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Di Nezio, PN, Timmerman, A, Tierney, JE et al. (7 more authors) (2016) The climate response of the Indo-Pacific warm pool to glacial sea level. Paleoceanography, 31 (6). pp. 866-894. ISSN 0883-8305

https://doi.org/10.1002/2015PA002890

The climate response of the Indo-Pacific warm pool to glacial sea level

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Submitted October 16, 2015

Revised March 5, 2016

Revised May 9, 2016 Accepted May 5, 2016

School of Ocean and Earth Science and Technology publication number 9633.

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Key Points.

- Shelf exposure is the main mechanism whereby glacial changes in sea level influence warm pool climate
- The climate response is initiated by surface cooling over the Sahul shelf caused by the increased albedo of the exposed land
- Coupled ocean-atmosphere dynamics akin to the Bjerknes feedback amplify the response

Abstract. Growing climate proxy evidence suggests that changes in sea level are important drivers of tropical climate change on glacial–interglacial time-scales. These paleodata suggest that rainfall patterns over the Indo-Pacific Warm Pool (IPWP) are highly sensitive to the landmass configuration of the Maritime Continent, and that lowered sea level contributed to large-scale drying during the Last Glacial Maximum (LGM, ca. 21,000 years before present). Using the Community Earth System Model Version 1.2 (CESM1) we investigate the mechanisms by which lowered sea level influenced the climate of the IPWP during the LGM. The CESM1 simulations show that, in agreement with previous hypotheses, changes in atmospheric circulation are initiated by the exposure of the Sunda and Sahul shelves. Ocean dynamical processes amplify the changes in atmospheric circulation by increasing the east-west sea-surface temperature (SST) gradient along the equatorial Indian Ocean. The coupled mechanism driving this response is akin to the Bjerknes feedback, and results in a large-scale climatic reorganization over the Indian Ocean with impacts extending from east Africa to the western tropical Pacific. Unlike exposure of the Sunda shelf, exposure of Sahul shelf and the associated changes in surface albedo play a key role because the positive feedback. This mechanism could explain the pattern of dry (wet) eastern (western) Indian Ocean identified in climate proxies and LGM simulations. However, this response also requires a strengthened SST gradient along the equatorial Indian Ocean, a pattern that is not evident in marine paleoreconstructions. Strategies to resolve this issue are discussed.

1. Introduction

Proxy data from the Indo-Pacific Warm Pool (IPWP) region suggest that on glacial–interglacial timescales, precipitation responds to the landmass configuration of the Maritime Continent, which is determined by changes in global sea level [*De Deckker et al.*, 2002; *Zhao et al.*, 2006; *Griffiths et al.*, 2009, 2013; *Tierney et al.*, 2012; *DiNezio and Tierney*, 2013]. Specifically, it is hypothesized that lowered sea level contributed to large-scale drying during the Last Glacial Maximum (LGM), the period ca. 21,000 years ago when ice sheets were at their maximum and sea level was 120 meters lower than present day. These studies have argued that this response resulted from changes in deep atmospheric convection over the Sunda Shelf, an area presently located underneath the Gulf of Thailand, the South China Sea, and the Java Sea, which became subaerially exposed as ice sheets grew and sea level dropped (Fig. 1).

The proxy evidence have several important implications. For one, they suggest that glaciation (via changes in sea level) is a major driver of tropical climate change on glacial–interglacial timescales. This contrasts with previously-proposed mechanisms, which have focused primarily on the role of greenhouse gas and orbital forcing on the east-west sea-surface temperature (SST) gradient across the tropical Pacificf [*Clement et al.*, 1996; *Tudhope et al.*, 2001; *Koutavas et al.*, 2002; *Clement et al.*, 2004; *Timmermann et al.*, 2007; *Koutavas and Joanides*, 2012] and therefore challenges our understanding of tropical climate change during the Pleistocene.

The climate proxy data from the LGM suggest that central Indonesia and Northern Australia were drier than present, while equatorial east Africa became wetter near the coast (Fig. 2). This pattern of a dry IPWP and wetter western Indian Ocean (IO) is consistent with large-scale changes of the Walker circulation, with decreased ascending motion (i.e. decreased convection) over the Maritime Continent and increased ascending motion (i.e. increased convection) over the western IO. *DiNezio and Tierney* [2013, hereafter DNT13] proposed that exposure of the Sunda Shelf drove this large-scale response mainly because deep convection was reduced over the exposed land in the Maritime Continent. However, all but one of the climate models participating in the Paleoclimate Modelling Intercomparison Project (PMIP) fail to simulate this proxy-inferred pattern of hydroclimate change and the associated changes in circulation [DNT13]. This result precludes a systematic exploration of the effect of lower sea level on IPWP climate using the PMIP simulations. In addition, the PMIP LGM experiments included all glacial boundary conditions (i.e. changes in greenhouse gases, ice sheets, and the Earth's orbit), making it difficult to isolate and diagnose the response to sea level.

