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Differences in the Indonesian seaway in a coupled climate model and their relevance to Pliocene climate and El Niño

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[1] A fully coupled general circulation model is used to investigate the hypothesis that during Pliocene times tectonic changes in the Indonesian seas modified the Indo-Pacific heat transport and thus increased the zonal sea surface temperature gradient in the equatorial Pacific to its large, current magnitude. We find that widening the Indonesian seaway by moving the northern tip of New Guinea south of the equator leads to an increased inflow of South Pacific waters into the Indian Ocean, but because of potential vorticity constraints on cross-equatorial flow it also leads to reductions in both the inflow of North Pacific waters and the total Indonesian throughflow transport. The reduced throughflow is matched by increased eastward transport of warm and fresh North Pacific surface waters along the equator to the central equatorial Pacific. As a result, the Intertropical Convergence Zone lies closer to the equator and the western Pacific Warm Pool expands farther east than for present-day conditions. This reduces the delayed oscillator component of El Niño–Southern Oscillation (ENSO) and enhances the role of stochastic perturbations. Thus, with a more open Indonesian seaway, ENSO becomes weaker and more irregular.

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1. Introduction

[2] Because nearly all water that enters the Pacific Ocean from the Southern Ocean passes out of the Pacific Basin between the islands of Indonesia, variability in the currents that pass through the Indonesian Sea and link the Pacific and Indian Oceans are often assigned a major role in modern climate [e.g., *Gordon*, 1986; *Hirst and Godfrey*, 1993; *Schneider*, 1998]. Moreover, because this region has been evolving rapidly over the past tens of millions of years, an evolving Indonesian seaway has long offered a target for attributing cause, via its effect on the Indonesian throughflow (ITF), to changing global climates on geological timescales [e.g., *Kennett et al.*, 1985; *Srinivasan and Sinha*, 1998; *Cane and Molnar*, 2001]. It follows that attributions of global climate change to an evolving Indonesian seaway require an understanding of current processes that govern its throughflow.

[3] Problematic to both modern and paleoclimate is the fact that these currents are governed by some of oceanography's least quantified and understood processes. Observations are made difficult by the myriad of islands and the large temporal variability from intraseasonal to interannual timescales [e.g., *Sprintall et al.*, 2003]. Theories about the ITF are difficult to develop because the weak Coriolis effect requires a balance, necessarily uncertain and difficult to quantify, between pressure gradients, wind stress, and viscous drag. Nevertheless, the currently dominating theory that explains the ITF as the result of mass continuity and Sverdrup balance ("Island Rule") [*Godfrey*, 1989] and direct observations of transport both arrive at a transport estimate for the ITF between 5 and 15 sverdrups (Sv) [*Godfrey*, 1996; *Hautala et al.*, 2001]. This agreement is reassuring, but the fact that most of the ITF water originates from the North Pacific [*Gordon and Fine*, 1996] and not from the South Pacific as required by the Island Rule leaves some doubt [*Nof*, 1996]. The theoretical and observational uncertainty surrounding the ITF makes the application of numerical tools a natural but possibly misleading step. Still, if nothing else, numerical studies highlight the importance of basin geometry [*Morey et al.*, 1999; *Rodgers et al.*, 2000], topography, and dissipation [*Wajsowicz*, 1993a, 1993b]. The uncertainties are even larger for the impact of the ITF on the surface ocean heat budget. Although it is of obvious relevance to climate it is much less understood, largely because of observational challenges. For example, the heat carried by the ITF may only have a minor effect on climate [e.g., *Vranes et al.*, 2002], because much of the ITF water is colder than 27°C, and not only originates but also returns to the surface outside the Indo-Pacific warm pool. Recently, however, *Jochum and Potemra* [2008] showed that the enhanced levels of diapycnal diffusivity in the Indonesian Seas can connect these subsurface waters to the surface and change tropical climate.

[4] The present study takes its motivation from the hypothesis that sometime during Pliocene times (approximately 3 million years ago, 3 Ma) the ITF changed its source waters, which triggered Northern Hemisphere glaci-

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ation [*Cane and Molnar*, 2001] (for a general review of theories on Pliocene climate see *Fedorov et al.* [2006]). *Cane and Molnar* [2001] speculate that the emergence of Halmahera and the northward movement of New Guinea blocked the New Guinea Coastal Current, which today supplies only a small amount of South Pacific water to the ITF, and allowed North Pacific water brought by the Mindanao Current to comprise the majority of the ITF. This, according to their hypothesis, drastically affected the inter-basin heat exchange, and led to a larger zonal SST gradient, and a cooling of the higher latitudes.

[5] Although neither Pliocene tectonic movement nor the development of Northern Hemisphere glaciation is disputed, their connection remains speculative; it could be that these two events happened coincidently during the Pliocene Epoch. *Haywood et al.* [2007] and *Lunt et al.* [2008] provide an extensive discussion of the uncertainties in tropical Pacific paleodata, and list possible causes for the onset of Northern Hemisphere glaciation. On the basis of numerical simulations of Pliocene climate they then conclude that it is unlikely that changes to the atmospheric trace gas concentration alone could trigger the glaciation. The present study should be viewed as a companion case to these two studies in that it investigates one more possible mechanism by which the tropical oceans could trigger climate change.

[6] The first part of *Cane and Molnar*'s [2001] hypothesis, the ITF-SST gradient connection, was supported by forced ocean model (OGCM) studies [*Hirst and Godfrey*, 1993; *Morey et al.*, 1999; *Rodgers et al.*, 2000] and the second part, the connection between SST gradient, ENSO and high-latitude cooling, can be examined with a coupled climate model (GCM) in a straightforward manner. Furthermore, the hypothesis and the aforementioned OGCM studies suggest a large response to changes in the Indonesian passages.

[7] This study describes possible differences in climate due to differences in positions of the Indonesian islands. The next section describes the GCM and the experiment, the third section discusses the differences in the ITF, and section four shows how ITF differences lead to differences in ENSO properties. A summary and discussion concludes the present manuscript.

2. Description of Model and Experiment

[8] The numerical experiment is performed using the National Center for Atmospheric Research (NCAR) Community Climate System Model version 3 (CCSM3), which consists of the fully coupled atmosphere, ocean, land and sea ice models; a detailed description can be found in the work by *Collins et al.* [2006].

[9] The ocean model has a zonal resolution that varies from 340 km at the equator to 40 km around Greenland, and a meridional resolution that varies from 70 km at the equator to 40 km around Greenland and 350 km in the North Pacific. This spatially varying resolution is achieved by placing the north pole of the grid over Greenland, and reflects the different relevant length scales of the 2 processes that are deemed most important to maintain a stable global climate: deep convection around Greenland and in the Arctic, and oceanic heat uptake at the equator. In the vertical there are 25 depth levels; the uppermost layer has a thickness of 8 m, the deepest layer has a thickness of 500 m. The atmospheric model (Community Atmosphere Model, CAM3) uses T31 spectral truncation in the horizontal (about 3.75° resolution) with 26 vertical levels. The sea ice model shares the same horizontal grid as the ocean model and the land model is on the same horizontal grid as CAM3. This setup (called T31x3) has been developed specifically for long paleoclimate integrations and its performance is described in detail by *Yeager et al.* [2006]. The most significant difference between the present model setup (CONT) and the one described by *Yeager et al.* [2006] is the new convection scheme, which leads to significant improvements in the simulation of ENSO [*Neale et al.*, 2008].

[10] While the coarse resolution allows long integration times, it limits the ability to represent the narrow passages in the Indonesian Seas. In particular the main passages, Makassar, Ombai and Timor Strait, are replaced by a one grid point throughflow between Borneo and New Guinea (Figure 1). Thus, the present study cannot claim to quantify the impact of a single island like Halmahera or Sulawesi, on climate. Rather, it seeks to identify the physical processes that are affected by island topography and are relevant to climate. For the sake of argument we will use the name Makassar Strait for the one grid point (or two grid points in the sensitivity run) that connects the Pacific and Indian ocean between Borneo and New Guinea.

