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Surface water processes in the Indonesian Throughflow as documented by a high-resolution coral $\Delta^{14}C$ record

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Abstract

To explore the seasonal to decadal variability in surface water masses that contribute to the Indonesian Throughflow we have generated a 115-year bi-monthly coral-based radiocarbon time-series from a coral in the Makassar Straits. In the pre-bomb (pre-1955) era from 1890 to 1954, the radiocarbon time series occasionally displays a small seasonal signal (10-15‰). After 1954 the radiocarbon record increases rapidly, in response to the increased atmospheric ^{14}C content caused by nuclear weapons testing. From 1957 to 1986 the record displays clear seasonal variability from 15 to 60‰ and the post-bomb peak (163 per mil) occurred in 1974. The seasonal cycle of radiocarbon can be attributed to variations of surface waters passing through South Makassar Strait. Southern Makassar is under the influence of the Northwest Monsoon, which is responsible for the high Austral summer radiocarbon (North Pacific waters) and the Southeast Monsoon that flushes back a mixture of low (South Pacific and upwelling altered) radiocarbon water from the Banda Sea. The coral record also shows a significant ^{14}C peak in 1955 due to bomb ^{14}C water advected into this region in the form of $CaCO_3$ particles (this implies that the particles were advected intact and then become entrapped in the coral skeleton - is this what we really mean? Wouldn't even fine particles settle out over the inferred transit time from Bikini to MAK?) or water particles with dissolved labeled CO_2 produced during fallout from the Castle tests in 1954.

Introduction

The Indonesian throughflow (ITF) is thought to play an important role in global thermohaline circulation and influence global climate by funneling Pacific Warm Pool

46 water into the Indian Ocean [*Broecker*, 1991; *Gordon*, 1986; *Gordon and Piola*, 1983;
47 *Hirst and Godfrey*, 1993]. Observations suggest the ITF is composed of North Pacific
48 subtropical and thermocline waters that flow through Makassar Strait [*Ffield and*
49 *Gordon*, 1992; *Fine*, 1985; *Gordon and Fine*, 1996; *Ilahude and Gordon*, 1996]. At the
50 southern end of Makassar Strait water flows through Lombok Strait to the south and
51 eastward into the Banda Sea ending up in the Indian Ocean (Figure 1). An additional
52 component of deep South Pacific water enters Eastern Indonesia and mixes in the Banda
53 Sea [*Ffield and Gordon*, 1992; *Hautala, et al.*, 1996]. The flow of waters through the
54 ITF is highly variable, estimates of –2 to 20 Sv have been reported [*Godfrey*, 1996;
55 *Gordon and Fine*, 1996; *Lukas, et al.*, 1996; *Myers*, 1996; *Potemra, et al.*, 1997]. The
56 atmospheric pressure gradient between the Pacific and Indian Ocean which results in a
57 dynamic height difference between the two oceans is the main driving force for the ITF
58 [*Wyrtki*, 1987]. The dynamic height difference is greatest during the Southeast Monsoon
59 (Southern Hemisphere winter) resulting in the strongest flow [*Gordon, et al.*, 1999;
60 *Murray and Arief*, 1988; *Wyrtki*, 1987]. During the Northwest Monsoon (Southern
61 Hemisphere summer) the gradient lessens, thereby weakening the throughflow [*Gordon,*
62 *et al.*, 1999; *Murray and Arief*, 1988; *Wyrtki*, 1987]. Additionally, the NW monsoon
63 decreases the surface salinity due to a large influx of freshwater via rainfall and runoff.
64 Throughflow variability and flow volume is correlated with the state of ENSO
65 [*Gordon, et al.*, 1999]. During La Niña events an increase in the sea surface height in
66 the Western Pacific Warm Pool region results in an increase in the dynamic height
67 gradient with the Indian Ocean and a corresponding increase in the throughflow volume,
68 with the converse occurring during El Niño events [*Gordon, et al.*, 1999]. Vertical

mixing influences the composition of the ITF. Heat and freshwater are transported down toward the thermocline and cool water mixes upward potentially influencing the atmosphere-ocean heat flux [*Ffield and Gordon*, 1992; 1996]. The ITF results in a significant export of heat and freshwater from the tropical Pacific into the Indian Ocean and may influence atmosphere-ocean coupling, tropical SST patterns and the Asian Monsoon [*Schneider*, 1998]. One way to examine the sources and variability of the throughflow is to use a water mass tracer. The radiocarbon ($\Delta^{14}C$) content of waters can be used to track ocean currents, vertical mixing, and air-sea CO_2 exchange [*Druffel*, 1985; *Druffel*, 1987; *Guilderson, et al.*, 2000a; *Guilderson, et al.*, 2000b; *Guilderson, et al.*, 1998; *Moore, et al.*, 1997].

Measurements of sub-annual samples from coral skeletal material provide a proxy time-series record of the $\Delta^{14}C$ of the dissolved inorganic carbon (DIC) of the surrounding seawater [*Druffel*, 1981; *Druffel and Suess*, 1983; *Guilderson, et al.*, 1998; *Moore, et al.*, 1997]. In this paper, we present the radiocarbon content in roughly bi-monthly samples from a coral in the Makassar Strait to study the processes governing water mass evolution in the Indonesian Seas over time scales long enough to study the linkage between processes in the Indonesian Seas and decadal modulation of ENSO, Asian Monsoon and global climate.

