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Seasonally resolved surface water Δ^{14}C variability in the Lombok Strait: A coralline perspective

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[1] We have explored surface water mixing in the Lombok Strait through a bimonthly resolved surface water Δ^{14}C time series reconstructed from a coral in the Lombok Strait that spans 1937 through 1990. The prebomb surface water Δ^{14}C average is −60.5‰, and individual samples range from −72‰ to 134‰. The annual average postbomb maximum occurs in 1973 at 122‰. The timing of the postbomb maximum is consistent with a primary subtropical source for the surface waters in the Indonesian seas. During the postbomb period, the coral records regular seasonal cycles of 5‰ to 20‰. Seasonal high Δ^{14}C occur during March−May (warm, low salinity), and low Δ^{14}C occur in September (cool, higher salinity). The Δ^{14}C seasonality is coherent and in phase with the seasonal Δ^{14}C cycle observed in Makassar Strait. We estimate the influence of high Δ^{14}C Makassar Strait (North Pacific) water flowing through the Lombok Strait using a two end-member mixing model and the seasonal extremes observed at the two sites. The percentage of Makassar Strait water varies between 16% and 70%, and between 1955 and 1990, it averages at 40%. The rich Δ^{14}C variability has a biennial component reflecting remote equatorial Indian Ocean forcing and a component in the ENSO band, which is interpreted to reflect Pacific forcing on the Δ^{14}C signature in Lombok Strait.

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1. Introduction

[2] The Indonesian Seaway is the conduit for surface to mid-depth cross-equatorial transocean exchange between the western Pacific and Indian Ocean. *Gordon* [1986] recognized the importance of the Indonesian Throughflow (ITF) as an important contributor to the global thermohaline circulation ultimately contributing to the formation of North Atlantic Deep Water. Enhanced vertical mixing in the Indonesian Seas drives large fluxes of heat and freshwater into the water column which is ultimately incorporated into the ITF and contributes to the regional freshwater and heat budget [*Gordon and Fine*, 1996; *Hautala et al.*, 1996] with concomitant influences on regional climate.

[3] Although much of the thermocline and surface water driven by the trade winds is recirculated within the Pacific, some enters the Indonesian Seaway and flows into the Indian Ocean (Figure 1). This Indonesian throughflow (ITF) is driven by the difference in sea level between the two oceans (average 16 cm) and much of the flow is in the upper 200 m [*Wyrtki*, 1987]. Current meters in the Makassar Strait indicate reduced flow below ∼400 m [*Gordon et al.*, 1999]. The principal path is considered to be through the Sulawesi and Java Seas via the Makassar Strait with the bulk of the throughflow derived from the North Pacific supplied by the Mindanao Current [*Ffield and Gordon*, 1992]. Another path is via the Halmahera Strait which enters the Banda Sea, with throughflow derived from the South Pacific and supplied by the South Equatorial Current [*Wajsowicz*, 1993]. Waters flowing southward through the Makassar Strait enter the Flores Sea. A portion of this water directly exits the Indonesian Seas into the Indian Ocean via the Lombok Strait [e.g., *Murray and Arief*, 1988] and the remainder appears to flow eastward to the Banda Sea. Water that enters the Banda Sea is modified by vertical dynamic processes (e.g., upwelling and entrainment) that when combined with exchanges of heat and freshwater with the atmosphere can alter the heat and salt content of the water masses prior to their export to the Indian Ocean via the Sumba, Savu, and Dao Straits, and the Timor Passage [*Hautala et al.*, 1996, 2001, and references therein].

[4] Estimates of the ITF span a wide range between −2 Sv to 22 Sv with *large* seasonal and interannual variability. Current meter measurements in the Makassar Strait during 1996 through 1998 document an average flow of 9.5 ±

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2.5 Sv with much of the error derived from how the surface flow is accounted [*Gordon et al.*, 1999]. Maximum flow is near 300 db at the depth of the salinity minimum associated with North Pacific Intermediate Water with decreasing velocities below 400 db [*Gordon et al.*, 1999]. In general, surface flow is weak in austral summer and strong (to the south) in austral winter (July–August–September) during the southeast monsoon. During this short observation there is a strong correlation with the state of ENSO with diminished flow observed during the 1997 El Niño event. Just how representative this ∼2 yr time series is of the long-term mean flow is unknown. It is equally unknown how the surface flow is coupled with the flow within the thermocline.

[5] Estimates of the (∼100 m) shallow flow derived from an array of pressure gauge pairs and ADCP profiles across the five principal straits that separate the eastern Indian Ocean from the interior Indonesian Seas for December 1995–May 1999 [*Hautala et al.*, 2001] have shown that there can be large differences in the timing of the peak influx of water into the Indonesian Seas (as inferred from the Makassar Strait current meter data) and that exported to the Indian Ocean. In an admittedly simple sensitivity test *Hautala et al.* [2001] explore how the offset in timing of peak outflow (equivalent to a 5 Sv imbalance) can lead to a ∼5°C difference in thermocline temperature, and thus a significant change in the heat content, storage, and export in the Banda Sea. A similar imbalance (maximum of 4 Sv) and a ∼3 month lag (relative to the shallow pressure gauge data) was inferred from satellite altimetry and an assessment of Ekman driven divergence in the Banda Sea using NCEP reanalysis data [*Gordon and Susanto*, 2001]. Therefore it is important to understand the processes within the Indonesian Seas governing the evolution and properties of the inter-ocean exchange waters.

