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Variability in Coral-Reconstructed Sea Surface Salinity Between the Northern and Southern Lombok Strait Linked to East Asian Winter Monsoon Mean State Reversals

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Abstract The Indonesian throughflow (ITF) impacts heat and buoyancy transport from the Pacific Ocean to the Indian Ocean, influencing air-sea heat exchange and Indo-Pacific climate. Nearly 80% of the total 15 sverdrups (1 Sv = 10⁶ m³/s) of ITF water moves through the Makassar Strait in the western Indonesian seas, with ~20% of total ITF transport subsequently entering the Indian Ocean through the Lombok Strait. During the East Asian winter monsoon (EAWM), buoyant South China Sea (SCS) waters obstruct southward surface ITF transport in the Makassar Strait, likely impacting surface variability throughout the Lombok Strait. Here we present two subannually resolved, multicentury records of coral-reconstructed sea surface salinity (SSS) from the northern (110 years) and southern Lombok Strait (193 years). Differences in boreal winter (January–March) SSS variability between the two sites suggest the influence of multiple source waters. Instrumental and reconstructed temperature-salinity (T-S) relationships indicate that SCS surface waters dominate the northern Lombok Strait, while Indian Ocean surface waters instead dominate the southern Lombok Strait before 1960. These dissimilarities are likely due to changes in monsoon-driven surface water advection. At the northern site, the EAWM consistently influences SSS variability. The EAWM influence at the southern site, however, reverses in direction (inverse to direct) coincidentally with a transition from a positive (strong) to negative (weak) EAWM state in 1960. Our records collectively reveal that changes in the strength and state of the EAWM impact Lombok Strait surface water circulation, likely interacting with southward ITF transport and thus Indo-Pacific climate.

1. Introduction

The Indonesian seas provide an oceanic pathway, the Indonesian throughflow (ITF), that transports water from the tropical Pacific into the Indian Ocean, affecting heat distribution and associated ocean and climate systems (Godfrey, 1996; Gordon et al., 2010; T. Lee et al., 2002; S.-K. Lee et al., 2015; McCreary et al., 2007; Sprintall et al., 2014; Wajsowicz, 2002). Estimates indicate that ~20% of the total ITF transport exits through the Makassar Strait (Murray & Arief, 1988; Sprintall et al., 2009), with maximum southward flow occurring during boreal winter (e.g., Hualta et al., 2001; Murray & Arief, 1988; Sprintall et al., 2009). During boreal winter, minimum southward, or even northward, flow occurs due to the northward advection of Indian Ocean waters into the Lombok Strait at both intraseasonal and seasonal timescales (e.g., Hualta et al., 2001; Murray & Arief, 1988; Sprintall et al., 2009).

Intraseasonal remote Indian Ocean wind forcing drives Kelvin waves northward through the Lombok Strait that obstruct southward ITF flow. During monsoon transition periods, westerly Indian Ocean wind bursts generate eastward propagating Kelvin waves that reach the Sumatra coastline after approximately 1 month (e.g., Chong et al., 2000; Clarke & Liu, 1993; Sprintall et al., 2000). These waves subsequently propagate...
southward along the southern coast of Sumatra and Java (Clarke & Liu, 1993, 1994; Iskandar et al., 2005), often moving northward through the Lombok Strait to reach the internal Indonesian seas (e.g., Gordon et al., 2010; Hautala et al., 2001; Potemra et al., 2002; Sprintall et al., 2000, 2003, 2009; Susanto et al., 2007, 2012; Wijffels & Meyers, 2004). Previous studies have found that westerly wind bursts occurring during the April/May

Figure 1. Map of surface water circulation through the Indonesian seas. (a) During boreal winter, surface waters from the South China Sea (solid arrows) block southward surface flow of the Indonesian throughflow (dotted arrows) in the Makassar and Lombok Straits. The green triangle indicates the Doangdoangan Besar (DD) coral site (Murty et al., 2017). The Lombok Strait, Ombai Strait, and Timor passage Indonesian throughflow outflow passages are indicated by the bold numbers 1, 2, and 3, respectively. (b) Both the southward Makassar throughflow (solid black line) and wind-driven Indian Ocean waters from the South Java Current (dotted black arrow) influence the Lombok Strait during boreal winter. At this time, eastward monsoon winds result in northward Ekman transport and geostrophic flow throughout the Lombok Strait. The blue and light brown diamonds indicate the Gili Selang (GS) and Nusa Penida (NP) coral sites in the Lombok Strait (this study) and the red circle shows the location of published coral records (Charles et al., 2003; Guilderson et al., 2009; Moore, 1995).
monsoon transition season most commonly generate Kelvin waves that reach the Lombok Strait, with a secondary occurrence in November (Hautala et al., 2001; Iskandar et al., 2006; Potemra et al., 2002; Sprintall et al., 2003). The presence of winter Kelvin waves in the Lombok Strait suggests that intraseasonal remote Indian Ocean wind forcing plays an important role in reversing winter Lombok Strait surface transport and thus obstructing the ITF (Potemra et al., 2002; Sprintall et al., 2003). However, monsoon-driven advection of Indian Ocean waters at longer (e.g., seasonal) timescales additionally plays a role in reducing surface ITF transport during boreal winter.

The Australasian Monsoon is characterized by seasonally reversing winds associated with both Northern Hemisphere (e.g., Asian Monsoon and East Asian Monsoon) and Southern Hemisphere (e.g., Indo-Australian Monsoon) monsoon systems (e.g., Chen & Guan, 2017; Webster et al., 1998). During boreal winter, these monsoon winds drive surface waters from the Indian Ocean (South Java Current, SJC) southeastward along the coastline of the Indonesian archipelago (Iskandar et al., 2006; Sprintall et al., 1999; Wijffels et al., 2008), bringing Indian Ocean waters in close proximity to the Lombok Strait. Previous work has shown that both Ekman dynamics and geostrophic flow can induce northward surface layer flow of these monsoon-driven Indian Ocean waters through the Lombok Strait, thus weakening southward transport in the surface layer (Hautala et al., 2001; Susanto et al., 2007; Wijffels et al., 2008). However, limited observational data prior to the early 1980s has prevented analysis of past changes in the monsoon influence. Paleoclimate archives, such as corals, are therefore crucial to examine the lower-frequency (e.g., seasonal to multidecadal) monsoon influence on Lombok Strait variability over the past few centuries.

Massive corals provide subannually resolved records of climate that can extend over several centuries due to their density banding and high linear extension rates (Corrège, 2006; Grottoli, 2001; Grottoli & Eakin, 2007). Throughout the growth process, corals incorporate chemical climate tracers into their aragonite skeletons that reflect seawater conditions (Gagan et al., 2000). The coral proxy δ18O represents a combined signal of both sea surface temperature (SST) and salinity (SSS) (Dunbar & Wellington, 1981; Epstein et al., 1953), while Sr/Ca ratios inversely correlate with SST (Smith et al., 1979). Within the Indonesian seas, these coral proxies collectively record changes in precipitation and ocean circulation related to a range of climate variability phenomena (e.g., El Niño–Southern Oscillation [ENSO], Indian Ocean dipole [IOD], and East Asian Monsoon).