[11] For the present study, we carried out five runs: 3 forced OGCM simulations, and 2 coupled T31x3 simulations; all are initialized with horizontally averaged [*Levitus et al.*, 1998] temperature and salinity fields. The forced integrations use a recently compiled climatology of seasonally varying surface fluxes [*Large and Yeager*, 2008] as an upper boundary condition, and the resulting subtropical and tropical circulation is shown in Figure 2. All forced OGCM runs, a control (CONTF) and two sensitivity runs (PLIOF, PLIOFvisc), are integrated for 250 years. PLIOFvisc is identical to PLIOF, but in PLIOFvisc horizontal friction along all western boundaries is increased by a factor of 10 [*Large et al.*, 2001]. It takes approximately 100 years to equilibrate the ITF transport, so that the coupled simulations (CONT, PLIO) are both integrated for 200 years. The presented results are based on the means of year 200 for the forced runs, and the means of years 160–200 for the coupled runs. The only difference between CONT(F) and PLIO(F) (here, and in what follows, the forms "CONT(F)" and "PLIO(F)" are short for "CONT and CONTF" and for "PLIO and PLIOF") is the island geometry in the Indonesian Sea. The removal of the northwestern tip of New Guinea approximates conditions in early Pliocene time, 3–5 Ma: the island of Halmahera has emerged approximately 1000 m since 5 Ma and New Guinea has moved northward approximately 200 km since 3 Ma [e.g., *Hall*, 2002, Figure 1]. It should be noted, though, that Pliocene island geometry is highly uncertain. The present coarse resolution ocean model cannot resolve Halmahera, or the details of the coastline, so we focused on two aspects of the island geometry that are deemed important [*Rodgers et al.*,

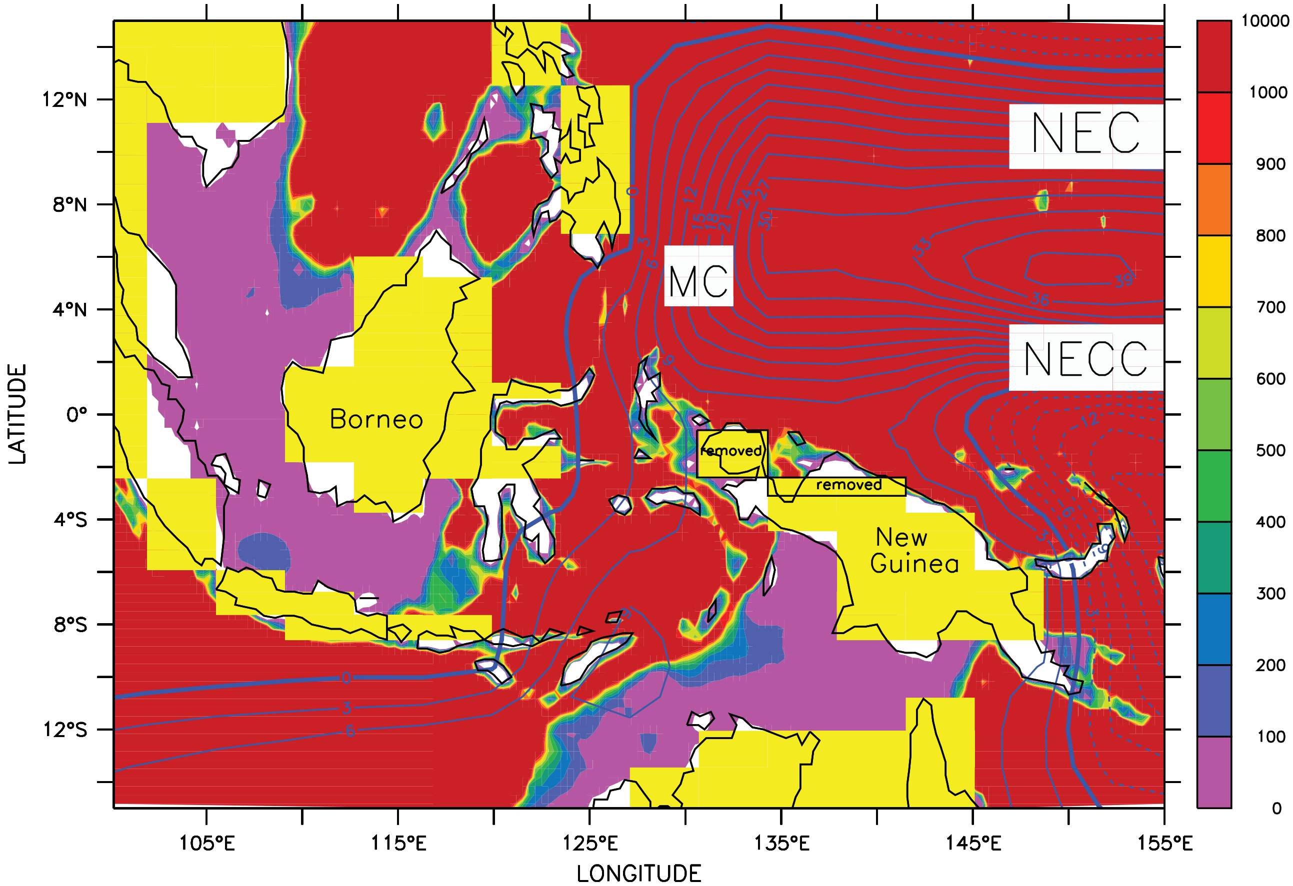

Figure 1. Observed present-day bathymetry (in color) and coastline (black contours), barotropic stream function from CONTF (blue, contour interval 3 Sv), and model land grid cells (bright yellow). The white within the black contours indicates land that has been turned into ocean because of the coarse model resolution. In PLIO(F) the northernmost part of New Guinea has been converted from land into ocean cells (marked "removed"). Note that neither the island of Halmahera (1°N, 127°E) nor the Makassar Strait (straddling the equator along approximately 119°E) is present in any of the simulations. The stream function indicates the major ocean currents in this area: the westward North Equatorial Current (NEC), which splits to feed the northward Kuroshio (north of this image's domain), and the southward Mindanao Current (MC). The MC mostly feeds into the eastward North Equatorial Countercurrent (NECC), but a fraction feeds the Indian Ocean via the ITF.

2000; *Cane and Molnar*, 2001]: the Indonesian seaway was wider, and extended into the Southern Hemisphere. Given these uncertainties, and the numerical complexities that are involved in turning land into ocean (in particular the changes required to the river runoff schemes), we decided not to change the land surface at all, but let the water flow below the land. Thus, the land surface and river runoff maps are identical all simulations, but in PLIO(F) water can flow below land in the four grid cells shown in Figure 1. This approach is relatively simple to implement, but the drawback is that the ocean does not receive atmosphere fluxes on these four points. The other possible approach, removing the land surface, would be more complicated, and would require us to invent a new river runoff map for Pliocene times. Thus, our calculations are only a first attempt to examine how differences in island geometry that affect the ITF might lead to differences in climate that resemble differences between Pliocene and present-day climates. To sum up, five simulations were performed: CONTF, an ocean simulation forced with present-day forcing; PLIOF, an ocean simulation forced with present-day forcing and a widened Indonesian seaway; PLIOFvisc, identical to PLIOF but with increased friction at the western boundaries; CONT, a fully coupled present-day simulation; and PLIO, a fully coupled simulation with a widened Indonesian seaway.

[12] The difference in thermocline temperature between CONTF and PLIOF is relatively small, but consistent with the results from *Rodgers et al.* [2000]: a warming of the Pacific and a cooling of the Indian ocean thermocline (Figure 2). These subsurface differences in the forced runs are also present in the coupled runs (not shown) where they translate into SST differences of less than 0.5°C (section 4) and a slightly more equatorward position of the Pacific