[6] The pressure gauge estimates of transport exhibit interannual variability that is of the same order as the long-term mean. *Hautala et al.* [2001] document intra-seasonal reversals where surface waters flow north from the Indian Ocean into the Indonesian Seas. It is thought that this reversal is a consequence of coastally trapped Kelvin waves generated on the west coast of Sumatra. The three ADCP (cruises) profiles indicate that when flow is to the south (March 1997, 1998) it occurs over the whole water column whereas when there is northward flow (December 1995), the north flow is constrained to the surface with no net flow at depth. Interestingly, although similar in mean transport, flow through the Lombok Strait had much larger variability than that in the Timor Passage with a strong correlation ($r^2 = 0.8$) between the Lombok Strait component and the total, as measured through all of the straits, shallow flow.

[7] It is thought that variations in the transport and modification of water properties due to vertical exchange within the Indonesian Seas have an interannual to decadal signature on climate [*Ffield and Gordon*, 1992; *Ffield*, 1994; *Hirst and Godfrey*, 1993; *Rodgers et al.*, 1999]. In addition to its influence on tropical Pacific climate through setting the equatorial Pacific thermocline the throughflow provides a direct conduit linking to Indian Ocean climate and circulation. Flow through the Lombok Strait is seasonally influenced by sea level height along the eastern boundary of the Indian Ocean [*Yamagata et al.*, 1996]. Higher sea level promotes downwelling and indirectly leads to a decrease in the southward flow out of Lombok Strait. The winds that promote the seasonal response are in turn influenced by ENSO which then leads to the potential of an interannual modulation of the seasonal cycle of the Lombok Strait throughflow. Thus we anticipate a complex interaction between local and remote forcing from both the Pacific and Indian Ocean on the ITF.

[8] Atmospheric nuclear weapons testing in the late 1950 s and early 1960 s resulted in an excess of ^{14}C in the atmosphere and as this signature has penetrated the ocean it has augmented the natural gradient between the surface and deeper waters. Isotopic equilibration with atmospheric $^{14}C/^{12}C$ is on the order of a decade [*Broecker and Peng*, 1982] and thus $\Delta^{14}C$ in surface waters can be used as a quasiconservative, passive advective tracer. Time series derived from archives such as hermatypic corals can augment historical, conventional (temperature, salinity) observations especially in times and regions where observations are sparse. Corals act like strip chart recorders continuously recording the radiocarbon content of the waters in which they live. Ocean dynamics can be reconstructed and studied from these biogenic archives.

[9] Measurements of coral skeletal material which accurately record the $\Delta^{14}C$ of ΣCO_2 [e.g., *Druffel*, 1981, among others] have added important information to water sampling programs like GEOSECS [*Östlund et al.*, 1987] and the World Ocean Circulation Experiment (WOCE [*Key et al.*, 1996]). There are notable limitations to shipboard sampling, primarily the inability to continuously monitor ocean conditions. For ^{14}C in the deep ocean, this is not a problem because the transport is relatively slow and the gradients are relatively low. For the surface ocean, where ^{14}C gradients are highest and transport is rapid, it has been demonstrated that temporal variability in surface $\Delta^{14}C$ can be of the same order as spatial variability [e.g., *Guilderson et al.*, 1998], an observation which is lost in discrete analyses like GEOSECS or WOCE whose "snapshots" of bomb radiocarbon are integrations of ∼20 and ∼40 yr (respectively) of ocean dynamics.

[10] We have chosen sites (Figure 1) to monitor variations in the transport of water out of the Pacific (Langkai) and into the Indian Ocean (Padang Bai). Our multidecadal continuous $\Delta^{14}C$ records complements the existing physical and chemical data sets from oceanographic expeditions of the Indonesian region and will prove to be a valuable tool for exploring circulation through the Indonesian Seas. The results presented here are placed in a dynamic context through comparison with a new record from the Makassar Strait [*Fallon and Guilderson*, 2008], and a previously published record from the northwest coast of Sumatra in the Mentawai Islands [*Grumet et al.*, 2004]. A simple end-member mixing model to elucidate the fraction of higher-^{14}C Makassar Strait (inferred to be of northern hemisphere subtropical origin) relative to lower-^{14}C Indian Ocean water is constructed from the data. Although a time series of percentage of water masses is not strictly flux, such a reconstruction provides a heretofore unknown view of the history of water mass mixing in the Indonesian Seas.