Lowered glacial sea level also exposed the Sahul shelf, the continental shelf extending over from the northern coast of Australia to the island of New Guinea underneath the Gulf of Carpentaria and the Timor Sea. The ocean over the Sunda and Sahul shelves has depths that rarely exceed 50 m and extensive areas that are less than 20 m (Fig. 1, left). Thus the reduction in sea level at the LGM fully exposed those shelves (Fig. 1, right). The areal extension of the shelves is rather limited; however their exposure could have sizable impact on IPWP climate given their central location between areas of deep convection of the IPWP.

Lowered glacial sea level could also have an effect on the Indonesian Throughflow (ITF) and tidal mixing over the Indonesian seas. Climate model simulations show that the eastern IO could become cold and dry (as seen in paleoreconstructions of LGM hydroclimate) if the flow of warm waters from the Pacific is shut down [*Schneider*, 1998; *Kajtar et al.*, 2015]. We do not expect a complete shutdown of the ITF at the LGM because some of its key passages, such as the Ombai and Timor passages, are deeper than 1000 m; however, other passages, such as as Makassar, Lombok, and Karimata straits are much shallower, and their flow could be substantially altered by lower sea level. Tidal mixing is also known to influence tropical climate, mainly by cooling sea-surface temperatures over the shallow seas surrounding Indonesia and the Maritime continent [*Jochum and Potemra*, 2008; *Brierley and Fedorov*, 2011; *Sprintall et al.*, 2014]. Lowered sea level could influence the magnitude and location of the tidal mixing, as the shelf break becomes exposed [*Montenegro et al.*, 2007; *Egbert et al.*, 2004].

Here we isolate and systematically explore the impact of lowered sea level on glacial IPWP climate by performing a series of simulations with the Community Earth System Model Version 1.2 (CESM1). CESM1 has new and improved physical parameterizations relative to its predecessor, the Community Climate System Model (CCSM4), a model which exhibited a muted response to shelf exposure and poor agreement with the climate proxies at the LGM [DNT13]. The differences in atmospheric physics between CESM1 and CCSM4 are extensive and involve every major physics parameterization except for deep convection. We will show that CESM1 simulates the pattern of dry eastern IO and wet western IO during the LGM, allowing us to perform simulations isolating the effect of: shelf exposure, changes in the Indonesian Throughflow (ITF), and tidal mixing, as well as the role of ocean dynamics amplifying the climate responses.

2. Experimental design and setup

2.1. Climate Model

The simulations were run using the Community Earth System Model Version 1.2 (CESM1), the most recent version of the global coupled model developed at the National Center for Atmospheric Research (NCAR). CESM1 was configured to simulate the coupled interactions between the atmosphere, ocean, land, and sea ice with prescribed vegetation, carbon cycle, and marine ecosystems. The atmospheric component is the Community Atmosphere Model Version 5 (CAM5), with extensively upgraded physics packages, including new schemes for the simulation of moist turbulence, a shallow convection, cloud microphysics, and aerosol–cloud–rainfall interactions [*Neale et al.*, 2012]. CAM5 was run on a finite volume (FV) grid at a nominal horizontal resolution of 2° with 30 pressure levels for the vertical coordinate. The land component is the Community Land Model Version 4 (CLM4) configured on the same $2° \times 2°$ horizontal grid as the atmosphere model.

CLM4's new capabilities include a prognostic carbon-nitrogen model, an urban canyon model, a prognostic land cover and land use, a crop model, a revised snow model with aerosol deposition of black carbon and dust, grain-size dependent snow aging, and vertically resolved snowpack heating [*Lawrence et al.*, 2011]. None of these features were active in our simulations with the exception of CLM's ability to pass dust mobilized by wind to the prognostic atmospheric aerosol module. This process is relevant for our study because dust emissions from the exposed Sahul shelf could increase overlying aerosol loading causing rainfall to decrease.