Methods

Coral

A 2.3m core was drilled from a *Porites lutea* coral off Langkai Island (5°02′ S, 119°04′E; Figure 1) [*Moore, et al.*, 1997]. The core was cut to a thickness of 9mm [*Moore, et al.*, 1997] and sonicated in distilled water. This study builds on the original

93 Langkai $\Delta^{14}C$ dataset that spanned 1970-1985 published by Moore, et al. [1997]. The
94 samples reported here were re-milled from the same coral core as Moore, et al. [1997]
95 along a main growth axis Samples were milled sequentially in 1.8 mm increments using
96 a manual low-speed mill with a 2mm diameter mill bit, corresponding to ~6-8 samples
97 per year. ^{14}C sample splits (8-10 mg) were reacted in individual chambers (BD
98 Vacutainer), evacuated, heated and acidified with orthophosphoric acid at 90°C
99 [*Guilderson, et al.*, 2000b; *Guilderson, et al.*, 1998]. The CO_2 was purified, trapped, and
100 converted to graphite using an iron catalyst following a method similar to that described
101 by Vogel, et al., [1987]. The graphite targets were analyzed at the Center for Accelerator
102 Mass Spectrometry, Lawrence Livermore National Laboratory. The ^{14}C results are
103 reported as age-corrected $\Delta^{14}C$ (‰) as defined by Stuiver and Polach [1977] and include
104 $\delta^{13}C$ correction for isotope fractionation, and a blank subtraction based on ^{14}C-free
105 calcite. Radiocarbon accuracy and precision is ±3.5‰ (1σ) based on long-term statistics
106 of secondary standards. A coral tertiary standard was analyzed with the unknown coral
107 samples with a resulting total precision of ±0.75‰ (1σ) (n=88) (Figure 2). This result is
108 consistent with the analytical uncertainty of the measurements. Additional sample splits
109 (~100 μg) were analyzed on a Finnigan MAT 252 with a Kiel carbonate device using
110 105% H_3PO_4 at 90ºC. Stable isotope data are reported relative to the Vienna Pee Dee
111 Belemnite (V-PDB) with an average analytical precision of ±0.03‰.
112 An initial age-model and time series was constructed using the seasonal
113 variability in coral $\delta^{13}C$. In some tropical reef locations the coral $\delta^{13}C$ is modulated by
114 seasonal light cycles. Higher light levels are associated with more positive $\delta^{13}C$ whereas
115 lower light levels are associated with more negative $\delta^{13}C$ [*Fairbanks and Dodge*, 1979;

116 *McConnaughey*, 1989]. For this project we are assuming that the coral $\delta^{13}C$ primarily
117 tracks variability in solar insolation and that $\delta^{13}C$ inflection points are interpreted as
118 markings of seasonal change. In this part of Indonesia, the maximum cloudiness and
119 maximum rainfall occur in February/March. To generate the primary time series we
120 pinned the minimum $\delta^{13}C$ to February of each year, and linearly interpolated the data
121 between marker points. We then used the seasonal cycle of $\delta^{18}O$, which is influenced by
122 water temperature and fresh water to fine-tune the time series [references: Michael
123 Moore's dissertation and the Charles et al., Mar Geol paper]. Coral $\delta^{18}O$ is negatively
124 correlated with water temperature, the summer $\delta^{18}O$ coral signal is heavily influenced by
125 rainfall and runoff, so we matched the coral $\delta^{18}O$ maxima to the winter SST minimum
126 from the GOSTA SST atlas [*Bottomley, et al.*, 1990]. With our sampling resolution this
127 results in a time series that is precise to ± 2 months.
128
129 ***Results and Discussion***
130 The coral $\Delta^{14}C$ time series spans 1870 to 1990. The pre-bomb interval (1870-
131 1950) has an average $\Delta^{14}C$ value of -56.5‰ (Figure 3a) and does not exhibit a consistent
132 seasonal cycle. During this period there is a very weak secular decrease of ~8‰ (based
133 on a least squares fit). $\Delta^{14}C$ begins to slowly increase in January 1954 with a more rapid
134 increase in February 1955 (Figure 3). The first large, sustained peak above the prebomb
135 average value occurs in July 1955 (Figure 3), where values jump from -50‰ in January
136 1955 to -7.2‰ in July 1955. After this early peak, $\Delta^{14}C$ values decrease until October
137 1957 after which time they begin to slowly rise until after 1962 when they rise rapidly to
138 a post-bomb max of 163.3‰ in Jan. 1974 (Figure 3). The peak occurs at approximately