2. Methods

[11] Coral cores were drilled in January 1990 from an exceptionally large *Porites* colony growing at a depth of

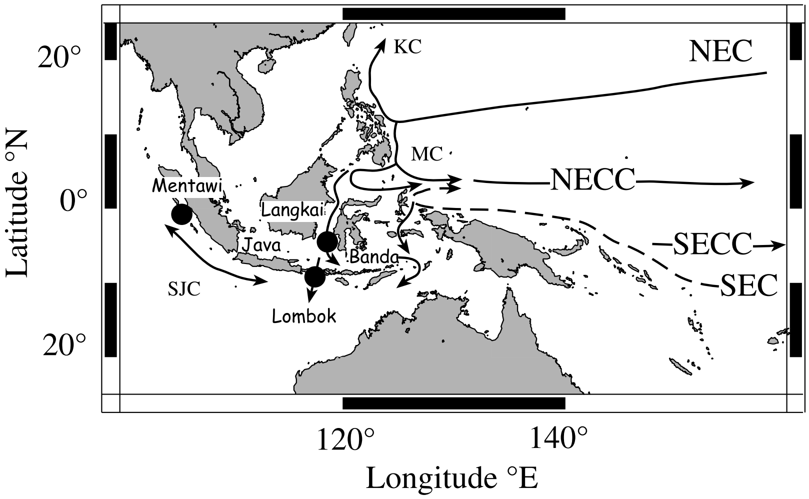

Figure 1. Map of the Indonesian region, and schematic representation of surface and near surface currents (adapted from *Lukas et al.* [1996], *Gordon et al.* [1999], and *Hautala et al.* [2001]). Coral locations (solid circles) discussed in the text are Lombok Strait (Padang Bai, Bali), Makassar Strait (Langkai), and Sumatra (Padang Is, Mentawai). Currents denoted include the North/South Equatorial Current (NEC/SEC), North/South Equatorial Counter Current (NECC/SECC), the Mindanao and Kuroshio Currents (MC, KC), and the seasonally reversing South Java Current (SJC).

5 meters at Padang Bai, Bali (8°15′S, 115°30′E) on the southern edge of the Lombok Strait. The recovered coral cores span nearly 300 yr of growth and a long (~1 mm resolved) stable isotope record can be found in the work of *Charles et al.* [2003]. We have focused our radiocarbon work on the uppermost ~70 cm of the coral which mainly covers the pre to postbomb period.

[12] The cores were cut into ~9 mm slabs, ultrasonically cleaned in distilled water, and air dried. After identifying the major vertical growth axis, the coral was sequentially sampled at 2 mm increments with a low-speed drill. Where necessary, we overlapped parallel sample tracts in order to adequately splice sections together. Splits (~1 mg) were reacted in vacuo in a modified autocarbonate device at 90°C and the purified CO_2 analyzed on a gas source stable isotope ratio mass spectrometer. Stable isotope data are reported in standard per mil notation relative to Vienna Pee Dee Belemnite [*Coplen*, 1993]. Analytical precision based on an in-house coral standard is better than ±0.05‰ (1 s) for both oxygen and carbon. The remaining sample splits (8–9 mg) were placed in individual reaction chambers, evacuated, heated, and acidified with orthophosphoric acid at 90°C. The evolved CO_2 was purified, trapped, and converted to graphite in the presence of iron catalyst and a stoichiometric excess of hydrogen [*Vogel et al.*, 1987]. Graphite targets were measured at the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory. Radiocarbon results are reported in conventional per mil (‰) notation as age corrected $\Delta(^{14}C)$ as defined by *Stuiver and Polach* [1977] and include a background correction using ^{14}C-free calcite, and the $\delta^{13}C$ correction obtained from the stable isotope results. Analytical precision and accuracy of the radiocarbon measurements is ±3.5‰ (1σ sd).

[13] Embedded in the oxygen isotopic composition of the coral's aragonite skeleton is information on temperature and the oxygen isotopic composition of the water (δ_w) in which the coral resided [e.g., *Epstein et al.*, 1951]. The δ_w of the water is directly related to salinity [*Craig and Gordon*, 1965] with low salinities having low $\delta^{18}O$ and the converse true for high salinity waters. Thus salinity is often used as a proxy for δ_w. The average Lombok Strait sea surface temperature is ~28.7°C with an annual range on the order of 2°C [*Sprintall et al.*, 2003; *Levitus and Boyer*, 1998; *Reynolds and Smith*, 1994], but not with what one would consider a clear annual cycle. Mean annual salinity is ~33.4 psu [*Sprintall et al.*, 2003; *Levitus and Boyer*, 1998] with a solid seasonal cycle in excess of 2 psu. In the Lombok Strait sea surface temperature maxima and salinity minima co-occur in late March–May and correspondingly highest salinities tend to occur during the lowest sea surface temperatures in September–October. At this location the seasonal signal in salinity and temperature are reinforced in the $\delta^{18}O_{coral}$ with the annual $\delta^{18}O_{coral}$ cycle being primarily driven by salinity (δ_w) variations.

[14] Coral chronology has historically relied upon the presence of annual high- and low-density band couplets [e.g., *Dodge and Vaisnys*, 1980] or the seasonal variability in coral $\delta^{13}C$, which is thought to reflect surface irradiance [e.g., *Shen et al.*, 1992]. Independent chronologies based on these two methods on the same coral specimen tend to agree within a few to 6 months. We created a preliminary age model based upon the seasonal structure within the $\delta^{13}C$ record and sclerochronology but in order to obtain the best timescale we have refined our age model by correcting the preliminary age model through coral $\delta^{18}O$ comparisons with instrumental records [e.g., *Guilderson and Schrag*, 1999]. Chronological assignments for the core were straightforward due to the clear seasonal cycle in $\delta^{18}O$. The seasonal extremes were anchored to March/April and September of each year consistent with instrumental observations of temperature, salinity, and cloudiness [*Charles et al.*, 2003;