The ocean model is the Parallel Ocean Program Version 2 (POP2) configured at the nominal horizontal resolution of 1°, with increased meridional resolution of about 1/3° on the equatorial wave guide, and 60 vertical levels. POP2 has parameterizations that simulate overflows, tidal mixing, and eddy mixing as described by *Smith et al.* [2010]. The reader is referred to a special collection of *Journal of Climate* (http://journals.ametsoc.org/page/CCSM4/CESM1) for a description of standard climate simulations performed with CESM1.

CESM1 simulates the present-day patterns of rainfall over the Indo-Pacific region realistically. The annual-mean rainfall climatology of the pre-industrial (0ka) simulation exhibits key large-scale features, including the Inter-Tropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), pronounced rainfall over the eastern equatorial IO, the center of the IPWP, and the western tropical Pacific, as well as dry conditions over the western IO and the central Pacific (Fig. 3a,b). Over the tropical oceans (30°N–30°S), the simulated monthly-mean rainfall climatology and the GPCPv2 observational product [*Adler et al.*, 2003], have a pattern correlation coefficient (r) of 0.78. This represents an improvement in the realism of the patterns of the simulated rainfall with respect to the predecessors of CESM1, versions 3.5 and 4 of the Community Climate System Model (CCSM3.5 and CCSM4), which had $r = 0.69$ and $r = 0.72$. The performance over tropical land is comparable to previous versions of the model, with r values of 0.81 for CESM1, 0.79 for CCSM4 and 0.78 for CCSM3.5. Last, the simulated sea-surface salinity (SSS) also shows realistic patterns, with fresh conditions around the Maritime Continent and salty conditions in the Arabian Sea (Fig. 3d) compared with the observations (Fig. 3d).

CESM1 exhibits common deficiencies in the simulation of rainfall and SSS over the Indo-Pacific region. Relative to observations, the annual-mean rainfall climatology shows much wetter conditions at the edges of the IPWP: the western IO, the ITCZ over the western Pacific, and the eastern edge of the SPCZ (Fig. 3a,b). Conversely, CESM1 simulates less rain over the eastern IO and central Pacific compared with observations. Over the Maritime Continent, CESM1 simulates less rain than observed over land (e.g. Borneo) and more rain than observed over the ocean, particularly over the Celebes and Banda seas. Additionally, averaged over the tropics, CESM1 simulates 20% more rainfall. This stronger rainfall results in much fresher SSS around the Maritime Continent (Fig. 3d) compared with observations (Fig. 3c).

2.2. Sea level boundary conditions

We focus on the effect of sea level on the Maritime Continent (MC) region defined as the box (30°S–30°N, 90°E–160°E). Outside this region all boundary conditions remain at pre-industrial (year 1850 AD) values. All other boundary conditions, such as, GHG concentrations, orbital configuration, and continental ice (i.e. ice sheets) are prescribed at pre-industrial values. In other words, we solely focus on the effects of glacial sea level over the Maritime Continent region. The following three subsections describe the implementation of the boundary conditions required to represent the three key mechanisms through which lowered sea level can influence IPWP climate, namely shelf exposure, ITF routing, and tidal mixing:

2.2.1. Shelf exposure

Representing the exposure of the Sunda and Sahul shelves required the following changes in the model setup. First, we defined a new land-sea mask over the Maritime Continent region based on a sea level drop of 120 m with respect to the present day (Fig. 4 top). Estimates of LGM sea level range from 120 m to 135 m below pre-industrial values [*Yokoyama et al.*, 2000; *Hanebuth et al.*, 2000; *Waelbroeck et al.*, 2002], thus our choice of 120 m represents the smallest perturbation within observational uncertainty. We defined the surface properties of the new land grid points as follows. Soil properties were extrapolated using a nearest neighbor algorithm to fill in the new land points. Over the Sunda and Sahul shelves, vegetation was prescribed as an equal mix of tropical deciduous tree and tropical grass (C_4) plant functional types (PFTs) (Fig. 5). This setup is based on evidence that the Sunda Shelf was a savanna (i.e. mainly tropical grass) environment during the LGM [*Bird et al.*, 2005]. We explored the sensitivity to other PFTs such as C_3 grass, bare soil, and tropical forest and we did not find substantial changes, indicating that the results are, to first order, insensitive to differences in vegetation cover. The remaining surface properties, such as albedo or surface roughness, are computed by CLM4.0 based on the soil and plant properties and passed to CAM5 model via the coupler. Run off water from the new land grid points was directed to the nearest ocean grid point.