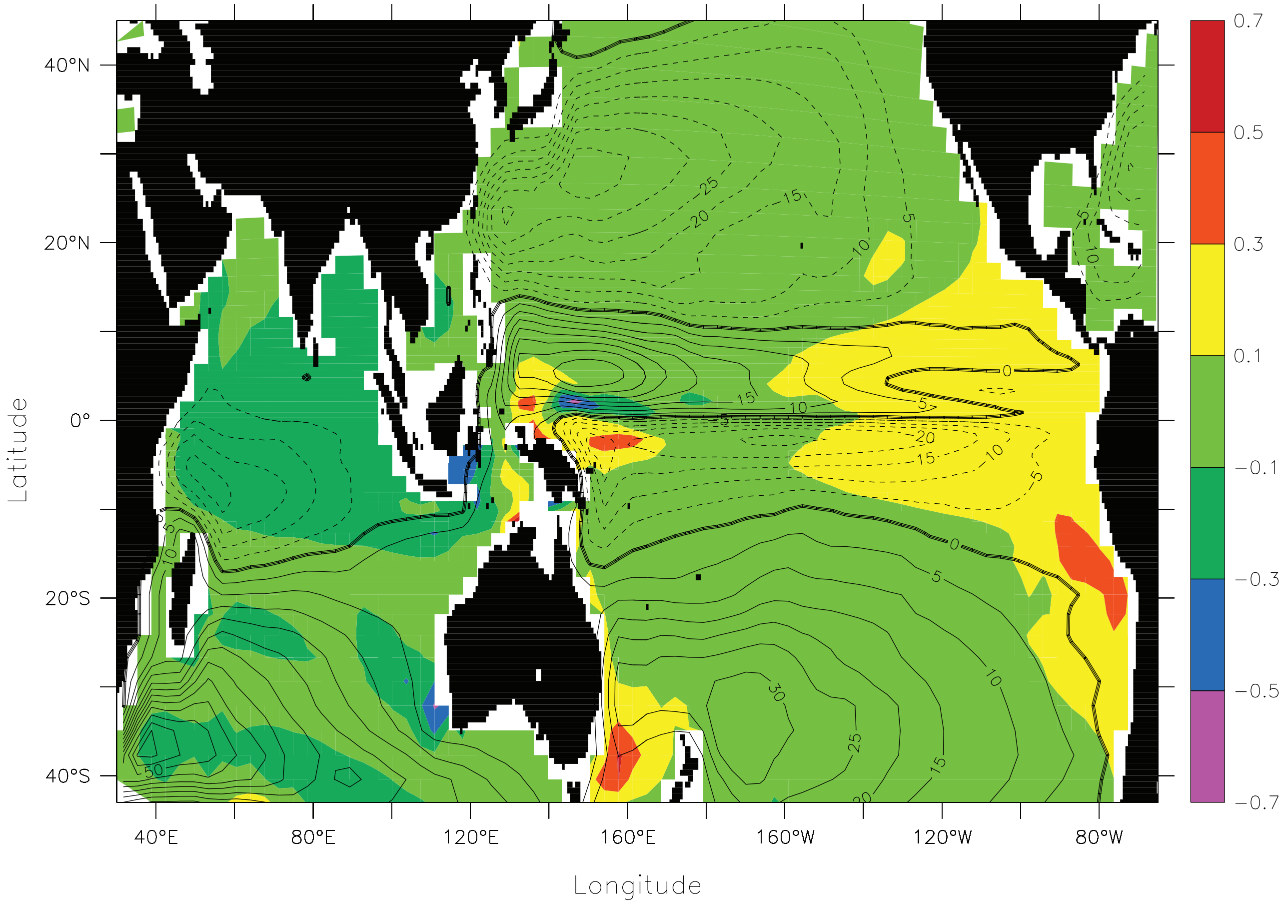

Figure 2. Temperature difference at 150 m depth between PLIOF and CONTF (PLIOF − CONTF). The contour lines (interval 5 Sv) illustrate the circulation by showing the barotropic stream function for CONTF. Positive values show that the temperature is larger in PLIOF than in CONTF.

Intertropical Convergence Zone (ITCZ) in PLIO (Figure 3). The causes behind these differences are discussed in the next 2 sections.

3. Indonesian Throughflow

[13] Virtually all large-scale flows in the ocean have large Reynolds and small Rossby numbers, so viscous and inertial effects are negligible. The dominant dynamics neglecting these effects (Ekman transport, geostrophy, and the Sverdrup balance) apply nearly everywhere and specify the depth-integrated flow given only the single boundary condition of no normal flow. In a closed basin, however, two boundary conditions are required: no normal flow through both the eastern and western boundaries. *Veronis* [1973] shows that a reasonable model with two boundaries can be formed by assuming that viscous and inertial effects are confined within a western boundary current. This current can be arbitrarily thin, but it always transports enough mass to close the circulation. *Godfrey*'s [1989] Island Rule takes advantage of this situation by considering only closed contours of depth integrated flow that avoid western boundary currents. This allows a calculation of the flow around an island based only on wind stress (independent of the details of viscous and inertial effects).

[14] The island rule has been employed to determine the flow around Australia [*Godfrey*, 1989], Hawaii [*Qiu et al.*, 1997; *Firing et al.*, 1999], and even around entire continents to estimate transport through the Bering Strait [*De Boer and Nof*, 2004]. The Island Rule transports for the ITF here are based on the line integral of the annual mean wind stress around the western and southern coast of Australia, along the latitude of Tasmania to South America, along the western coast of South America to the latitude of the northern tip of New Guinea, to New Guinea, and along the eastern coast of New Guinea and northern coast of Australia (as in the work by *Godfrey* [1996]).

[15] The time to spin up the ITF transport is approximately 100 years (Figure 4), consistent with the expectations from the Island Rule (travel time of Rossby waves between Tasmania and South America [*Godfrey*, 1996]). The ITF transports, too, are roughly consistent with the Island Rule (Table 1). Note that the larger transports in the coupled runs are due to a westerly wind bias in the southern midlatitudes, a well documented but still not resolved bias in CCSM [*Boville*, 1991], which forces more water into the Pacific basin and hence requires a larger ITF transport. The

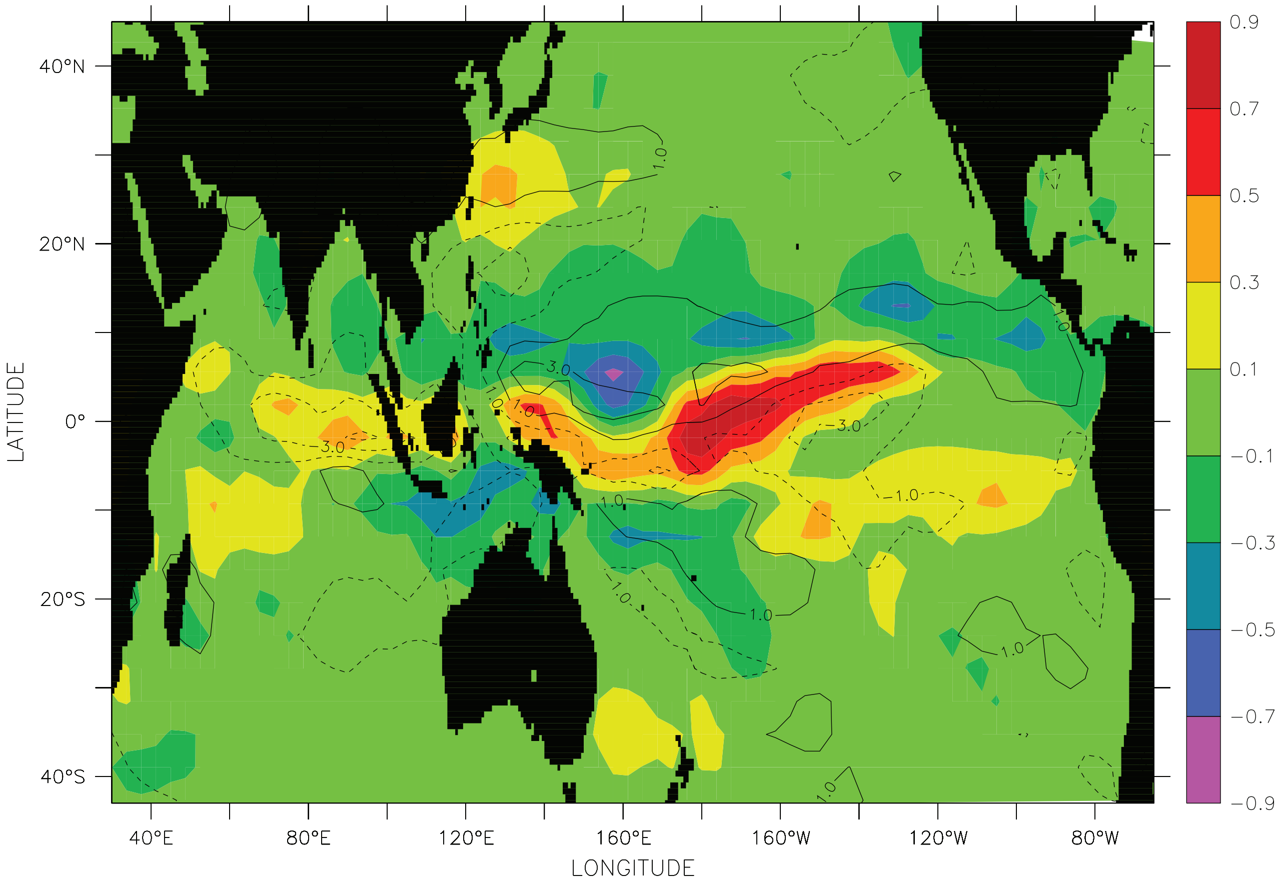

Figure 3. Precipitation difference between PLIO and CONT (color; in mm/d) and relative difference in surface wind speed (contour interval: 2% increase in wind speed in PLIO compared to CONT). For clarity only differences over sea are shown. The red color in the central and western tropical Pacific indicates that the ITCZ is shifted equatorward, which weakens the winds south of the ITCZ and strengthens them to the north of it.

