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

Markus Jochum, Baylor Fox-Kemper, Peter Molnar and Christine Shields

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Corresponding author’s address:
National Center for Atmospheric Research
PO Box 3000
Boulder, CO, 80307
1-303-4971743
markus@ocean.mit.edu
Abstract: 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 (SST) 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. Because of potential vorticity constraints on cross-equatorial flow, the inflow of North Pacific waters and the total Indonesian throughflow transport is reduced. The reduction in transport is matched by an increase in eastward transport of North Pacific waters along the equator, and the resulting shift of warm and fresh water to the central equatorial Pacific leads to an equatorward shift of the Intertropical Convergence Zone (ITCZ) and an eastward enlargement of the western Pacific Warm Pool. These changes reduce the coupling between equatorial SST and off-equatorial wind stress in the eastern Pacific, which reduces the delayed oscillator component of ENSO and enhances the role of stochastic perturbations. The larger warm pool for Pliocene-like conditions, however, is small compared to paleoceanographic data that suggest a negligible zonal SST gradient across the Pacific.
1 Introduction

Because nearly all water that enters the Pacific Ocean from the Southern Ocean passes out of the Pacific Basin between 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 1996; 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 time scales [e.g., Kennet 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.

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 $Sverdrup \ [Sv, \ Godfrey \ 1996; \ Hautila \ et \ al. \ 2001]$. This agreement is reassuring, but the fact that most of the ITF water originates from the North Pacific 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 1993ab].

Thus, although a considerable amount of time and effort has been spent to quantify and understand the ITF, large uncertainties still remain. For that reason, the present work takes a step back and asks what particular aspects of the ITF might be relevant for climate. For example, it is not obvious how important the ITF mass transport is for climate. Even 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 warmpool. Rather than asking what determines strength and composition of the ITF, we ask what aspect of the ITF most affects climate, and specifically precipitation and its variability. There are, of course, many different sensitivity studies one could conceive of, but a reasonable starting point is the hypothesis that sometime during Pliocene times [approximately 3 million years ago, 3 Ma, see Fedorov et al. 2006 for a review] the ITF changed its source waters, which triggered northern hemisphere glaciation [Cane and Molnar 2001]. They 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.
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] provide an extensive discussion of the uncertainties in tropical Pacific paleo-data, and list possible causes for the onset of northern hemisphere glaciation. Based on numerical evidence 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 Haywood et al. [2007] in that it investigates one more possible mechanism by which the tropical oceans could trigger climate change.

The first part of Cane and Molnar’s hypothesis, the ITF - SST gradient connection, has support from 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.

This study describes possible differences in climate due to differences in positions of Indonesian islands and adjacent ocean floor topography. 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

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 Collins et al. [2006].

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 paleo-climate 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 in Yeager et al. [2006] is the new convection scheme, which leads to significant improvements in the simulation of ENSO [Neale et al. 2008].

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 Timour Strait, are replaced by a one grid-point throughput between Borneo and New Guinea (Figure 1, note that because of the staggered grid 2 tracer grid points make one active velocity grid point). 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 climate relevant. 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.

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 upper boundary condition, and the resulting subtropical and tropical circulation is shown in Figure 2. All forced OGCM runs, a control (CONT(F)) and two sensitivity runs (PLIOF, PLIOFvisc), are integrated for 250 years. PLIOFvisc is identical to PLIOF, but 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 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 to 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 time island geometry is highly uncertain. The present modifications are inspired by the work of Rodgers et al. [2000] and Cane and Molnar [2001], which suggests that the ocean is very sensitive to this particular aspect of the island geometry. Thus, our calculations are an 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.

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

3 The Indonesian Throughflow

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 a single no-normal flow boundary condition. However, in a closed basin two boundary conditions are required: no-normal flow through both the eastern and western bound-
aries. 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 Island Rule [Godfrey 1989] takes advantage of this situation by only considering 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.

The island rule has been successfully employed around Australia to determine the ITF transport Godfrey [1989], the flow around 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]. 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 [like in Godfrey 1996].

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, 1993]. 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
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% indicate that additional, secondary processes also effect the flow. These secondary effects are larger in PLIO(F) than in CONT(F).

In principle the ITF transport could be modified by Pacific upwelling [Stommel and Aarons 1961; 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 did not change in the forced runs, and only by 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 CONT, 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, pers. communication, 2008]. The Torres Strait is shallow and Wyrtki [1961] estimates its transport to be less than 1 \( Sv \). The value in CONT is larger than that, but not large enough to jus-
tify 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 the North Pacific water that in CONT(F) is destined
for the Indian Ocean is forced to retroect 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.

Theory relating to the topic of cross-equatorial flow emphasizes that conserva-
tion of potential vorticity severely constrains inviscid flow across the equator
[Killworth 1991; Edwards and Pedlosky 1998]. Applied to the present case this im-
plies that dissipation of potential vorticity is necessary to allow a non-trivial 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 fric-
tion along western boundaries increased by a factor of 10 (PLIOFvisc). The Makas-
sar 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 espe-
cially 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, 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 gridpoints 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 retroect and stay in the Pacific. Note that
the return flow of the ITF is under the same potential vorticity constraint: the
New Guinea Coastal Current does not cross the equator and joins the ITF directly
but retroects 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 et al. 1993]. This is analogous
to the tropical Atlantic where the northward flowing North Brazil Current has to
retroect into the Atlantic Equatorial Undercurrent to adjust its potential vortic-
ity. In the Atlantic, however, some southern hemisphere water is trapped in ed-
dies, which travel northwest 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].