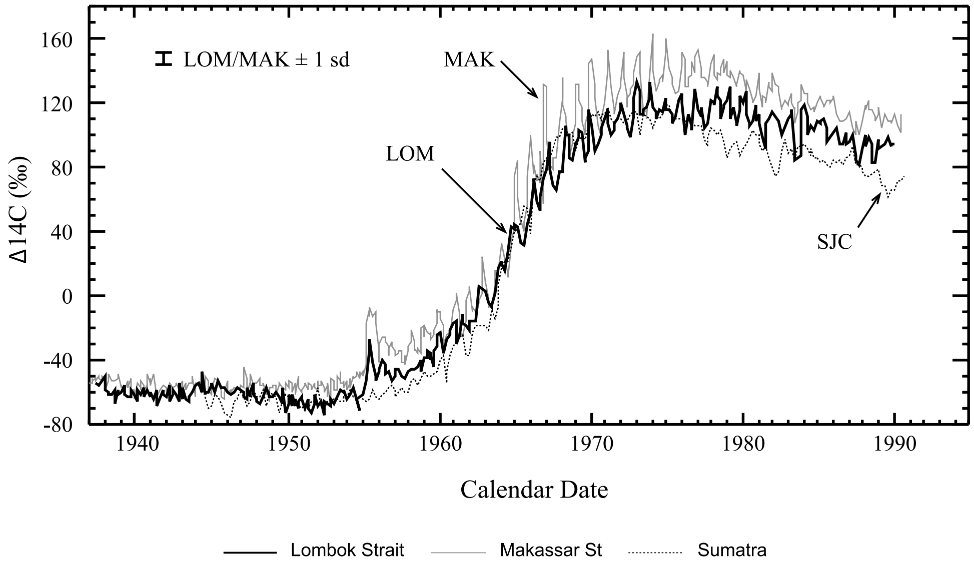

Figure 2. Surface ocean Δ^{14}C as reconstructed from reef-building hermatypic corals from Lombok Strait (LOM; thick solid line), Makassar Strait (MAK; thin gray line), and Sumatra (SJC; thin dotted line). The Lombok and Makassar data have a 1σ SD of ±3.5‰. Coral chronologies were derived from independent δ^{18}O records anchored to seasonal extremes in sea surface temperature and salinity ($\delta^{18}O_w$).

Sprintall et al., 2003]. Age errors are estimated to be approximately one to two months.

3. Results

[15] The Δ^{14}C time series spans $\sim$1937.5–1990. The $\sim$14 mm/a apparent linear growth rate yielded seven to eight Δ^{14}C samples per year with an average resolution of $\sim$1.5 months. Individual $\sim$bimonthly Δ^{14}C values range from −72‰ to 134‰ (Figure 2). The average Δ^{14}C prebomb (1937–1950) value is −60.5 ± 4.2‰, or 501 radiocarbon years. Over the prebomb interval the Lombok coral surface water Δ^{14}C is intermediate in value between Makassar Strait and lower Δ^{14}C Indian Ocean surface water as recorded at Mentawai. Between 1947 and 1954 the data trends to slightly more negative values: $\sim$−68‰ in 1951 and 1952. In 1954 Δ^{14}C values average −62.5‰ and are followed by a rise of $\sim$30‰ during 1955. Values decrease slightly in 1956 and into 1957 before rising toward the postbomb peak. The mean annual postbomb maximum occurred in 1973 (122‰). Mean annual Δ^{14}C values remain between $\sim$110‰ and $\sim$120‰ through 1982 and then begin a slow decrease until the end of the record in 1990.

[16] In the prebomb portion of the record seasonality is irregular. Seasonality is distinct in nearly all of the post-bomb (post-1955) years and ranges between $\sim$5 and $\sim$20‰, with most years $\sim$10‰ (Figure 2). Seasonal Δ^{14}C maxima occur coincident with warm temperatures and low salinity (more negative $\delta^{18}O_{coral}$) and conversely Δ^{14}C minima occur with more positive $\delta^{18}O_{coral}$ values (cooler, saltier water).

4. Discussion

[17] Prior to atmospheric weapons testing and oceanic uptake of "bomb-^{14}C" the mid to low latitude surface water Δ^{14}C gradients were small. In the western equatorial Pacific and eastern Indian Ocean the gradient is on the order of 17‰: from a high of $\sim$−47‰ for the North Pacific subtropics [e.g., *Druffel et al.*, 2001] to −64‰ at Penang Island (0°0.8′S, 98°31′E) off the northwest coast of Sumatra in the Mentawai Islands [*Grumet et al.*, 2004]. We use the seasonal cycle in Δ^{14}C in Lombok Strait to infer dynamic (mixing) processes and potential transport as reflected as mixing through the Lombok Strait. Admittedly, by doing so we simplify the complicated seasonal changes in surface currents that occur in the eastern Indian Ocean: the South Java Current flows southeast during the monsoon transitions (May and November) whereas during the rest of the year it flows north westward [*Tomczak and Godfrey*, 2002; *Schott and McCreary*, 2001]. The low ^{14}C water observed at Penang Island is brought to the surface by upwelling along the Sumatra coast which occurs when the South Java Current flows west and when local winds lead to divergence. Using the Penang Island record as a pseudo-endmember is probably reasonable during the prebomb interval where Δ^{14}C seasonality and gradients in the eastern Indian Ocean are small. In the postbomb era we use the intrinsic seasonal Δ^{14}C in the Lombok Strait coral record in our assessment of the evolution and mixing of Lombok Strait surface water.