139 the same time as that in the northern hemisphere subtropics [*Guilderson, et al.,* 2000b]
140 which implies a lateral mixing processes dominated by N. Pacific source waters. The pre
141 to post bomb $\Delta^{14}C$ amplitude is 220‰. A strong $\Delta^{14}C$ seasonal signal occurs after 1957
142 with varying amplitude ranging from 15 to 65‰ (Figure 3b).
143
144 From 1957 to 1986 the record displays clear seasonal variability of 15 to 65‰.
145 The seasonal cycle of radiocarbon can be attributed to variations in the ^{14}C of surface
146 waters passing the coral in Makassar Strait. During the NW Monsoon (Jan. – Mar.)
147 *Gordon, et al.* [2003] showed that low salinity surface waters from the Java Sea (due to
148 high rainfall/runoff) can lower the temperature and salinity of water entering the Indian
149 Ocean (Figure 4A). The NW monsoon wind pushes the low salinity surface water from
150 the Java Sea into South Makassar and possibly even northward into the strait itself. This
151 low salinity surface water also flows into the Banda Sea and ultimately the Indian Ocean.
152 During the opposite season (Southeast Monsoon) the winds push the Banda Sea surface
153 waters westward back toward South Makassar and perhaps even slightly northward due
154 to the northward flowing meridional winds [*Gordon, et al*., 2003]. Sea surface
155 temperature and salinity records indicate the Banda Sea is an area of intense vertical
156 mixing (upwelling), resulting in lower surface seawater ^{14}C content [*Cresswell and Luick*,
157 2001; *Ffield and Gordon*, 1996]. We attribute the coral seasonal radiocarbon signal to
158 this seesaw effect of the monsoon.
159 The strong seasonal $\Delta^{14}C$ cycle observed in this coral displays high radiocarbon
160 values during the austral summer (Feb.). The seasonal maximum $\Delta^{14}C$ values are
161 associated with the maximum eastward component of the Java Sea Zonal wind and the

fresh/warm water signal preserved by the coral $\delta^{18}O$ record [*Moore, et al*., 1997; Charles *et al*., 2000]. At first this seems to fit nicely with the Gordon et al. (2003) idea of fresh warm water from the Java Sea during the NW monsoon being advected into the southern end of Makassar Strait. However, the high coral radiocarbon values are more indicative of unmodified North Pacific source waters entering Makassar Strait and flowing southward past the Langkai coral site and not waters arriving via the Java Sea. We suggest that there may be northward flowing low salinity water from the Java Sea into Makassar Strait (as per *Gordon et al*., 2003) but it is tightly constrained along the western boundary of the Makassar Strait. (Figure 4A).

Seasonally low radiocarbon values occur during the austral winter when the zonal wind component is at its westward maximum, the meridional component is northward, and the coral $\delta^{18}O$ indicates cool/dry conditions (Figure 5)[*Moore, et al*., 1997]. We infer that the low radiocarbon waters are sourced from the Banda Sea (East Indonesian Seas) (Figure 4B), which has inputs of South Pacific water (lower in $\Delta^{14}C$ than N. Pacific water) modified by mixing with upwelled, subsurface, lower-^{14}C water. During the NW monsoon the Banda Sea acts as a reservoir, filling up and deepening the thermocline by ~55m [*Cresswell and Luick*, 2001; *Ffield and Gordon*, 1996]. The shoaling of the thermocline during the austral winter SE monsoon and the extensive upwelling in the Banda Sea can account for the low ^{14}C surface waters [*Wyrtki*, 1961]. These data imply that surface water flows from the Banda Sea back into the southern confines of Makassar Strait during the SE monsoon. (Figure 4B). This supports the ideas of the "flush back" of Banda Sea waters of *Gordon, et al*. (2003). Therefore our coral record observes the movement of high radiocarbon surface waters through Makassar Strait during the

warm/wet NW monsoon and the "flush back" of low-^{14}C "upwelling altered" surface waters from the Banda Sea during the cooler SE monsoon. Our interpretation is slightly different to the one proposed by *Moore, et al.* (1997). They argue that the seasonal cycle is due to seasonal radiocarbon variability in the open waters of the Pacific as inferred from a 4-year ^{14}C-coral series at Guam. The "seasonal" cycle at Guam is not very clear or consistent and does not fit the timing of warm water temperature and high $\Delta^{14}C$ (See Figure 3 in [*Moore, et al.*, 1997].

Early Bomb Peak

This coral $\Delta^{14}C$ record exhibits an interesting set of radiocarbon peaks commencing in February 1955. The coral/seawater $\Delta^{14}C$ DIC increases by 40‰ from –50‰ to –10‰ in 3 months with high values continuing until Nov. 1955 (Figure 6). This is a tremendous change and a significant source of ^{14}C is needed to alter the seawater $\Delta^{14}C$ DIC. During the first part of 1954 the United States of America conducted a series of thermonuclear weapons tests at Bikini Atoll (11° 35N, 165° 23E). This started with the 15 MT "Bravo" test on March 1, 1954, followed by tests on Mar. 27 (11 MT), Apr. 26 (6.9MT) and May 5 (13.5 MT) [*Yang, et al.*, 2000]. A significant labeling of waters around the Marshall Islands and to the west must have take place either by gas phase CO_2 exchange or particulate fallout (adsorbed onto $CaCO_3$ particles). The large rise observed in this coral suggests that fallout and not air/sea CO_2 exchange must be the source of the ^{14}C signature in the surface waters observed at Langkai. General open ocean air-sea CO_2 exchange is too slow to account for the rapid rise in coral/seawater $\Delta^{14}C$.

