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Key Points:
- The Maritime Continent area has increased 60% since 5 Ma
- More islands of the Maritime Continent strengthened the Walker Circulation
- More exposed basalt drew atmospheric pCO_2 down

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Growth of the Maritime Continent and its possible contribution to recurring Ice Ages

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Abstract The areal extent of the Maritime Continent (the islands of Indonesia and surrounding region) has grown larger by ~60% since 5 Ma. We argue that this growth might have altered global climate in two ways that would have contributed to making recurring Ice Ages possible. First, because rainfall over the islands of the Maritime Continent not only is heavier than that over the adjacent ocean but also correlates with the strength of the Walker Circulation, the growth of the Maritime Continent since 5 Ma may have contributed to the cooling of the eastern tropical Pacific since that time. Scaling relationships between the strength of the Walker Circulation and rainfall over the islands of the Maritime Continent and between sea surface temperature (SST) of the eastern tropical Pacific and the strength of easterly wind stress suggest that the increase in areal extent of islands would lead to a drop in that SST of 0.75°C. Although only a fraction of the 3–4°C decrease in SSTs between the eastern and western tropical Pacific, the growth of the Maritime Continent may have strengthened the Walker Circulation, increased the east-west temperature gradient across the Pacific and thereby enabled ice sheets to wax and wane over Canada since 3 Ma. Second, because the weathering of basaltic rock under warm, moist conditions extracts CO_2 from the atmosphere more rapidly than weathering of other rock or of basalt under cooler or drier conditions, the increase in weathering due to increasing area of basalt in the Maritime Continent may have drawn down enough CO_2 from the atmosphere to affect global temperatures. Simple calculations suggest that increased weathering of basalt might have lowered global temperatures by 0.25°C, possibly important for the overall cooling.

1. Introduction

During the past 5 Ma, the Earth has cooled gradually, with that cooling punctuated by recurring Ice Ages manifested as huge ice sheets on Canada and Fennoscandia. Stable isotopes in benthic foraminifera imply global cooling, with measurements of $\delta^{18}O$ indicating cooling of the abyssal ocean and increases in continental ice volume (Figure 1) [e.g., *Lisiecki and Raymo*, 2005; *Mudelsee and Raymo*, 2005; *Zachos et al.*, 2001], and more sparsely sampled Mg/Ca ratios also indicating cooling of the abyssal ocean [*Lear et al.*, 2000]. Alkenones, TEX_{86}, and Mg/Ca ratios have been used to infer changes in ocean temperatures above the thermocline; widespread and monotonic cooling has been found in the eastern tropical Pacific [*Dekens et al.*, 2008; *Groeneveld et al.*, 2006; *Lawrence et al.*, 2006; *Wara et al.*, 2005; *Y. G. Zhang et al.*, 2014], off the coast of Peru [*Dekens et al.*, 2007], off Southern California [*Dekens et al.*, 2007], in the Caribbean [*O'Brien et al.*, 2014], in the North Atlantic [*Lawrence et al.*, 2009], off the west coasts of northern Africa [*Herbert and Schuffert*, 1998] and southern Africa [*Etourneau et al.*, 2009; *Marlow et al.*, 2000; *Rosell-Melé et al.*, 2014] (though for this last case, *Leduc et al.* [2014] suggest that cooling applies only to the warm season), in the southeastern Indian Ocean [*Karas et al.*, 2011], and in the South China Sea [*O'Brien et al.*, 2014]. In the western equatorial Pacific and eastern Indian Ocean, long-term cooling is more ambiguous. Using Mg/Ca ratios, *Wara et al.* [2005] and *Karas et al.* [2009, 2011] inferred little change in temperatures in the mixed layer in the past 5 Ma, whereas *Karas et al.* [2009, 2011] reported cooling in subsurface water at 10°S, which they inferred to be Indonesian Throughflow from the Pacific. Collectively, these observations indicate gradual cooling of much of the Earth.

Although the initiation of recurring Ice Ages may require some kind of instability and the crossing of some threshold in the climate system, the gradual cooling implies that a protracted geologic process must underlie the transition from hundreds of million years without continental ice sheets to recurring Ice Ages. Beginning from G. Philander's (personal communication, 2000) aphorism, "Ice is incidental to the Ice Ages," we ignore the recurrence of Ice Ages and explore a geologic process that might underlie that gradual cooling

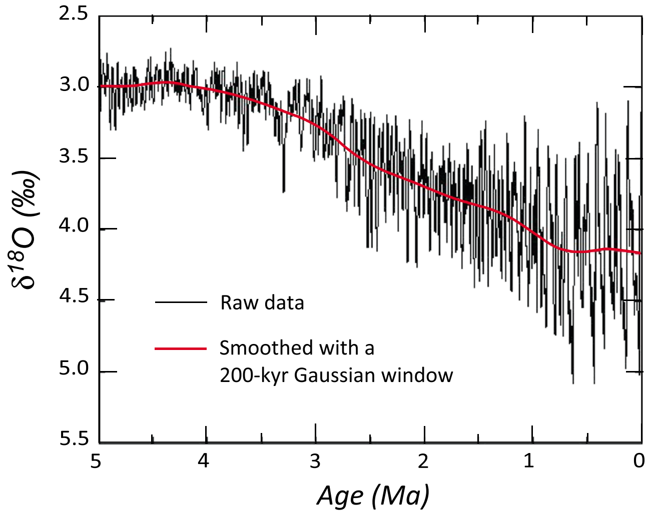

Figure 1. Values of δ^{18}O from benthic foraminifera, from *Zachos et al.* [2001], with red line showing data smoothed using a Gaussian filter with an averaging length of 200 kyr.

(Figure 1): the growth of the Maritime Continent, the islands of Indonesia and surrounding regions (Figure 2). We discuss two parallel mechanisms by which growth of the Maritime Continent could have led to gradual global cooling and an increasingly favorable climate for inception of an ice sheet in Northern Canada: strengthening of the Walker Circulation related to the increasing fraction of islands and drawdown of global CO_2 concentrations by weathering of emergent terrain rich in basalt. We recognize that many consider the closing of the Central American Seaway to have played a crucial role in the development of recurring Ice Ages [e.g., *Bartoli et al.*, 2005; *Berggren and Hollister*, 1974; *Haug and Tiedemann*, 1998; *Kaneps*, 1979; *Keigwin*, 1982; *Sarnthein et al.*, 2009; *Weyl*, 1968]. We discuss this possibility no further here, however, and merely suggest that the mechanisms that we discuss could have aided whatever role the closing of the Central American Seaway might have played.

1.1. Maritime Continent, Walker Circulation, and Laurentide Ice Sheets

Among possible links between gradual changes with distinct regional structure and the possibility of Ice Ages, that between the eastern tropical Pacific and Canada is particularly clear. When the eastern Pacific warms during El Niño events, teleconnections to the southern half of Canada lead to warmer surface temperatures [e.g., *Halpert and Ropelewski*, 1992; *Horel and Wallace*, 1981; *Kiladis and Diaz*, 1986, 1989; *Rasmusson and Wallace*, 1983; *Ropelewski and Halpert*, 1986; *Trenberth et al.*, 2002]. Although such teleconnections are strongest in winter, warmer winters make for longer summers throughout Canada, as measured with positive degree days, which are conducive to melting the previous winter's snowfall [e.g., *Huybers and Molnar*, 2007]. Moreover, general circulation model (GCM) simulations with a warm eastern tropical Pacific consistently show a warmer Canada with more positive degree days than for normal, average, present-day conditions [e.g., *Barreiro et al.*, 2006; *Brierley and Fedorov*, 2010; *Shukla et al.*, 2009, 2011; *Vizcaíno et al.*, 2010]. Thus, the cooling of the eastern tropical Pacific over the past 4 Ma may have enabled the accumulation of perennial snow and the inception of Canadian Ice Sheets [e.g., *Barreiro et al.*, 2006; *Fedorov et al.*, 2006; *Huybers and Molnar*, 2007; *Molnar and Cane*, 2002, 2007; *Philander and Fedorov*, 2003; *Ravelo et al.*, 2004, 2006].

The inferred climate of 3–5 Ma, wherein the eastern equatorial Pacific was warmer than today by ~3–4°C, has been described as a "permanent El Niño" state [e.g., *Barreiro et al.*, 2006; *Brierley and Fedorov*, 2010; *Brierley et al.*, 2009; *Fedorov et al.*, 2006, 2010, 2013; *Goldner et al.*, 2011; *Lawrence et al.*, 2006; *Molnar and Cane*, 2002, 2007; *Ravelo and Wara*, 2004; *Shukla et al.*, 2009, 2011; *Vizcaíno et al.*, 2010; *Wara et al.*, 2005]. Recently, *Y. G. Zhang et al.* [2014] reported SSTs, inferred using the TEX_{86} temperature proxy, that suggest a 3° in the difference between SSTs the eastern and western equatorial Pacific until ~ 5–7 Ma, after which the difference gradually increased to 5–6°C, which might be taken to mean that the present-day La Niña like SST distribution has always existed, but become stronger since 4–5 Ma. The analogy with El Niño derives also from the suggestion that the thermocline in the western Pacific was shallower at 5 Ma than today [*Chaisson*, 1995; *Chaisson and Leckie*, 1993], as well as the suggestion, from δ^{18}O, that the temperature gradient across the thermocline in the eastern equatorial Pacific increased at approximately the same time [*Cannariato and Ravelo*, 1997; *Chaisson and Ravelo*, 2000; *Ford et al.*, 2012] and qualitative similarities of differences between El Niño air temperature and precipitation teleconnections with differences between regional climate at 3–6 Ma from those today [*Molnar and Cane*, 2002, 2007]. Although the term "permanent El Niño" is imprecise,

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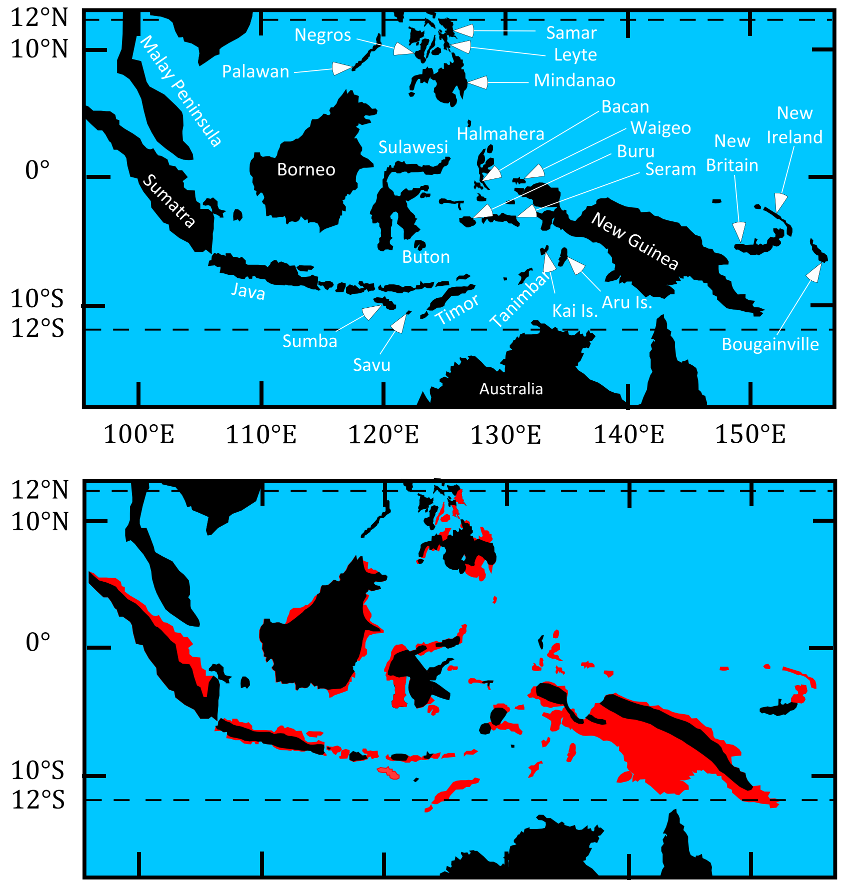

Figure 2. Maps of the Maritime Continent showing (top) present-day land and that for 5 Ma, where (bottom) submerged terrain is show in red and islands have been moved to their positions at 5 Ma using maps of *Hall* [2002].

failing to distinguish between changes in the basic state of the tropical Pacific or changes in the variability (e.g., frequency or intensity) of El Niño events themselves, we continue with the view that regardless of the character of the SST anomalies, the teleconnections associated with any El Niño-like SST anomaly pattern provide a framework with which to consider pre-Ice Age climates.

If cooling of the eastern tropical Pacific underlies the transition to recurring Ice Ages, the question then becomes: What caused sea surface temperatures (SSTs) specifically in that region to decrease over the past ~5 Ma? One possibility is that an overall thermocline shoaling, although global in extent, would lead to regionally concentrated cooling in ocean upwelling regions; *Philander and Fedorov* [2003] suggested that as the thermocline shoaled, easterly winds across the tropical Pacific drew that cold water to the surface and created a background La Niña state. Then, the Bjerknes feedback enhanced zonal winds along the equator and, with an increased SST zonal gradient across the Pacific, strengthened that gradient. In regions of marked upwelling of deep water, apparently warm, nutrient-rich water upwelled in early Pliocene time, consistent with a more recent shoaling of the thermocline [*Dekens et al.*, 2007]. We cannot reject the idea that shoaling of the thermocline contributed to the present-day zonal asymmetry of the Pacific SST, but we pursue another possible contributor to an El Niño-like state.