[18] Between 1938 and 1944 the Lombok Strait Δ^{14}C data are 5–10‰ more negative than the corresponding Makassar Strait record. Between 1944 and 1947 Lombok and Makassar Strait coral Δ^{14}C are indistinguishable from each other and values are 10–15‰ more positive than surface waters off the west coast of Sumatra at Penang Island. From 1947 to 1954 Lombok Δ^{14}C values are equivalent to Penang Island and both records are 10–15‰ more negative than values recorded in the Makassar Strait. When Lombok and Penang Island Δ^{14}C are similar we infer enhanced upwelling on the west coast of Sumatra and Java and assume that there is a concomitant northward

flow through the Lombok Strait. When Lombok and Makassar Strait Δ^{14}C are more similar we infer the converse: less upwelling, less potential influence of Indian Ocean water backfilling the Indonesian Seas through Lombok, and potentially more flow through Makassar and Lombok Straits.

[19] The distinct ~30‰ transient Δ^{14}C in 1955 is a unique feature, until recently previously unrecognized in circum-Pacific coral-based Δ^{14}C reconstructions [*Fallon and Guilderson*, 2008]. The feature occurs well before a significant rise in atmospheric ^{14}C concentrations [e.g., *Manning and Melhuish*, 1994; *Stuiver and Quay*, 1981] and is therefore unlikely to be the result of air-sea ^{14}C exchange given the ~decadal time delay for isotopic equilibration [*Broecker and Peng*, 1982]. *Toggweiler and Trumbore* [1985] documented a ^{90}Sr (a weapons fallout product) peak in the skeletal material corresponding to 1955 in a coral from Cocos Island in the Indian Ocean. Given the spatial distribution and timing of individual atomic weapons tests and the constraints of air-sea isotopic exchange, the path of least resistance to reconcile the ^{90}Sr and rapid ^{14}C increase is for waters from near Bikini Atoll (11°35′N, 165°23′E, where the early atmospheric weapons tests were conducted) to be transported via the North Equatorial Current to the Mindanao Current to be transported into the Indian Ocean through the Lombok Straits.

[20] The timing of the postbomb Δ^{14}C peak is consistent with air-sea isotopic equilibration of ~10 yr relative to the atmospheric peak in the early 1960s [*Druffel*, 1987] and a subtropical origin of much of the surface water. One could argue that there is not an individual single "peak" but that between 1973 and 1982 Δ^{14}C values are equivalent before they begin to turn down. The elevated value reflects the slow penetration and dilution of bomb-^{14}C with interior waters that upwell in the tropical Pacific and Indian Ocean. Penetration of bomb-^{14}C laden water into newly subducted subtropical mode waters will (*eventually*) be entrained and upwelled at low latitudes to recycle its ^{14}C signature. A "stretched" bomb peak, such as observed at Lombok Strait, reflects the mixing of subtropical "enriched" ^{14}C waters with a source of low ^{14}C water that is also evolving toward more positive values [*Guilderson et al.*, 2000]. The stretching integrates the ventilation time of the upwelled water mass and the water mass' ^{14}C evolution which includes the Δ^{14}C value when it was initially subducted and mixing and entrainment along the subsurface pathway.

[21] To characterize the visually striking variability that we see in the postbomb era we determined the spectral characteristics of an equal (0.2 yr^{-1}) interval linear interpolated record after passing the record through a high pass (21 weight, tukey-cosine, 10 yr half amplitude) filter to remove the long-term "bomb-^{14}C" trend. Spectral analysis of the Δ^{14}C time series confirms the visually striking seasonal cycle, biennial and interannual periodicity (Figure 3). This is in contrast to the Mentawai Δ^{14}C record which has a much more dominant ~ 3 yr periodicity [*Grumet et al.*, 2004]. The 2-yr periodicity is likely a reflection of the SE Asian monsoon which has a strong biennial component [*Meehl*, 1997] and remote wind-forcing in the Indian Ocean which influences the sea surface temperatures and potentially the passage of waters through the Lombok Strait [*Wijffels and Meyers*, 2004]. It is interesting that the Δ^{14}C record exhibits a biennial period because the Padang Bai δ^{18}O record does not [*Charles et al.*, 2003]. This is an example of the important and subtle difference between a true water mass tracer such as Δ^{14}C versus δ^{18}O in corals which is a combination of temperature and $\delta^{18}O_w$ (salinity): two tracers that over a few months can be significantly modified by air-sea processes and thus are not completely conservative water mass tracers. Although the Indonesian region is impacted by ENSO the Δ^{14}C (and δ^{18}O) of Lombok Strait surface water is not, *sensu strictu*, a simple recorder of ENSO events. Interannual "events" although consistent with ENSO periodicities do not bear a simple linear correspondence with, for example, the Southern Oscillation Index modulated by the SE Asian monsoon, or events in the Indian Ocean [see also *Charles et al.*, 2003]. This is because Δ^{14}C in the Lombok Strait reflects not only the mixing of Indonesian Seas and Indian Ocean water, but the temporal evolution of the Δ^{14}C at the source regions of the individual water masses.