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 reduc-
tion 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 sen-
sitivity study PLIOFvisc. It is noteworthy that, assuming identical wind fields and
the validity of the Island Rule, the more southerly latitude of New Guinea appro-
priate 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,
the Island Rule, previous OGCM studies, and the current simulations all suggest that widening the Indonesian passages near the equator leads to increased non-linearity and 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 \cite{Jochum2008}. With the caveat that the presently modeled differences in the ITF are sensitive to resolution and parameterized dissipation, the next section discusses their climate impact.

4 Climate response

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 (Figure 7). The frequency peak of NINO3 SST (SST anomaly averaged between 150°W-90°W and 5°S-5°N) is shifted from 2.4 years in CONT to 3.3 years in PLIO, and its standard deviation is reduced by more than 10% from 0.74°C to 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.

The western warm pool is buttressed against the Indonesian islands, which are surrounded by some of the world's warmest and freshest surface waters (Figure 8). These waters are advected eastward by the retroflected part of the Mindanao Current as part of the northern Pacific tropical gyre. A stronger retroreflection makes stronger eastward flow in the western Pacific and weaker westward flow in the east
(Figure 8, bottom), which extends the western Warm Pool eastward and reduces the extent of the cold tongue in the eastern equatorial Pacific (Figure 9). 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 9). This fresher and warmer water leads to increased upper ocean stratification in the western Pacific (not shown), and can reduce entrainment of subsurface water [Lukas and Lindstroem 1991; Yeager et al. 2006].

How does this more equatorially centered position of the ITCZ in PLIO affect ENSO? There are numerous dynamical regimes proposed to explain ENSO behaviour [see Wang and Picault 2004 for an overview], but on the most fundamental level the question is whether ENSO is 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 inconsistent with present-day observations [Kessler 2002], but could feasibly have been a natural mode in the past [e.g., Garcia-Herrara et al. 2008].

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 windbursts 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 windbursts 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 windburst activity is almost non-existent 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 10), 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

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 weak-
ening the delayed oscillator regime and thus creating a weaker and less regular
ENSO.

Cane and Molnar [2001] suggested that the northward movement of New
Guinea would have blocked relatively warm water in the Pacific south of the equa-
tor, 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 In-
dian 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. Based on this Cane and Mol-
nar [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.

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]. The explanation for the warmer central equatorial Pacific associated with the more open seaway, however, is very different from what Cane and Molnar imagined. Indeed, 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. Moreover, although its inertia carries this North Pacific water across the equator, it turns back northward (conserving potential vorticity) and then retroreflects eastward to join the North Equatorial Countercurrent and continue to the central Pacific. This surface water is relatively warm, and thus, with a more open seaway than today, warm water not from south of the equator, but from north of it, creates the mildly warmer 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 Figure 7, the average ENSO period is longer for a more open Indonesian Seaway, and the amplitude is smaller.

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 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 2004ab]. However, one could attempt to arrive at an upper bound 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 mid-latitude atmosphere [Gordon 1986]. Testing this hypothesis would require a thousand year long integration of a coupled climate model that represents the Agulhas Current realistically. This is not impossible, but with the current computing resources is 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 mid- 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 times as strong.
as the response in PLIO, with a maximum of 3 \( \text{mm/day} \) 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.

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 requires less computing power.

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Table 1: ITF transports (in $Sv$) from CCSM3 and from Godfrey’s Island Rule.

<table>
<thead>
<tr>
<th>Transports</th>
<th>Island Rule</th>
<th>Model ITF</th>
<th>source: north/south Pacific</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTF</td>
<td>11.8</td>
<td>8.8</td>
<td>7.7/1.1</td>
</tr>
<tr>
<td>PLIOF</td>
<td>13.2</td>
<td>8.0</td>
<td>6.4/1.6</td>
</tr>
<tr>
<td>CONT</td>
<td>17.1</td>
<td>15.1</td>
<td>12.1/3.0</td>
</tr>
<tr>
<td>PLIO</td>
<td>18.1</td>
<td>12.9</td>
<td>8.4/4.5</td>
</tr>
</tbody>
</table>
Figure 1: Sea surface salinity from CONTF, which, apart from highlighting the large horizontal salinity gradients in this area, illustrates the ocean grid, and the real (black lines) and model (white rectangles) land surface. In PLIO(F) the northernmost part of New Guinea has been converted from land into ocean cells (marked 'removed').

Figure 2: Temperature difference at 150 m depth between PLIOF and CONTF (PLIOF - CONTF). The contourlines (interval 5 Sv) illustrate the circulation by showing the barotropic streamfunction for CONTF.

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

Figure 4: Timeseries of ITF transport, solid:CONTF, dashed:PLIOF, dash-dotted: CONT, broken: PLIO.

Figure 5: Barotropic Streamlines for CONTF (left) and PLIOF (right). Contour interval is 1 Sv; real land is shown in black and the ocean model landmask is shown in white. Note the retroreflection of the Mindanao Current as it crosses the equator in PLIOF.

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

Figure 7: Statistical analysis for NINO3 SST anomalies of the years 101-200 for CONT (top) and PLIO (bottom). The boxes on the left show the variance per frequency band, the middle boxes show the autocorrelation, and the boxes on the
right side show the seasonal distribution of the variance. The solid lines across the
power spectra mark the variance for a AR(1) process (noise with memory), and the
parts of the spectra that are above the uppermost dashed lines are significant at
the 99\% level.

Figure 8: Mean surface salinity and velocity in the western tropical Pacific re-
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velocity between PLIO and CONT (bottom). Surface is here defined as upper 50 m.
Note that the velocity changes happen in a region of a large east-west buoyancy
gradient in temperature as well as in salinity.

Figure 9: Differences between PLIO and CONT in SST (shades), precipitation
(contour interval: 0.2 \(mm/day\)), and surface wind stress.

Figure 10: Correlation between NINO3 SST anomalies and zonal wind stress
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