Other suggested changes that might affect eastern Pacific SSTs include possible roles of the Maritime Continent. *Cane and Molnar* [2001] suggested that the steady, gradual northward movement of New Guinea and Australia so as to restrict flow through the Indonesian Seaway might have helped create the warm pool and, aided by Bjerknes feedback, the cold tongue in the eastern Pacific. Ocean GCM simulations with different amounts of closing of that seaway, but with present-day winds, suggest that the closing of the seaway could have led to a difference in SSTs between the eastern and western Pacific of 0.5°C [*Rodgers et al.*, 2000]. Allowing for Bjerknes feedback and an adjustment to the winds, that contribution might be greater, but more recent coupled GCM runs have challenged this interpretation [*Jochum et al*., 2009]. Thus, although a restricting of the Indonesian seaway seems likely to have enhanced the zonal SST differences between the eastern and western Pacific, the likely enhancement seems to be too small by itself to account for the apparent drop of ~3–4°C.

Also with the Maritime Continent in mind, *Dayem et al*. [2007] noted that rainfall in that region tends to occur preferentially over islands, a result corroborated by closer investigation [e.g., *As-syakur et al*., 2013; *Sobel et al*., 2011]. They showed that variability in the strength of the Walker Circulation correlated with variability of rainfall over the Maritime Continent, but not with that over the warm pool. They suggested that the growth of land area in the Maritime Continent would have enhanced rainfall over the region, which in turn strengthened the Walker Circulation, and again enhanced by Bjerknes feedback, played a role in the cooling of eastern equatorial SSTs in the Pacific over the past 5 Ma. We develop further this last idea.

We argue that the fraction of land area comprising the Maritime Continent has increased by ~60% in the past 5 Myr. Then by exploiting the relationship of rainfall over the Maritime Continent and the strength of the Walker Circulation determined by *Dayem et al*. [2007], we estimate a strengthening of easterly wind stress across the Pacific associated with the expected increase in island area and greater rainfall there. Finally, using numerical model calculations that relate easterly wind strength to eastern equatorial Pacific SSTs, we estimate how much the eastern Pacific might have cooled as the land area of the Maritime Continent grew.

1.2. Weathering of Basalt in the Maritime Continent, CO_2 Drawdown, and Global Cooling

The growth of the Maritime Continent may have played a second role in facilitating Ice Ages. One candidate for explaining global warmth at 5 Ma is higher concentrations of greenhouse gases. Paleoatmospheric pCO_2 proxies extending to ~5 Ma include carbon stable isotope ratios in alkenones [e.g., *Pagani et al*., 2010; *Seki et al*., 2010] and planktic foraminifera [*Raymo et al*., 1996], boron stable isotope ratios in carbonate sediment [*Pearson and Palmer*, 2000] and planktic foraminifera [*Bartoli et al*., 2011; *Seki et al*., 2010], B/Ca ratios in foraminifera [*Tripati et al*., 2009], and stomatal density indices of well-preserved remains of leaves from continental sediments [*Kürschner et al*., 1996]. Uncertainty margins within and across this set of studies are large [e.g., *Seki et al*., 2010] but generally support the notion that any real decline in pCO_2 over the past 5 Ma has been relatively small, with atmospheric pCO_2 not much greater than ~400 ppm at 5 Ma, declining slowly to ~280 ppm during interglacials.

Is this decline by ~120 ppm enough to explain the global cooling of the past 5 Ma? Studies using global climate models answer with a qualified yes. *Haywood et al*. [2013] describe results from the Pliocene Model Intercomparison Project, where a set of coupled atmosphere-ocean models were run with boundary conditions designed to mimic the mid-Pliocene. These boundary conditions include higher pCO_2 of 405 ppm, elimination of the West Antarctic Ice Sheet and much of the Greenland Ice Sheet, accompanied by a ~25 m rise in global sea level and a reduction of total land area, and changes in vegetation cover, including poleward shift of tree line at high latitudes, and a reduction of desert extent in the subtropics. The global mean temperature difference between Pliocene and Modern, averaged across eight climate models, was found to be 2.66°C, with nearly 10°C of warming at 80°N. *Lunt et al*. [2010] argued that the inclusion of slowly responding surface changes—such as vegetation and land ice—are in part responsible for a climate sensitivity of the Earth System to pCO_2 that is ~50% larger than has been found for simulations of future climate change. These slow changes, however, must be represented as a prescribed boundary forcing, for they are not feedbacks in the models. This is a relevant distinction because it means that changes in land cover, which are important for the large global climate sensitivity in the results summarized by *Haywood et al*. [2013], cannot be unambiguously viewed as feedbacks in a climate forced primarily by changes in pCO_2. Thus, although decreasing atmospheric pCO_2 has likely been important in the transition to a cooler global climate, work to date cannot definitively attribute the fraction of cooling that has been caused by the drop in pCO_2. Furthermore, the fundamental question of how much higher atmospheric pCO_2 was at 5 Ma is still plagued by large uncertainties in the proxy reconstructions.

Nonetheless, the potential importance of declining pCO_2 for the gradual cooling of global climate raises the question of why pCO_2 might have declined over the past several million years. Part of the decrease in atmospheric pCO_2 may have occurred abruptly, for example, as a consequence of changes in atmosphere/ocean partitioning of carbon during glaciation, but the timing of abrupt changes inferred from various studies varies from 3.2 Ma [*Tripati et al*., 2009] to 2.8 Ma [*Seki et al*., 2010] to 2 Ma [*Bartoli et al*., 2011]. Thus, although a step change in pCO_2 associated with glaciation at 2.7 Ma is plausibly a part of the long-term decline, in the absence of agreement on the timing of abrupt changes, it seems likely that at least part of the decline is steady and gradual.

We suggest that increases in silicate weathering due to an increasing area of islands in the Maritime Continent contributed to part of this pCO_2 drop. Rates of chemical weathering and consequent CO_2 consumption per unit area in the Maritime Continent are among the highest in the world, owing to the combination of warmth, high rainfall, lithology, and relief. In consideration of the world's 60 largest rivers in terms of discharge, *Gaillardet et al*. [1999] found that four of the top five of them in terms of CO_2 consumption rate per unit area drain basins on the island of New Guinea (#3 is the Irrawaddy of Southeast Asia). Weathering rates per unit area in excess of 50 times the global mean have been found in basins in the Philippines [*Schopka et al*., 2011]. *Dessert et al*. [2003] noted the importance of small basaltic regions in the total global CO_2 consumption by weathering and estimated that basalt in Southeast Asia and the Maritime Continent contributes ~9% of the total global silicate weathering consumption of CO_2. *Kent and Muttoni* [2013] discussed the paleoatmospheric CO_2 concentrations over the past 120 million years, noting that the migration of South Indochina and the Maritime Continent into the "humid tropical belt" between 5°N and 5°S could have contributed to the decline of CO_2 over the past 20 million years. In section 3 we develop further the idea that changes in island area in the Maritime Continent have played a large role in global silicate weathering and the gradual decline in pCO_2 over the past ~5 Ma; we develop a scaling argument that relates changes in the area of the Maritime Continent to changes in pCO_2 and global mean temperature.

2. Growth of the Maritime Continent

Two factors have contributed to the growth of the Maritime Continent: (1) the northward movement of Australia and New Guinea, from high southern latitudes, into the equatorial region [e.g., *Hall*, 2002], and (2) the emergence of islands or portions of islands from below sea level. We summarize briefly here, but discuss in much more detail in Appendix A, the geologic evidence for emergence of terrain since 5 Ma (Figure 2). Toward this end, we have relied on syntheses by *Hall* [2001, 2009, 2012] and *Wilson* [2008], and where possible, we have relied on quantitative paleodepth indicators [e.g., *De Smet et al*., 1989, 1990; *Fortuin and De Smet*, 1991; *Fortuin et al*., 1988, 1990, 1994, 1997].

Most of the emergence of land above sea level has occurred because of crustal shortening and isostatically compensated crustal thickening, in some places enhanced by sedimentation due to erosion of growing high terrain, though a global sea level fall would also contribute where elevations are within tens of meters of sea level. This combination of processes accounts for the emergence of northeastern Sumatra [*Barber and Crow*, 2005; *Barber et al*., 2005; *De Smet and Barber*, 2005] and Java [e.g., *Clements et al*., 2009; *Hall*, 2002, p. 371; *Hall*, 2009, p. 152; *Hamilton*, 1979, pp. 39–44]. *Hall* [2009] wrote: "Most of Sumatra and Java were elevated above sea level and emerged to their present size only since 5 Ma." Similar processes have operated in northern and eastern Borneo [e.g., *Chambers et al*., 2004; *Hamilton*, 1979, p. 99; *Hall and Nichols*, 2002; *Morley et al*., 2003], but this island has grown only slightly since 5 Ma.

The islands along the Banda arc, from east of Java to the small Kai and Aru Islands, seem to have emerged largely in response to the subduction of the Australian continent beneath the arc. By contrast, Timor consists of rock once part of the ancient margin of Australia and scraped off the top of that margin to build an island [e.g., *Audley-Charles*, 1986; *Bowin et al*., 1980; *Carter et al*., 1976; *Veevers et al*., 1978]. New Guinea also has grown as the Australian continent plunged beneath a subduction zone along its northern edge, but in this case subduction has ceased and a major mountain range has been built [e.g., *Baldwin et al*., 2012; *Crowhurst et al*., 1996; *Cloos et al*., 2005; *Dow*, 1977; *Dow and Sukamto*, 1984; *Hall*, 2009, p. 156; *Hall*, 2012, p. 57; *Hill and Hall*, 2003; *Smith and Davies*, 1976]. Sedimentation on the flanks of the high axial range also contributed to emergence, particularly on the southern flank of the range [e.g., *Pigram and Symonds*, 1991; *Quarles van Ufford and Cloos*, 2005].

Between Borneo and New Guinea, Sulawesi has been built by rapid westward movement and Pliocene accretion of small continental fragments [e.g., *Charlton*, 1986; *Davies*, 1990; *Hall*, 2002, p. 416; *Hall*, 2009, p. 153; *Hall*, 2012, p. 54; *Hamilton*, 1979, pp. 159, 173–174, 181–183; *Silver et al*., 1983; *Smith and Silver*, 1991]. Concurrently, subduction beneath Seram and Buru has elevated these islands [*Charlton*, 2000; *Pairault et al*.,

Table 1. Names of Islands in the Maritime Continent, Centroid Locations, Areas at Present and at 5 Ma, and Pertinent References

Name	Latitude	Longitude	Area Present (km^2)	Area at 5 Ma (km^2)	References[a]
Alor	8.30°S	124.75°E	2,120	0	H01
Ambon	3.64°S	128.19°E	806	0	P
Atauro	8.25°S	125.60°E	147	0	CV, H01
Bacan	0.57°S	127.58°E	1,900	0	MH96
Bali	8.40°S	115.20°E	5,416	1,000	H01
Banggai	1.59°S	123.53°E	261	0	D90
Bangka	1.34°S	106.00°E	11,413	11,413	BCD05
Basu	0.34°S	103.59°E	409	0	H01
Batam	1.05°N	104.05°E	399	0	H01
Batanta	0.86°S	130.67°E	456	0	Ch91
Belitung	2.92°S	107.96°E	4,478	4,478	BCD05
Bengkalis	1.45°N	102.30°E	929	0	H01
Biak	1.00°S	136.00°E	1,904	0	H01
Bintan	1.05°N	104.50°E	1,173	0	H01
Borneo	1.00°N	113.00°E	748,168	673,351	90%: Bo
Buru	3.45°S	126.56°E	8,473	4,500	Ch00, FD91, F88
Buton	5.00°S	122.96°E	4,408	2,200	FD91, F90, SS91
Dolak	7.92°S	138.50°E	11,742	0	H01
Enggano	5.38°S	102.25°E	397	0	BCD05, S97
Flores	8.70°S	121.00°E	14,154	8,500	60%: H01
Halmahera	0.50°N	128.00°E	18,040	2,700	15% BM96, H88
Java	7.50°S	110.00°E	138,794	104,000	JAVA
Kabaena	5.25°S	121.94°E	873	0	H01
Kai Besar	5.60°S	133.00°E	550	0	Like Kai Kecil
Kai Kecil	5.75°S	132.75°E	399	0	Ch91, vMD
Kangean	6.90°S	115.35°E	430	0	H01
Karakelong (Talaud Islands)	4.28°N	126.84°E	846	0	M81
Kasiruta	0.39°S	127.19°E	473	0	H01
Kobroor (Aru Islands)	6.13°S	134.55°E	1,723	0	H01
Komodo	8.55°S	119.45°E	330	0	H01
Komoren	8.26°S	138.80°E	695	0	H01
Larat (Tanimbar Islands)	7.17°S	131.81°E	216	0	Ch91, FD91
Laut	3.70°S	116.20°E	2,057	0	vLM05
Lingga	0.15°S	104.69°E	889	0	H01
Lomblen	8.40°S	123.50°E	1,270	0	H01
Lombok	8.60°S	116.36°E	4,625	0	H01
Madura	7.00°S	113.35°E	4,429	0	H01
Maikoor (Aru Islands)	6.20°S	134.27°E	398	0	H01
Mangole (Sula Islands)	1.84°S	125.85°E	1,229	0	H01
Maya	1.15°S	109.55°E	992	992	H01
Misool	1.87°S	130.17°E	2,034	0	S09
Morotai	2.34°N	128.50°E	2,266	0	H01
Moyo	8.25°S	117.55°E	330	0	H01
Muna	5.00°N	122.59° E	2,889	1,400	H01.
New Guinea	6.00°S	140.50°E	785,753	196,000	25%: NG
Nias	1.10°N	97.55°E	4,048	0	BCD05, M80, S95, S97
Niur	0.46°S	103.50°E	342	0	H01
Numfor	1.03°S	134.88°E	335	0	H01
Obi	1.57°S	127.78°E	2,542	2,000	BM96
Padang	1.15°N	102.35°E	1,109	0	H01
Pagai Utara	2.67°S	100.10°E	622	0	BCD05, S97
Pantar	8.40°S	124.10°E	720	0	H01
Peleng	1.38°S	123.25°E	2,346	1,200	50%(?): S83
Rangsang	1.00°N	102.95°E	908	0	H01
Rantau	0.90°N	102.60°E	1,598	0	H01
Rote	10.70°S	123.15°E	1,227	0	H09
Rupat	1.85°N	101.60°E	1,490	0	H01

