fresher over the past 60 years, while the dry and saline southern Indian Ocean is getting drier and more saline. The consistent pattern of the climatological mean and the long-term trend of the E-P flux indicates a continuous
and progressively stronger freshwater input into the Indonesian seas and accumulation in the ITF outflow region, which leads to surface freshening and increased DH (Du et al., 2015) that acts to intensify the Indo-Pacific pressure gradient. Buoyancy forcing originating from the freshening Indonesian seas thus spreads into the ITF outflow region and across the 114°E section, modulating the meridional salinity and temperature distributions and intensifying the zonal currents. Collectively, this acts to intensify the halosteric component of the ITF.

Large-scale wind forcing is considered to be the driving factor of the Indo-Pacific pressure gradient (Wyrtki, 1987; Tillinger and Gordon, 2009; Susanto et al. 2012; Feng et al., 2016; Shilimkar et al., 2022). The climatological mean trade wind in the western Pacific Ocean results in the accumulation of water and increasing sea level in the Indonesian seas, while the trade wind in the southern Indian Ocean drives water westward and so decreases the sea level in the ITF outflow region (Fig. 9b). This pattern thus illustrates the wind-forced Indo-Pacific pressure gradient that drives the ITF. However, the long-term trend indicates a different pattern from the climatological mean over the period 1958–2017 (Fig. 9d). The equatorial trade wind is weakened in both the western Pacific Ocean and the Indian Ocean, while the trade wind of the southern hemisphere is significantly enhanced. Weakened equatorial trade wind in the Pacific Ocean would reduce the accumulation of water in the Maritime Continent. Subsequently, it appears that the large-scale wind forcing would weaken the Indo-Pacific pressure gradient and lead to a weaker ITF on the multi-decadal time scale.

Correlations between precipitation and the 4-month lagged geostrophic ITF transport of the ensemble mean across the 114°E section are shown in Fig. 10a. The ITF geostrophic transport is significantly negatively correlated to precipitation in the western Pacific Ocean and the Indonesian seas, and positively correlated to precipitation in the southern Indian Ocean. The positive correlation regions in the Indian Ocean are separated by the SEC, which is fed by the ITF. The mean precipitation (Fig. 10b) of the Indonesian seas is found to lead the ITF transport by 4 months, with a significant negative correlation coefficient of -0.45. Precipitation in the Maritime Continent indicates different trends from the ITF transport over different periods. Before the “climate regime shift” in 1976/77 when the PDO turned to a positive phase, the precipitation in the Maritime Continent and the ITF increased, and from 1976/77 to the mid-1990s, the precipitation and the ITF decreased (Fig. 10b), and then, the precipitation in the Maritime
Continent significantly intensified and the ITF strengthened, with a maximum geostrophic transport occurring around 2013.

![Image](image.png)

**Fig. 10.** (a) Correlation coefficient between precipitation and 4-month lagged ITF geostrophic transport of the ensemble mean, and (b) precipitation in the Indonesian seas (blue lines, thick for 7-year low-pass and thin for 13-month running mean) and the anomalous ITF geostrophic transport (black lines, thick for 7-year low-pass and thin for 13-month running mean) over the period 1958–2017. The black box (Indonesian seas) in panel a represents the area where the regional mean precipitation for panel b is calculated. Correlation coefficients significant at the 95% confidence level are hatched in panel a. Dashed lines in panel b indicate linear trends of the ITF (black) and precipitation (blue) over 1958–1976, 1977–1993, and 1994–2013.

e. The role of the PDO in multi-decadal variability

The transitions between PDO phases modulate the large-scale distribution of the Indo-Pacific precipitation and sea surface wind field (Sprintall et al., 2019; Jyoti et al., 2019). Fig. 11 shows the composited precipitation and surface wind field during the negative and positive PDO phases over 1958–2017. Precipitation in the Indonesian seas and the western Pacific Ocean is significantly stronger during the negative PDO phases (Fig. 11a), and weaker over the positive PDO phases (Fig. 11b). The Indo-Pacific surface wind field is convergent during the negative
PDO phases and divergent over the positive PDO phases. That is, stronger buoyancy forcing due to heavy rainfall in the Indonesian seas and sea level rise because of the wind convergence would intensify the Indo-Pacific pressure gradient over the negative PDO phases and hence lead to an intensified ITF with vice-versa over the positive PDO phases. Perhaps more interestingly, the convergence wind field over the Maritime Continent could act to intensify the ascending branch of the Walker Circulation found over the Maritime Continent and subsequently lead to intensified precipitation (Fig. 11a).