The general westward flow of the North Equatorial Current (NEC) must have brought this labeled water toward the Philippines and into the ITF via the Mindanao

208 Current (Figure 1). Seawater ^{90}Sr measurements in the Philippines, show high levels (850
209 dpm/ 100L of seawater) by the spring of 1955 suggesting that weapons test labeled water
210 was there in high concentration [*Miyake and Saruhashi*, 1958]. Since the waters of the
211 Philippines and the Mindanao Current are sourced from the NEC it follows that this
212 labeled water entered the ITF and flowed past the coral site at Langkai. In an elegant
213 study, Toggweiler and Trumbore [*1985*] reported elevated ^{90}Sr measurements in coral
214 skeletal material from Cocos Island (12.5 S, 97E Indian Ocean) band counted to the year
215 1955. Konishi, et al., [*1982*] observed a 40‰ increase in Δ^{14}C in a coral from Okinawa
216 from 1955 to 1956. All of these observations suggest a 1 yr. transit time for waters
217 originating in the Marshall Islands to reach the bottom of Makassar Strait.
218 Another interesting feature of this part of the record is that the high Δ^{14}C values
219 occur during May-Nov. 1955 (the SE monsoon) and not during the NW monsoon as the
220 post 1958 record (Figures 2,5). The subsequent year, 1956, has 2 cycles peaking in Apr.
221 and Nov. (Figure 6). The time period 1954 to 1957 was a moderately strong La Niña,
222 suggesting that the surface wind and sea level pressure field during the La Niña event
223 (more North Pacific water) and not the Indonesian monsoon was the dominant surface
224 flow during this time period. Since there was no testing in 1955 the two peaks in 1956
225 indicate that some type of "storage" or new input of labeled (high Δ^{14}C) water occurred.
226 The peaks in 1957 could be due to further testing in 1956 (Figure 6).
227 **Pre-bomb era**
228
229 The Langkai pre-bomb (1870-1950) Δ^{14}C record exhibits an overall secular
230 decrease of ~8‰. Although consistent with the dilution of atmospheric ^{14}C by the
231 combustion of ^{14}C-free fossil fuel carbon (the so called Suess effect) and its subsequent

232 transfer into the surface ocean, we do not believe that the time-series is only influenced
233 by the Suess Effect. $\Delta^{14}C$ values between 1870-1890 are ~5‰ higher than the 1891-1905
234 average (Figure 7A), thus most of the decrease occurs prior to a significant decrease in
235 atmospheric $\Delta^{14}CO_2$, particularly when considering the decadal time constant of air-sea
236 isotopic equilibrium. We hypothesize that the reason for the decrease is a change in
237 either the proportion of Banda Sea water 'back flushing' the Southern Makassar Strait, or
238 an increase in the upwelling in the Banda Sea that are coincident with an overall decrease
239 of higher $\Delta^{14}C$ waters sourced from the Northern hemisphere subtropics.
240
241 Superimposed on this shift are interannual $\Delta^{14}C$ oscillations that are consistent
242 with the redistribution of Pacific surface waters driven by ENSO. The coral $\delta^{18}O$ record
243 (not shown) suggests that 1876-1878 was a significant drought on the same order of
244 magnitude as the 1982/1983 El Niño drought. This time period corresponds to the low
245 $\Delta^{14}C$ seen in Figure 7A (arrow). This low period is surrounded by higher $\Delta^{14}C$ in 1874/75
246 and 1879/80, which may be due to stronger La Niña conditions bringing more N. Pacific
247 water through the ITF and less dilution by Banda Sea water during the S.H. winter. In
248 the instrumental record, stronger than average El Niño events often are terminated by
249 stronger than average La Niña events. In the context of the weaker pre-bomb spatial
250 surface ocean $\Delta^{14}C$ gradients, only strong events are likely to provide a consistent signal
251 in our surface water reconstruction. The pre-bomb interval is dominated by an
252 inconsistent seasonal $\Delta^{14}C$ cycle, although some years do appear to follow the $\delta^{18}O$
253 cyclicity similar to the post-bomb era (data not shown). The lack of a distinct seasonal

254 cycle is due to the lower spatial $\Delta^{14}C$ gradient between the water masses prior to nuclear
255 weapons testing.

256 To examine the coral record in relation to ENSO frequency fluctuations of ITF
257 water we applied a (2.6-8 yr.) band-pass filter to the coral $\Delta^{14}C$ data and to the Southern
258 Oscillation Index (Figure 7B,C). The comparison implies a gross coupling between the
259 two records (Figure 7B). Higher $\Delta^{14}C$ occurs during positive SOI events, which is
260 analogous to the S.E. Monsoon and greater throughflow (more North Pacific water)
261 (Figure 7B). Also seen are low $\Delta^{14}C$ and negative SOI (El Niño events), analogous to the
262 NW Monsoon or less throughflow (less North Pacific water) (Figure 7B). We remind the
263 reader that the amplitude of surface $\Delta^{14}C$ is not a direct proxy for the strength and
264 intensity of the ENSO event, particularly during the post-bomb interval [*Guilderson et*
265 *al.,* 1998]. The Makassar record is a reflection of changes in the $\Delta^{14}C$ signature of waters
266 that are sourced into the ITF. These dynamic processes are imprinted upon an evolving
267 $\Delta^{14}C$ background where, in the post-bomb period, the bomb-^{14}C transient is being mixed
268 into and with waters with less ^{14}C.