Table 1. (continued)

Name	Latitude	Longitude	Area Present (km^2)	Area at 5 Ma (km^2)	References[a]
Salawati	1.15°S	130.92°E	1,623	0	F78, S09
Sanane (Sula Islands)	2.20°S	125.92°E	558	0	H01
Sangihe	3.55°N	125.56°E	552	0	M03
Savu	10.55°S	121.85°E	380	0	H09
Seram	3.26°S	129.50°E	17,454	0	FD91, F88, H02, P
Siberut	1.40°S	98.97°E	3,829	0	BCD05, S97
Simeulue	2.65°N	96.10°E	1,754	0	BCD05, S97
Sipura	2.20°S	99.67°E	601	0	BCD05, S97
Sulabesi (Sula Islands)	2.21°S	125.97°E	558	0	H01
Sulawesi	2.00°S	121.00°E	180,681	145,000	80%: BR, H02, SS91, Su, vLM
Sumatra	0.50°S	102.00°E	443,066	332,000	75%: BCD05, Sum
Sumba	6.65°S	120.00°E	10,711	0	F94, F97
Sumbawa	8.50°S	118.00°E	14,386	8,600	60%
Supiori	0.85°S	135.58°E	659	0	H01
Taliabu (Sula Islands)	1.83°S	124.88°E	2,913	1,700	60%: H01
Tanahbala	0.42°S	98.40°E	468	0	BCD05, S97
Tanahmasa	0.17°S	98.46°E	344	0	BCD05, S97
Timor	9.30°S	125.50°E	28,418	0	H02, Tim
Trangan (Aru Islands)	6.53°S	134.28°E	2,149	0	H01
Utara Selatan	3.03°S	100.17°E	900	0	BCD05, S97
Waigeo	0.22°S	130.84°E	3,154	0	Ch91
Wetar	7.82°S	126.13°E	3,600	0	H01
Wokam (Aru Islands)	5.79°S	134.53°E	1,604	0	H01
Workai (Aru Islands)	6.84°S	134.72°E	152	0	H01
Wuliaru (Tanimbar Islands)	7.45°S	131.06°E	151	0	Ch91, FD91
Yamdena (Tanimbar Islands)	7.57°S	131.44°E	4,350	0	Ch91, FD91
Yapen	1.85°S	136.34°E	2,278	0	H01
		Papua New Guinea			
Bougainville	6.20°S	155.50°E	9,318	8,800	BM67
Buka	5.30°S	154.70°E	682	0	BM67
Goodenough	9.50°S	150.15°E	687	0	BDFL, H02
Karkar	4.67°S	146.00°E	400	0	H79
Kiwai	8.58°S	143.45°E	359	0	BDFL, H02
Lavongai (New Hanover)	2.30°S	150.15°E	1,227	0	EM, Ho78
Long	5.33°S	147.08°E	500	0	H79
Manus	2.05°S	146.90°E	1,940	0	EM, Fr
Moratau	9.60°S	150.60°E	1,437	0	BDFL, H02
Mussau	1.50°S	149.66°E	400	0	EM
Muyua	9.10°S	152.80°E	874	0	BDFL, H02
New Britain	5.70°S	150.90°E	35,145	23,430	BR, L06, RC
New Ireland	3.70°S	152.50°	7,405	0	EM, Ho78, SS88
Normanby	10.00°S	151.00°E	1040	0	BDFL, H02
Sudest	11.70°S	153.60°E	866	0	BDFL, H02
Umboi	5.70°S	148.00°E	930	0	H79
		Philippines			
Balabac	7.95°N	117.50°E	319	319	M91, RS91
Basilan	6.50°N	122.00°E	1,266	1,266	H01, R89
Bohol	9.80°N	124.20°E	3,821	0	F03
Cebu	10.30°N	123.75°E	4,468	0	F03
Jolo	5.97°N	121.17°E	869	869	H01, R89
Leyte	10.80° N	125.00°E	7,368	3,700	D09, Saj
Mindanao	7.50°N	125.00°E	97,530	49,000	50%. Min, Saj
Negros	10.00°N	123.00°E	13,075	13,075	F03
Palawan	10.00°N	118.70°E	12,189	12,189	M91, RS91
Panay	11.10°N	122.60°E	12,011	12,011	Y09
Samar	11.90°N	125.30°E	12,849	6,400	T78
Siargao	9.90°N	126.05°E	416	416	H02, Min

Table 1. (continued)

Name	Latitude	Longitude	Area Present (km^2)	Area at 5 Ma (km^2)	References[a]
Siquijor	9.21°N	123.60°E	334	334	F03
Tawi-Tawi (Sulu Archipelago)	5.20°N	120.00°E	581	0	H01, R89
Melville	11.60°S	130.80°E	5,765	0	H02
North Central Australia	11.80°S	132.90°E	7,180	0	H02
NE Australia	11.60°S	143.00°E	12,850	0	H02
South Vietnam	9.00°N	105.50°E	133,000	133,000	(Ignoring Mekong Delta growth)
Malay Peninsula	6.00°N	101.00°E	187,000	187,000	
Sum			3,102,400	1,944,800	
Region as a whole 95°E to 154°E and 12°S to 12°N: Area = 17.38×10^6 km^2					
Percent of region as a whole			17.9%	11.2%	

[a]BCD05: *Barber et al.* [2005]; BDFL: *Baldwin et al.* [2012], *Davies and Warren* [1988], *Fitz and Mann* [2013], and *Little et al.* [2011]; BM67: *Blake and Miezitis* [1967]; BM96: *Baker and Malaihollo* [1996]; Bo: *Chambers et al.* [2004], *Hall et al.* [2008], *Hamilton* [1979, p. 99], *McClay et al.* [2000], *Moss and Chambers* [1999], *Moss et al.* [1997], *Satyana et al.* [1999], and *van de Weerd and Armin* [1992]; BR: *Bromfield and Renema* [2011]; Ch91: *Charlton et al.* [1991a]; Ch00: *Charlton* [2000]; CV: *Chappell and Veeh* [1978]; D90: *Davies* [1990]; D09: *Dimalanta et al.* [2009]; EM: *Exon and Marlow* [1981]; F78: *Froidevaux* [1978]; F88: *Fortuin et al.* [1988]; F90: *Fortuin et al.* [1990]; F94: *Fortuin et al.* [1994]; F97: *Fortuin et al.* [1997]; F03: *Faustino et al.* [2003]; FD91: *Fortuin and De Smet* [1991]; Fr: *Francis* [1988]; H79: *Hamilton* [1979, p. 289]; H88: *Hall et al.* [1988a, 1988b] and *Nichols and Hall* [1991]; H01: *Hall* [2001] and *Wilson* [2008]; H02: *Hall* [2002] (In many areas, his map, Figure 24, shows this to be deep marine.); H09: *Harris et al.* [2009] and *Roosmawati and Harris* [2009]; Ho78: *Hohnen* [1978]; JAVA: *Burckle* [1982], *Clements and Hall* [2007], *Clements et al.* [2009], *P. Lunt et al.* [2009], *Saint-Marc and Suminta* [1979], *Umbgrove* [1946], and *Van Gorsel and Troelstra* [1981]; L06: *Lindley* [2006]; M80: *Moore et al.* [1980]; M81: *Moore et al.* [1981]; M91: *Müller* [1991]; M03: *Macpherson et al.* [2003]; MH96: *Malaihollo and Hall* [1996]; Min: *Pubellier et al.* [1991], *Queaño* [2005], and *Quebral et al.* [1996]; NG: *Abbott* [1995], *Abbott et al.* [1994], *Bailly et al.* [2009], *Davies et al.* [1996], *Dow* [1977], *Dow and Sukamto* [1984], *Pieters et al.* [1983], *Pigram and Symonds* [1991], and *Pubellier and Ego* [2002]; P: *Pairault et al.* [2003a, 2003b]; R89: *Rangin* [1989]; RC: *Riker-Coleman et al.* [2006]; RS91: *Rangin and Silver* [1991]; Saj: *Sajona et al.* [1994, 2000]; S83: *Silver et al.* [1983]; S95: *Samuel et al.* [1995]; S97: *Samuel et al.* [1997]; S09: *Sapin et al.* [2009]; SS88: *Stewart and Sandy* [1988]; SS91: *Smith and Silver* [1991]; Su: *BouDagher-Fadel* [2002], *Calvert* [2000], *Grainge and Davies* [1985], *Mayall and Cox* [1988], *Sudarmono* [2000], *van Leeuwen and Muhardjo* [2005], and *van Leeuwen et al.* [2010]; Sum: *Barber and Crow* [2005] and *De Smet and Barber* [2005]; Tim: *Audley-Charles* [1986], *Bowin et al.* [1980], *Carter et al.* [1976], *De Smet et al.* [1990], *Haig*, 2012, *Keep and Haig* [2010], *Johnston and Bowin* [1981], *Nguyen et al.* [2013], *Quigley et al.* [2012], *van Marle* [1991], and *Veevers et al.* [1978]; T78: *Travaglio et al.* [1978]; vLM: *van Leeuwen and Muhardjo* [2005]; vMD: *van Marle and De Smet* [1990]; and Y09: *Yumul et al.* [2009].

2003a, 2003b]. East-west crustal shortening and thickening have elevated eastern Halmahera, while volcanism has built the western part [e.g., *Baker and Malaihollo*, 1996; *Hall et al.*, 1988b; *Nichols and Hall*, 1991].

Crustal shortening and thickening within Mindanao [*Pubellier et al.*, 1991, 1996; *Sajona et al.*, 1994, 2000] and across smaller islands of the Philippines have elevated these regions [e.g., *Faustino et al.*, 2003; *Travaglio et al.*, 1978]. Palawan, however, seems to have emerged by middle Miocene time (between ~15 and 10 Ma) [e.g., *Holloway*, 1982; *Rangin and Silver*, 1991].

As summarized in Table 1, the region between 95°E and 154°E (from just west of Sumatra to just west of Bougainville) and between 12°N and 12°S, comprises 17.38×10^6 km^2. Today, 17.9% of this area or 3.10×10^6 km^2 stands above sea level, but at 5 Ma, only 11.2% or 1.94×10^6 km^2 had emerged (Figure 2). Thus, the emergent fraction has increased by 60% or 1.16×10^6 km^2; this new land is roughly equal to the combined area of France and Spain, or the combined area of the four-corners states in the USA (Utah, Colorado, Arizona, and New Mexico).

In most cases the emergence of land above sea level has also added orography to the region. Although we have made no attempt to quantify changes in mean elevations, as discussed in Appendix A, mountain ranges seem to have grown in many regions, like New Guinea, Halmahera, Seram, Sulawesi, Borneo, Java, and Sumatra, in the past 5 Ma.

3. Silicate Weathering and Maritime Continent Area

Atmospheric pCO_2 evolution on long time scales is determined by a competition between volcanic outgassing and the combined effects of silicate weathering and burial of organic sediment:

$$\frac{dC}{dt} = \mathcal{V} - \mathcal{W} - \mathscr{B}, \quad (1)$$

where C is the reservoir of carbon in the atmosphere, oceans, and biosphere (units: Pg), $\mathcal{V}$ is the volcanic outgassing rate, $\mathcal{W}$ is the sink from silicate weathering, and $\mathscr{B}$ the burial rate of organic sediment (all Pg/yr). If the volcanic input is fixed, then equilibration of atmospheric pCO_2 requires that the sinks adjust to make the right-hand side of (1) zero. This adjustment is dominated by the increase of silicate weathering under warmer and wetter conditions—a byproduct of increased pCO_2—and the interaction between the long-term carbon and climate systems acts to stabilize both [e.g., *Kump et al.*, 2000; *Walker et al.*, 1981]. In models of atmospheric pCO_2 evolution over tens or hundreds of millions of years, global silicate weathering $\mathcal{W}$ is typically represented as a product of several functions that relate rates of weathering to factors including atmospheric pCO_2, global mean temperature and runoff, and total land area [e.g., *Berner and Kothavala*, 2001].