Island rule predicts larger transports in PLIO(F) than in CONT(F): Because of the more southerly northern edge of New Guinea in PLIO(F), the Island Rule transports are based on the stronger Trade winds at this latitude. It is a key result that the actual transports in PLIO(F) are smaller than those in CONT(F), although the Island Rule predicts a larger transport. From this we conclude that the ITF transport is well approximated by the Island Rule, but that the disagreements of up to 30% (Table 1) indicate that additional, secondary processes also affect the flow. These secondary effects are larger in PLIO(F) than in CONT(F).

[16] In principle the ITF transport could be modified by Pacific upwelling [*Stommel and Arons*, 1960; *Gordon*, 1986], topography [*Wajsowicz*, 1993a], viscosity [*Wajsowicz*, 1993a], and advection of momentum or vorticity (nonlinearities) [*Inoue and Welsh*, 1993]. Note that, together with the curvature of the flow, viscosity determines the dissipation of momentum and vorticity. Thus, larger viscosity leads to reduced nonlinearities, whereas increased nonlinearities are not necessarily the result of reduced viscosity. Because the mean atmospheric forcing does not differ in the forced runs, and only by a little in the coupled runs, one can rule out the contribution of upwelling. Also, by design, topography and viscosity were held constant in all runs, which leaves nonlinearities to explain the differences between CONT(F) and PLIO(F). In CONTF, 7.7 Sv of North Pacific water join the ITF through the model's Makassar Strait, and 1.1 Sv of South Pacific water enter through the Torres Strait between Australia and New Guinea (Figure 5, left). The total transport is consistent with observations [e.g., *Hautala et al.*, 2001], and the Torres Strait transport is only poorly constrained by observations (A. Gordon, personal communication, 2008). The Torres Strait is shallow and *Wyrtki* [1961] estimates its transport to be less than 1 Sv. The value in CONTF is larger than that, but not large enough to justify drastic measures like closing the Torres Strait in the simulations. In PLIOF, in spite of a wider Makassar Strait, the North Pacific inflow is reduced from 7.7 to 6.4 Sv, and the Torres Strait transport is increased from 1.1 to 1.6 Sv (Figure 5, right). The differences are the same in structure but larger in magnitude for the coupled runs: 12.1 Sv through Makassar and 3.0 Sv through Torres in CONT, versus 8.4 Sv and 4.5 Sv in PLIO (Figure 6). It appears that by reducing the northernmost extent of New Guinea, part of

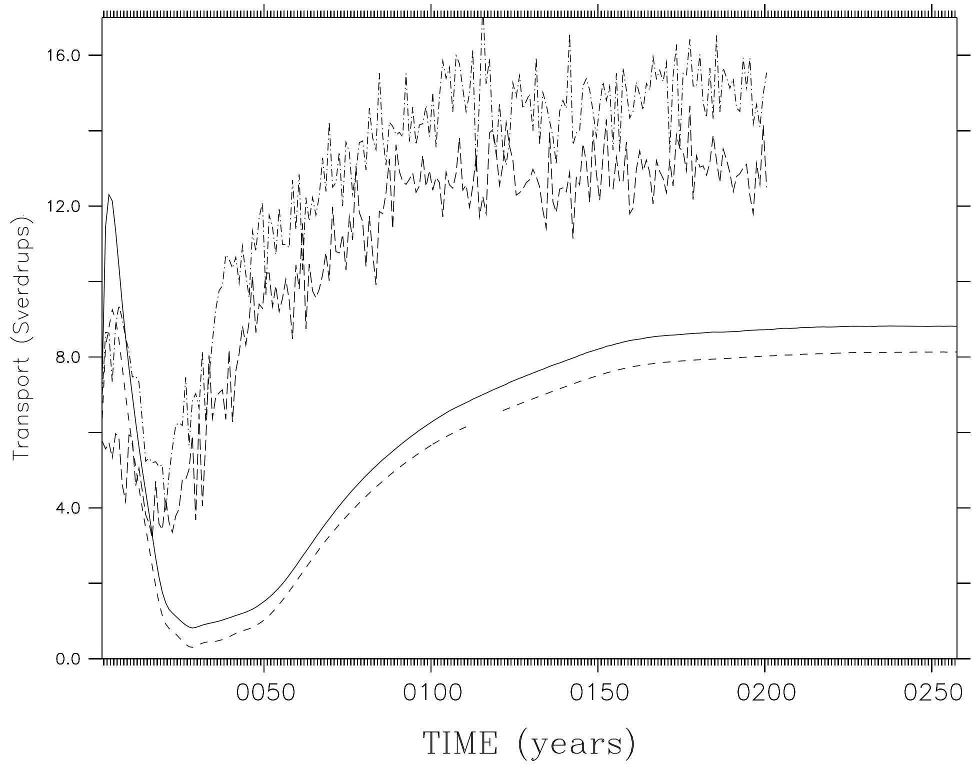

Figure 4. Time series of ITF transport: solid, CONTF; dashed, PLIOF; dash-dotted, CONT; and broken, PLIO.

the North Pacific water that in CONT(F) is destined for the Indian Ocean is forced to retroflect across the equator and remain in the Pacific. This increases the total and relative contribution of South Pacific water to the ITF (Table 1). The latter is consistent with the studies by *Morey et al.* [1999] and *Rodgers et al.* [2000], although the path of the South Pacific water is different in their studies.

[17] Conservation of potential vorticity severely constrains inviscid flow across the equator [*Killworth*, 1991; *Edwards and Pedlosky*, 1998]. Applied to the present case this implies that dissipation of potential vorticity is necessary to allow a nontrivial ITF transport of North Pacific water, and in the absence of dissipation the application of the Island Rule should be limited to islands confined to a single hemisphere. To demonstrate the importance of dissipation, PLIOF is repeated with horizontal friction along western boundaries increased by a factor of 10 (PLIOFvisc). The Makassar Strait transport increases from 6.4 to 7.2 Sv and the Torres Strait transport decreases from 1.6 to 1.2 Sv. Thus, increasing viscosity increases the ITF and especially the Makassar Strait transport. Of course, changing the size of New Guinea does not affect the value of viscosity along the western boundary, but it does affect the dissipation of momentum, the product of viscosity and the curvature (or second derivative) of velocity. In CONT(F) the Makassar Strait is treated using only one active velocity grid point, with the adjacent grid points being set to zero by the no-slip boundary conditions. The Makassar Strait in PLIO(F) is wider, thereby reducing the total dissipation and forcing more water to retroflect and stay in the Pacific. Note that the northward return flow of the ITF in the South Pacific is under the same potential vorticity constraint: the New Guinea Coastal Current does not cross the equator to join the ITF directly but retroflects to feed the eastward flowing Equatorial Undercurrent. This water can cross into the Northern Hemisphere only in the diabatic Ekman layer where potential vorticity is dissipated [see also *Godfrey*, 1996]. This is analogous to the tropical Atlantic where the northward flowing North Brazil Current has to retroflect into the Atlantic Equatorial Undercurrent to adjust its potential vorticity. In the Atlantic, however, some Southern Hemisphere water is trapped in eddies, which travel north-

Table 1. ITF Transports From the Simulations and From Godfrey's Island Rule[a]

	Transports		
	Island Rule	Model ITF	Source
CONTF	11.8	8.8	7.7/1.1
PLIOF	13.2	8.0	6.4/1.6
CONT	17.1	15.1	12.1/3.0
PLIO	18.1	12.9	8.4/4.5

[a]ITF transports are in Sv. The simulations are listed in the first column, the second column shows the transports based on the Island Rule, the third column shows the actual transports, and the fourth column shows the split between Northern Hemisphere and Southern Hemisphere source waters in the simulations (cannot be predicted from the Island Rule).