269 The two records are coherent during the 1889 and 1905 ENSO events; the strong
270 events of 1878 also appear in the $\Delta^{14}C$ record (Figure 7B). Throughout the record there
271 are mismatches, especially in the post-bomb (1950-1990) period, that may be due to the
272 variable nature of ENSO and a temporally varying $\Delta^{14}C$ pattern. We also cannot discount
273 the effect of the monsoon winds and their interactions, both positive and negative with
274 ENSO, causing different surface circulation in southern Makassar resulting in misfits
275 between the two datasets. Thus the $\Delta^{14}C$ time series reinforces that the Indonesian Seas'
276 surface water is a complex interaction between seasonally varying local winds, and

externally forced variability that incorporates both ENSO and the Asian Monsoon. This complex interaction excludes of the common and simplistic ENSO-centric interpretation of surface ocean ^{14}C time-series. Additional and strategically located coral-based records in concert with data-model comparisons [eg., Rodgers et al., 2004] will provide a strong tool to explore the seasonal to decadal controls on the mixing and dynamics of the Indonsesian Seas.

Implications for ITF Surface Circulation

The surface coral radiocarbon record presented here documents a complicated surface water circulation pattern of North Pacific waters arriving during the SH summer NW Monsoon and a flush back of mixed (S. Pacific and upwelling altered) waters from the Banda Sea during the SE Monsoon. Recent studies have shown that the main flow of the ITF is at depths of 200-400m and that surface flow can act independently of the subsurface flow [*Gordon, et al*., 2003]. Although not a pancea to place all variability in a dynamical context, El Niño and La Niña events do manifest themselves as surface water mass variations that are often times discernible in the $\Delta^{14}C$ time-series. A potential measurement of surface water transit speeds in the West Pacific is provided by the arrival of a ^{14}C signature resulting from early weapons tests (in 1955-57) that documents a transit time from the Marshall Islands through the ITF of approximately one-year.

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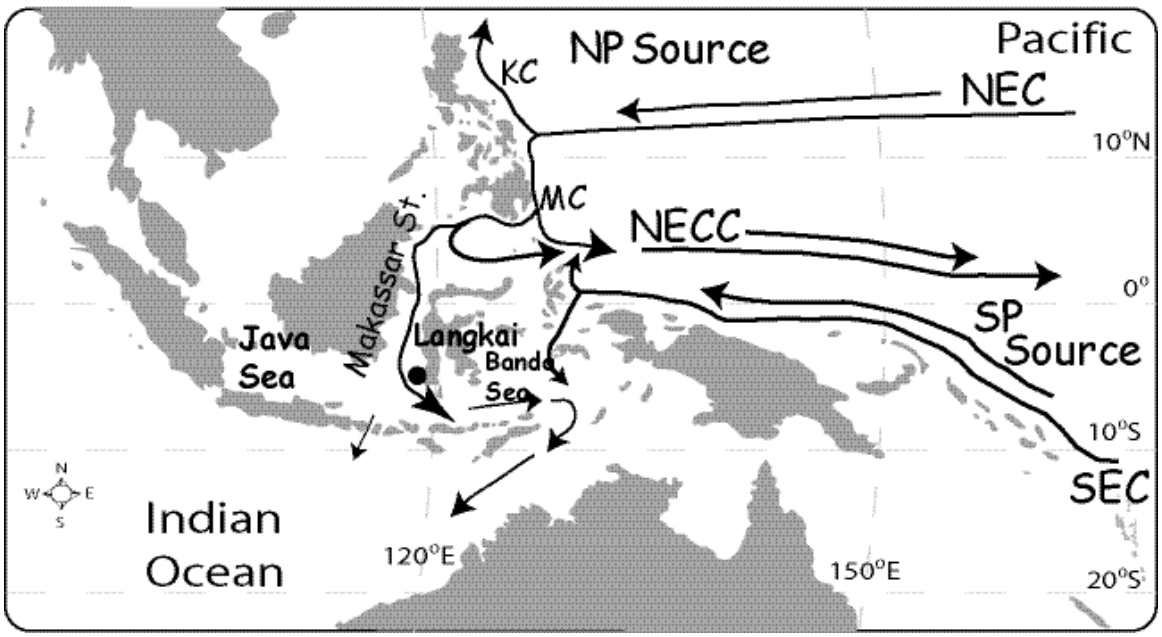
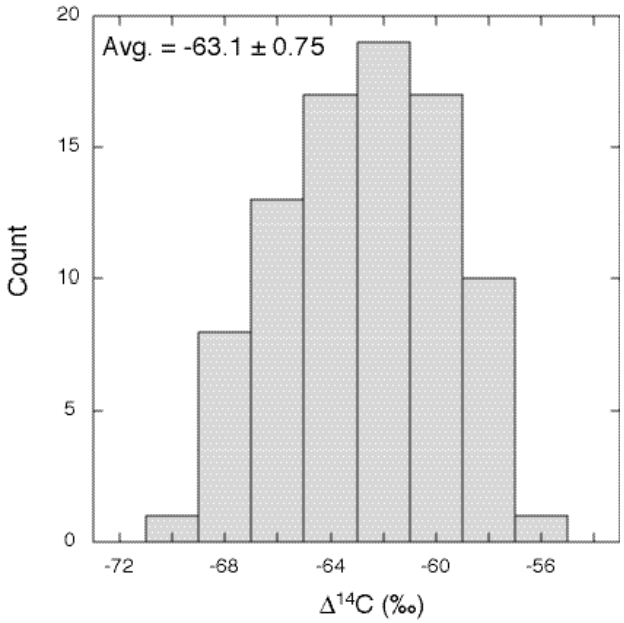


Figure 1. Indonesian archipelago showing sample site Langkai, and major currents and water sources in the area.