The simplest form that will suffice for us relates the relative change in weathering flux $\delta\mathcal{W}/\mathcal{W}_0$ to relative changes in weathering-rate-weighted land area $\left(\delta A^*/A_0^*\right)$ and to global temperature (δT):

$$\frac{\delta\mathcal{W}}{\mathcal{W}_0} = \frac{\delta A^*}{A_0^*} + \alpha\ \delta T. \tag{2}$$

In (2), we have assumed that the influence of both runoff and temperature can be subsumed into a single coefficient, α, that scales relative changes in weathering to changes in temperature; values given in *Berner and Kothavala* [2001] imply $\alpha = 0.12\ \mathrm{K}^{-1}$. We consider fractional changes in weathering-rate-weighted land area, because addition or removal of land with very high or low weathering rates can affect the global value of $\mathcal{W}$ more or less than what one would expect from simple fractional changes in land area.

We now assume that the volcanic source of carbon has been relatively constant over the past 5 Ma and that the quasi-steady limit of (1) can be taken: $\mathcal{V} - \mathcal{W} - \mathcal{B} = 0$. This then implies $\delta\mathcal{W}/\mathcal{W} \approx 0$. It follows that we can estimate quantitatively, if only roughly, how increases in the area in the Maritime Continent—where weathering is especially rapid—have been compensated by declining global temperature since 5 Ma. We justify constant volcanism because it is thought that seafloor-spreading rates largely determine $\mathcal{V}$. It is difficult to reject the null hypothesis that seafloor-spreading rates have not varied significantly in time even over time scales as long as the past 180 Ma, based on seafloor area-age relationships [e.g., *Parsons*, 1982; *Rowley*, 2002]. Moreover, changes in rates since ~6 Ma have been negligible [*Krijgsman et al.*, 1999]. We justify the quasi-steady limit of (1) because the time scale of changes we consider (~5 Ma) is much larger than the relaxation time scale of C in (1), which is on the order of ~0.5 Ma for the current climate. Applying $\delta\mathcal{W}/\mathcal{W} \approx 0$ in (2) gives the scaling relationship

$$\delta T = -\frac{1}{\alpha}\frac{\delta A^*}{A_0^*}. \tag{3}$$

Dessert et al. [2003] estimated that weathering of basalt in SE Asia and Indonesia consumes 1.03×10^{12} mol C/yr; this is ~9% of the total silicate weathering flux of 11.7×10^{12} mol C/yr [*Gaillardet et al.*, 1999]. Our use of weathering-rate-weighted area would thus give $\delta A^*/A_0^* = -9\%$, if all of this basaltic region were instantaneously submerged. The question is how much of this basalt has emerged since 5 Ma?

We make a crude estimate, starting with Figure 1 of *Amiotte Suchet et al.* [2003], which clearly shows that the vast majority (about ~5/6) of the basaltic area in the "SE Asia and Indonesia" province is on islands—including a large contribution from the Philippines and particularly Mindanao—and not on the mainland of Asia. In the absence of more geographically explicit information, we estimate that changes in the area of island basalts have scaled similarly to changes in the total island area, i.e., that it increased by 60% from 5 Ma to the present or that it was 40% smaller than the present at 5 Ma. Along with previous assumptions, this gives $\delta A^*/A_0^* = 0.09 \times 5/6 \times 0.4 = 0.03$ in (3). Then, using $\alpha = 0.12\ \mathrm{K}^{-1}$ gives a decrease in global temperature of 0.25°C since 5 Ma, due to the drawdown of pCO_2 by weathering from newly exposed basalt in the Maritime Continent. Although small compared to the apparent temperature change between the Pliocene and the present, this cooling would likely be amplified by a factor of ~2 at the poles, or perhaps more if feedbacks in high-latitude land cover are important.

We may also use this 0.25°C estimate of temperature change due to decreased pCO_2 in the atmosphere to infer how much the concentration of CO_2 has changed. Using the top-of-atmosphere energy balance between the radiative forcing of additional CO_2 and the increased longwave emission from a warmer planet, we may write

$$\gamma \ln\left(1 + \frac{\delta C}{C_0}\right) = \lambda \delta T, \tag{4}$$

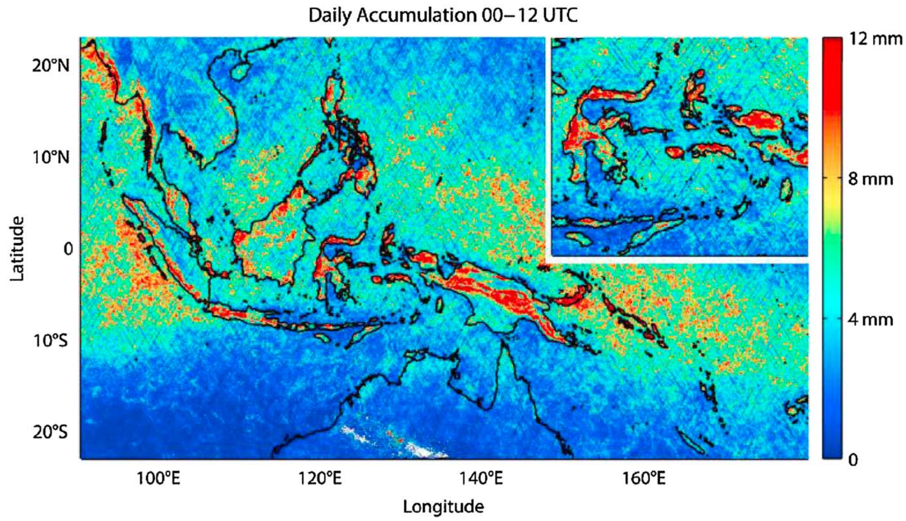

Figure 3. Daytime rainfall over the Maritime Continent, based on data from the Tropical Rainfall Measurement Mission (TRMM) [from *Sobel et al.*, 2011]. Note the concentration of rainfall over islands.

where $\gamma \approx 5.35\ \mathrm{W\,m^{-2}}$ is the radiative forcing in associated with an *e*-fold change in CO_2 concentration [*Ramaswamy et al.*, 2001], δC is the change in concentration of CO_2, C_0 is the reference-state concentration of CO_2 (both in ppm), λ is the total climate feedback parameter ($\mathrm{W\,m^{-2}\,K^{-1}}$), and $\delta T \approx -0.25°C$ is the change in temperature. Our estimate of the decline in pCO_2 thus depends on the choice of climate feedback parameter; values of $\lambda \sim 0.75\ \mathrm{W\,m^{-2}\,K^{-1}}$ are implied by "Earth System Sensitivity" estimates of 5°C in *Haywood et al.* [2013], but larger values of $\lambda \sim 1.2\ \mathrm{W\,m^{-2}\,K^{-1}}$ are given by *Soden and Held* [2006] for a set of simulations of future climate change due to elevated CO_2. Taking $\lambda \sim 1\ \mathrm{W\,m^{-2}\,K^{-1}}$ and $C_0 \approx 400$ ppm, the estimated decrease in the concentration of CO_2 would be 19 ppm, albeit a small, but perhaps not negligible, fraction of the plausible 120 ppm decrease from 400 to 280 ppm.

4. Rainfall Over the Maritime Continent

As noted above, rain falls more over the islands than over the open ocean. *Holland and Keenan* [1980, p. 225] wrote about rainfall over the Maritime Continent on one day, 10 December 1978: "This afternoon convection provides a near perfect map of the region; every island and mountain range is delineated by one or more cumulonimbi and there is a rash of towering cumulonimbi interspersed with smaller cumuli over northern Australia." A particularly striking example is the recurring thunderstorm known as "Hector," which forms nearly every day during certain seasons over the Tiwi islands, just north of the Australian mainland [*Carbone et al.*, 2000; *Keenan et al.*, 1989, 2000]. For an 8 year period from December 1997 to November 2005, *Dayem et al.* [2007] calculated an average of 6.8 mm/d for rainfall over the islands of the Maritime Continent and 4.3 mm/d for the surrounding ocean, corresponding to 58% more rain over land than sea. Similarly, for 1998–2000, *As-syakur et al.* [2013] reported a mean rainfall rate of 7.62 mm/d over the Indonesian islands, but only 5.47 mm/d over the adjacent oceans or ~40% more rain over islands than ocean. *Sobel et al.* [2011] reported a 25% enhancement of rain over the islands with surface areas of 315 $\mathrm{km^2}$ to 6150 $\mathrm{km^2}$ in the Maritime Continent, relative to the nearby surrounding ocean.

The enhancement of rainfall over islands is intimately tied to the diurnal cycle of surface heating over islands [e.g., *Carbone et al.*, 2000; *Cronin et al.*, 2015; *Keenan et al.*, 2000; *Qian*, 2008; *Qian et al.*, 2013; *Saito et al.*, 2001; *Sato et al.*, 2009]. The imprint of the diurnal cycle is particularly apparent when rainfall during daytime hours is plotted on a map, as in Figure 3, from *Sobel et al.* [2011]. In addition, although rainfall enhancement can occur over flat islands for reasons related to the diurnal cycle alone, observations and numerical simulations also suggest that orography can further enhance rainfall over tropical islands [e.g., *Liberti et al.*, 2001; *Saito et al.*, 2001; *Sobel et al.*, 2011; *Zhou and Wang*, 2006].

To understand how the diurnal cycle alone might lead to island rainfall enhancement, *Cronin et al.* [2015] inserted a circular region with low heat capacity, to represent an island, into a cloud-resolving model with a

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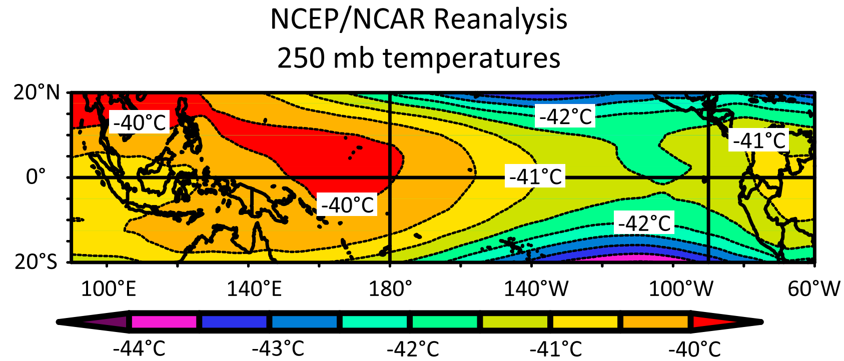

Figure 4. Annual average upper troposphere (250 mb) temperature across the equatorial Pacific from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis [*Kalnay et al*., 1996]. Image provided by the NOAA-Cooperative Institute for Research in Environmental Sciences Climate Diagnostics Center, Boulder, Colorado, from their Web site at http://www.cdc.noaa.gov/.

slab ocean and performed several 250 day simulations of radiative-convective equilibrium across a range of island sizes. For a reference island of radius 48 km, they found that rainfall over the low heat capacity island region, 6.17 mm/d was greater than that over the surrounding ocean, 2.94 mm/d. The islands may have a second impact that is less obvious than average rainfall but perhaps more important for large-scale circulation. In their numerical experiments, *Cronin et al*. [2015] found that as they increased the fraction of area with low heat capacity, the mean temperature of the upper troposphere over the entire region increased. The asymmetry of hot moist air rising during the day, but cooler air simply spreading out near the surface at night led to a rectification of the troposphere's response to periodic heating, such that with increasing island fraction, up to 20%, the upper troposphere over the entire model atmosphere warmed by up to ~1.5°C.

Because of the weak Coriolis effect in the tropics, lateral differences of 1°C within the tropical free troposphere are difficult to maintain. The mean temperature difference across the equatorial Pacific today is only ~1.5°C (Figure 4). The pressure gradient associated with thermal anomalies drives a large-scale circulation with ascent and divergence aloft in anomalously warm regions and with convergence and descent in anomalously cool regions. Thus, based on the findings of *Cronin et al*. [2015] for upper tropospheric warming, the strength of the Walker Circulation might be dependent on the fraction of land within the Maritime Continent. Such a suggestion must remain speculative, however, because of many limitations: consideration of only small regions in the cloud-resolving model, idealized characterization of islands without consideration of many differences between land and ocean (e.g., albedo), neglect of ocean circulation, the Coriolis effect, extratropical interactions, etc. Nevertheless, we pursue the idea that the fraction of island area in the Maritime Continent could play a role in the strength of the Walker Circulation.