![Composite precipitation and surface wind during the negative and positive PDO phases over the period 1958–2017.](image)

**Fig. 11.** Composited precipitation and surface wind during (a) the negative and (b) the positive PDO phases over the period 1958–2017. Composited precipitation significant at the 95% confidence level is hatched, and only the composited surface wind vector that is significant at the 95% confidence level is shown.

On multi-decadal time scales, the depth-integrated thermosteric ITF indicates similar variability to the total geostrophic ITF transport and is mostly intensified in the positive phase of the PDO and weakened in the negative phase, while the halosteric ITF indicates an opposite variability over the period 1958–2017 (Fig. 12a). Unsurprisingly then, the thermosteric ITF zonal velocity anomaly with depth (Fig. 12c) also indicates a very consistent pattern with that of the total geostrophic ITF velocity anomaly (Fig. 12b), while the halosteric ITF velocity indicates an opposite pattern (Fig. 12d). The intensified thermosteric component of velocity/transport has a
stronger variability than the weakened halosteric component. Velocity anomalies of both the thermosteric and the halosteric components reach 500 m and are mostly strongest at surface. Overall, the PDO seems to be the dominant factor of the multi-decadal ITF variability by modulating the thermosteric and the halosteric components.

![Fig. 12. Anomalous time series of the (a) 7-year low-pass depth-integrated total (black), thermosteric (red), and halosteric (blue) ITF transport; and the zonal velocity anomaly with depth of the (b) total geostrophic ITF, (c) thermosteric ITF and (d) halosteric ITF across the 114°E section over the period of 1958–2017. Shaded colors in panel a represent positive (red) and negative (blue) phases of the PDO.](image)

The consistency between the total geostrophic ITF and the thermosteric ITF component influenced by the PDO is likewise significant across the 114°E section, showing similar patterns correlated to the PDO (Figs. 13a, b), while the halosteric ITF velocity shows a different and roughly opposite pattern (Fig. 13c). The thermosteric component of the ITF in the upper 300 m is weakened during the negative phase of the PDO, while the halosteric component is intensified, but the correlation coefficients between PDO and the halosteric component are much smaller than that between PDO and the thermosteric component.
Fig. 13. Correlation coefficients between the PDO index and the (a) ITF geostrophic velocity, (b) thermosteric component of ITF geostrophic velocity, and (c) halosteric component of ITF geostrophic velocity across the 114°E section over the period of 1958–2017. Correlation coefficients significant at the 95% confidence level are hatched. Note the change in depth scale below 300 m.

4. Conclusions and discussion

In this paper, the ITF transport across 114°E is investigated using the climatological mean WOA18 dataset and the ensemble mean of five monthly products (IAP, Ishii, ORAS5, SODA, and OFES). Contributions of salinity and temperature are separated into the halosteric and thermosteric components of the transport. The climatological mean ITF geostrophic transport is -15 Sv in WOA18 and that of the ensemble mean is (-9.1 ± 1.5) Sv over the depth range 0–1100 m. Due to a lack of long-term continuous observations, understanding of multi-decadal variability of the ITF transport relies on data products based on different data assimilation and mapping methods, but the inter-dataset difference of the absolute mean transport is a challenge. Nonetheless, the ITF transport from the WOA18 and the other data products exhibits a consistent multi-decadal variability and agrees with previous studies (Zhuang et al., 2013; Liu et al., 2015; Feng et al., 2016, 2017). Our results indicate that the ITF transport was weakened after the “climate regime shift” in 1976/77 and then intensified after the mid-1990s until 2013 when a maximum in transport occurred.

Over the period 1958–2017, the ITF geostrophic transport shows a significant intensification trend in the WOA18, while the ensemble mean has a weaker intensification trend. The ITF transport is intensified in IAP, ORAS5, and SODA products, but weakened in Ishii and OFES. However, all data sets consistently showed that the halosteric ITF is intensified. The halosteric component contributes (44.1 ± 15.6)% of the total ITF variability in the ensemble mean and ~50%
in the WOA18. The salinity trend across the 114°E section shows significant meridional differences over the period 1958–2017, which act to intensify the meridional pressure gradient and subsequently the zonal currents. In contrast, the temperature trend indicates a vertical two-mode structure, and a meridional gradient at 18–20°S in the upper 200m. The long-term trend of the ITF is determined by the halosteric component and the thermosteric component together, but the role of the halosteric component might be increasingly important over the past few decades.