[22] In the Hautala et al., SPGA (1995–1998) study only 20% of the variability was in the seasonal cycle, and 50% was intraseasonal. In a modeling study forced with 14 yr (1985–1999) of ECMWF (observed) 3-day winds, *Poterma et al.* [2003] estimate that over the 14-yr span 48% of the variability is associated with the seasonal cycle. Within the bimonthly resolution, our record supports surface water export and/or mixing processes that are dominated by the seasonal cycle with a subdominant intraseasonal periodicity. The spectral analysis documents a biennial and interannual (ENSO) component which indicates a strong link between the Δ^{14}C signature in Lombok Straits and a combination of remote forcing from both the Pacific and Indian basins. *Wijffels and Meyers* [2004] analyzed XBT data from transects, starting in ~1983, across the Banda Sea from Australia to Java (PX2 line) and in the Indian Ocean along a line from Freemantle Australia to Sumatra (IX1 line). In this study they specifically explored the influence of Pacific, Indian, and remote winds on the observed temperature variability. To the north of Lombok Strait sea surface temperature exhibits seasonal variability whereas ENSO variability is focused in the thermocline and deeper. Off the southwest coast of Java seasonal variability is weak in the surface waters and only becomes strong at depth. This is in contrast to the biennial variance which occurs in the mixed layer and penetrates through the thermocline to depth. In the thermocline off Freemantle there is significant variance in the ENSO band. Wijffels and Meyers then used a linear regression model to document the relative influence of remote Indian, Pacific, and local wind-forcing on the temperature variability in the XBT transects. They came to the conclusion that Indian Ocean winds have a significant role in the interannual variability on the temperature characteristics and posit that the same forcing influences flow through the western portion of the Indonesian Seas. Our results support their conclusion of a remote equatorial Indian Ocean forcing on the hydrography and mixing in the Lombok Strait region.

[23] To explore the relationship between surface waters in Makassar and Lombok Straits we passed the respective δ^{18}O and Δ^{14}C records individually through a Gaussian filter centered on the annual cycle (1 ± 0.3 hz). The (visual) correspondence of the individual strait's δ^{18}O (temperature/

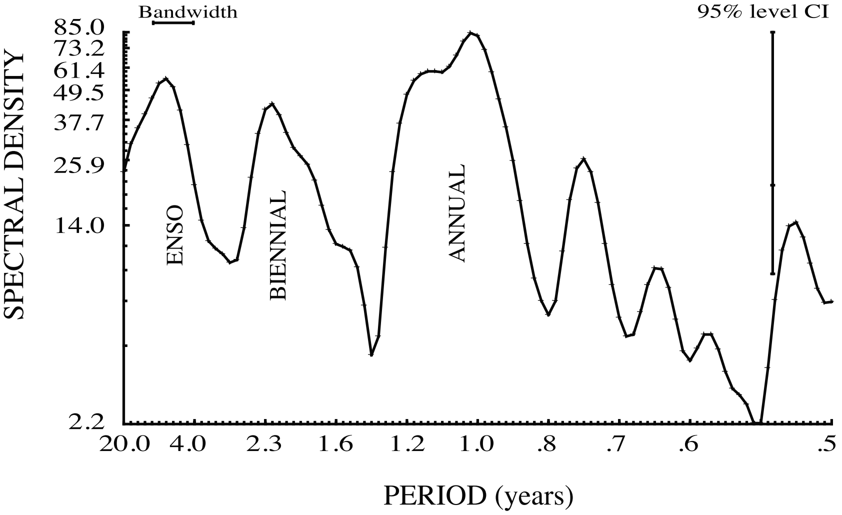

Figure 3. Spectral analysis of the post-1950 Lombok coral Δ^{14}C time series as determined from a one-third lagged autocovariance function using the ARAND statistical package [*Howell et al.*, 2006]. The time series was the first equal interval linearly interpolated (0.2 yr^{-1}) to remove the longer-term increase in ^{14}C and then passed through a high-pass tukey-cosine filter with a half amplitude at 10 yr.

salinity) and Δ^{14}C is confirmed: they are coherent and nearly always in phase (Figure 4). There are instances where the Lombok Strait's seasonal high Δ^{14}C value is lagged relative to the δ^{18}O by one sample. Not surprisingly and entirely due to the mechanics of creating the age model, δ^{18}O between Makassar and Lombok Straits is in phase. The seasonal cycle in radiocarbon between the two sites is also highly coherent and in phase. Although we have a priori fixed the calendar month (using δ^{18}O) there was no guarantee that the coral Δ^{14}C seasonal cycles would be in phase. If we make the logical first order assumption that Makassar and Lombok Strait share a common source of high Δ^{14}C surface water, then the fact that the Δ^{14}C seasonal cycles are in phase implies, within the resolution of our sampling, little to no lag in the transport through the two "coral-based observation platforms". The implicit assumptions used in the construction of the age model does force some amount of correspondence at least within the sample resolution (~bimonthly). If we were able to derive a completely independent coral calendar age model we might be able to tease out a lag smaller than several months.