Coral δ18O, Sr/Ca, and radiocarbon (Δ14C; used as a water mass tracer) proxy records from the eastern Indian Ocean and the eastern Indonesian seas document influences of global and regional climate phenomena (Abram et al., 2007, 2008, 2009, 2015; Charles et al., 2003; Fairbanks et al., 1997; Fallon & Guilderson, 2008; Grumet et al., 2004; Guilderson et al., 2009; Linsley et al., 2017). For example, records from the Sumatran coastline reveal that ENSO and the Asian Monsoon can influence the eastern Indian Ocean region of the IOD (Abram et al., 2007; Grumet et al., 2004). When compared with western Indian Ocean variability, Sumatran coral records document changes to the ENSO-IOD and Asian Monsoon-IOD relationships through time, with implications for the role of the IOD in Indo-Pacific climate change under anthropogenic warming (Abram et al., 2008). A coral dipole index constructed from Lombok Strait and western Indian Ocean δ18O records strongly correlates with the Niño3.4 ENSO index, further supporting the influence of ENSO-IOD teleconnections across the Pacific and Indian Oceans (Charles et al., 2003). Notably, recent site comparisons suggest that Sumatran corals are more optimally located than those from the Lombok Strait to examine IOD variability (Abram et al., 2015). Though Charles et al. (2003) use a Lombok Strait coral to reconstruct the IOD, proxy records throughout the Lombok and Makassar Straits additionally reveal ENSO and East Asian Monsoon influences at interannual to decadal timescales.

Coral paleoclimate records from the Lombok and Makassar Straits reveal that surface layer circulation responds to both ENSO and South Pacific Convergence Zone zonal events over the past ~300 years (Charles et al., 2003; Fallon & Guilderson, 2008; Guilderson et al., 2009; Linsley et al., 2017). In the southern Makassar Strait, a Δ14C record from Langkai reveals seasonal contributions of equatorial Pacific Ocean waters that are strongly influenced by ENSO variability (Fallon & Guilderson, 2008). A δ18O record from Kapoposang nearby further reveals that South Pacific Convergence Zone zonal events drive high-salinity Pacific Ocean waters into the Makassar Strait during boreal winter that reduce the seasonal influence of monsoon-driven
South China Sea (SCS) waters (Linsley et al., 2017). Farther along the ITF pathway in the Lombok Strait, ENSO continues to influence coral δ\(^{18}\)O and Δ\(^{14}\)C variability, as evidenced by interannual spectral power and comparisons with the Darwin sea level pressure anomaly record (Charles et al., 2003; Guilderson et al., 2009). These studies collectively document that surface waters throughout the main ITF pathway consistently respond to ENSO. However, the seasonally reversing East Asian Monsoon further influences surface ocean variability throughout the Indonesian seas.

Seasonally resolved climate reconstructions indicate that the East Asian winter monsoon (EAWM), defined as the pressure difference between the Siberian High and Aleutian Low (Wen et al., 2000), has influenced boreal winter surface water advection from both the Indian Ocean and SCS into the Indonesian seas over the past century. A 40-year coral Δ\(^{14}\)C record from the Lombok Strait reveals significant variability at biennial timescales, indicating influences of both the EAWM and remote Indian Ocean wind forcing (Guilderson et al., 2009). Lombok Strait coral δ\(^{18}\)O records similarly reveal the influence of eastern Indian Ocean forcing, though the impact of the monsoon is less consistent (Charles et al., 2003; Moore, 1995). In addition, SCS surface waters influence the Indonesian seas during boreal winter. An 87-year record of coral-reconstructed SSS from the southern Makassar Strait indicates that mixing of SCS source waters from the Luzon Strait and Java Sea varied over the past century with the strength of EAWM winds (Murty et al., 2017). The results of Murty et al. (2017) suggest that SCS waters dominate boreal winter surface ocean variability in the southern Makassar Strait throughout much of the twentieth century. However, no study has yet examined the influence of both SCS and Indian Ocean source waters on Lombok Strait surface water circulation. Given the role of the Lombok Strait as an outflow passage for the Makassar Strait throughflow, it is likely that the SCS surface waters are advected southward to the Lombok Strait en route to the Indian Ocean during boreal winter. Thus, EAWM-driven surface water mixing in the Lombok Strait likely occurs between multiple source waters from both the Indian Ocean and the SCS.

We present records of coral-reconstructed SSS from the northern and southern Lombok Strait to examine monsoon-driven surface water circulation and mixing. The two sites collectively allow us to assess the influence of southward flowing SCS waters, building upon the work of Murty et al. (2017). The northern and southern sites additionally allow us to examine the extent to which Indian Ocean surface waters intrude into the Lombok Strait to modify surface water mixing patterns.

2. Materials and Methods

2.1. Study Sites

Two Porites spp. corals from the Lombok Strait, Indonesia (Figure 1), were cored using an underwater hydraulic drill in May 2016 (Gili Selang [GS]: 8.38°S, 115.71°E; Nusa Penida [NP]: 8.67°S, 115.51°E; drilled at 10 and 5 m depths, respectively). One core was collected from the GS colony, located off the northeastern edge of Bali, while two cores were collected from the same coral in the southern Lombok Strait (NP). Seasonal SST ranges from 27.3 to 29.4 °C at GS and from 26.9 to 29.4 °C at NP, while SSS ranges from 32.8 to 34.3 psu at both sites (see section 2.3 for SST and SSS information).

2.2. Sampling and Analysis

The 10-cm diameter cores were cleaned in freshwater, cut parallel to the axis of maximum growth, and subsequently cut into ~0.7-cm thick slabs. The slabs were X-rayed using 69 kV, 5 mAs, and a focal distance of 90 cm to identify sampling pathways along the major growth axes (Figure S1 in the supporting information). The second NP core (NP2) was used to replace the bioeroded section of the NP1 core and ensure full temporal coverage of the NP record. Prior to sampling, each slab was sonicated in deionized water a minimum of three times and dried overnight in a 50 °C oven. Continuous subsamples were drilled at 0.5-mm (~biweekly) resolution and 1-mm depths using a manual drill press with a 1-mm diameter tungsten carbide drill bit. The final records for GS and NP span 110 and 193 years, respectively. Roughly 250–450 and 20–70 μg of each sample were used for Sr/Ca and δ\(^{18}\)O analyses, respectively.