5. The Strength of the Walker Circulation and SSTs in the Cold Tongue

Dayem et al. [2007] correlated an indicator of Walker Circulation strength and the rainfall over the Maritime Continent and nearby Pacific Warm Pool. To characterize Walker Circulation strength they used the difference in westerly winds between 250 hPa and 850 hPa in the Central Pacific or wind shear: $\Delta u = u_{250} - u_{850}$ (m/s); a larger Δu implies a stronger Walker Circulation. Whereas they found a negligible correlation of such wind shear with rainfall over the warm pool, for rainfall over the Maritime Continent, P_I (mm/d), the sensitivity of wind shear was large:

$$\frac{\mathrm{d}\Delta u}{\mathrm{d}P_I} \approx 3\,\frac{\mathrm{m/s}}{\mathrm{mm/d}}, \tag{5}$$

over the Pacific between approximately 140°E and 140°W (Figure 5a). A climatological view of equatorial winds, between 10°S and 10°N, over the Pacific (Figure 5b), shows approximately equal magnitudes of westerly winds at 250 hPa and easterly winds at 850 hPa. The average wind shear increases from a negligible value (possibly even negative) at 140°E, to a maximum of ~16 m/s near 140°E, with an average from 140°E-140°W of $\Delta u \approx 8$ m/s, and then with a decrease farther east.

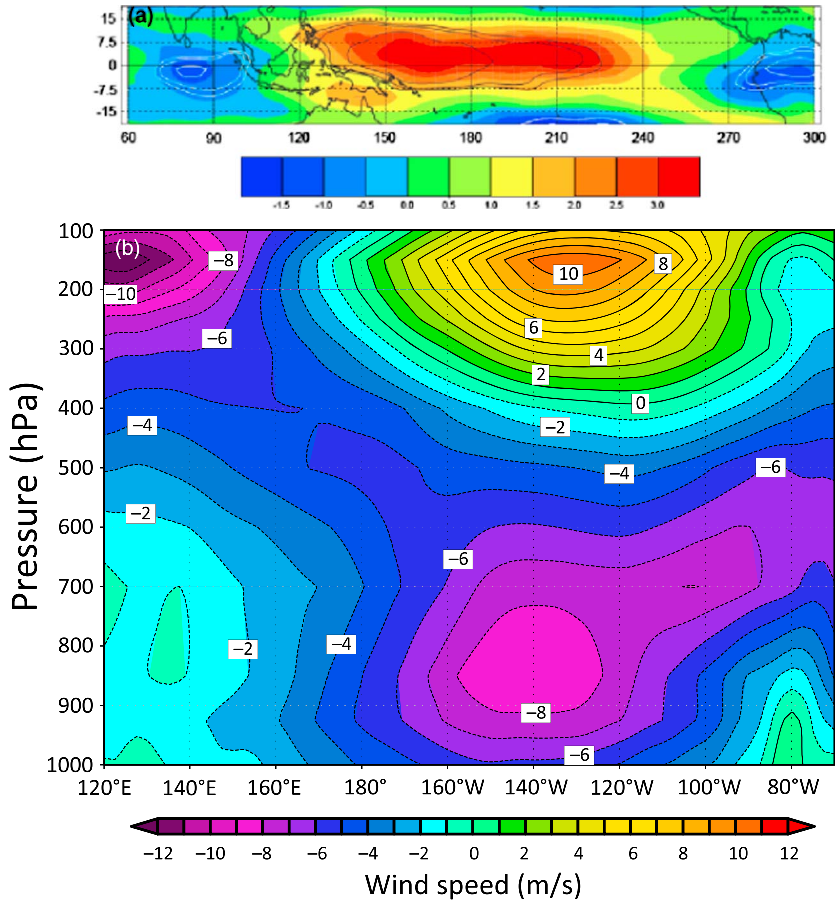

Figure 5. (a) Regression of wind shear ($\Delta u = u_{250} - u_{850}$, where u_{250} is westerly wind speed aloft at 250 hPa and u_{250} is low-level easterly wind speed at 850 hPa) against precipitation, P, over the islands of the Maritime Continent, and (b) east-west cross section of zonal winds along the equatorial Pacific.

Because rain falls over or near the islands, a logical step is to assume that an increase in the fraction of the region occupied by islands would lead to an increase in precipitation, which in turn would strengthen the Walker Circulation. We seek a scaling for the change in the wind shear, $\delta\Delta u$, equal to $\delta\Delta u = (d\Delta u/dP_I)\delta P_I$, for a change of δP_I in rainfall over the islands. Substituting space for time, we then suppose that the fraction of area occupied by land increased by δf. It follows that we would predict an increase in wind shear of

$$\delta\Delta u = \delta f \frac{d\Delta u}{dP_I} \Delta P, \tag{6}$$

where ΔP is the difference in rainfall over islands and surrounding ocean. Using the values of Maritime Continent island fraction of 11.2% at 5 Ma as compared to 17.9% today, then $\delta f = 6.7\%$. Using (5) and $\Delta P \approx 2.5$ mm/d, (6) gives a change in the wind shear of 0.5 m/s, and therefore a reduction of ~6% at 5 Ma as compared to today.

The difference in time mean midtropospheric vertical velocity between island and ocean in the results of *Cronin et al.* [2015] provides another way to estimate how island area has affected changes in Δu. By mass continuity, air that ascends over the Maritime Continent must be balanced by descent elsewhere; idealized dynamics of an equatorial beta-plane suggest that most of the descent from zonally asymmetric ascent near the equator occurs on the equator to the east and resembles the Walker Circulation more than the Hadley Circulation [*Gill*, 1980]. Given the horizontal half width of the Maritime Continent region, $L \sim 3000$ km, and the pressure scale between the surface and maximum vertical velocity, $\Delta p \sim 500$ hPa, changes in vertical velocity over the Maritime Continent would translate into changes in westerly wind difference as

$$\delta\Delta u = 2L \frac{\delta\omega}{\Delta p}. \tag{7}$$

If we assume that changes in the vertical velocity over the Maritime Continent result from changes in island fraction, multiplied by a land-ocean contrast in vertical pressure velocity of $\Delta\omega \sim 100$ hPa/d (consistent with the findings of *Cronin et al.* [2015]), then we obtain $\delta\Delta u = 2L \;\; \delta f \;\; \Delta\omega/\Delta p$ or an increase in westerly wind difference of ~0.9 m/s since 5 Ma. This is somewhat larger than the estimate of 0.5 m/s above, which used the observationally based regressions of *Dayem et al.* [2007], but is still of the same order of magnitude.

We next exploit results of *Fedorov and Philander* [2001], who used *Zebiak and Cane*'s [1987] simple model to examine how sea surface temperature differences depend on wind stress, and of *von der Heydt et al.* [2011], who used a variant of that model. Both showed that a ~40% drop in easterly wind stress over the Pacific would lead to an increase of ~2.5°C in the eastern Pacific SST.

Recall that wind stress, τ, depends quadratically on wind speed, u:

$$\tau = \rho_a C_D u^2, \tag{8}$$

where ρ_a (≈1.2 kg/m^3) is the density of air and C_D (≈1.5×10^{-3}) is a dimensionless drag coefficient. Assuming that surface winds scale with winds somewhat above the surface, then relative changes in surface wind speeds are equal to relative changes in Δu (Figure 5b). Taking Δu at 5 Ma to be ~6% smaller than that today, the easterly wind stress would have been smaller by ~12%. Assuming that differences in eastern Pacific temperatures vary linearly with wind stress, as calculations by *von der Heydt et al.* [2011] suggest, this simple calculation suggests that the eastern Pacific SST would have been warmer by 0.75°C (= 2.5°C × 12%/40%) at 5 Ma.

This 0.75°C decrease in temperature in the eastern Pacific might translate to a global cooling as well, through interactions between global temperatures and atmospheric circulation. *Pierrehumbert* [1995] noted that overturning circulations in the tropics create very dry regions of the atmosphere, allowing the planet to emit more longwave radiation to space than an atmosphere with horizontally uniform humidity would. This finding would lead us to expect that long-term strengthening of the Walker Circulation could lead to global cooling by drying out the atmosphere of the tropical Pacific. On shorter time scales, the logical inverse of this proposition holds: a weakening of the Walker Circulation warms the Earth. *Chiang and Sobel* [2002] pointed out that the entire tropics warmed by ~1°C in March–May 1998, following the 1997–1998 El Niño, the largest recent El Niño, and the one whose teleconnections best match differences between pre-Ice Age and present-day climates [e.g., *Molnar and Cane*, 2007]. On a global scale, GCM simulations of both future climate [e.g., *Held and Soden*, 2006; *Vecchi and Soden*, 2007] and idealized Walker Circulation [*Merlis and Schneider*, 2011] show that large-scale atmospheric circulation weakens with global warming. Moreover, this weakening follows from a simple theoretical argument that depends on the temperature dependence of evaporation via the Clausius-Clapeyron relationship [*Held and Soden*, 2006; *Merlis and Schneider*, 2011]. From GCM runs, *Vecchi and Soden* [2007] estimated that the strength of the Walker Circulation should increase by 5%–10% per 1°C decrease in global mean temperature. Thus, a strengthening of the Walker Circulation not only should cool of the eastern tropical Pacific but also should contribute to a global cooling, which by positive feedback would contribute to further weakening of that circulation.

Obviously, a drop of 0.75°C, even if enhanced by a positive feedback between global cooling and strengthened Walker Circulation, is only a fraction of the apparent drop of 3–4°C in eastern Pacific SSTs discussed above in section 1. We cannot assert that the emergence of islands in the Maritime Continent played the principal role in altering eastern Pacific SSTs. At the same time, with an additional small contribution from oceanographic consequences of the northward movement of New Guinea [e.g., *Cane and Molnar*, 2001; *Rodgers et al.*, 2000], perhaps the growth of the Maritime Continent played an essential role in that change.

6. Discussion

Because the logic exploited here is speculative at all levels, it seems worth reviewing each step. *Hall* [2001], on whose work we have relied heavily, stated, "I believe that the maps of Figures 7 to 10 are generous in assessing areas of possible land and shallow sea." Thus, perhaps we have underestimated the area below sea level at 5 Ma. Conversely, however, his Figure 10 for 5 Ma, Figure 14 of *Hall* [2009], and Figure 3.12 of *Hall* [2012] all show the region between Borneo and the Malaya Peninsula as emergent. With an area of ~300,000 km^2, inclusion of this emergent territory would reduce the difference between 5 Ma and the present from 6.7% to ~5%, and therefore the estimated effect of the growth of the Maritime Continent on

Eastern Pacific SSTs to 0.56°C. We ignore this possibility, for he did not discuss it, but obviously its inclusion would call for a smaller difference between eastern Pacific SSTs today and at 5 Ma.

In the opposite direction, we have ignored the evidence for growth of high terrain since ~5 Ma, but as discussed in Appendix A, most of the high terrain seems to be young. All high belts and islands seem to be younger than 10 Ma, and most have grown since 5 Ma. *Sobel et al*. [2011] reported an orographic enhancement of rainfall over tropical islands, and calculations of rainfall over islands consistently suggest an orographic enhancement of rainfall [e.g., *Liberti et al*., 2001; *Saito et al*., 2001; *Sobel et al*., 2011; *Zhou and Wang*, 2006]. Thus, we may have underestimated the impact of islands and their topography on rainfall, and therefore on the strength of the Walker Circulation.

We have treated all land as equal, but for that affecting the Walker Circulation, the islands in the eastern part of the Maritime Continent might be the more important ones. The eastern part of the Maritime Continent has emerged more than the western part since 5 Ma (Figure 2).

The correlation of wind shear with precipitation over the islands of the Maritime Continent is stronger in the western part of the equatorial Pacific than the eastern part (Figure 5a). Yet the part of the tropical Pacific that seems likely to have the greatest impact on Ice Ages is the eastern part. The wind shear and low-level easterlies also are strongest in the eastern part (Figure 5b). Thus, for the growth of islands in the Maritime Continent to have affected SSTs in the eastern Pacific, there must be an additional process that links wind shear in the western equatorial Pacific to SSTs in the east.

Last, although the simple calculations of *Fedorov and Philander* [2001] and *von der Heydt et al*. [2011] rest on a strong theoretical foundation, the precise links between wind stress and equatorial SSTs depend on many other model parameters. The quantitative link used here—a 6% decrease in easterly wind strength and therefore a 12% reduction in wind stress leading to a ~0.75°C increase in eastern Pacific SST—is qualitative in the sense that assigning an uncertainty to that temperature difference requires subjectivity.

With regard to the influence of island area the drawdown of atmospheric CO_2 since 5 Ma, our estimate that this has decreased global mean temperatures by ~0.25°C also is subject to several uncertainties. First, our estimate of the change in weathering-rate-weighted area has been crude and has not relied on geographically matching the recently emerged areas with basalt areas of *Amiotte Suchet et al*. [2003]. On the one hand, it is possible that we have overestimated the fractional change in area of basalt. On the other, however, we have not attempted to account for weathering from other rock types in newly emerged segments of the Maritime Continent, which may contribute additionally to an increase in weathering rate since 5 Ma, because weathering rates in the region are high even in granitic terrain. We have also not accounted for the growth of high terrain since 5 Ma, and the role it may have played in increasing CO_2 drawdown, due to the tight coupling between physical and chemical erosion rates [e.g., *Gaillardet et al*., 1999; *Larsen et al*., 2014].