2.2.2. Routing of the Indonesian Throughflow

Representing the effect of a 120 m sea level drop required the modification of the bathymetry of POP2 over the MC (Fig. 4c,d). We implemented these changes in a two-step process: 1) land masses were added to represent the exposed shelves, 2) the depth of ocean floor was raised by 120 m in those grid boxes where the vertical resolution of POP2 allowed it. These changes have an impact on the Indonesian Throughflow (ITF) since they result in total or partial closure of some of its passages (Fig. 6c,d). We do not expect a complete shutdown of the ITF because some of its passages, such as the Ombai and Timor passages, are deeper than 1000 m. However, other key ITF passages, such as as Makassar, Lombok, and Karimata straits are shallower, and their flow could be altered by LGM sea level. For instance the sill of Lombok straits, presently 220 m deep, would have been less than 100 m deep at the LGM, potentially affecting the regional flow of warm waters into the eastern IO.

Therefore the main changes are: 1) blocked flow between the South China Sea (SCS) and the Indonesian Seas through Karimata straits (modern sill depth of 50 m), 2) no flow through the Java Straits due to Sunda exposure, and 3) a 120 m shoaling of the sill of Makassar and Lombok straits (modern sill depths of 670 m and 220 m respectively). Note however, that the $1° \times 1/3°$ horizontal grid of POP2 only allows for a very crude representation of other key ITF channels, such as the Lombok, Ombai, and Timor passages. In these passages, we changed the sill depth proportionally with respect to the model's present day sill depth. We also raised the sills of Mindoro Strait and Sibutu passages. Raising the sills of these passages leads to a slower thermocline flow, which is compensated by faster surface flow, resulting in small changes in depth-integrated transport.

2.2.3. Tidal mixing

Tidal mixing is caused by the breaking of internal tides in places where the topography is steep, such as the continental shelf break or within narrow straits. Recent estimates of dissipation and vertical diffusivity reveal hotspots of mixing in the Banda Sea, with high diffusivity values of the order 1 to 10 cm^2s^{-1} in the thermocline and at the base of the mixed layer [*Koch-Larrouy et al.*, 2007; *Hatayama*, 2004]. Lowered sea level could influence the magnitude and location of this mixing, as the shelf break becomes exposed, changing the spatial distribution of tidal mixing [*Montenegro et al.*, 2007; *Egbert et al.*, 2004]. Climate models show that when tidal mixing is included, SSTs over the Banda Sea are cooled by about 0.5 °C, reducing overlying deep convection by as much as 20% [*Sprintall et al.*, 2014]. The effect of this mechanism on LGM climate was explored by *Montenegro et al.* [2007] based on the parametrization of tidal mixing of *Jayne and St. Laurent* [2001] and estimates of tidal dissipation by *Egbert et al.* [2004]. We largely follow their approach since POP2 uses the same tidal mixing parametrization. We implemented the tidal dissipation rates used by *Montenegro et al.* [2007], which result in increased mixing over the Banda Sea and decreased upper ocean mixing over the Timor Sea (Fig. 6 bottom).

2.3. Simulations and mechanisms

The simulations in our experiment aim to explore different combinations of the sea level boundary conditions. Table 1 lists all the simulations performed and the changes in their boundary conditions with respect to the control simulations. We ran additional simulations with CAM5, the atmospheric component of CESM1, forced with prescribed climatological SSTs, in order to isolate atmospheric from ocean dynamical processes. These simulations, named 0kaAtm, 21kaAtmShelves, 21kaAtmSunda, 21kaAtmSahul, were run with climatological SSTs from the coupled pre-industrial control (0ka) and the only perturbation is the change of land mask with exposed shelves. In these simulations, the ocean is unable to respond to atmospheric changes, thus allowing us to isolate the *uncoupled* response to shelf exposure. Comparing these simulations with the fully *coupled* simulations allows us to isolate the effect of ocean dynamics on the response. Each of the proposed mechanisms can be isolated by differencing specific simulations from our experiment. Table 2 lists the procedure we followed to compute the climate responses driven by each mechanism. For instance, the response to tidal mixing is computed by differencing the simulation with all LGM sea level boundary conditions (21kaSL) minus a simulation with only shelf exposure and changes in bathymetry included, and tidal dissipation rates set at pre-industrial values (21kaSLnoTM).