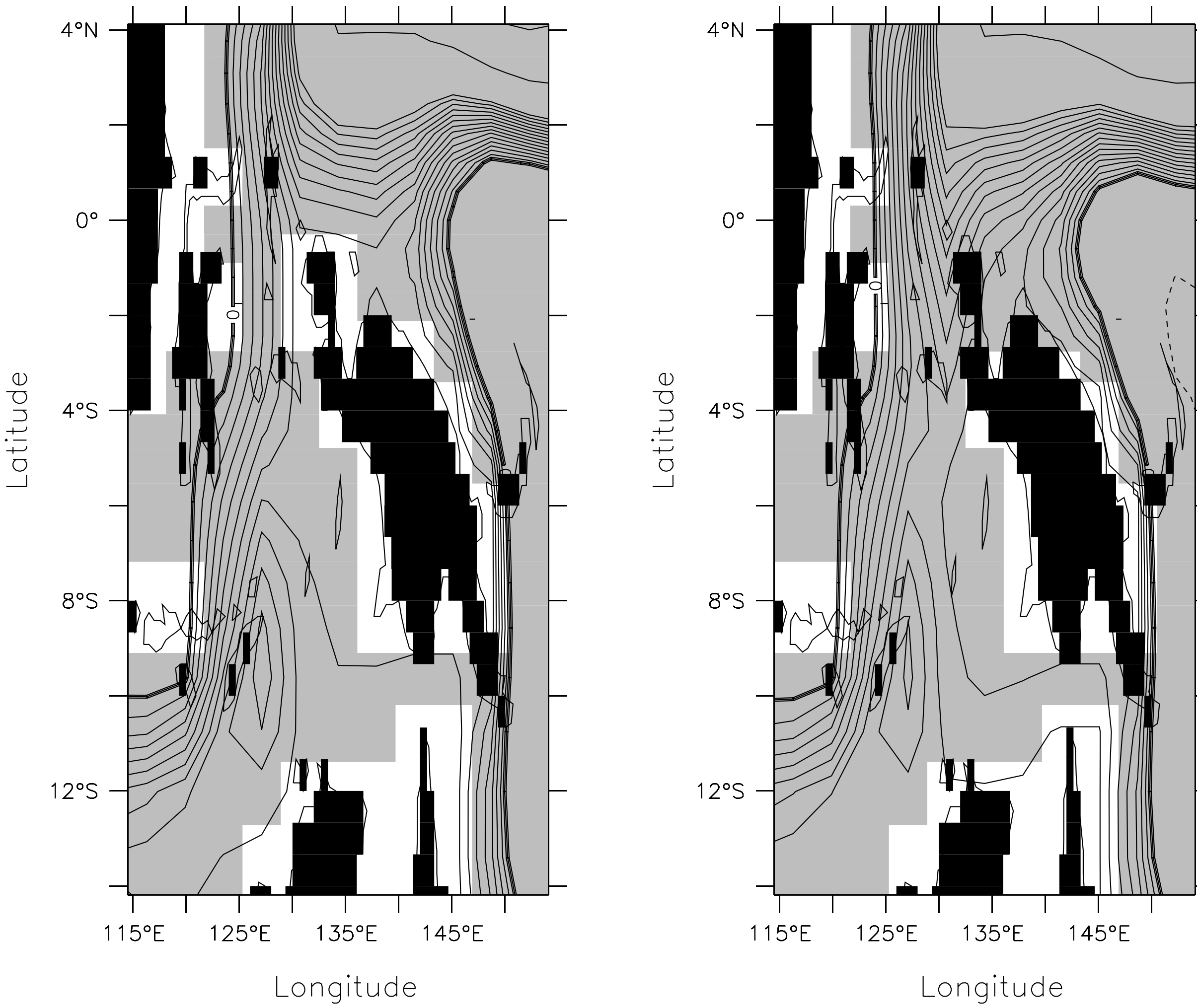

Figure 5. Barotropic streamlines for (left) CONTF and (right) PLIOF. Contour interval is 1 Sv for the first 15 Sv and is 5 Sv after that; real land is shown in black, and the ocean model land mask is shown in white (see Figure 1 for a more detailed depiction of the domain). The central island is New Guinea, and the northern tip of Australia can be seen in the bottom of the images. Note the retroflection of the southward Mindanao Current (crowding of streamlines entering the image at 4°N) as it crosses the equator in PLIOF.

west along the coast. This eddy transport is possible in the Atlantic because of the weakness or absence of a low-latitude western boundary current there [*Jochum and Malanotte-Rizzoli*, 2003].

[18] We are aware of only two OGCM studies that considered a land-sea geometry in the Indonesian region similar to that of PLIOF, and both also found a reduction in the total ITF transport, albeit by only 4% [*Morey et al.*, 1999; *Rodgers et al.*, 2000]. These reductions are smaller than the 9% and 15% that we found for PLIOF and for the coupled run PLIO, but similar to the 5% reduction in the sensitivity study PLIOFvisc. These differences are probably too small to be verified with paleo-observations [*Huybers et al.*, 2007]. It is noteworthy that, assuming identical wind fields and the validity of the Island Rule, the more southerly latitude of New Guinea appropriate for Pliocene time calls for a larger ITF transport. This increase is predicted because the northernmost edge of New Guinea is shifted from the equator, where Trade winds are weak, to a more southern latitude, where they are stronger. Thus, previous OGCM studies and the current simulations all suggest that widening the Indonesian passages near the equator leads to increased nonlinearity and, in contrast to the Island Rule, to reduced ITF transport. The amount of reduction will depend on the details of island geometry, topography and the strength of the diabatic processes acting on the Mindanao Current. In an OGCM the nonlinearity will depend on viscosity, diffusion, and boundary conditions, the numerical values of which are only poorly constrained by theory or observations [*Jochum et al.*, 2008]. With the caveat that the presently modeled differences in the ITF are sensitive to resolution

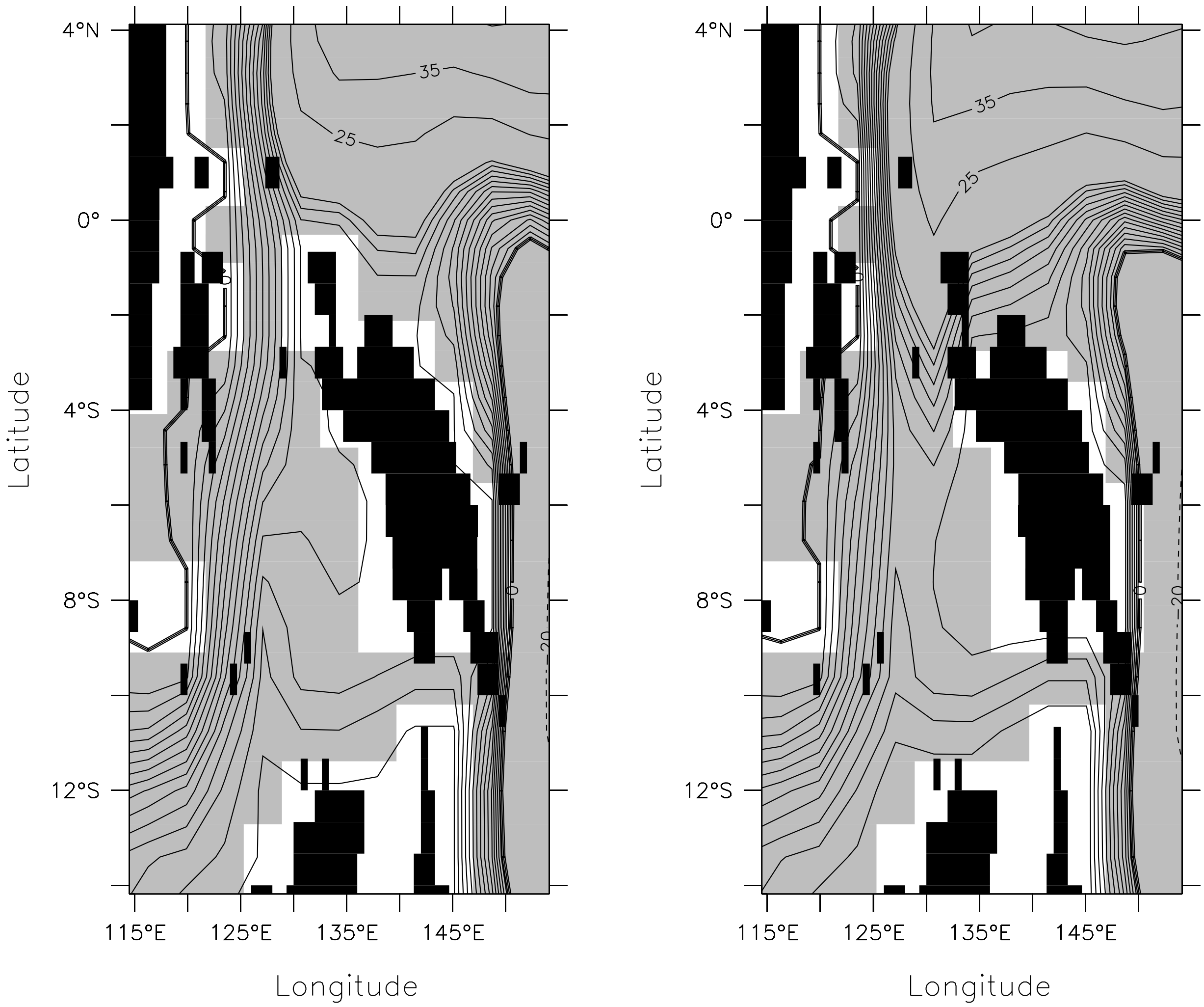

Figure 6. Same as Figure 5 but for (left) CONT and (right) PLIO.

and parameterized dissipation, the next section discusses their climate impact.