The value of δT we have estimated depends on the fractional sensitivity of weathering to global mean temperature change, which may be smaller than the value of $\alpha = 0.12\ \mathrm{K}^{-1}$ that we have assumed. The parameter α represents the sum of fractional changes in weathering rate with temperature (α_T), and fractional changes in global runoff with temperature (α_R). At the microscopic scale, $\alpha_T \sim 0.1\ \mathrm{K}^{-1}$ is justified by measurements of activation energies of silicate dissolution reactions of $E_a \sim 60\text{–}80$ kJ/mol [*Kump et al*., 2000]; $\alpha_T \approx E_a / \left(RT_0^2\right)$ (where R is the universal gas constant and T_0 is a reference temperature in Kelvin). Empirical estimates from a set of basin-scale weathering rates across a range of temperatures, however, tend to be slightly lower, and also making such estimates becomes difficult due to the covariance of runoff and temperature across basins. For the sake of confidence in these microscopic estimates of α_T, it is disconcerting that *Gaillardet et al*. [1999] noted "On a global scale, there is no general correlation of weathering rates (total or cationic) with temperature." By focusing on the basins with the highest weathering rates at each temperature, however, they obtained an envelope suggesting $\alpha_T \sim 0.08\ \mathrm{K}^{-1}$. *Dessert et al*. [2003] more carefully controlled for runoff and calculated $\alpha_T \sim 0.06\ \mathrm{K}^{-1}$ for a number of basaltic regions. Estimates of $E_a \sim 50$ kJ/mol and thus $\alpha_T \sim 0.07\ \mathrm{K}^{-1}$ have also been obtained from a survey river chemistry in granitic basins [*Oliva et al*., 2003].

The fractional change in global runoff with temperature, α_R, can be estimated from climate models or historical observations. Observational studies are controversial and challenging to conduct cleanly; the significance of the finding of *Labat et al*. [2004] that global runoff has increased by $0.04\ \mathrm{K}^{-1}$ was disputed by

Legates et al. [2005] and *Dai et al*. [2009], both of whom found no statistically significant correlation between historical runoff and temperature changes. Climate modeling studies in which increasing greenhouse gas concentrations force temperature change find $\alpha_R \sim 0.03\ \mathrm{K}^{-1}$; the multimodel mean estimate of α_R does not depend strongly on the specific scenario of greenhouse gas concentrations, but estimates across models range from $>0.02\ \mathrm{K}^{-1}$ to nearly $0.05\ \mathrm{K}^{-1}$ [*X. Zhang et al*., 2014]. Taken together, these results suggest that a central estimate of α should be closer to $0.1\ \mathrm{K}^{-1}$ ($\alpha_T \sim 0.07\ \mathrm{K}^{-1}$ and $\alpha_R \sim 0.03\ \mathrm{K}^{-1}$) than the value of $\alpha = 0.12\ \mathrm{K}^{-1}$ used above, which would increase δT slightly to ~0.3°C.

There is also a more fundamental structural uncertainty associated with the estimation of α, to which *Kent and Muttoni* [2013] alluded: if temperatures in the low-latitude regions where a large fraction of global weathering takes place are much less sensitive to pCO_2 than is the global mean temperature, then global weathering can be much less sensitive to global temperature than either microscopic or basin-scale estimates would indicate. For instance, suppose the tropical half of global surface area from 30°S–30°N contributes 75% of the silicate weathering consumption of CO_2 and the extratropical half of global surface area poleward of 30° contributes the remaining 25%. Then, if the tropics warm at 0.5 times the global average, and the extratropics warm at 1.5 times the global average, global weathering rates will increase with global average temperature at only 75% the rate one would expect from a globally uniform warming. Because local weathering rates scale exponentially with temperature, this issue is directly analogous to the response of global mean atmospheric water vapor to global mean temperature, in that the local exponential scaling can be wrong if applied globally, due to the negative correlation between temperature and temperature change [*Back et al*., 2013].

Along with this latitudinal covariance of weathering rates and temperature changes, covariance between regional weathering rates and runoff changes, particularly over areas with high weathering rates, could conceivably alter the sensitivity of global weathering to global temperature. For example, we have suggested that the Walker Circulation may have increased in strength as the Maritime Continent has expanded; this would imply greater runoff over the Maritime Continent and would thus lead to an increase in per-area weathering rates that could also have a global influence on pCO_2. Following this line of reasoning to quantitative conclusions would require too many speculative assumptions to pursue, but the importance of small basaltic areas for global weathering rates underscores the importance of constraining regional changes in both temperature and hydrology.

7. Conclusions

From a synthesis of published geological observations, we infer that the fraction of land area in the Maritime Continent has increased by ~60%, from $1.94 \times 10^6\ \mathrm{km}^2$ at 5 Ma to the present-day $3.10 \times 10^6\ \mathrm{km}^2$. This increased area could have facilitated the shift from a largely ice-free Northern Hemisphere to recurring Ice Ages in two ways.

First, previous work correlated the strength of the Walker Circulation, as quantified by the difference in zonal winds at 250 hPa and 850 hPa, with precipitation over the islands of the Maritime Continent [*Dayem et al*., 2007]. Because rain falls preferentially over islands [e.g., *As-syakur et al*., 2013; *Dayem et al*., 2007; *Holland and Keenan*, 1980; *Sobel et al*., 2011], the growth of island area could have strengthened the Walker Circulation. Scaling relationships between precipitation and wind shear [*Dayem et al*., 2007] and between wind stress and eastern Pacific SSTs [e.g., *Fedorov and Philander*, 2001; *von der Heydt et al*., 2011], with additional assumptions, allow for the increase in islands in the Maritime Continent to have reduced the eastern Pacific SST by ~0.75°C. Such a reduction in eastern Pacific SST, if only a fraction of the 3–4°C change in the difference between eastern and western Pacific equatorial SSTs since ~5 Ma, might have contributed to a cooling of Canada and contraction of summer warmth. Then shorter summers throughout Canada could have prevented winter snow from melting and allowed Ice Ages to become recurring phenomena since that time [e.g., *Barreiro et al*., 2006; *Fedorov et al*., 2006; *Huybers and Molnar*, 2007; *Molnar and Cane*, 2002, 2007; *Philander and Fedorov*, 2003; *Ravelo et al*., 2004, 2006].

Second, increased area in the Maritime Continent since 5 Ma exposed basaltic rock to weathering. Because the weathering of basalt, particularly in warm, moist conditions, extracts CO_2 from the atmosphere rapidly, this increased exposure of basalt could have lowered pCO_2 in the atmosphere. Simple calculations suggest a lowering of the concentration of CO_2 by ~20 ppm and a drop in global mean temperature of 0.25°C since 5 Ma. Uncertainties in the requisite parameters needed to make this estimate of temperature change render 0.25°C more likely an underestimate than overestimate. Moreover, because the poles respond more strongly than equatorial regions, a small global average should call for a greater high-latitude temperature drop.

We recognize that the many assumptions make these links between islands in the Maritime Continent and Ice Ages speculative. At the same time, we have tried to be conservative in quantifying the links, and we may have underestimated them.

Appendix A: Growth of the Maritime Continent

We discuss the evidence used to infer the area of islands or peninsulas that reach into the Maritime Continent at 5 Ma, starting in the west and, crudely, moving east. Numerical values of both present-day areas and estimates of those at 5 Ma are given in Table 1.

A1. Sumatra and Malay Peninsula

It appears that little change in land area has occurred in the Malay Peninsula or in southern Southeast Asia, except maybe a progradation of the Mekong Delta, which we ignore here. Thermochronologic studies suggest little erosion of the Malay Peninsula for tens of millions of years [e.g., *Cottam et al*., 2013a]. *Hall* [2001, 2009, 2012] shows the Malay peninsular emergent at 5 Ma, and we assume no change in area of those regions.

A consensus suggests that more than a third and perhaps as much as half of the island of Sumatra was still submerged as recently as at 5 Ma. The flat terrain northeast of the Barisan Mountains, which lie just inland of the southwest coast, emerged since ~5 Ma [*Hall*, 2001, Figure 10; *Hall*, 2009, Figure 14] or 7 Ma [*Wilson*, 2008]. Relying on *Barber and Crow*'s [2005] comprehensive discussion of individual basins, *Barber et al*. [2005] inferred that roughly half the island lay below sea level in Late Miocene time, but by Early Pliocene time, that fraction was only a quarter of the present-day area of the island.

The northeastern half of the island underwent largely mild tectonic activity in Cenozoic time, with a bit of crustal extension and subsidence. The Barisan Mountains apparently began to form in Late Oligocene time, in response to crustal shortening, but they seem to have been a minor feature, at least a minor source of sediment, until mid-Miocene time [*Barber et al*., 2005; *De Smet and Barber*, 2005]. The regression of the sea from the northeastern half of the island occurred in large part because of abundant fluvial sediment from the mountains. In describing the late Miocene to early Pliocene transition from shallow marine to sublittoral and then to deltaic sediment, *De Smet and Barber* [2005] stated: "The climax of uplift and erosion of the Barisans occurred in the Late Pliocene…" At that time reverse slip became common, and mild NE-SW shortening of the crust under northeastern Sumatra began. We presume that the small islands northeast of Sumatra—Basu, Batam, Bengkalis, Bintan, Lingga, Niur, Padang, Rangsang, Rantau, and Rupat—emerged at the same time.

This emergence of the northeastern side of the island since ~5 Ma does not seem to apply to the islands of Bangka or Belitung, which lie just east of the southeastern end of Sumatra. *Barber et al*. [2005] show them to be either eroding or collecting fluvial sediment in late Miocene and early Pliocene time.

It appears that the islands southwest of Sumatra, in its forearc, emerged in Pliocene time [*Barber et al*., 2005; *Moore et al*., 1980; *Samuel et al*., 1995, 1997]. These include Enggano, Nias, Pagai Utara, Siberut, Simeulue, Sipura, Tanahbala, Tanahmasa, and Utara Selatan. Although *Moore and Karig* [1980; *Moore et al*., 1980] and *Samuel et al*. [1995, 1997] disagree on the structural history of Nias, and on details of when Paleogene sediment on the island rose from deep water, they agree that the island emerged in Pliocene time. Although data are sparse from other islands, a concurrent emergence seems likely [*Barber et al*., 2005; *Samuel et al*., 1997].

A2. Java

Much of the northeastern side of Java seems to have emerged since Pliocene time [e.g., *Hall*, 2009, Figure 14]. Built on the southern edge of the Eurasian landmass, the island stands above sea level in part because of late Cenozoic thrust faulting and crustal thickening and because of volcanism [e.g., *Hall*, 2002, p. 371; *Hamilton*, 1979, pp. 39–44]. This elevated terrain has then shed abundant sediment that has filled adjacent basins.

In a synthesis of sediment from western Java, *Clements and Hall* [2007] showed that the northern and southern quarters of that part of the island were submerged in late Miocene time. Then in a more comprehensive study, *Clements et al*. [2009] summarized evidence of middle Miocene and younger thrust faulting and north-south crustal shortening across the axis of the island, which flexed the terrain to the north down to form a basin [see also *Hall et al*., 2007; *Yulianto et al*., 2007]. This basin is developed best in the east and only modestly developed in the west. *Van Gorsel and Troelstra* [1981] described "an E-W-trending anticlinorium, about 250 km long and 20 km wide [that] during Miocene-Pliocene … was a deep 'foreland basin', situated between the relatively stable Sunda Shelf in the north and the volcanic arc (axial ridge of Java) in the south." They inferred north-south shortening as recent as Pleistocene and emergence at only 2.5 Ma of what had been deep water, 1000 m, at 5 Ma.

Elsewhere, regarding the date of the emergence of Java, *P. Lunt et al*. [2009] reported that "widespread, shallow marine limestone," dated at 5.5–7.5 Ma using strontium isotope stratigraphy, overlies a middle-late Miocene unconformity in central and northeastern Java, suggesting that emergence occurred in Pliocene time. *Burckle* [1982] and *Saint-Marc and Suminta* [1979] reported marine diatoms and planktonic foraminifera from exposed late Miocene and Pliocene sediments in northeastern Java. *Umbgrove* [1946] described a rich assemblage of early Pliocene corals from central Java.

From these observations of Pliocene submergence, and following *Hall* [2001, Figure 10, 2009, Figure 14], we assume that approximately the northern quarter of Java and all of Madura and Kangean (the large and the much smaller islands to the northeast of Java) were submerged at 5 Ma.

A3. The Banda Arc, Sumba, and Timor

We found little specific information about the islands along the Banda Arc, but *Hall* [2001, Figure 10, 2009, Figure 14] shows many of the islands east of Java as totally or partly submerged at 5 Ma, as does *Wilson* [2008, Figure 1f] at 7 Ma. From west to east these are Bali, Lombok, Moyo, Sumbawa, Komodo, Flores, Lomblen, Pantar, Alor, Wetar, Atauro, the Taninbar Islands of Wuliaru, Yamdena, and Larat, and the Kai Islands of Kai Besar and Kai Kecil.