3. Results

3.1. Climate response to LGM sea level

The simulated annual-mean climate response to LGM sea level exhibits large scale patterns extending beyond the MC influencing the entire IO, coastal equatorial east Africa, and northern Australia (Fig. 7). The simulated changes in rainfall are characterized by a dipole of wetter conditions over the western IO, and drier conditions over the eastern IO (Fig. 7a). The rainfall changes extend away from the ocean influencing continental areas, such as easternmost equatorial Africa, which becomes wetter, as well as the MC and northern Australia, which become drier. The climate response to LGM sea level is dominated by the changes over the IO, however the Pacific Ocean also exhibits zonally asymmetric changes, with wetter conditions over the western Pacific and SPCZ region, and drier conditions in the central Pacific ITCZ region.

The main features of the annual-mean changes in SST are: 1) warmer conditions over the western IO and 2) colder conditions over the eastern IO, particularly off the coast of Sumatra (Fig. 7c). These changes represent a strengthened zonal gradient over the IO, with an east-west contrast of about 1 K. The changes in surface winds exhibit anomalous easterly wind stress along the equatorial IO consistent with the strenghtened SST gradient. The rainfall response is also consistent with these changes. Rainfall increases over the western IO where surface winds converge over the warmer SSTs, causing ascending motion and increased convection. Conversely, rainfall decreases over the eastern IO where surface winds diverge, causing anomalous descending motion and drying, over the colder SSTs.

The changes in sea-surface salinity (SSS) broadly follow the changes in rainfall, with fresher conditions over the western IO and Arabian Sea, and saltier conditions over the eastern IO (Fig. 7b). The western IO freshening is relatively muted given the co-located increase in rainfall, and is much smaller in magnitude than the large increase in salinity further east. In addition, the patterns of SSS change show evidence of changes in the routing of the ITF. For instance the model simulates a pronounced freshening of the SCS (in excess of 1 psu). This SSS change takes place because the lower sea level closes the Karimata Straits, shutting down the export of fresh water out of the SCS southward into the Java Sea.

CESM1 simulates ocean dynamical changes in the IO that are consistent with the equatorial adjustment to an easterly wind anomaly. The annual-mean changes in thermocline depth show a stronger east–west tilt, with a shallower thermocline in the east and a deeper thermocline in the west (Fig. 8a). Simulated ocean currents show increased westward surface velocity (Fig. 8b) and increased equatorial upwelling (Fig. 8c). A shallow thermocline in the eastern IO would make climatological upwelling more effective at cooling the surface. Stronger equatorial upwelling would also cool the eastern IO, whereas the stronger westward currents would act to cool the eastern IO and warm the western IO. All these processes act to reinforce the anomalous zonal SST gradient, leading to stronger easterly winds and a stronger ocean response and the SST gradient. Together these SST, wind, and ocean changes suggest that the Bjerknes feedback [*Bjerknes*, 1969] could be playing a key role in our simulations. We explore this in more detail in Section 3.2.2 where we analyze a set of simulations performed to isolate the effect of coupled ocean–atmosphere interactions.

3.2. Mechanisms

In this subsection we explore the role played by different mechanisms in the climate response of the IPWP to LGM sea level. First we isolate the effect of the different boundary conditions associated with the change in sea level, namely: shelf exposure, closure of ITF passages, and tidal mixing. Second, we focus on the role played by ocean dynamics.

3.2.1. Role of boundary conditions

First we focus on the response to exposure of each shelf in isolation. The response to Sunda Shelf exposure shows drying over the MC and wetter conditions over the western equatorial IO (Fig. 9a). The Sunda shelf exposure does not, however, explain the full sea level response mainly because it fails to simulate the dipole of wet/dry warm/cold western/eastern IO (compare Fig. 9a vs. Fig. 7a). Exposure of the Sahul Shelf, in contrast, shows this larger-scale dipole of rainfall and SST change (Fig. 9c–d), similar to the full sea level response. What is the cause of this difference? Both responses exhibit anomalous easterlies along the equatorial IO (Figs. 9b and 9d, vectors), however, only the response to Sahul Shelf exhibits a strengthened zonal SST gradient there. The absence of an altered SST gradient indicates that the Bjerknes feedback is not activated in our Sunda simulation.