For Sr/Ca analysis, coral samples were dissolved in 4 ml of 5% HNO\(_3\), vortexed for ~10 s, and then left to continue dissolving overnight. Sr and Ca concentrations for both coral samples and solution standards (to correct for matrix and drift effects; Schrag, 1999) were measured using an inductively coupled plasma optical emission spectroscopy (ICP-OES) system.
spectrometer (ICP-OES) at the Asian School of the Environment (ASE). Analytical precision and accuracy were determined through repeat measurements of the coral reference material JCp-1 throughout each run. The reproducibility of JCp-1 at ASE is 0.20% relative standard deviation (RSD) ($n = 2,287$), with an average value of $0.01935 \text{ g/g (± 0.000038, 1σ)}$, which is within error of the accepted value (Hathorne et al., 2013).

Measurements of $\delta^{18}O$ were also conducted at ASE, where samples were acidified with 105% H$_2$PO$_4$ at 70 °C using an automated Kiel IV carbonate device coupled with a ThermoFisher MAT-253 Isotope Ratio Mass Spectrometer (IRMS). Isotopic measurements were calibrated relative to Vienna Peedee belemnite (VPDB) using National Bureau of Standards (NBS) 19 ($\delta^{18}O = -2.20‰$; Stichler, 1995). The reproducibility of NBS 19 at ASE is 0.05‰ (1σ, $n = 42$). Estremoz and Carrara standards were also repeatedly measured, with average values of $-5.93‰$ (±0.09‰, $n = 353$) and $-1.95‰$ (±0.07‰, $n = 273$), respectively.

Due to the length of the GS and NP records, we were unable to analyze $\delta^{18}O$ for all samples outside the calibration period (1987–2011; period of comparison to local or gridded SST and SSS). We instead used the Sr/Ca records to select a subset of samples for which to measure $\delta^{18}O$. The maxima and minima of the Sr/Ca seasonal cycles coincide with those of the $\delta^{18}O$ records (Figure 2). Based on the Sr/Ca seasonality, we selected three to four consecutive samples representing the maximum and minimum values for each year in order to examine interannual variability. We analyzed additional samples in areas where the Sr/Ca seasonality was unclear to ensure the maxima and minima were captured. The resulting $\delta^{18}O$ records contain approximately six to eight samples per year prior to 1987.

2.3. Data Sources

Monthly reanalysis Simple Ocean Data Assimilation SSS (SODA, v. 2.2.4; Carton & Giese, 2008) from 0.5° grids (centered at 8.25°S, 115.75°E and 8.75°S, 115.75°E for GS and NP, respectively; Figure S2) was used to develop the age models and calibrate the $\delta^{18}O$ proxy. We chose this salinity product due to its length and the availability of the data throughout the region. We were unable to use other SSS products because of their limited spatial (e.g., Delcroix et al., 2011) or temporal (e.g., Aquarius, Lagerloef et al., 2008; HYCOM output, http://www.hycom.org) ranges. However, SODA SSS at GS and NP is strongly correlated with observational salinity data from the Lombok Strait (Sprintall et al., 2003), suggesting that SODA is robustly representing SSS variability in the Lombok Strait ($r = 0.82$ and 0.80 for GS and NP, respectively; $p < 0.0001$ and root-mean-square residual (RMSR) $< 0.02$ psu for both; Figure S3).

To evaluate temperature, we used monthly resolution 1° gridded SST from the Hadley Center for Sea Ice and Sea Surface Temperature (HadISST, v. 1.1, Rayner et al., 2003). Grid points were centered at 7.5°S, 115.5°E and 8.5°S, 115.5°E for GS and NP, respectively (Figure S2). Because we use HadISST to directly examine low-frequency temperature variability in the early twentieth century, we were limited to using an SST product with longer temporal coverage. However, the HadISST product reveals robust correlations with SST from SODA and the Integrated Global Ocean Services System Products Bulletin (IGOSS) (Reynolds et al., 2002) for both the GS ($r = 0.92$ and 0.90, respectively; $p < 0.0001$ for both) and NP locations ($r = 0.91$, $p < 0.0001$; Figure S4). It is important to note that we were unable to compare 4-km gridded Advanced Very High Resolution Radiometer SST (AVHRR) (http://pathfinder.nodc.noaa.gov) due to large errors likely related to cloud cover complications (Zhang et al., 2009).

We used a record of rainfall amount (mm) from 2008 to 2013 collected by the Department of Meteorology, Climatology and Geophysics, Region III, in Bali to evaluate the influence of precipitation on SSS variability in the Lombok Strait (Figure S2). This rainfall record indicates average maximum rainfall from November to January and minimum rainfall from May to July. We chose this rainfall record as it represents a local signal nearby to the coral sites instead of a spatially averaged rainfall product, such as the 2.5° gridded Global Precipitation Climatology Project (https://www.ersl.noaa.gov/psd).
2.4. Age Model Development and Statistical Analysis

Subannual age models were developed for the calibration period (1987–2011) by aligning the minima, maxima, and inflection points of the δ18O records to those of SODA SSS using Analyseries version 2.0 (Paillard et al., 1996). The inflection points were included as tie points for the age model to account for seasonal bias in the coral growth rate (DeLong et al., 2014), similar to the approach of several previous studies (e.g., Bolton et al., 2014; DeLong et al., 2007, 2012, 2014; Murty et al., 2017, 2018; Ramos et al., 2017). The δ18O time series were then resampled using linear interpolation into a monthly resolution record.

Prior to 1987, the maxima and minima of the lower-resolution (six to eight samples/year) coral δ18O time series were aligned to a salinity climatology developed based on SODA SSS variability during the calibration period. We were unable to align the inflection points due to the lower sampling resolution of the record. However, this reduction in tie points is unlikely to impact our analysis due to our focus on boreal winter (seasonal minima) interannual to interdecadal variability. Following age model development, the δ18O records were linearly interpolated into bimonthly time series. The δ18O age models were then applied to the Sr/Ca records for each core. Calibrations were performed using ordinary least squares regressions. Significance levels were determined using the effective degrees of freedom (Thomson & Emery, 2014).

Cross-correlation spectral analysis was performed using a multitaper method with adaptive weighting (e.g., Huybers & Denton, 2008). Confidence intervals were calculated using a Gaussian process (Amos & Koopmans, 1963) where the degrees of freedom used were calculated by multiplying the number of windows used in the multitaper method by 2 and subtracting 1.