4. Climate Response

[19] The differences in circulation that are caused by modifications in Indonesian island geometry lead to small differences in mean temperature (Figure 2), precipitation and surface winds (Figure 3), but clear differences in ENSO properties (Figures 7 and 8). The spectral peak of NINO3 SST (SST anomaly averaged between 150°W–90°W and 5°S–5°N) shows that the average recurrence interval of El Niño is shifted from 2.4 years in CONT to 3.3 years in PLIO, and its amplitude is reduced by more than 10% from a standard deviation of NINO3 SST of 0.74°C–0.65°C. Moreover, compared to CONT, ENSO in PLIO has more energy at periods longer than 5 years. The causes of the differences in ENSO properties are discussed in the present section.

[20] The western warm pool is buttressed against the Indonesian islands, which are surrounded by some of the worlds warmest and freshest surface waters (Figure 9). These waters are advected eastward by the retroflected part of the Mindanao Current as part of the northern Pacific tropical gyre. A stronger retroflection makes stronger eastward flow in the western Pacific and weaker westward flow in the east (Figure 9, bottom), which extends the western Warm Pool eastward and reduces the extent of the cold tongue in the eastern equatorial Pacific (Figure 10). This warming leads to an increase in rainfall and surface wind convergence on the equator. Thus, 3.7 Sv of warm and fresh Mindanao Current waters, which join the ITF in CONT, instead enter the western equatorial Pacific, where they lead to a southward shift in the ITCZ (Figure 10). This fresher and warmer water leads to increased upper ocean stratification in the western Pacific (not shown), and can reduce entrainment of subsurface water into the mixed layer [*Lukas and Lindstroem*, 1991; *Yeager et al.*, 2006].

[21] How does this more equatorially centered position of the ITCZ in PLIO affect ENSO? There are numerous dynamical regimes proposed to explain ENSO behavior (see *Wang and Picault* [2004] for an overview), but on the most fundamental level the question is whether ENSO is

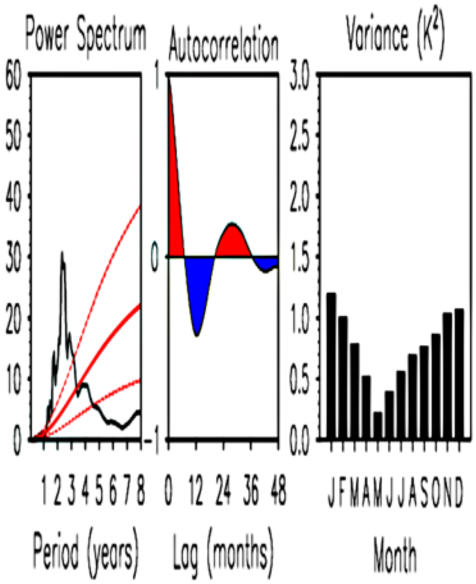

Figure 7. Statistical analysis for NINO3 SST anomalies of the years 101–200 for CONT. (left) The variance per frequency band, (middle) the autocorrelation, and (right) the seasonal distribution of the variance. The solid line across the power spectrum marks the variance for an AR(1) process (noise with memory), and the part of the spectrum that is above the uppermost dashed line is significant at the 99% level.

a series of events or a delayed oscillator. In the "series of events" regime, the tropical Pacific is in equilibrium, and it takes strong stochastic forcing to trigger an El Niño event. In the "delayed oscillator" regime, equatorially trapped Kelvin waves travel eastward across the tropical Pacific Ocean, and reflect into westward propagating Rossby waves. Being reinforced by atmospheric feedbacks, these 2 oceanic planetary waves grow into successive El Niño and La Niña events. In general, the delayed oscillator leads to a more regular ENSO, which is inconsistent with present-day observations [*Kessler*, 2002], but could feasibly have been a natural mode in the past [e.g., *Garcia-Herrera et al.*, 2008].

[22] *Neale et al.* [2008] showed that when implemented into the high-resolution version of CCSM, their modifications to the convection scheme transform ENSO from a delayed oscillator into a series of independent events, with little memory of previous events. Their modifications achieve this by reducing the spuriously strong off-equatorial ocean-atmosphere coupling in the central and eastern Pacific of the model, and by creating westerly wind bursts of realistic strength in the western Pacific. The former weakens the deterministic nature of the delayed oscillator, the latter adds stochastic forcing, and the two together make ENSO more irregular and less frequent. Compared to CONT, ENSO in PLIO becomes more irregular, weaker, and less frequent. Thus, to explain the shifted spectrum in PLIO, the westerly wind bursts or the off-equatorial coupling must have become weaker. Inspection of the model fields show that, because of the coarse resolution in the present GCM, westerly wind burst activity is almost nonexistent in either run (not shown), but the off-equatorial coupling in the eastern and central Pacific, between 5°N and 10°N and between 10°S and 20°S (Figure 11), is weaker in PLIO than in CONT. Thus, the more equatorial position of the ITCZ in PLIO affects the response of the Trade winds to an El Niño event, which reduces the strength of off-equatorial Rossby waves (not shown). As an integral part of the delayed oscillator, these weaker Rossby waves reduce the role of that mechanism in starting and ending El Niño events, making ENSO more irregular and weaker.

5. Summary and Discussion

[23] A fully coupled general circulation model is used to investigate the hypothesis that in Pliocene time tectonic changes in the Indonesian Seas led to a different Indo-Pacific heat transport, and an increased zonal SST gradient in the equatorial Pacific. A more open Indonesian seaway with the northern edge of New Guinea 200 km south of its current position, as it was in Pliocene time, does lead to a greater flow of South Pacific waters into the Indian ocean, which is consistent with previous, forced ocean model studies. Because of potential vorticity constraints to cross-equatorial flow, ITF transport during Pliocene time may have been smaller than today. The resulting greater flow of warm and fresh North Pacific water to the central equatorial Pacific leads to an equatorward shift of the ITCZ. This reduces the coupling between equatorial SST and off-equatorial wind stress, thereby weakening the delayed oscillator regime and thus creating a weaker and less regular ENSO.

[24] *Cane and Molnar* [2001] suggested that the northward movement of New Guinea would have blocked relatively warm water in the Pacific south of the equator,

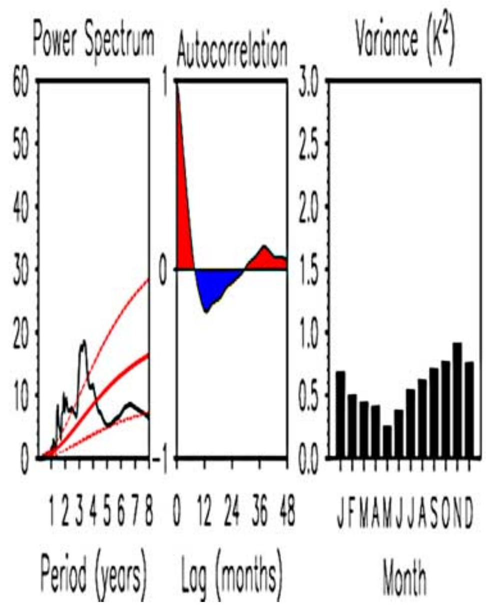

Figure 8. Same as Figure 7 but for PLIO.