The partial closure of ITF passages and changes in tidal mixing drive weaker and more localized changes than the response to shelf exposure. Changes in the ITF cause only small changes in rainfall and SST (Figs. 9e and 9f) suggesting that a 120 m drop in sea level does not alter the ITF sufficiently to have an impact on IPWP climate. CESM1 simulates volume, heat, and freshwater transports that approximately agree with observational estimates. The volume transport is slightly underestimated by CESM1, whereas the heat and freshwater transports are overestimated (Table 3). This disagreement is not unexpected given the coarse resolution of the ocean model and the biases in the wind fields common to coupled climate models.

Sea level causes a reduction in volume transport of about 1.5 Sv, mainly due to closure of Karimata straits. The transport through Makassar straits, the main pathway of the ITF, decreases from 7.7 Sv in the PI control to 7.4 Sv. The surface flow increases (Figs. 6c), possibly due to the increased inter-basin pressure gradient (Fig. 10d). Conversely, the transport decreases in the thermocline, because of the reduction in sill depth of Makassar straits. This compensation explains the rather muted ITF response similarly to the conclusion from previous studies [*Kuhnt et al.*, 2013]. The weak ITF changes could explain the negligible SST changes over the eastern Indian Ocean, downstream from the ITF.

Closure of ITF passages does, in contrast, have a massive impact on SSS, causing a pronounced freshening of the SCS (Fig. 10c and 10d). The large-scale patterns of SSS changes over the IO, however, are largely driven by the changes in rainfall in response to shelf exposure (Fig. 10a). The changes in ocean circulation do appear to play a role in the equatorial IO, where the anomalous westward currents increase the advection of freshwater from the eastern IO to the western IO, thus enhancing the dipole of fresher (saltier) western (eastern) IO. The tidal mixing response suggests a localized impact from enhanced mixing over the Banda Sea driving colder SSTs (Fig. 9h) and associated drying (Fig. 9g). Conversely, a reduction in tidal mixing over Australia's North West Shelf causes warmer SSTs and increased rainfall. The magnitude and spatial scale of these responses does not suggest an active role for the Bjerknes feedback.

3.2.2. Role of ocean dynamics

So far we have shown that shelf exposure is the main mechanism whereby lower glacial sea level influences IPWP hydroclimate. The climate response suggests an active Bjerknes feedback, including a strengthened SST gradient, anomalous easterlies along the equatorial IO (Fig. 7), as well as changes in thermocline depth, ocean currents, and upwelling (Fig. 8). Here we explore this hypothesis more rigorously using a set of simulations performed with the objective of isolating the influence of ocean dynamics on the coupled response. We performed a coupled simulation with both shelves exposed, but no changes in ITF or tidal mixing (21kaShelves). We then performed a similar simulation, but with an inactive ocean (21kaShelvesAtm). In this simulation CAM5 was run with prescribed climatological SSTs from the pre-industrial coupled simulation (0ka) and the land sea mask with exposed Sunda and Sahul shelves.

The atmosphere-only simulation shows drying over the MC (Fig. 11c), but without the full large-scale pattern seen in the coupled response. This suggests that shelf exposure alone cannot drive a large-scale pattern similar to the coupled response. Moreover, the 21kaShelvesAtm simulation shows anomalous easterlies along the equatorial IO (Fig. 11d, vectors), which would drive anomalous westward currents, stronger upwelling, and a stronger east-west thermocline tilt if the ocean was active. This ocean response would drive a strengthened SST gradient amplifying the initial easterly wind change, i.e. an active Bjerknes feedback. This effect is more clearly seen in the difference of 21ka minus 21kaShelvesAtm, which shows that inclusion of air-sea coupling enhances the drying over the east and causes the wetter conditions over the west (Fig. 11e).

Last, we explore the interaction between the shelves in the coupled responses. We added the rainfall and SST changes in response to exposure of each individual shelf (21kaSunda + 21kaSahul) in order to estimate the linear response. Both the changes in rainfall (Fig. 11g) and SST (Fig. 11h) are weaker than the changes to exposure of the combined shelves (Fig. 11, top). This suggests a constructive effect between the two shelves in the full response. Thus, even though we found that the Sahul Shelf plays a prominent role in the response (section 3.2.1), this effect is enhanced by the exposure of the Sunda Shelf. The uncoupled wind response to Sahul exposure is strongest over the SE IO, off the coast of Sumatra and Java (Fig. 12c, vectors). In contrast, the response to Sunda exposure is strongest over the NE IO (Fig. 12b, vectors). This inter hemispheric asymmetry could explain why exposure of the Sahul shelf activates the Bjerknes feedback, whereas Sunda exposure does not.