A bit is known about a few of the smaller islands. Atauro, the small island just north of Timor, is rising rapidly today, at 0.47 mm/yr [*Chappell and Veeh*, 1978], and at that rate it must have emerged in last couple of Ma. Coral reefs at 700 m overlie volcanic rock older than 3.3 Ma [*Ely et al*., 2011]. The small Tanimbar and Kai Islands, just east of the Banda Arc, also appear to have emerged in Pliocene or Quaternary time. Late Miocene marine sediment on the Taninbar Islands was folded in Pliocene time, presumably in response to the same collision of Australia with the Banda Arc as in Timor, and then overlain unconformably with Pleistocene sediment [*Charlton et al*., 1991b]. From microorganisms *Fortuin and De Smet* [1991] inferred that sediment currently exposed on those islands lay at depth greater than 2 km at 5 Ma and merged at 2 Ma. Similarly for one of the Kai Islands, *van Marle and De Smet* [1990] reported exposures of Quaternary sediment whose microorganisms commonly live at depths of 1000 m, suggesting emergence since ~1 Ma. Finally, *Hall* [2001, Figure 10, 2009, Figure 14] shows the Aru Islands of Kobroor, Maikoor, Trangan, Wokam, and Workai as having been submerged at 5 Ma.

Abundant evidence demonstrates that beginning near 3 Ma, the northern margin of the Australian continent met the Banda Arc. As subduction of that margin began, slices of it became detached from the underlying lithosphere and were stacked to create the island of Timor [e.g., *Audley-Charles*, 1986; *Bowin et al*., 1980; *Carter et al*., 1976; *De Smet et al*., 1990; *Johnston and Bowin*, 1981; *Veevers et al*., 1978]. Moreover, in a number of cases, volcanism along the arc seems to have shut down at ~3 Ma [*Abbott and Chamalaun*, 1981]. Elsewhere, the chemistry of the volcanic rock shows contamination by Australian crust underthrust beneath the islands [*Herrington et al*., 2010]. This same history seems to characterize many of the islands both along the arc and south of it.

Regarding the emergence of Timor, recent work with improved dating techniques and more detailed mapping has refined the history of the island, but not changed the inference that the island emerged since ~3 Ma or maybe a little before that time [e.g., *Haig*, 2012; *Keep and Haig*, 2010; *Nguyen et al*., 2013; *Quigley et al*., 2012; *van Marle*, 1991]. Similarly, islands just to the west of Timor, like Savu and Rote, seem to have emerged more recently, closer to 1 Ma [*Harris et al*., 2009; *Rigg and Hall*, 2011; *Roosmawati and Harris*, 2009], like that of Sumba, yet farther west. *Fortuin et al*. [1994, 1997] inferred that Sumba lay as deep as 4500 m at

~6 Ma, and subsequently rose at an average rate of ~0.7 mm/yr. It did not emerge above sea level before 3 Ma, though deformation beneath it seems to have begun near 7 Ma [*Rutherford et al*., 2001].

Thus, it appears that most of the islands east of Java along or south of the Banda Arc emerged since ~5 Ma.

A4. Borneo

Although Borneo has not been immune to late Cenozoic tectonics, it appears that the areal extent of the island has increased little since 5 Ma. Despite high erosion rates, inferred from thick sediment surrounding the island [*Hall and Nichols*, 2002], and the resulting flux of sediment that has rapidly filled accommodation space, folding and thrusting of the sedimentary layers seems to have compensated for an expansion of subaerial terrain. In fact, both *Hall* [2001, Figure 10, 2009, Figure 14] and *Wilson* [2008, Figure 1f] show Borneo and the small island of Maya, connected to the Malay Peninsula at 5 Ma and 7 Ma, respectively, suggesting that a larger area may have been emergent at those times than now.

Shortening of the sedimentary cover along the north side of the island has occurred since middle Miocene time [e.g., *Hesse et al*., 2009; *Hinz et al*., 1989; *Hutchison*, 1996; *Morley*, 2009; *Morley et al*., 2003; *Rangin et al*., 1990; *Wannier*, 2009], though some have questioned whether convergence continues [e.g., *Cottam et al*., 2013b; *Hall*, 2013]. Seismic reflection profiles attest to abundant folding, which is exposed clearly on the northwestern margin of the island [*Chambers and Daley*, 1997; *Franke et al*., 2008; *Hesse et al*., 2010; *Hinz et al*., 1989], and much of the sediment exposed on land is marine and late Cenozoic in age [*Back et al*., 2005; *Balaguru and Nichols*, 2004; *Hall*, 2013; *Hutchison*, 1996]. Following, if not concurrent with, this folding, the highest mountain on the island, Mount Kinabalu, seems to have grown by intrusion of granite at 7–8 Ma [*Cottam et al*., 2010, 2013b]. Moreover, GPS measurements show that shortening across this region continues at several millimeters per year [e.g., *Simons et al*., 2007; *Socquet et al*., 2006].

On the eastern side of the island, a large sedimentary basin, the Kutai Basin, has grown eastward largely by east-west crustal shortening, the creation of eroding terrain on the inland side, and progradation of deltas [*Chambers et al*., 2004; *Hamilton*, 1979, p. 99; *McClay et al*., 2000; *Moss and Chambers*, 1999; *Moss et al*., 1997; *Satyana et al*., 1999].

South of the Kutai Basin, the Meratus Mountains, which form a range parallel to the southeast coast, have grown since late Miocene time, presumably in response to north-south crustal shortening [*Hall et al*., 2008; *Satyana et al*., 1999; *van de Weerd and Armin*, 1992; *Witts et al*., 2011]. GPS measurements suggest modest shortening of a few mm/yr across the range [*Simons et al*., 2007].

Although some eastward and northward expansion of Borneo may have occurred since 5 Ma, the amount seems to be small, no more than 10%. Similarly, although some of the high terrain, Mount Kinabalu in the north and the Meratus Mountains in the southeast, may have grown higher since 5 Ma, we have no way to quantify the amount.

A5. Sulawesi, the Makassar Strait, and the Sula Islands

Sulawesi has undergone a complicated history, with amalgamation of its various arms only in the past few million years. Moreover, the westward movement of Sulawesi toward Borneo has narrowed the Makassar Strait, though which most of the present-day Indonesian Throughflow passes [e.g., *Gordon et al*., 2003].

Seismic reflection profiles show clear late Miocene and younger folding on the eastern margin of the Makassar Strait [*Bergman et al*., 1996]. *Calvert* [2000] reported late Miocene-Pliocene shallow marine sediment now cropping out on the west coast. In a series of schematic cross sections, *van Leeuwen and Muhardjo* [2005] show Plio-Quaternary emergence of much of the northerly trending belt of mountains that form the western side of the island, and the small island of Laut. They also report Quaternary reef limestone as high as 1000 m both along the western margin and on the northern arm of Sulawesi. In the north arm of Sulawesi, *Hennig et al*. [2014] reported late Miocene to Pliocene granitic intrusions and volcanic rock that has been exhumed rapidly, ~1–4 mm/yr. They infer rapid emergence of this region and cite Plio-Pleistocene sediment adjacent to the region as supporting evidence of rapid recent erosion.

Hall [2002, p. 416] stated "from about 5 Ma that there is clear evidence for contraction and uplift throughout the whole of Sulawesi and the beginning of the rise of the mountains in western Sulawesi," an inference corroborated by *Bellier et al*. [2006]. GPS measurements of control points on the east and west sides suggest as much as 10 mm/yr of NW-SE convergence between the margins, and therefore ~7 mm/yr of narrowing of the Makassar Strait [*Rangin et al*., 1999; *Simons et al*., 2007]. If this rate operated for the past 5 Myr, then the strait would have been ~15% wider at that time than today.

For the southern arm of the island, *BouDagher-Fadel* [2002] reported late middle Miocene to early Pliocene shallow water deposits and early Pliocene emergence, though the areal extent of this emergence is not clear. Similarly, *Bromfield and Renema* [2011], *Grainge and Davies* [1985], and *Mayall and Cox* [1988] presented evidence of a late Miocene widespread transgression and reef development at the southern end of this arm of Sulawesi, followed by a Pliocene emergence of the region. *Sudarmono* [2000], *van Leeuwen et al*. [2010], and *Camplin and Hall* [2014] reported similar histories—Miocene submergence and Pliocene emergence—for the Bone Basin, which lies between the southern and southeast arms of the island, and the Bone Mountains along the eastern side of the southern arm.

For the southeastern arm, a submerged terrain (Tukang Besi platform) seems to have collided in middle Miocene time with the island of Buton [*Smith and Silver*, 1991], which lies east of the southeast arm. The island is covered by only mildly deformed younger rock, and in some areas that younger rock consists of "fine-grained pelagic foraminiferal chalk" [*Smith and Silver*, 1991]. *Fortuin et al*. [1990] reported abundant marine microorganisms dated at 6 Ma from Buton, some coming from as deep as 1000 m [*Fortuin and De Smet*, 1991]. Thus, they concluded that at least part of the island emerged since ~5 Ma, and we presume the same for the smaller adjacent islands of Kabaena and Muna.

For the eastern arm, rapid westward displacement of small slivers of crust, including the islands of Peleng and Banggai, has led to their collision with the rest of the arm and their accretion on the eastern end of the peninsula [e.g., *Charlton*, 1996; *Ferdian et al*., 2010; *Hamilton*, 1979, pp. 159, 173–174, 181–183; *Silver et al*., 1983]. *Van Leeuwen et al*. [2010], *Villeneuve et al*. [1998, 2000, 2001], and *Watkinson et al*. [2011] date this collision as Pliocene. *Davies* [1990] showed that one block (Banggai) collided with the peninsula between 5.3 and 3.8 Ma, when clastic sedimentation on limestone began.

Despite all of the evidence of post-Miocene deformation and emergence, only a relatively small fraction of the island and adjacent islands seems to have been below sea level at 5 Ma. *Hall* [2001, Figure 10, 2009, Figure 14] shows parts of all of the arms as submerged at 5 Ma, but the fraction is only 10–20% of the present-day emergent region. By contrast, *Wilson* [2008, Figure 1f] shows the entire northern arm and much of the eastern arm submerged at 7 Ma. In a cross section across the northern arm of Sulawesi, *van Leeuwen and Muhardjo* [2005] show most of its emergence to be Plio-Quaternary.

Farther east, *Hall* [2001, Figure 10] shows roughly half of the area of the Sula Islands of Mangole, Sanane, Sulabesi, and Taliabu submerged at 5 Ma.

A6. Buru, Seram, and Surrounding Islands

Buru and Seram lie between Sulawesi and New Guinea (Figure 2). Although Buru has undergone only mild folding, and Seram has been caught in a subduction zone, they share similar histories of pre-Pliocene sedimentation [e.g., *Charlton*, 2000].

For Buru, mild deformation apparently has elevated the island [*Charlton*, 2000]. *Fortuin et al*. [1988] showed emergence at 6 Ma, but in a more recent summary, *Fortuin and De Smet* [1991] reported depths of 1000 m at 3 Ma. In his synthesis, however, *Hall* [2001, Figure 10] shows only the axial part of the island submerged at 5 Ma.

Much, if not all, of Seram and the small island of Ambon to its southeast seem to have emerged since 5 Ma. Although recent work by *Pownall et al*. [2013, 2014] suggests that much of the high-grade rock exposed on Seram has been exhumed by normal faulting, in some places deeper rock has undergone melting in the past few million years [*Linthout and Helmers*, 1994; *Linthout et al*., 1996, 1997]. At present, the seafloor north of Seram is underthrusting the island, and from folding of sediment offshore, *Pairault et al*. [2003a, 2003b] inferred a Pliocene onset of that subduction. Consistent with this timing, *Charlton* [2000] reported that the volcanism on Seram and Ambon is Pliocene. Interpretative cross sections given by *Pairault et al*. [2003a, 2003b] show Seram as having emerged entirely in Pliocene time. From sediment cropping out on the southwest side of Seram, *De Smet et al*. [1989] inferred depths of 500 m between ~3.5 and ~1 Ma. *Fortuin and De Smet* [1991] reported this as depths of 2000 m at 2 Ma.

Although data seems sparse, we follow others [e.g., *Hall*, 2002. p. 417; *Pairault et al*., 2003a, 2003b] and assume that Seram emerged entirely since 5 Ma, and so did much of Buru.

A7. New Guinea and Adjacent Islands

Much of New Guinea and the islands immediately adjacent to it—Batanta, Biak, Dolak, Kiwai, Komoren, Misool, Numfor, Salawati, Supiori, Waigeo, and Yapen—seem to have emerged since 5 Ma [*Hall*, 2001, Figure 10, 2009, Figure 14, 2012, Figure 3.12] or since 7 Ma *Wilson* [2008, Figure 1f]. Moreover, although the high axial chain of mountains began to grow before that time [e.g., *Baldwin et al*., 2012; *Cloos et al*., 2005; *Davies*, 2012], high topography also seems to postdate 5 Ma [*Hall*, 2009, p. 156, 2012, p. 57].

In western New Guinea, *Dow and Sukamto* [1984] reported that deformation began no later than late Miocene, but with the oldest-coarse clastic sediment as early Pliocene, they inferred that mountains developed in Pliocene time. They stated that 8000 m of Pliocene and Quaternary clastic sediment accumulated in the bay east of the Bird's Head, which clearly requires deep erosion of material from nearby. In a synthesis of sedimentary rock from western New Guinea, *Pieters et al*. [1983] showed that limestone covered much of the region until late Miocene time. On the southwest side of the Bird's Head of western New Guinea, late Miocene to early Pliocene folded rock is overlain by Pliocene and younger shallow marine sediment [*Bailly et al*., 2009; *Pubellier and Ego*, 2002]. Thus, much of this region seems to have been below sea level until Pliocene time.