411
412 Figure 2. Histogram of secondary coral standard UCI-2, the average value is -63.1 ± 0.75
413 ‰, n=88.
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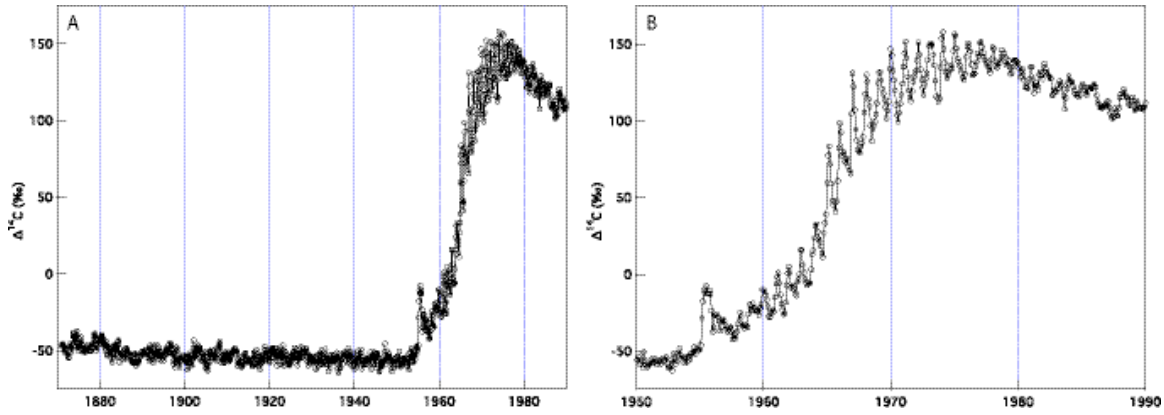

Figure 3. Coral radiocarbon record from Langkai. A) Time series spans the time frame 1870-1990. B). Close up of time frame 1950-1990.

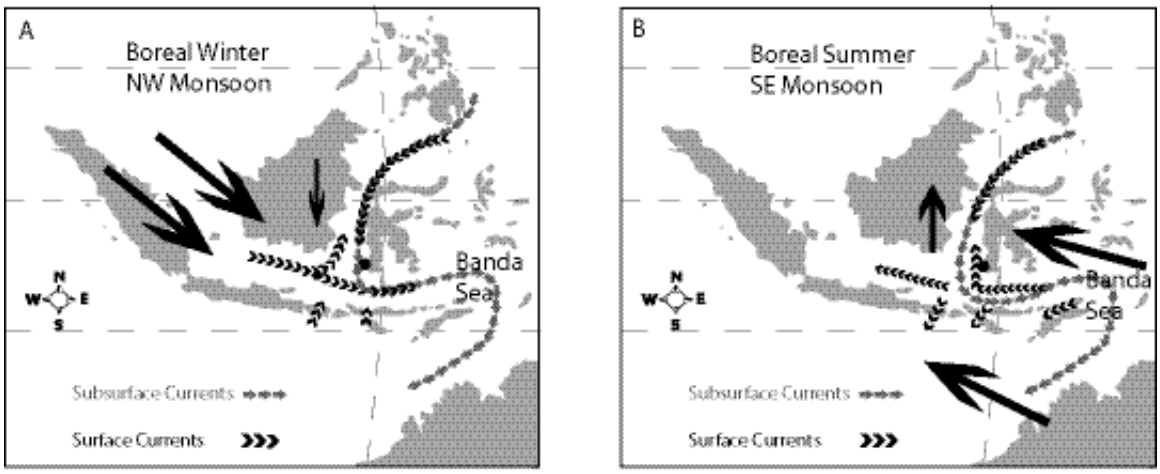

Figure 4. Indonesian Sea surface currents during Northwest Monsoon (A) and Southeast Monsoon (B). Winds are shown in large arrows. Diagram adapted from Gordon et al. [*2003*].