3.3. Seasonality of the response to shelf exposure

The simulated climate response to exposure of the Sahul shelf exhibits marked seasonal features. The changes in rainfall show drying over the Banda Sea and the Sahul shelf during the March–April–May (MAM) season (Fig. 13a). This season is also characterized by wetter conditions over the equatorial Indian and Pacific oceans. By the June–July–August (JJA) season, the region of reduced rainfall shifts westward and equatorward towards the eastern equatorial IO (Fig. 13c). The climate response during September–October–November (SON) exhibits the dipole of drier (wetter) eastern (western) IO (Fig. 13e) similar to the annual-mean response (e.g. Fig. 7a). In addition to this dipole, wetter conditions are simulated over the Banda and Timor seas. This pattern weakens in magnitude by December–January–February (DJF) (Fig. 13g) leading to the completion of the seasonal cycle. The east–west SST gradient along the Indian Ocean emerges during JJA, peaks in SON, and decays by DJF (Figs. 13d, 13f, and 13h respectively). The seasonal growth and decay of this gradient is directly related to the seasonality of the dipole of dry (wet) eastern (western) IO (Fig. 13, left). The seasonality of the SST changes is consistent with the critical role played by ocean dynamics discussed in the previous subsection.

The seasonal changes in surface wind stress show additional evidence for an active Bjerknes feedback. The MAM season appears to be particularly important for the initiation of the coupled response because it exhibits anomalous easterlies off the coast of Sumatra (Fig. 13a, vectors). These anomalous surface easterlies appear to be driven by the divergent circulation associated with the drying over the Banda Sea and the Sahul shelf (Fig. 13a). This initial easterly wind anomaly generates an ocean response characterized by stronger upwelling, strong westward currents and a more tilted equatorial thermocline (similar to the changes shown in Fig. 8), leading to a stronger east–west SST gradient in the following JJA and SON seasons (Fig. 13d and Fig. 13h).

Analysis of the seasonal changes in the uncoupled response reveals further details on the mechanisms initiating the coupled response. Our atmosphere–only Sahul exposure simulation (listed as 21kaSahulAtm in Table 1) exhibits reduced rainfall over the Sahul shelf and Banda Sea during MAM (Fig. 14a). This pattern is quite similar to that of the coupled response indicating that ocean–atmosphere coupling does not influence the response during this season. Moreover, during MAM the Sahul shelf is colder (Fig. 14b) suggesting that this could play a role in the drying, and therefore on the initiation of the response. During JJA, the anomalous drying shifts westward and equatorward relative to the shelves (Fig. 14c) as in the coupled response (Fig. 13c). Our model simulates anomalous easterly winds over the eastern IO during this season (Fig. 14d, vectors) consistent with a divergent circulation driven by the drying over the Sahul shelf. This response weakens in the subsequent seasons in contrast to the coupled simulations where the easterly wind changes strengthen and peak in SON. This confirms that ocean dynamics are required for the full response to develop and persist throughout the SON and DJF seasons.

Both the coupled and the uncoupled simulations show pronounced seasonal changes in temperature over the exposed Sahul shelf, which cools down by more than 2 K during DJF and MAM and warm in excess of 2 K during SON (Figs. 13 and 14, right). Reduced thermal heat capacity of land (relative to the ocean mixed layer) could explain this amplified seasonal cycle of land surface temperature. The role of these seasonal changes in shelf surface temperature will explored in the next subsection.

3.4. Physics of the response to shelf exposure

We have extensively explored the *dynamics* of the climate response to shelf exposure. One outstanding question regarding its *physics* remains; namely, how does the exposure of the shelf drive the initial atmospheric response. The initial drying over the shelves could be explained by a reduction in relative humidity over the shelf due to the lower evaporative capacity of land vs. ocean (Fig. 15a). Conversely, it could be caused by the surface cooling seen in both the coupled and uncoupled simulations during MAM and JJA (e.g. Fig. 13b and Fig. 14b). Are both processes essential?

We separated the effect of relative humidity and surface temperature by performing an additional simulation where the Sahul shelf is set to wetland (listed as 21kaSahulWet in Table 1). Wetlands have about the same evaporative capacity than the ocean. This results in negligible changes in surface relative humidity (Fig. 15c). In addition, we set the albedo of the wetland (Fig. 15d) equal to that of the simulations with dry vegetated land (Fig. 15b) so that its cooling effect over the exposed land is the same.