3. Results

3.1. δ18O–SSS Calibrations

The average seasonal cycle of monthly δ18O in the Lombok Strait is 0.5‰ and 0.8‰ for the GS and NP sites, respectively (Figure S5). To calculate the salinity contribution to the δ18O signal, we directly used HadISST to remove the temperature component of the δ18O record (see supporting information Text S1 and Figure S6). At both sites, δ18Osw is similar to the δ18Ocoral proxy (Figure S5), suggesting that salinity drives the majority of the variability in the δ18Ocoral signal. IGSS- and SODA-derived δ18Osw records, as well as δ18Osw calculated using the δ18O-SST relationship from Gagan et al. (1998) of −0.18‰ °C−1, further support the dominance of SSS, revealing similar variability to the HadISST-derived δ18Osw record (Figure S7). Notably, we were unable to use Sr/Ca to remove the temperature contribution to δ18Ocoral because of weak Sr/Ca–SST calibrations (r2 = 0.07 and 0.20 for GS and NP, respectively) likely due to low Sr/Ca (0.12 and 0.15 mmol/mol for GS and NP, respectively) and SST variability that increase the noise-to-signal ratio.

Another methodological approach for determining the influence of SST on the δ18Ocoral signal is to employ a multivariate calibration that accounts for the influence of both temperature and salinity. Multivariate regressions of δ18Ocoral with each temperature product and SODA SSS reveal low δ18O–SST slopes that continue to suggest small temperature contributions to δ18Ocoral (Table S1). These multivariate regressions, when considered with the similarity of the δ18Osw and δ18Ocoral records, collectively indicate that salinity dominantly contributes to the δ18Ocoral signal. Previous studies have similarly noted the importance of δ18Osw (salinity and rainfall) variability in influencing the δ18Ocoral proxy in the Lombok Strait (Charles et al., 2003; Guilderson et al., 2009). We therefore directly compare δ18Ocoral to SSS for the rest of the study, particularly given that monthly and winter interannual relationships between Bali rainfall and both SODA SSS and δ18Osw at GS and NP are insignificant (p > 0.3 for all; see section 2.3). Doing so is further necessary as all SST time series are too short to be used to remove the temperature signal for the entire NP record.

Least squares linear regressions of monthly δ18Ocoral against SODA SSS reveal robust relationships (Figure 3 and equations (1)–(3)):

\[
\delta^{18}O_{\text{coral,GS}}(1987 – 2011)\, (\%) = 0.255 \pm 0.014 \times SS\text{S (psu)} – 13.971 \pm 0.487
\]

\[(r = 0.72, p<0.0001, \text{RMSR} = 0.008 \text{ psu}, n = 288, \text{effective degrees of freedom (n')} = 80)\]

\[
\delta^{18}O_{\text{coral,NP}}(1987 – 2011)\, (\%) = 0.365 \pm 0.015 \times SS\text{S (psu)} – 17.59 \pm 0.514
\]

\[(r = 0.82, p<0.0001, \text{RMSR} = 0.001 \text{ psu}, n = 288, \text{n'} = 53)\]
Notably, the NP1 and NP2 regressions have slopes and y intercepts that are within error of each other, suggesting that SSS similarly influences δ¹⁸Ocoral in both cores. Least squares linear regressions of boreal winter (January–March) interannual δ¹⁸Ocoral and SODA SSS also reveal robust relationships (r = 0.71, 0.68, and 0.68 for GS, NP1, and NP2, respectively; p < 0.001 for all), where the NP1 and NP2 relationships continue to be within error of each other (Figure 3 and Table S2). We therefore replaced the bioeroded section of the NP1 core (1972–1987) with the NP2 record due to the similarity of the NP calibrations at monthly and winter interannual timescales. Least squares linear regressions of boreal summer (July–August) interannual δ¹⁸Ocoral and SODA SSS reveal insignificant relationships (p > 0.2 for all) that may be related to augmented noise-to-signal ratios associated with low summer SSS variability (~0.5 psu for both sites).

We applied the GS and NP1 monthly δ¹⁸Ocoral-SSS calibrations (equations (1) and (2)) down core to reconstruct SSS for 110 and 193 years in the northern and southern Lombok Strait, respectively (Figure S8). We then focused on SSS variability during the boreal winter monsoon (Figure 4) to examine the influence of the EAWM on SSS and surface water circulation throughout the Lombok Strait, building upon the work of Murty et al. (2017). We chose to focus on winter variability as the seasonal reversal in monsoon winds and surface circulation (e.g., Fang et al., 2010; Susanto et al., 2013) primarily drives the advection of SCS surface waters into the Indonesian seas during boreal winter. In addition, the greater SSS variability during boreal winter (compared with summer) and the strong winter interannual calibrations suggest that records of winter coral-reconstructed SSS should be robust.

### 3.2. Boreal Winter Reconstructed SSS

Reconstructed boreal winter SSS in the Lombok Strait reveals variability between the northern (GS) and southern (NP) sites. The GS SSS record (RMSR = 0.36 psu) shows a rapid shift in the midtwentieth century, where SSS decreases from ~34.3 psu in 1945, reaching a minimum of ~32.4 psu in 1958, and subsequently rebounding to ~33.9 psu by 1965. This pattern coincides with a similar shift previously observed in the southern Makassar Strait (Doangdoangan Besar [DD]; Murty et al., 2017; Figure 4). Murty et al. (2017) attributed this salinity shift to monsoon-driven variability in surface water mixing between SCS waters from the Luzon Strait (higher salinity) and Java Sea (lower salinity). However, the NP SSS record in the southern Lombok Strait (RMSR = 0.17 psu), does not reveal the 1945–1965 salinity feature (Figure 4). The difference in SSS

$$\delta^{18}O_{\text{c}}(1987 - 2011) \text{ (‰)} = 0.385 \times SSS \text{ (psu)} - 18.175 \text{ (±0.634)}$$

$$r = 0.77, p < 0.0001, \text{RMSR} = 0.07 \text{ psu, } n = 288, n^* = 66$$

Figure 3. The $\delta^{18}O_{\text{c}}$-SSS calibrations of $\delta^{18}O_{\text{c}}$ to SODA SSS from 1987 to 2011 for NP1 (light brown), NP2 (black), and GS (light blue) for (a) monthly and (b) boreal winter (January–March) interannual averages. Calibrations are robust at both monthly ($r = 0.82, 0.77, 0.72$, respectively; $p < 0.001$ for all) and winter interannual timescales ($r = 0.67, 0.68, 0.71$, respectively; $p = 0.08$ for all). See equations (1)–(3) of the main text and Table S2 for equations. SSS = sea surface salinity; SODA = Simple Ocean Data Assimilation; NP = Nusa Penida; GS = Gili Selang.
variability at NP suggests that additional source waters may be influencing the site other than the Luzon Strait and Java Sea surface waters that are advected via the Makassar throughflow. We therefore examined instrumental temperature-salinity (T-S) mixing relationships to identify the source waters that contribute to SSS variability in the northern and southern Lombok Strait.