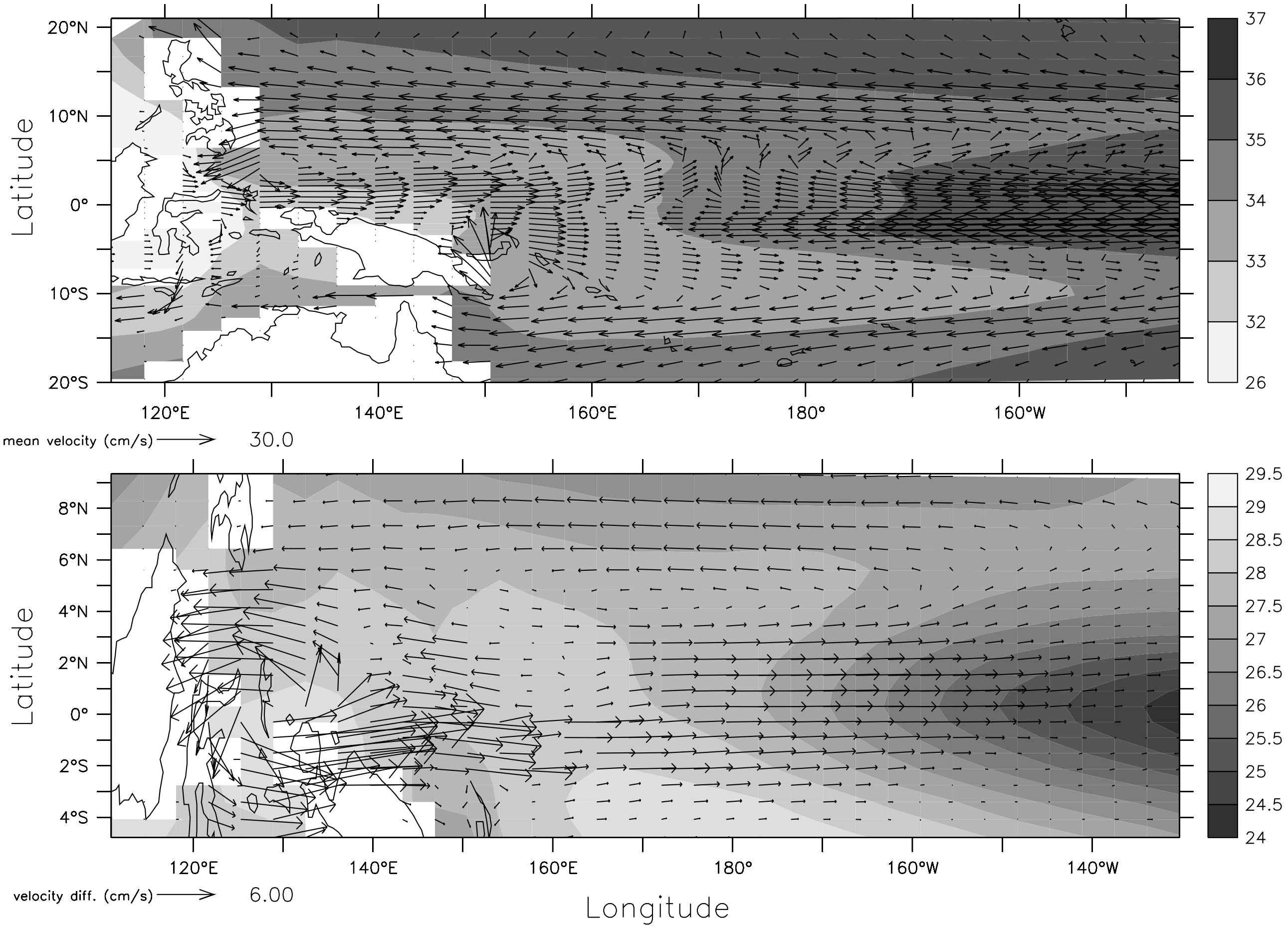

Figure 9. (top) Mean surface salinity and velocity in the western tropical Pacific region of CONT. (bottom) Mean surface temperature of CONT and difference in surface velocity between PLIO and CONT. Surface is defined here as the upper 50 m. Note that the velocity changes happen in a region of a large east-west buoyancy gradient in temperature and in salinity.

so that ITF would have come largely from the cooler water from the Pacific north of the equator. OGCM runs by *Rodgers et al.* [2000], indeed, showed that with a distribution of islands in Indonesia mimicking those today, water entering the Indian Ocean would be approximately 2°C cooler than if the northern part of New Guinea were removed, and a direct passage from the Pacific to the Indian Ocean were opened south of the equator. *Rodgers et al.* [2000] also calculated that with present-day winds and an island geometry appropriate for 3–5 Ma, the temperature at a depth of about 100 m in the central equatorial Pacific would be about 0.5°C cooler than for present-day conditions. On the basis of this *Cane and Molnar* [2001] speculate that the narrowing of the Indonesian seaway blocked warm water in the Pacific south of the equator to form, or strengthen, the Western Pacific Warm Pool. This in turn strengthened the Walker Circulation, and transformed equatorial Pacific climate from one resembling that during El Niño events, with weak zonal SST gradients, to the present state with a strong SST gradient. They speculate further that, like present-day teleconnections during El Niño events, a warm eastern Pacific would have maintained a warm North America and would have prevented ice sheets, or ice ages.

[25] The present results show subtle tendencies in the direction that *Cane and Molnar* [2001] suggested, but they offer little support for the mechanisms that they used to argue for a weaker zonal temperature gradient across the Pacific in Pliocene time. A more open seaway leads to an SST in the central Pacific that is slightly warmer (0.2–0.3°C) than for CONT, and hence to a slightly more easterly extent of the Warm Pool. This effect, however, is miniscule compared to the 3–4°C difference between Pliocene and present-day SSTs in the eastern equatorial Pacific [e.g., *Wara et al.*, 2005; *Lawrence et al.*, 2006]. Moreover, the explanation for the warmer central equatorial Pacific associated with the more open seaway is different from what Cane and Molnar imagined. With the northern edge of New Guinea at 2°S, compared to its present-day equatorial position, less water from north of the equator is calculated to pass through Indonesia into the Indian Ocean. Although its inertia carries this North Pacific water across the equator,

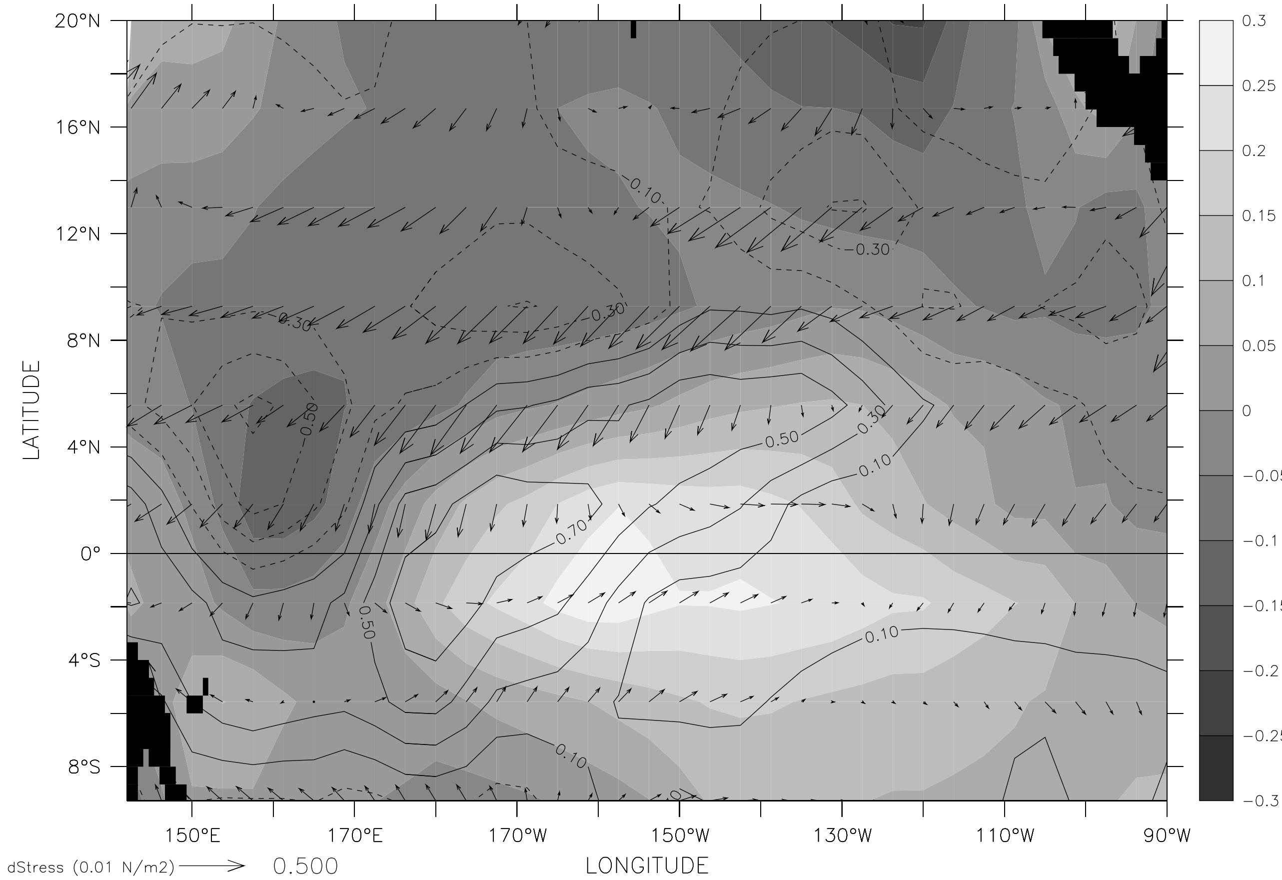

Figure 10. Differences between PLIO and CONT in SST (shades), precipitation (contour interval: 0.2 mm/d), and surface wind stress.