[24] Following the idea that the Makassar and Lombok Straits' high Δ^{14}C is sourced from the same North Pacific water that seasonally flows southward through the Makassar Strait [*Fallon and Guilderson*, 2008; *Gordon et al.*, 1999, 2003], we explore the influence of this North Pacific water with Indian Ocean water. Admittedly the relative dilution of North Pacific (Makassar Strait high Δ^{14}C water) with lower Δ^{14}C "Eastern Indian Ocean" water is not a quantitative measure of the flux. It should however provide a sense of past fluxes and dynamics that would otherwise be unattainable. Indeed, if the Lombok Strait's Δ^{14}C looked exactly like that of the Makassar Strait this would imply no Indian Ocean water at all nor any modification of water masses in the Indonesian Seas. In this case, one could infer that the surface "transport" was in effect a "sloshing" back and forth like water in a bathtub going from end to end. We point out that in detail the postbomb Lombok Δ^{14}C time series does not exactly look like that of the South Java Current, at least as reconstructed in the Mentawai coral record (Figure 2). For much of the period of study the Mentawai coral Δ^{14}C is lower than that of the corresponding seasonal Δ^{14}C low as reconstructed at Padang Bai. There is an interval during the late 1960s when Mentawai Δ^{14}C is higher than Lombok Strait. For the intervals where the Lombok seasonal low is higher than Mentawai this means that surface waters have entrained some amount of higher Δ^{14}C water prior to flowing northward back into the Lombok Strait, and conversely the periods where the seasonal lower there is additional upwelling or entrainment of low-^{14}C water that is not captured in the Mentawai record. An alternative interpretation is that the seasonal Δ^{14}C low at Lombok is a signature from the Banda Sea.

[25] To estimate the relative percent of Makassar Strait water we use the seasonal high Δ^{14}C from the Langkai coral record (MAK end-member) and the corresponding seasonal high Padang Bai coral Δ^{14}C value (Lombok measured). The "Eastern Indian Ocean" end-member is selected from the preceding seasonal low Δ^{14}C in the Padang Bai record. A potential confound to this analysis is the influence of Banda Sea water. Due to upwelling and vertical entrainment of subthermocline waters [e.g., *Ffield and Gordon*, 1992; *Hautala et al.*, 2001] we expect Banda Sea Δ^{14}C to be lower than that observed in the Makassar Strait, at least until significant incorporation of bomb-^{14}C into subthermocline waters in the Banda Sea has taken place. During the southwest monsoon Banda Sea surface water gives the appearance of "back filling" the Java Sea [*Gordon et al.*, 2003]. If Banda Sea water passes through Lombok Strait it would influence the Padang Bai Δ^{14}C signal that we have measured, and our reference point would be Banda Sea water and not Eastern Indian Ocean water: i.e., we would be assessing mixing within the Indonesian Seas not between the ITF and the Indian Ocean. This could be empirically explored through the acquisition