3.3. Instrumental Surface Water Mixing

HadiSST-SODA SSS T-S relationships from the GS and NP sites confirm that the SCS surface water mixing signal is advected southward from the Makassar Strait to the Lombok Strait (Figure 5). During boreal winter (January–March), the T-S relationships at both sites indicate mixing between two source waters. Three-month averages of HadiSST and SODA SSS from the Java Sea (January–March; HadiSST: 4.5°S, 112.5°E; SODA SSS: 4.75°S, 112.75°E) and the Luzon Strait (May–July; HadiSST: 18.5°N, 118.5°E; SODA SSS: 18.25°N, 118.25°E) suggest that surface water from both regions likely mix throughout the Lombok Strait. Notably, we are unable to resolve the impacts of Indian Ocean intrusions into the Java Sea via the Sunda Strait independently of the Java Sea source water (Susanto et al., 2016; see Figure 1 and supporting information).

A similar pattern of Luzon Strait and Java Sea source water mixing was observed at DD in the southern Makassar Strait (Murty et al., 2017) suggesting that SCS waters continue to flow southward from the

Figure 4. Boreal winter (January–March) reconstructed SSS records. Coral-reconstructed SSS records for the boreal winter monsoon for DD in the southern Makassar Strait (green; Murty et al., 2017), GS in the northern Lombok Strait (blue; this study), and NP in the southern Lombok Strait (light brown; this study). Shading indicates RMSR = 0.13, 0.36, 0.17 psu, respectively. The black box indicates the rapid salinity shift at DD and GS. SSS = sea surface salinity; DD = Doangdoangan Besar; GS = Gili Selang; NP = Nusa Penida; RMSR = root-mean-square residual.
Makassar Strait to the Lombok Strait during the winter monsoon. As mentioned by Murty et al. (2017), the offset in Luzon Strait timing may be due to the residence time of Luzon Strait waters circulating in the SCS prior to moving southward during boreal winter. However, the southward advection of SCS waters throughout the Lombok Strait may not be consistent in the midtwentieth century during the period of rapidly shifting SSS at GS. We therefore examined T-S relationships using reconstructed SSS to constrain past changes in the southward advection of SCS source water mixing between the northern and southern Lombok Strait. We focus on boreal winter variability, as boreal summer (July–September) instrumental T-S relationships indicate that SCS source waters do not dominate at GS and NP during the summer monsoon (Figure S9).

3.4. Reconstructed Surface Water Mixing

T-S relationships using reconstructed SSS and instrumental HadISST at GS and NP reveal differences in relative source water mixing during the rapid salinity shift of the midtwentieth century (Figure 6). At GS, decadal bins of the boreal winter coral record (centered on the start of each decade; RMSR = 0.37 psu) indicate mixing between Luzon Strait and Java Sea source waters (source waters based on decadal averaged HadISST and SODA SSS). From 1940 to 1960, both the GS reconstructed mixing diagram (Figure 6a) and seasonal time series (Figure 6c) show salinity characteristics that shift from being similar to those of the Luzon Strait (higher salinity) to those of Java Sea waters (lower salinity). This shift suggests a change in the relative contribution of the two source waters, where the contribution of lower salinity Java Sea water increases relative to the higher salinity Luzon Strait water from 1940 to 1960. The pattern we observe at GS is similar to the change in SCS source water mixing in the southern Makassar Strait (Murty et al., 2017), supporting that southern Makassar Strait monsoon-driven surface water mixing moves southward into the Lombok Strait during boreal winter.

Unlike GS in the northern Lombok Strait, reconstructed T-S relationships at NP do not reveal a large change in relative mixing between source waters in the midtwentieth century. Decadal binned HadISST and
reconstructed SSS at NP (centered on the start of each decade; RMSR = 0.19 psu) are overall similar to decadal-binned Indian Ocean South Java Current (SJC) HadISST (8.5°S, 108.5°E) and SODA SSS (8.25°S, 108.25°E), particularly prior to 1950 (Figures 6b and 6d). From 1940 to 1960, the NP record shows a small shift toward lower salinities, suggesting a slight increase in the relative contribution of Java Sea waters. However, the shift toward lower salinities at NP is less (0.2 psu decrease) than that of GS (1.1 psu decrease), suggesting that Java Sea waters do not dominantly influence the NP site in the southern Lombok Strait in the midtwentieth century.

3.5. EAWM Influence on SSS and Surface Water Mixing

We compared the decadal averaged winter interannual SSS reconstructions for GS and NP to an index of the EAWM (D’Arrigo et al., 2005) to examine the EAWM influence on surface water mixing in the Lombok Strait. The EAWM is defined as the pressure difference between the Siberian High and the Aleutian Low in the Northern Hemisphere (Wen et al., 2000). D’Arrigo et al. (2005) used global mean sea level pressure (SLP) data (v2.1f; UK Met Office-Hadley Centre, 5°×5°; Basnett & Parker, 1997) to develop an EAWM index by summing normalized zonal SLP differences (between 110° and 160°E) at 5° latitude intervals over the 40°–65°N band, following the methods of Wu and Wang (2002). The resulting EAWM index extends from 1871 to 1994. Positive (negative) values indicate a strong (weak) EAWM state characterized by strong (weak) boreal winter (December–February) meridional wind strength. It is important to note that we focus on the EAWM due to the dominant contribution of EAWM-driven surface waters from the SCS into the northern and southern Makassar Strait during boreal winter (Murty et al., 2017). In addition, Australian Monsoon indices
(e.g., Kajikawa et al., 2010) are overall limited to the mid- to late- twentieth century, thus hindering our evaluation of monsoon influences over the full length of the coral records. Within the mid- to late-twentieth century, cross-correlation spectral analysis of an index for the Australian Monsoon (Kajikawa et al., 2010) further reveals no significant coherence to GS or NP reconstructed SSS (1948–2015; see supporting information).

Reconstructed SSS at GS is significantly correlated to the EAWM (with a 3-year lag in the coral record), revealing a consistent positive relationship over the twentieth century ($r = 0.71$, $p = 0.03$, $n = 9$; Figure 7a). This positive correlation extends over the rapid salinity shift from 1940 to 1960 and is similar to the relationship for DD in the southern Makassar Strait (Murty et al., 2017). At NP, however, the EAWM influence on reconstructed SSS reverses from a significant negative ($r = -0.84$, $p = 0.02$, $n = 7$, 1890–1950) to a significant positive ($r = 0.89$, $p = 0.10$, $n = 4$, 1960–1990) relationship around 1960, with the coral lagging by 3 years (Figure 7b). The 3-year lag of reconstructed SSS behind the EAWM at both sites is consistent with the relationship at DD in the southern Makassar Strait (Murty et al., 2017). Murty et al. suggest that the lag could be due to a residence time of the surface waters in the northern and southern SCS gyres before they enter the Indonesian seas, though the mechanism remains unclear. The consistency of the lag between DD, GS and NP, however, suggests a similar regional influence impacting the corals’ response to monsoon-driven circulation.