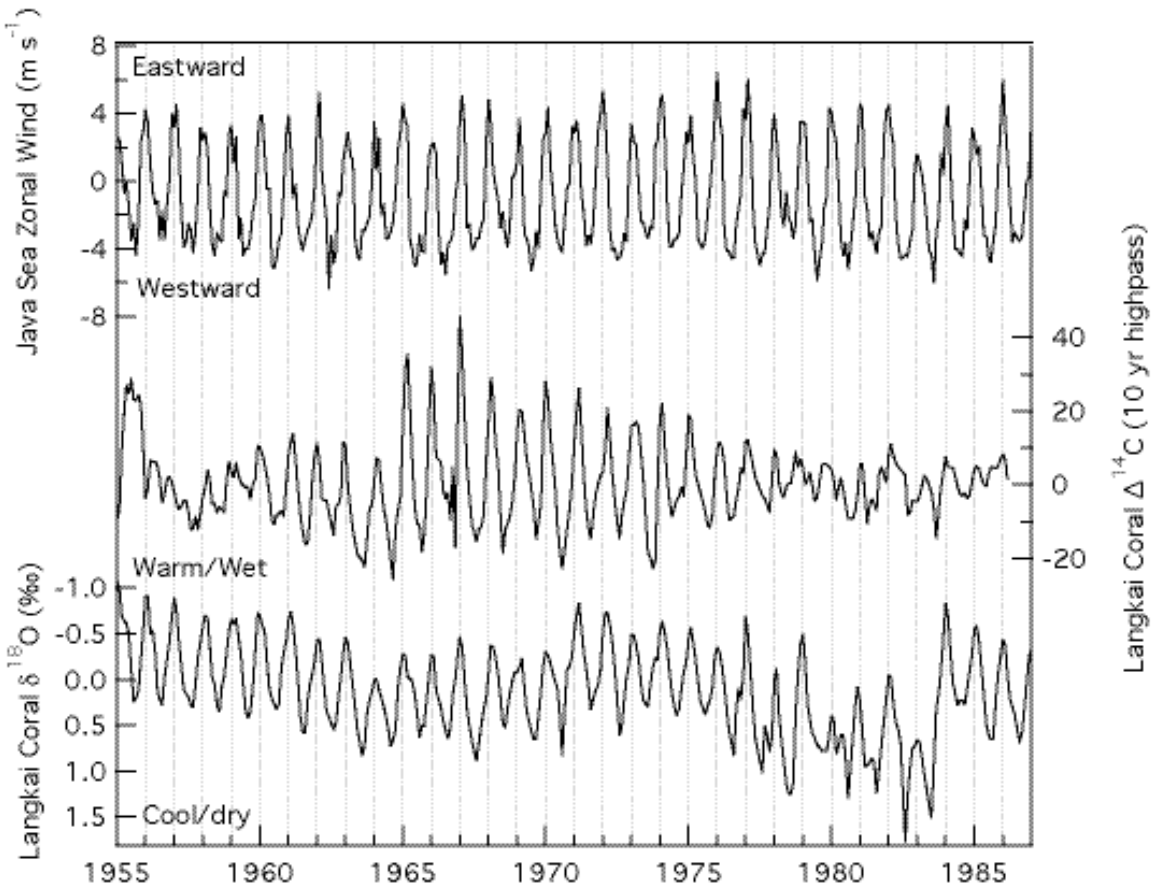

Figure 5. A sub-section of the Langkai coral $\Delta^{14}C$ time series after filtering with a high-pass filter (Tukey-cosine with a half-amplitude at 10 years) in relation to the coral $\delta^{18}O$ and the Java Sea (110° - 118° E; 4° - 7° S) Zonal wind speed.