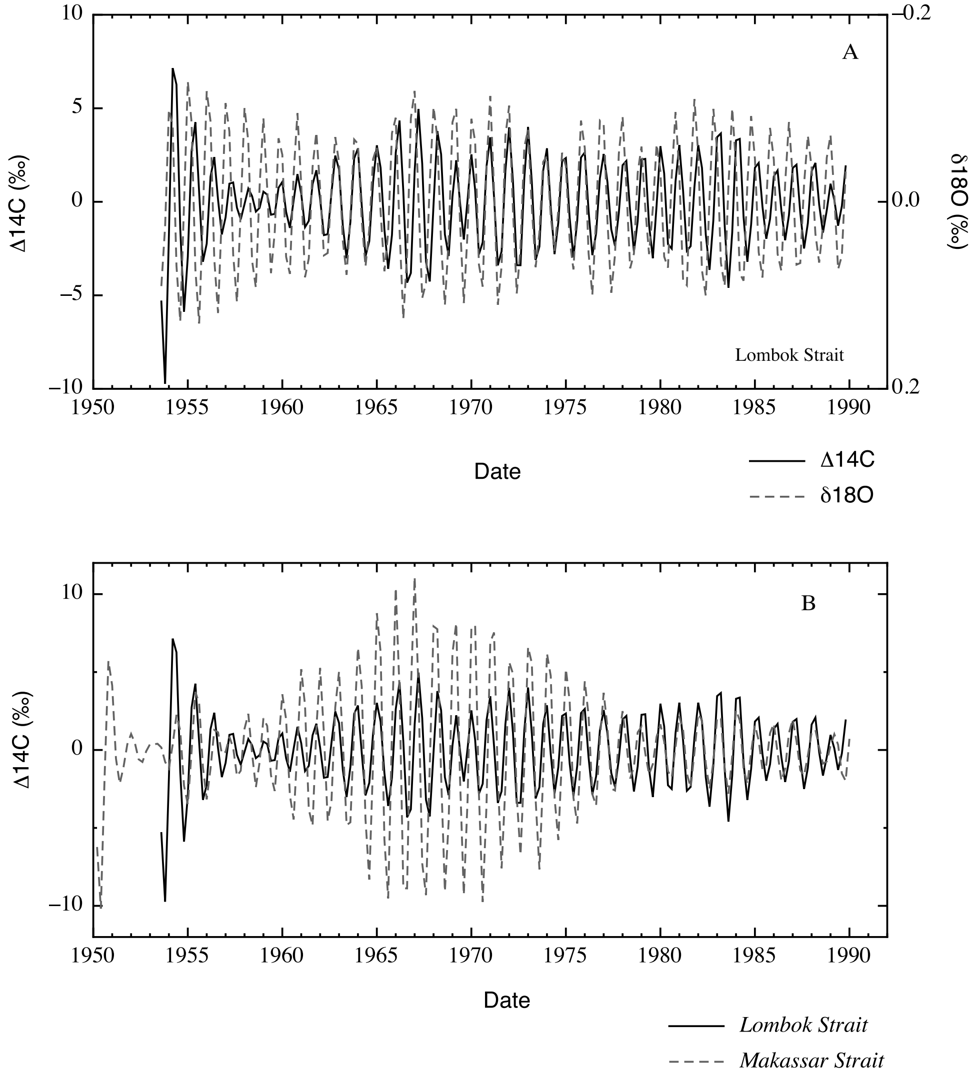

Figure 4. (a) Lombok coral δ^{18}O (dotted gray line) and Δ^{14}C (solid black line) passed through a 1-yr Gaussian filter (1 ± 0.3). Note that Δ^{14}C and δ^{18}O are coherent and nearly always in phase. (b) Similarly filtered Lombok (solid black line) and Makassar (dotted gray line) Strait Δ^{14}C. There is little lag between the Δ^{14}C seasonal cycle at the two locations.

of a Δ^{14}C time series from a well-chosen and representative site in the Banda Sea.

[26] We do not estimate the percent of MAK water in 1961 and 1962; these are the years when atmospheric Δ^{14}C is rapidly increasing and surface water Δ^{14}C is strongly influenced by air-sea isotope exchange which during these years can confound using Δ^{14}C as a surface water mass tracer. For all other years we calculate the percent of MAK water in the measured Lombok water using the simple two end-member mixing model. The results of this simple mixing model experiment are presented in Figure 5. The percent of MAK water averages 40% and ranges from a low of 16% to a high of 70%. Visually there is a hint of a long-term decrease in the (seasonal) influence of MAK water in the Lombok Strait. On either side of 1975 the average percentage of MAK water is not quite statistically different (44 ± 15, n = 18 versus 35 ± 15, n = 15; *t* test 1.716 and p 0.0961). The record is too short to confirm a multidecadal component to the variability. The interannual variability is

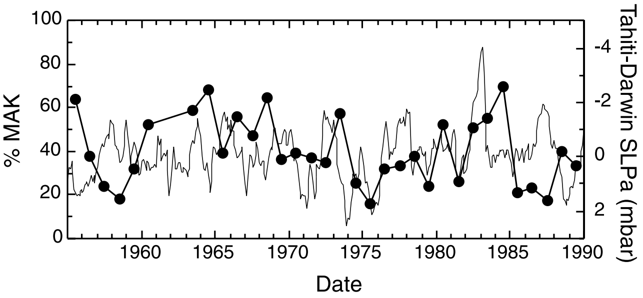

Figure 5. Seasonal influence of high Δ^{14}C "Makassar Strait" water in the Lombok Strait derived from a two-component mixing model. The average over the time series is 40% and ranges from 16 to 70%. The thin solid line is the seasonal Tahiti-Darwin sea level pressure anomaly (SLPa), the Southern Oscillation Index, with a 3-month running mean filter applied. A positive SOI value is usually correlative with a La Niña event, and the converse is true with El Niño events.

sometimes expressed as increased influence of MAK water during the strong La Niña events that in general terminate strong El Niño events. The sense of the influence of MAK water is consistent with our understanding of transport through the ITF based on present-day observations. During strong La Niña events there is a buildup of water and sea level height in the western equatorial Pacific which can lead to increased transport through the ITF. Particularly important in modulating the transfer of water from the Pacific to Indian Ocean is the establishment of buoyant, low salinity plugs that provide resistance to flow [*Gordon et al.*, 2003] as well as the surface winds in the Indian Ocean basin [*Yamagata et al.*, 1996; *Wijffels and Meyers*, 2004]. This interplay between competing and complementary dynamics influences the shallow water mixing and transport, the Δ^{14}C signature, and ultimately the heat and salt budget of the throughflow.

5. Conclusions

[27] To infer surface water mixing in the Lombok Strait over the last ~50 yr we have reconstructed the surface water Δ^{14}C history using a reef-building hermatypic coral cored off Padang Bai, Bali. The ~bimonthly record exhibits strong seasonality that is coherent and in phase with a similar data set acquired in the Makassar Strait [*Fallon and Guilderson*, 2008]. Using admittedly simplistic assumptions regarding the seasonal transport of high Δ^{14}C water through Makassar and Lombok Strait, we estimate the percentage of high Δ^{14}C water over the time frame common to the two records. The percentage of high Δ^{14}C "Makassar Strait" water varies between 16 and 70% with a mean of 40%. In addition to interannual variability there is a hint of multidecadal variability.

[28] The Lombok Strait Δ^{14}C time series is remarkably rich data set that also exhibits biennial variability. The biennial variability reflects the influence of the southeast Asian monsoon on the regional dynamics and the movement of surface waters through the Lombok Strait. Indeed, it is very enheartening that from what many oceanographers would consider an "unconventional" tracer time series we find results that are consistent in the spectral domain with more conventional hydrographic studies using XBTs and wind observations. The Δ^{14}C time series has characteristics reflective of Indian Ocean (biennial) and Pacific (ENSO) forcing. In time, we plan to expand the spatial coverage of time series such as the one presented here.

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