it turns back northward (conserving potential vorticity) and then retroflects eastward to join the North Equatorial Countercurrent and continue to the central Pacific. This surface water brings in additional heat to the western equatorial Pacific (in the hypothesis of Cane and Molnar the warmer western Pacific is caused by reduced inflow of thermocline water from the South Pacific) and thus, with a more open seaway than today, warm water not from south of the equator, but from north of it, creates the warmer SST in the central Pacific. The main climate impact of different distributions of islands in the Indonesian region that we find is a difference in ENSO. As shown in Figures 7 and 8, the average ENSO period is longer for a more open Indonesian seaway, and the amplitude is smaller.

[26] The present results depend on numerical details of the interaction between western boundary currents and topography, and they depend on the dominant ENSO regime. This leads to two obvious questions: How dissipative are the Indonesian islands now and were they during Pliocene times? How much of the sensitivity of ENSO to ITF depends on the dominating ENSO regime? It may never be possible to answer the first question, for even with increased horizontal and vertical resolution answers to fundamental questions about boundary conditions and viscosity remain challenges at the forefront of physical oceanography [e.g., *Pedlosky*, 1996; *Fox-Kemper and Pedlosky*, 2004a, 2004b]. One could attempt to arrive at an upper bound, however, for the possible climate response to changes in the ITF. Both *Hirst and Godfrey* [1993] and *Lee et al.* [2002] compared two global OGCM simulations, one with, and one without ITF. Their results suggest that the response to ITF changes are strongest below the thermocline and may have only little bearing on tropical SST or precipitation. In their calculations, the largest impact of the ITF blocking is seen in the Agulhas Current, the western boundary current of the southern Indian Ocean [*Lutjeharms*, 2006]. Thus, it is possible that the ITF affects global climate not directly through the tropical atmosphere, but indirectly through the Agulhas Current and the midlatitude atmosphere [*Gordon*, 1986]. Testing this hypothesis would require a 1000-year-long integration of a coupled climate model that represents the Agulhas Current realistically. This is not impossible, but with the current computing resources beyond the means of most. *Schneider* [1998] also investigates the effect of blocking the ITF, albeit in a coupled GCM, but integrated for only 10 years after closing the ITF, so that midlatitude and high-latitude responses could not be investigated. For the equatorial Pacific, however, his results should be meaningful. He finds a maximum warming of less than 1°C in the central Pacific, a result consistent with the present results given that PLIO only has a partial reduction of ITF transport. The rainfall response, too, is about three

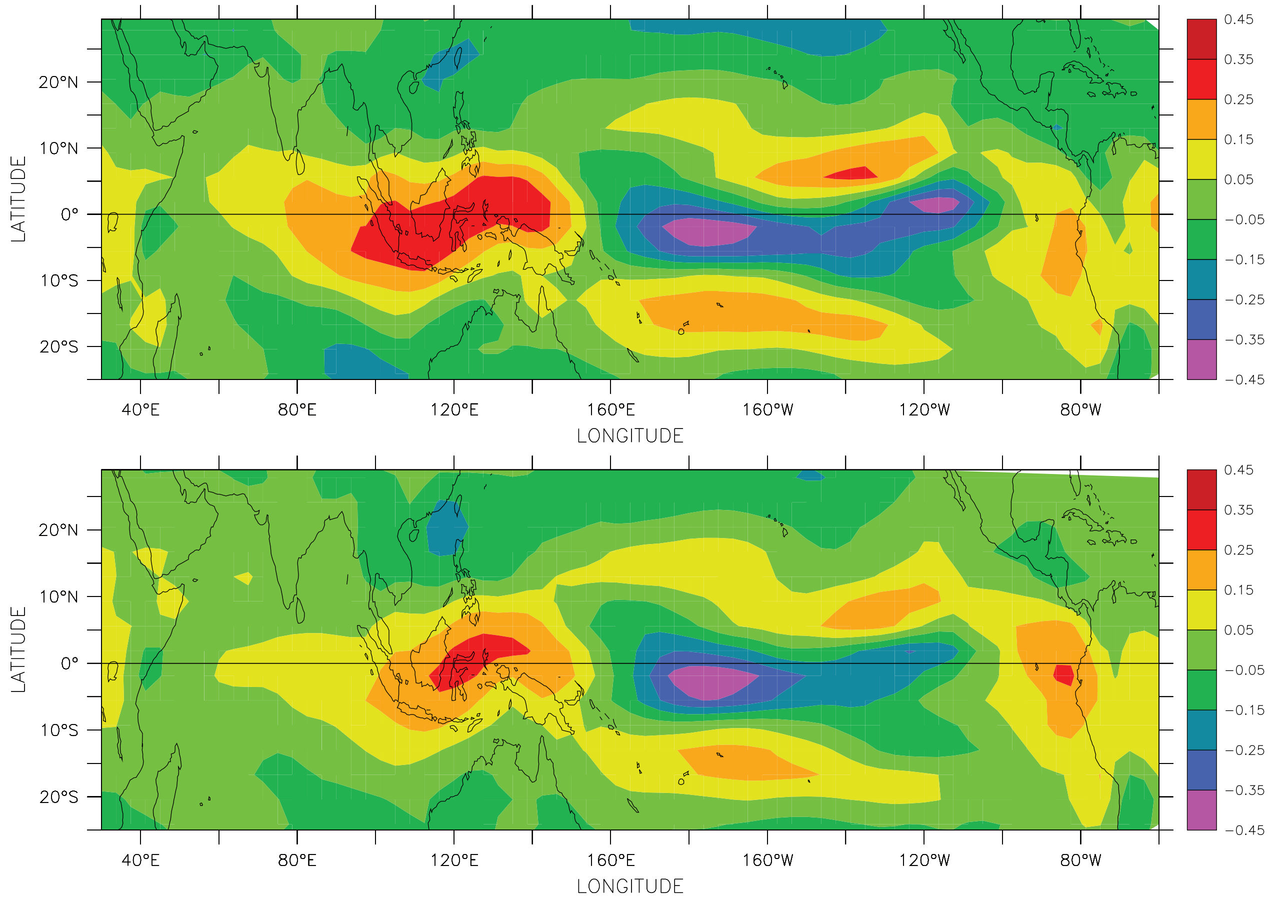

Figure 11. Correlation between NINO3 SST anomalies and zonal wind stress anomalies lagging 3 months for (top) CONT and (bottom) PLIO. The weaker correlation in the eastern equatorial Pacific, a key region for the delayed oscillator [*Neale et al.*, 2008], indicates a weaker coupling between equatorial SST and off-equatorial wind stress.

times as strong as the response in PLIO, with a maximum of 3 mm/d increase at the western Pacific equator. *Song et al.* [2007] also close the ITF in a fully coupled GCM, but integrate for several centuries. Their changes to mean SST and precipitation are, like Schneider's, consistent with ours.

[27] Regarding the ENSO regime, it is fairly straightforward to repeat the current experiment with the new high-resolution version of CCSM whose ENSO is more realistic and already in the series of events regime. However, even for today it is difficult to determine which regime dominates ENSO, and there is no information about ENSO variability during Pliocene times. Thus, a reasonable conclusion for this study is that details of the ITF can influence tropical variability, but they seem unlikely to affect the mean global climate directly. It is still possible that over centuries the ITF will affect climate through a modification of the northward heat transport, but for the time being it may be wise to develop hypotheses that are easier to test and that require less computing power.

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