Fig. 14 shows a schematic representation of the mechanism responsible for the enhancement of the ITF due to the intensified precipitation over the Maritime Continent over the period of 1958–2017. The climatological mean and the long-term trend of the E-P flux indicate continuously increased freshwater input into the Indonesian seas, while the salinity in the southern Indian Ocean increased over the period 1958–2017. Enhanced freshwater input and strong tidal mixing led to an increase in DH in the outflow region of the ITF, while the DH in the central Indian Ocean decreased due to salinification. Subsequently, the intensified Indo-Pacific pressure gradient strengthened the ITF, leading to intensified heat and freshwater transport into the Indian Ocean. The warm and fresh water transported by the ITF from the western Pacific Ocean leads to significant westward warming and freshening in the Indian Ocean (Figs. 7e, f). The warming trend deepens from ~100 m at 120°E to ~300 m at 100°E, and to ~500 m at 80°E (Fig. 7e). The freshening to the west contains two parts, stronger surface freshening limited to the upper 100 m that disappears westward by 90°E, and a weaker freshening that deepens from ~300 m at 120°E to ~400 m at 100°E, and to ~500 m at 80°E (Fig. 7f).
Fig. 14. Schematic representation of the mechanism responsible for the enhancement of the ITF due to the intensified precipitation in the Maritime Continent over the period of 1958–2017. (a) The linear trend of precipitation (cm/yr·dec$^{-1}$), (b) linear trends of the halosteric and thermosteric DH (cm·dec$^{-1}$), and (c) linear trends of the halosteric and thermosteric ITF (cm/s·dec$^{-1}$) of the ensemble mean along 114°E.

The decadal variability is related to the PDO, as suggested by previous studies (e.g., Hu and Sprintall, 2017). We find that the ITF transport is weakened in the positive phase of the PDO while intensified in the negative phase of the PDO. Over the negative PDO phase, heavy rainfall in the Maritime Continent and the western Pacific Ocean, and convergent Indo-Pacific winds would intensify the Indo-Pacific pressure gradient and the ITF, and vice-versa over the positive PDO phase. The thermosteric ITF velocity contributes more on the multi-decadal scale to the total ITF and thus determines the phase of the ITF velocity anomaly. The halosteric ITF velocity presents opposite patterns to the thermosteric component, including velocity with depth and correlation coefficient to the PDO. Both the thermosteric and halosteric components are influenced by the PDO state, and together modulate the low-frequency ITF variability and long-term trend.

It should be noted that the long-term trend of the ITF estimated from the various data sets contains significant uncertainty, implying that it is necessary to further improve the mapping method and
sustain comprehensive in situ ocean observations. The long-term trends of the ITF estimated with WOA18 and other data products are different from climate model results. For example, previous studies suggested that climate models tend to overestimate global warming (Hausfather et al., 2022; Voosen, 2022) and thus overestimate the thermosteric component of the ITF variability and suggest a weakened ITF in the future (Feng et al., 2017; Sun et al., 2020). This study indicates that correct relative contributions of halosteric and thermosteric components might be crucial in reproducing a correct ITF transport for the climate models. Given the increasing importance of the halosteric ITF transport and salinity effect in the context of global climate warming, we need to assess the ability of climate models to better simulate the salinity-related processes.

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Data Availability: The WOA18 objectively analyzed data is available from https://www.ncei.noaa.gov/access/world-ocean-atlas-2018/. The IAP data (version 1.0) can be accessed from http://www.ocean.iap.ac.cn/. The Ishii objectively analyzed data is available from https://climate.mri-jma.go.jp/pub/ocean/ts/v7.3.1/. The ORAS5 and ERA5 reanalysis datasets can be accessed from the ECMWF’s website (https://www.ecmwf.int/en/forecasts/datasets/browse-reanalysis-datasets). The SODA reanalysis data can be downloaded from https://www2.atmos.umd.edu/~ocean/. The OFES data can be downloaded from https://www.jamstec.go.jp/ofes/.

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