Salawati Island, just west of the Bird's Head, was connected to the New Guinea in Miocene time, and Pliocene separation seems to have occurred during modest Pliocene rifting between them [*Froidevaux*, 1978]. Extensive Miocene to lower Pliocene reefs cover Salawati island, and we presume that emergence of them is young. Similarly, in summary cross sections based on extensive seismic reflection profiling between Misool and Salawati, *Sapin et al*. [2009] showed this area emerging in Pliocene time.

Waigeo Island and the smaller island of Batanta, north of the Bird's Head of western New Guinea, also appear to be young. On Waigeo, ophiolite is capped by 2000 m of Miocene limestone, which shows coral reefs at the base but higher in the section it is rich in benthic foraminifera and then planktonic foraminifera [*Charlton et al*., 1991a]. *Charlton et al*. [1991a] allow for some of the limestone to be Pliocene in age. Pliocene folding created two broad anticlines and presumably elevated the island above sea level.

The main axial range of New Guinea seems to have started to grow in middle to late Miocene time in the northern part, and the locus of deformation migrated southward [*Cloos et al*., 2005; *Hill and Hall*, 2003; *Quarles van Ufford and Cloos*, 2005; *Sapiie and Cloos*, 2004]. Thermochronological studies show rapid exhumation of northern New Guinea in its western part at 14–12 Ma [*Hill and Hall*, 2003] and between 8 and 5 Ma it is eastern part [*Crowhurst et al*., 1996]. On the south flank, however, thermochronology shows exhumation as young as 4 Ma [*Hill and Gleadow*, 1989; *Hill and Raza*, 1999; *McDowell et al*., 1996] and deep incision since 2 Ma [*Weiland and Cloos*, 1996]. From the pattern of exhumation and its timing, *Weiland and Cloos* [1996] inferred that although exhumation had begun by 7 Ma along the crest of the range and south of it, elevations high enough to affect orographic precipitation did not occur until 5 Ma or more recently. Coinciding with this rapid Pliocene exhumation, a "pronounced sedimentological change occurred in the Pliocene-Pleistocene (~5 Ma), when the coarseness of deposits along the flanks of the Central Range increased dramatically for hundreds of kilometers along strike" [*Quarles van Ufford and Cloos*, 2005]. Relying on *Hill*'s [1991] estimate of 100 km of shortening across the southern part of the range, *Crowhurst et al*. [1996] inferred that that part of the belt formed in Pliocene time. The rise of these mountains then shed sediment southward, and a late Pliocene and Quaternary southward migration of sediment buried a carbonate platform to the south of the mountain belt [*Pigram and Symonds*, 1991].

On the northeastern coast, abundant coral reefs on the Huon Peninsula attest to rapid emergence of this region [*Chappell*, 1974]. Moreover, the sudden deposition of coarse clastic material southwest of the Adelbert and Finisterre Ranges in what had been a deep trough, and derived from those mountains, shows that they emerged at 3.0–3.7 Ma [*Abbott*, 1995; *Abbott et al*., 1994]. Farther west a large ophiolitic body, the Cyclops ophiolite was emplaced at ~20 Ma, but apparently Pliocene thrust faulting and crustal shortening elevated the region above sea level [*Monnier et al*., 1999].

Along the Papuan Peninsula of southeastern New Guinea, rapid deformation in Pliocene time seems to have elevated much of this region above sea level [e.g., *Davies et al*., 1996; *Dow*, 1977]. Plio-Quaternary coral reefs surrounding the island can be found at heights of 600 m [*Smith and Davies*, 1976]. Paleogeographic maps show the nearby D'Entrecasteaux islands (Goodenough, Moratau, Muyua, Normanby, and Sudest) just north of the Papuan Peninsula to have been submerged at 5 Ma [*Hall*, 2001, Figure 10] and at 7 Ma *Wilson* [2008, Figure 1f], but the geologic history of these islands is sufficiently complicated that the history of emergence of these islands, or perhaps submergence adjacent to them, is surely not simple [e.g., *Baldwin et al*., 2012; *Davies and Warren*, 1988; *Fitz and Mann*, 2013; *Little et al*., 2011]. *Davies and Warren* [1988] and *Little et al*. [2011] inferred that domal uplifts brought deeply exhumed rock above sea level in Pliocene time and since 6 Ma, respectively.

A8. Islands East and Northeast of New Guinea

We have found less information about islands east of New Guinea, perhaps because little change has occurred on these islands. Relatively flat Miocene limestone caps parts of New Britain [e.g., *Lindley*, 2006], but accurate dates of vertical movements seem to have been studied in few places. *Bromfield and Renema* [2011] discuss limestone in eastern New Britain, the bottom part of which they dated at 13–14 Ma, but which seems to be as young as 4 Ma in places. Thus, at least part of eastern New Britain seems to have emerged since 5 Ma. *Riker-Coleman et al*. [2006] reported a series of terraces each with corals atop them, and they inferred a present-day rate of emergence of this region between 0.4 and 2.1 mm/yr. Although the rate is not constrained well, the succession of terraces reaching 270 m above sea level attests to sustained emergence. Mostly as a guess, we assume that the island was two thirds its present size at 5 Ma, though neither *Hall* [2002] nor *Wilson* [2008] indicates an island much smaller than today at 5 and 7 Ma, respectively. Also, *Hamilton* [1979, p. 289] stated that Quaternary volcanoes built the islands of Umboi, Long, and Karkar to the west of New Britain.

Early Miocene limestone was deposited on earlier Cenozoic volcanic rock, apparently marking a period of sustained subsidence of the region that includes New Ireland, Lavongai (New Hanover), Mussau, and Manus. Marine deposition, of largely limestone, continued through Mio-Pliocene time, until the sequence that has emerged was tilted apparently in Plio-Quaternary time [*Exon and Marlow*, 1988; *Francis*, 1988; *Hohnen*, 1978; *Lindley*, 2006; *Stewart and Sandy*, 1988]. Moreover, raised marine terraces consisting of coral reach hundreds of meters in elevation [*Hohnen*, 1978]. Thus, we presume that essentially the entire region that includes these islands remained submerged until more recently than 5 Ma.

Although much of Bougainville seems to have submerged in early Miocene time, volcanic rock covers the much of the island and gives little evidence of post-Miocene emergence [*Blake and Miezitis*, 1967]. Only the northern end of the island and the adjacent island of Buka, which are covered by Quaternary reef limestone seem to have emerged since 5 Ma.

We ignore the region farther east, the Solomon Islands, but we note that abundant evidence suggests emergence of islands in this region since 5 Ma [*Bruns et al*., 1989; *Cowley et al*., 2002; *Hughes and Turner*, 1977; *Mann and Asahiko*, 2004; *Phinney et al*., 2004].

A9. Halmahera and Surrounding Islands

Most of Halmahera and the surrounding islands of Obi, Bacan, Kasiruta, and Morotai seem to have emerged since 5 Ma.

Halmahera consists of two different parts: a long narrow belt of volcanoes and volcanic rock in the west and ophiolitic mélange and oceanic crust with carbonate platforms in the east [*Hall*, 1987; *Hall et al*., 1988a]. For Bacan, which lies southwest of Halmahera, the ophiolite has been thrust atop rock possibly as old as Precambrian, but otherwise, its history seems to be similar to that of Halmahera [*Malaihollo and Hall*, 1996]. Volcanism seems to have begun at 11 Ma on Obi and progressed north to Bacan and Halmahera [*Baker and Malaihollo*, 1996]. Ages from Bacan and Halmahera begin at 7.5 Ma and 7.8 Ma, respectively. Thus, much of these islands did not exist at all until late Miocene time. *Hall et al*. [1988b] reported late Miocene reef limestone at an elevation of 1000 m over the eastern part of Halmahera. The Pliocene sequence includes marl at the base, with sandstone, siltstone, and conglomerate above, and *Hall et al*. [1988b] describe an early Pliocene "change from the stable conditions of carbonate deposition across east and central Halmahera with a transition from limestones to marls, followed rapidly by an increase in the amount of siliciclastic debris which was deposited as submarine fan turbidites." *Nichols and Hall* [1991] described a major phase of folding that requires 60 km of shortening across the eastern part of the island. In a preliminary summary of the structure of Halmahera, *Hall and Nichols* [1990] stated that this deformation began at 3 Ma. Later, *Hall* [2000] showed this to represent 60%

shortening, which almost surely would have raised a surface at sea level, with reefs, to an elevation of 1–2 km. Also, the geology of Morotai, just northeast of Halmahera, resembles that of Halmahera [*Hall et al.*, 1991], and *Hall* [2001, Figure 10, 2009, Figure 14] showed the island as submerged at 5 Ma.

There seems little doubt that most of Halmahera and surroundings underwent Pliocene crustal shortening [e.g., *Hall*, 2002, p. 382] and that most of the region emerged since 5 Ma.

A10. Philippines, Palawan, and Adjacent Islands

North of Halmahera, only small islands rise above sea level: the Sangihe Islands along the Sangihe volcanic arc and the Talaud Islands, of which Karakelong is the biggest. Karakelong seems to have emerged only in Pleistocene time [*Moore et al.*, 1981]. *Macpherson et al.* [2003] report that the Sangihe Islands are Quaternary in age.

We have found little information that places quantitative constraints on the emergence above sea level of Mindanao or other islands in the southern part of the Philippines. Nevertheless, several observations suggest that much of the Mindanao may have emerged since the latest Miocene Epoch, if not since 5 Ma. *Pubellier et al.* [1991] described the main late Cenozoic geologic event as a suturing of two blocks that began at circa 5 Ma [*Sajona et al.*, 1994, 2000]. Suturing was completed by early Pliocene time. The western block seems to be underlain continental crust that had undergone extension in early and middle Miocene time, and therefore presumably crustal thinning and submergence. The eastern block includes the northward continuation of the Sangihe Arc but seems to have been emergent during much of the Miocene Epoch. The collision of the two blocks led to folding, some overturned, and thrust faulting that attest to kilometers, if not much more, of crustal shortening [*Pubellier et al.*, 1996]. Moreover, *Quebral et al.* [1996] describe Pliocene unconformities in several places, consistent with erosion of rock that had undergone deformation just before that time. Also, *Queaño* [2005] described late Miocene-Pliocene turbidites that crop out on the western side of the Pujada Peninsula, southeastern Mindanao, but this is a small area, and it is not clear how much of the region has emerged since ~5 Ma. Based in part on *Hall*'s [2001, Figure 10, 2009, Figure 14] syntheses, we assume that half of the island, plus the small island of Siargao to its northeast, emerged since 5 Ma.

Regarding other islands, *Sajona et al.* [1997] associated a late Miocene to Pliocene unconformity on Leyte with the initiation of volcanic activity associated with subduction at the Philippine trench. Moreover, *Dimalanta et al.* [2009] noted that much of southwestern Leyte is covered by Plio-Quaternary limestone. It is difficult to know what fraction of the island has emerged since 5 Ma, but it would seem that at least half of it has. Similarly, *Faustino et al.* [2003] point out that most of Bohol is covered by Plio-Quaternary limestone and that although the island did not rise uniformly in Quaternary time, it seems to have emerged since ~5 Ma. *Faustino et al.* [2003] also noted that a northerly trending anticline passing through Cebu then began to grow in late Miocene time and exposes the same Plio-Quaternary limestone. Thus, it seems that Cebu, and presumably Siquijor to its south, also emerged since 5 Ma. To the west of Bohol, Los Negros also shares some of the same history as Bohol [*Faustino et al.*, 2003], but it is not apparent that the entire island emerged in the past 5 Ma. Farther north, for Samar Island, *Travaglio et al.* [1978] reported that much of the island was submerged in early Miocene time, when coralline limestone was deposited on sandstone and shale. They reported that emergence began late Miocene time, with different parts of the island rising different amounts at different times. With doubts about the precise date when the island merged, we assume that half of it rose by 5 Ma.

We suspect that as much as half of Mindanao and adjacent islands might have been submerged until as recently as 5 Ma. *Hall* [2001, Figure 10, 2009, Figure 14] showed much of southern Mindanao submerged at that time, but gave no discussion; after considering presently submerged surrounding terrain that he considers to have been above sea level at 5 Ma, we estimate approximately 30% of the current land above sea level was submerged. By contrast, but again without discussing the region, *Wilson* [2008, Figure 1f] and *Wilson and Moss* [1999, Figure 10] showed virtually all of southern Philippines, Mindanao, Leyte, and adjacent islands, as below sea level at 7 and 8 Ma, respectively.

Discussions by *Holloway* [1982], *Müller* [1991], and *Rangin and Silver* [1991] suggest that Palawan had emerged by middle Miocene time. Cursory study of Balabac to its southwest [*John*, 1963] suggests the same. Similarly, *Rangin* [1989] shows volcanoes on the islands of Basilan and Jolo along the Sulu Ridge as built on islands dating from early Miocene or earlier, like rock exposed on Tawi-Tawi. Farther east in the Philippines, the geologic history of Panay seems to resemble that of the Palawan and shows no evidence of post 5 Ma emergence [*Yumul et al.*, 2009].

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