4. Discussion

Coral-reconstructed SSS indicates differences in surface water circulation between the northern and southern Lombok Strait over the past two centuries. In the northern Lombok Strait, SSS variability corresponds to surface water mixing between high-salinity Luzon Strait and low-salinity Java Sea surface waters throughout the twentieth century (Figures 5a, 6a, and 6c), revealing a consistent direct influence of the EAWM (Figure 7a). In the southern Lombok Strait, SSS variability is strikingly different, particularly in the mid-twentieth century where no rapid freshening trend is apparent (Figures 4, 6b, and 6d). Prior to 1960, NP SSS variability corresponds to the influence of Indian Ocean surface waters from the SJC (Figures 6b and 6d), differing from the SCS mixing signal at the GS site (Figures 6a and 6c). The EAWM influence in the southern Lombok Strait also differs from the northern site, revealing a transition from an inverse to direct relationship with NP SSS around 1960 (Figure 7b). The dissimilarities between the two coral sites suggest that the EAWM differently influences surface water advection throughout the Lombok Strait over periods of the past two centuries. We propose the following:

1. The Makassar throughflow consistently transports SCS surface waters to the northern Lombok Strait.
2. Indian Ocean enhanced surface water advection inhibits surface throughflow transport in the southern Lombok Strait when the EAWM mean state is positive.
3. When the EAWM mean state is negative, weak monsoon winds no longer obstruct the throughflow in the southern Lombok Strait.

Figure 7. Reconstructed SSS-EAWM relationships. Decadal binned boreal winter reconstructed SSS for (a) GS (blue; shading indicates RMSR = 0.37 psu) and (b) NP (light brown; shading indicates RMSR = 0.19 psu), plotted with an index for the EAWM (EAWMI; D’Arrigo et al., 2005). EAWMI values above the zero lines indicate a strong (positive) monsoon state, and vice versa for EAWMI values below the zero lines. The EAWMI is shown with solid lines during periods of positive correlation at both GS and NP and a dotted line during a period of negative correlation, corresponding with a strong (positive) EAWM state. SSS = sea surface salinity; EAWM = East Asian winter monsoon; GS = Gili Selang; NP = Nusa Penida; RMSR = root-mean-square residual. 

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4.1. The Makassar Throughflow Consistently Transports SCS Surface Waters to the Northern Lombok Strait

The EAWM consistently influences low-frequency SSS variability in the northern Lombok Strait through an advected mixing signal from the southern Makassar Strait. Throughout both the Indonesian seas and the SCS, winds from the seasonally reversing East Asian Monsoon are the primary driver of surface ocean circulation (Fang et al., 2009; Wyrtki, 1961; D. Xu & Malanotte-Rizzoli, 2013). EAWM winds drive surface waters from the SCS (Java Sea and Luzon Strait) into the Makassar Strait that obstruct surface ITF flow during boreal winter (Fang et al., 2010; Gordon et al., 2003, 2014; Tozuka et al., 2007, 2009). A regional finite volume coastal ocean model estimates monsoon-driven contributions of ~3.6 and ~2.9 Sv from the Java Sea and Luzon Strait, respectively (D. Xu & Malanotte-Rizzoli, 2013), suggesting that both source waters play an important role in influencing Makassar Strait variability. However, the relative contributions of the Java Sea and Luzon Strait source waters vary with fluctuating EAWM wind strength (Murty et al., 2017). Murty et al. (2017) suggest that weaker winds reduce the relative contribution of Luzon Strait (higher SSS) waters to the southern Makassar Strait, possibly due to a weaker cyclonic tendency in the northern SCS (D. Xu & Malanotte-Rizzoli, 2013). Surface salinities in the southern Makassar Strait therefore decrease with reduced wind strength, and vice versa during periods of enhanced wind strength.

The GS SSS record indicates that monsoon-driven SCS surface waters consistently advect southward as part of the Makassar throughflow to influence the northern Lombok Strait. Instrumental T-S relationships at GS confirm that Luzon Strait (high-salinity) and Java Sea (low-salinity) surface waters influence the northern Lombok Strait (Figure 5a), similar to the DD site in the southern Makassar Strait (Murty et al., 2017). Throughout the twentieth century, winter reconstructed SSS at GS continues to coincide with the DD record, revealing a similar midcentury (1945–1965) rapid salinity shift (Figure 4). Between the 1940s and 1960s, reconstructed T-S relationships at GS reveal changes in relative mixing toward dominantly Java Sea waters (Figures 6a and 6c), again similar to DD. The similarity of the two sites suggests that monsoon-driven SCS surface waters that mix in the southern Makassar Strait consistently advect southward to the northern Lombok Strait throughout the twentieth century.

Given the consistent mixing patterns at both sites, it is no surprise that GS is similarly positively correlated with the EAWM (D’Arrigo et al., 2005; Figure 7a). The significant correlation at GS supports that decreasing monsoon wind strength in the early twentieth century (S. Song et al., 2012; M. Xu et al., 2006; D’Arrigo et al., 2005; Figure 7a) reduces the contribution of high-salinity Luzon Strait waters in the southern Makassar Strait that then advect to the northern Lombok Strait (Murty et al., 2017; Figures 8a and 8b). On the other hand, the positive EAWM influence at GS also supports that strengthening monsoon winds enhance the transport of high-salinity Luzon Strait waters to the southern Makassar and northern Lombok Straits (Figures 8a and 8b). It is important to note that these monsoon-driven changes in Luzon Strait advection influence the northern Lombok Strait throughout the twentieth century, even though the EAWM transitioned from a positive (strong) to a negative (weak) mean state around 1960 (Figure 7a; D’Arrigo et al., 2005). Thus, while water column-integrated studies recognize a small influence of SCS waters in the Lombok Strait (e.g., D. Xu & Malanotte-Rizzoli, 2013), our results suggest that SCS waters dominantly influence boreal winter northern Lombok Strait surface water variability throughout the twentieth century.

4.2. Indian Ocean Enhanced Surface Water Advection Inhibits Surface Throughflow Transport in the Southern Lombok Strait when the EAWM Mean State Is Positive

In contrast with the northern Lombok Strait, SSS variability in the southern Lombok Strait suggests that monsoon-driven source water advection varies through time. From 1945 to 1965, winter reconstructed SSS at NP in the southern Lombok Strait does not reveal a rapid salinity shift, differing from the GS and DD sites farther upstream (Figures 4, 6b, and 6d). The differing salinity response in the NP record suggests that Makassar throughflow waters that carry the SCS freshening signal do not reach the southern Lombok Strait in the midtwentieth century. Instead, monsoon-driven surface waters from the Indian Ocean likely advect into the Lombok Strait during this time, inhibiting the southward Makassar throughflow.