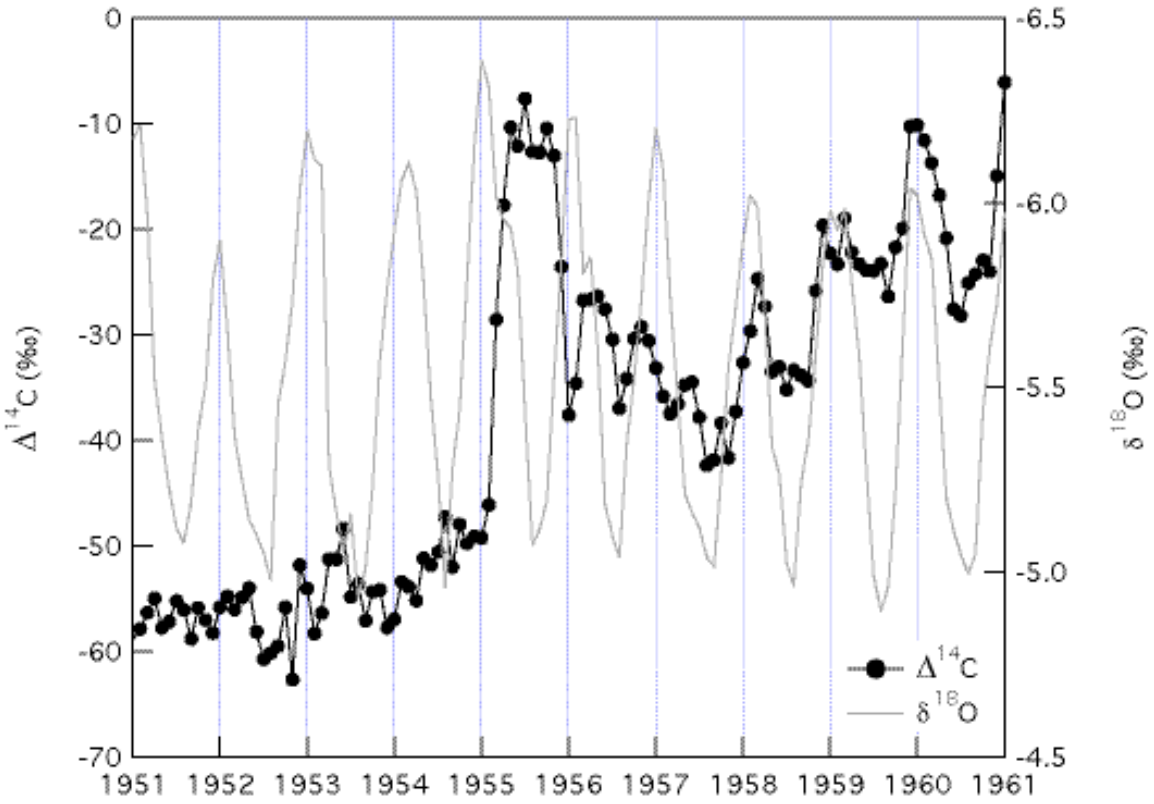

Figure 6. Close-up of mid-1950s bomb-^{14}C induced coral/seawater DIC Δ^{14}C increase. Superimposed is the coral δ^{18}O to show seasonality.

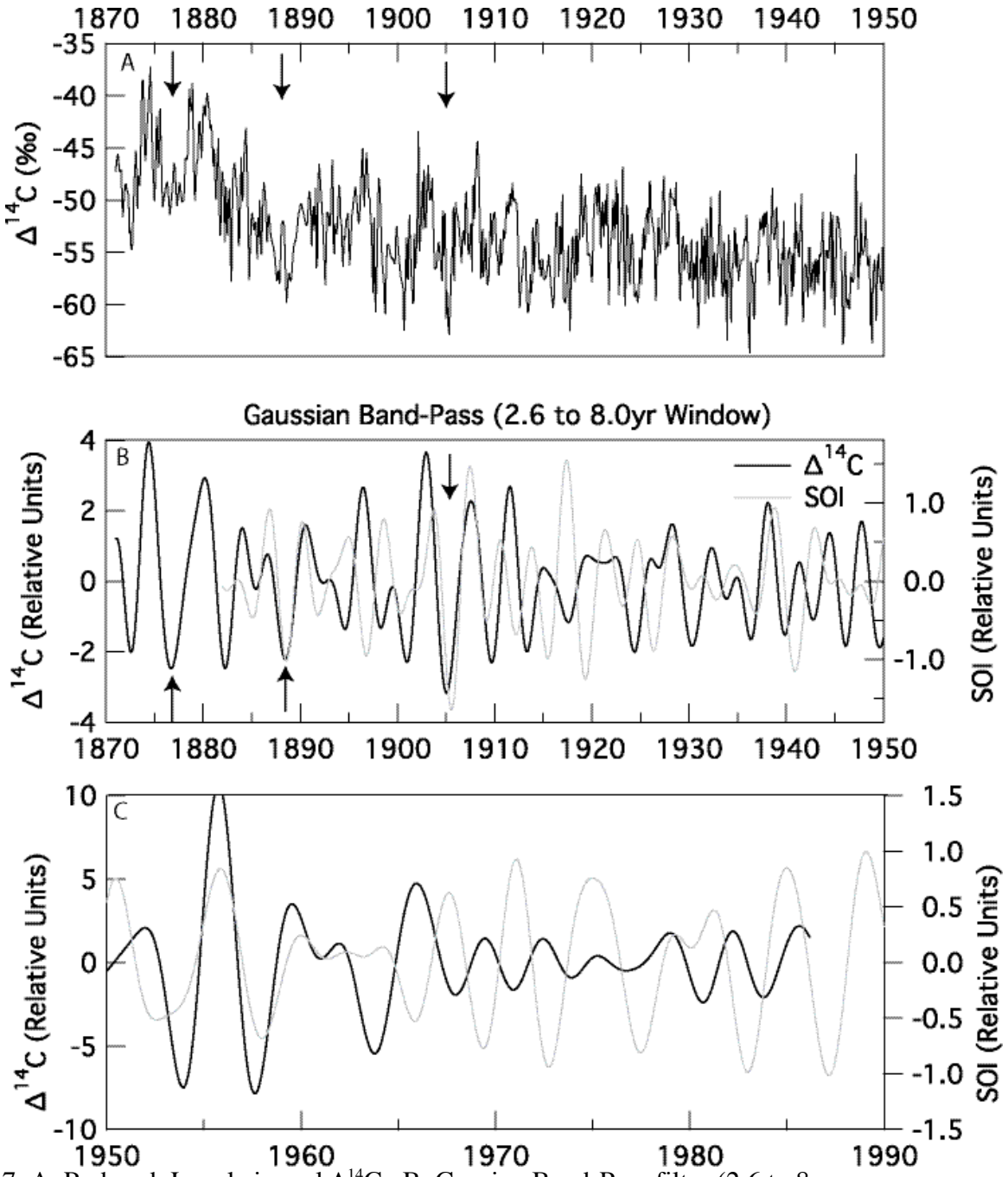

Figure 7. A. Prebomb Langkai coral $\Delta^{14}C$. B. Gausian Band-Pass filter (2.6 to 8 yr. window) of Langkai coral $\Delta^{14}C$ and Southern Oscillation Index Langkai coral $\Delta^{14}C$ pre-bomb era. C. Gausian Band-Pass filter (2.6 to 8 yr. window) of Langkai coral $\Delta^{14}C$ and Southern Oscillation Index Langkai coral $\Delta^{14}C$ post-bomb era. Arrows indicate selected El Niño events.