Coral proxy records from the Lombok Strait document the influence of Indian Ocean surface waters over the past few centuries. For example, Moore (1995) calculated an estimated salinity range of 0.48 psu from a central Lombok Strait δ18O_coral record, revealing decreased seasonal variability compared to that expected from the Java Sea or Makassar Strait (~1.54 psu). Instead, the seasonal variability of the record reveals similarity to
that of the Indian Ocean (~0.31 psu, Moore, 1995), suggesting that Indian Ocean waters may obscure the throughflow signal. A seasonally resolved coral radiocarbon record similarly documents varying Makassar throughflow contributions to the central Lombok Strait ranging from 16% to 70% over the past century (Guilderson et al., 2009). The authors attribute periods of reduced Makassar throughflow to enhanced Indian Ocean (SJC) advection, finding the strongest SJC influence from 1947 to 1954. Our reconstructed T-S relationships at NP additionally reveal a dominant SJC influence prior to 1960 (Figures 6b and 6d), coinciding with the timing of strongest Indian Ocean influence in the Guilderson et al. (2009) record. The similarity in timing of these records from the southern and central Lombok Strait suggests that increased Indian Ocean surface water advection into the southern Lombok Strait inhibits southward transport of the Makassar throughflow in the surface layer prior to 1960. This surface layer obstruction likely extends the width of the southern Lombok Strait, given the comparable surface layer transport estimates for either

Figure 8. Schematic of proposed monsoon-driven surface ocean circulation. (a) Throughout the twentieth century, strong EAWM winds enhance southward transport of Luzon Strait waters (red arrows) into the Makassar Strait, resulting in higher SSS in the southern Makassar Strait that then advects to the northern Lombok Strait via the Makassar throughflow (black arrow). (b) Weakening monsoon winds instead reduce southward Luzon Strait transport, resulting in a greater relative contribution of lower salinity Java Sea surface waters (orange arrows) to the southern Makassar Strait that then advect to the northern Lombok Strait via the Makassar throughflow. (c) When the mean state of the EAWM is positive (i.e., 1890s through 1950s), strong monsoon winds drive Indian Ocean surface waters from the SJC (dotted purple arrow) into the southern Lombok Strait. These waters inhibit the throughflow from reaching the southern Lombok Strait in the surface layer. (d) During a negative EAWM state (i.e., 1960s to 1990s), weak monsoon winds reduce the Indian Ocean contribution to the southern Lombok Strait, resulting in consistent southward throughflow advection throughout the Lombok Strait. The green triangle represents DD (Murty et al., 2017), the blue diamond represents GS (this study), and the light brown diamond represents NP (this study). Note that arrow thickness is used schematically to indicate the relative contribution of the source waters to the Makassar and Lombok Straits. EAWM = East Asian winter monsoon; SSS = sea surface salinity; SJC = South Java current; DD = Doangdoangan Besar; GS = Gili Selang; NP = Nusa Penida.
side of the Nusa Penida island (Arief, 1992). However, additional SSS records are needed from the eastern coast of Nusa Penida to fully constrain the SJC influence across the southern Lombok Strait.

Variability in the contribution of Indian Ocean waters to the southern Lombok Strait is likely influenced by EAWM wind strength. East Asian Monsoon winds are an important driver of surface ocean circulation along the southern coastline of the Indonesian archipelago (e.g., Iskandar et al., 2006; Wyrtki, 1961). The EAWM, for example, drives southeastward transport of the SJC along the southern coast of Sumatra and Java, subsequently moving past the Lombok Strait (e.g., Iskandar et al., 2006; Sprintall et al., 1999; Wijffels et al., 2008). Due to Ekman transport, these alongshore winds drive perpendicular northward flow that increases sea level south of the island chain (Iskandar et al., 2005; Potemra et al., 2002; Susanto et al., 2007). Ekman dynamics simultaneously reduce sea level north of the archipelago, resulting in a reverse pressure gradient that draws Indian Ocean water into the Lombok Strait and inhibits southward transport in the surface layer (e.g., Susanto et al., 2007; Wijffels et al., 2008). Notably, eastward transport of the SJC has increased with strengthening EAWM winds over periods of the past 80,000 years (e.g., Gingele et al., 2002), while enhanced EAWM wind strength in the late twentieth century coincides with reduced southward transport in the Lombok Strait (D. Xu & Malanotte-Rizzoli, 2013). These results imply that fluctuations in monsoon wind strength, and possibly overall mean state, over the past few centuries may alter the advection of SJC waters into the southern Lombok Strait that inhibit the southward surface Makassar throughflow transport.

Changes in the state of the EAWM likely alter the influence of Indian Ocean waters in the southern Lombok Strait over the past two centuries. At multidecadal timescales, the EAWM has revealed reversals between positive and negative mean states. From the 1890s to the 1950s, the D’Arrigo et al. (2005) EAWM index indicates a positive (strong) EAWM state that coincides with an inverse EAWM-SSS relationship at NP (Figure 7b). In the 1960s, however, the EAWM transitioned to a negative (weak) mean state, coinciding with a transition to a positive EAWM-SSS relationship (Figure 7b). Notably, the inverse relationship in the southern Lombok Strait from the 1890s to the 1950s is opposite to that observed at GS in the northern Lombok Strait (Figure 7a). The dissimilarities between the two sites suggest that monsoon-driven circulation differs throughout the Lombok Strait when the mean state of the EAWM is positive. Taken together with the consistent influence of SJC waters at NP prior to 1960 (Figures 6b and 6d), our results suggest that SJC surface waters obstruct the Makassar throughflow under a positive EAWM state (Figure 8c). Our results thus indicate that surface water circulation in the southern Lombok Strait differs from the northern Lombok Strait during periods of consistently strong monsoon winds due to the northward advection of SJC surface waters.

4.3. When the Mean State of the EAWM Is Negative, the Makassar Throughflow Reaches the Southern Lombok Strait

Weak EAWM winds after 1960 may reduce the northward advection of Indian Ocean surface waters into the southern Lombok Strait that block the Makassar throughflow. Around 1960, the EAWM transitioned from a positive to a negative mean state, characterized by overall weak monsoon winds (D’Arrigo et al., 2005; Figure 7b). D. Xu and Malanotte-Rizzoli (2013) used a finite volume coastal ocean model to examine regional surface water circulation under weak monsoon conditions. The authors identified increased southward Makassar throughflow transport in the Lombok Strait under weak monsoon conditions that only occurred during boreal winter. Their results suggest an important role of weakening EAWM wind strength in enhancing southward surface water transport. Given that increased EAWM wind strength can augment the eastward advection of the SJC (e.g., Gingele et al., 2002), overall weaker monsoon winds after 1960 may have reduced the influence of the SJC in the southern Lombok Strait that previously inhibited the Makassar throughflow (Figure 8d).

The EAWM influence on NP SSS variability reversed to a positive relationship after 1960, coinciding with the transition to a negative mean state of the EAWM (D’Arrigo et al., 2005; Figure 7b). The NP EAWM-SSS relationship at this time is similar to that observed at GS in the northern Lombok Strait (Figure 7a), suggesting that Indian Ocean waters no longer dominate the southern Lombok Strait. NP reconstructed SSS further confirms that variability in SJC surface waters do not fully describe southern Lombok Strait variability after 1960 (Figure 6d). Instead, instrumental T-S relationships at both sites (Figure 5) indicate that the Makassar throughflow transports mixing SCS surface waters throughout the Lombok Strait when the mean state of the EAWM is negative (Figure 8d). Monsoon-driven changes in relative mixing between the Luzon Strait and Java Sea surface waters would in this case more strongly influence the EAWM-SSS relationship in the southern Lombok Strait.

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**Figure 7b.** D. Xu and Malanotte-Rizzoli (2013) used a finite volume coastal ocean model to examine regional surface water circulation under weak monsoon conditions. The authors identified increased southward Makassar throughflow transport in the Lombok Strait under weak monsoon conditions that only occurred during boreal winter. Their results suggest an important role of weakening EAWM wind strength in enhancing southward surface water transport. Given that increased EAWM wind strength can augment the eastward advection of the SJC (e.g., Gingele et al., 2002), overall weaker monsoon winds after 1960 may have reduced the influence of the SJC in the southern Lombok Strait that previously inhibited the Makassar throughflow (Figure 8d).

**Figure 8d.** The EAWM influence on NP SSS variability reversed to a positive relationship after 1960, coinciding with the transition to a negative mean state of the EAWM (D’Arrigo et al., 2005; Figure 7b). The NP EAWM-SSS relationship at this time is similar to that observed at GS in the northern Lombok Strait (Figure 7a), suggesting that Indian Ocean waters no longer dominate the southern Lombok Strait. NP reconstructed SSS further confirms that variability in SJC surface waters do not fully describe southern Lombok Strait variability after 1960 (Figure 6d). Instead, instrumental T-S relationships at both sites (Figure 5) indicate that the Makassar throughflow transports mixing SCS surface waters throughout the Lombok Strait when the mean state of the EAWM is negative (Figure 8d). Monsoon-driven changes in relative mixing between the Luzon Strait and Java Sea surface waters would in this case more strongly influence the EAWM-SSS relationship in the southern Lombok Strait.
Strait (Murty et al., 2017; Figures 8a and 8b), owing to the lower relative contribution of Indian Ocean waters (Figure 8d). Our results therefore identify changes in SSS and surface water circulation throughout the Lombok Strait due to variability in Indian Ocean surface water advection under varying EAWM states.

Fluctuations in Lombok Strait surface water mixing related to variability in EAWM strength and state have widespread implications for ITF variability and Indo-Pacific climate. Previous studies have shown that the wind-driven injection of buoyant SCS surface layer water via the Java Sea into the western Indonesian seas reduces the transport of warm surface water within the ITF, which would in turn reduce the heat transport into the Indian Ocean (Gordon et al., 2003; Qu et al., 2005; Tozuka et al., 2007). Changes in monsoon winds and precipitation would thus be expected to affect the intensity of the SCS factor. For example, increased injection of SCS buoyant water into the Java Sea should intensify the surface buoyancy plug and lead to a cooler ITF (Gordon et al., 2003). Our results indicate that weakening EAWM winds enhance the transfer of SCS surface waters via the Java Sea (relative to Luzon Strait waters) into the Makassar Strait (Murty et al., 2017), northern Lombok Strait, and even southern Lombok Strait (post-1960), potentially intensifying the freshwater plug and reducing heat transport to the Indian Ocean. Invoking this mechanism, our study thus suggests a possible intensification of the buoyancy plug and reduced ITF heat transport under anthropogenic warming, given future projections of weakening monsoon circulation and enhanced precipitation (e.g., Allen & Ingram, 2002; Meehl et al., 2007). Such reduced heat distribution to the Indian Ocean would likely alter the heat budgets of both the Indian and Pacific Oceans and thus Indo-Pacific climate systems including ENSO, the Asian Monsoon, and the IOD (Saji et al., 1999; Susanto et al., 2001).

5. Conclusions

In this study, we present records of coral-reconstructed SSS from the northern and southern Lombok Strait to examine surface water advection over the past two centuries. Reconstructed SSS from Lombok Strait Porites corals documents differences in EAWM-driven SSS and surface water circulation variability between the northern and southern sites that coincide with changes in the state of the EAWM. We present a three-part hypothesis to describe the differences in Lombok Strait surface water circulation over the past two centuries:

Hypothesis 1.1: The Makassar throughflow consistently transports SCS surface waters to the northern Lombok Strait. SSS variability in the northern Lombok Strait corresponds with surface water mixing between high-salinity Luzon Strait and low-salinity Java Sea surface waters throughout the twentieth century, revealing a consistent direct influence of the EAWM that is similar to DD in the southern Makassar Strait (Murty et al., 2017). The similarity of the GS and DD sites indicates that monsoon-driven SCS surface water mixing consistently advects southward from the Makassar Strait to influence the northern Lombok Strait.

Hypothesis 1.2: Indian Ocean enhanced surface water advection inhibits surface throughflow transport in the southern Lombok Strait when the EAWM mean state is positive. In the southern Lombok Strait, SSS variability is strikingly different, particularly in the midtwentieth century. NP SSS variability prior to 1960 corresponds to that of Indian Ocean surface waters from the SJC, differing from the SCS mixing signal at the GS site. Notably, the EAWM inversely influences NP SSS before 1960, unlike the monsoon influence at GS. This inverse relationship at NP coincides with a positive EAWM state that is characterized by consistently strong winds. Thus, during a positive mean state, strong monsoon winds likely drive Indian Ocean waters into the southern Lombok Strait, obstructing the southward Makassar throughflow.

Hypothesis 1.3: When the EAWM mean state is negative, weak monsoon winds no longer obstruct the Makassar throughflow in the southern Lombok Strait. After 1960, the EAWM transitioned to directly influencing NP SSS, coinciding with a transition from a positive to negative EAWM state that is characterized by weak monsoon winds. The post-1960 weak monsoon winds likely reduced northward Indian Ocean surface water transport into the southern Lombok Strait, resulting in an enhanced southward Makassar throughflow throughout the entire Lombok Strait.

The reconstructed SSS records from GS and NP collectively indicate that changes in monsoon-driven surface water advection likely drive differing salinity responses to EAWM wind strength between the northern and southern Lombok strait. Given that SCS surface waters obstruct surface ITF flow (Gordon et al., 2003, 2012, 2014; Tozuka et al., 2007, 2009; D. Xu & Malanotte-Rizzoli, 2013), variability in the EAWM influence on surface
circulation likely impacts regional air-sea fluxes, the distribution of heat and buoyancy, and precipitation patterns. Thus, our study underscores the importance of resolving past surface ocean circulation in key ITF outflow passages to better constrain variability in the Indo-Pacific climate system.

References