The seasonal evolution of rainfall, SST, and surface winds in the 21kaSahulWet simulation (Fig. 16) is nearly identical to the simulation analyzed in the previous section, in which the shelf is covered with dry vegetated land (simulation 21kaSahul, Fig. 13). Therefore we conclude that the change in surface relative humidity does not play a role initiating the response. Instead, the 21kaSahulWet simulation exhibits cooling over the shelves throughout all four seasons (Fig. 16, left). This indicates that the cooling of the shelves is the main the driver of the initial atmospheric response. Wetlands have a thermal capacity similar to that of the ocean mixed layer (not shown), thus explaining the lack of seasonal swings in surface temperature as seen in the standard Sahul exposure simulation (Fig. 13, right). Therefore these seasonal fluctuations in shelf temperature do not appear to play a critical role on the seasonality of the coupled response. The one exception is the warming of the shelf during SON (Fig. 13f), which could explain the positive rainfall anomalies over the shelf during that season (Fig. 13e–f vs. Fig. 16e–f).

For timescales involved in the response to shelf exposure, tropical deep convection is expected to be controlled by the distribution of subcloud layer entropy [*Emanuel et al.*, 1994]. We diagnosed the changes in moist entropy in terms of the equivalent potential temperature, θ_e, using the formula of *Bolton* [1980]. We estimated the subcloud-layer entropy as the θ_e values on a terrain-following model level about 20 hPa above the surface as in *Boos and Kuang* [2010]. We focused on our set of uncoupled simulations because they allow us to study the initiation mechanism in isolation. We show results for the MAM season because this is the season when the changes in atmospheric circulation appear to initiate the coupled response. Similar conclusions could be obtained if we focused on the following JJA season.

The changes in low-level entropy are virtually identical between the simulation with and without relative humidity changes (Fig. 17, left). Furthermore, the changes in entropy and rainfall have strikingly similar spatial pattern (Fig. 17, right), consistent with the link between these two quantities under convective quasi-equilibrium [*Emanuel et al.*, 1994]. The entropy diagnostics suggests that during MAM rainfall is reduced over the Sahul shelf due the reduction in entropy associated with the surface cooling of the Sahul shelf.

Last, we performed two additional simulations to explore the effect of different vegetation types and associated albedo on the rainfall response. One of them has bare soil covering the Sahul shelf, to explore the effect of high albedo, and the other has a 100% C_4 grass coverage, to consider the possibility that the Sahul shelf was fully covered by savanna. The "bare soil" simulation shows stronger cooling and drying over the Sahul shelf during MAM, confirming that changes in shelf albedo are key for the response (Fig. 18). The changes in the full savanna case are similar to our standard case (Fig. 18 vs. Fig. 13), suggesting that these differences in coverage have a minor influence in albedo and therefore in the response. Both simulations show SON changes in large-scale circulation and rainfall (Fig. 18) similar to those in the standard case (Fig. 13,e–f) suggesting that differences in the coverage type are not fundamental for exciting the coupled response.

4. Discussion

We have performed a series a simulations with the objective of isolating the mechanisms whereby lowered glacial sea level influenced tropical climate during the LGM. Our simulations show the main driver of changes in IPWP hydroclimate consists of a reduction in atmospheric convention initiated by surface cooling of the exposed Sahul which is then amplified by air-sea interactions over the Indian Ocean. As hypothesized by DNT13, this mechanism explains the pattern of dry eastern IO and wet western IO identified in paleoproxy data and PMIP simulations of LGM hydroclimate. The effect of lowered sea level on the ITF has an impact on SSS, particularly over the SCS and the western Pacific, which become fresher mainly due to reduced freshwater export caused by closure of Karimata Strait (Fig. 10). Changes in tidal mixing have a more localized effect restricted to the Banda Sea. Critically, our simulations show that, to first-order, the climate response to LGM sea level is initiated by shelf exposure and amplified by coupled ocean–atmosphere processes in the Indian Ocean.

Our uncoupled simulations show that shelf exposure drives an initial easterly wind anomaly over the eastern IO along the coast of Sumatra. In these simulations the ocean is not interactive, therefore there are no SST changes that could amplify this change and drive large-scale changes over the IPWP. As a result the response, which already bears some resemblance to the fully coupled response, remains localized over NW Australia and the Banda Sea. These wind changes would drive the following ocean dynamical changes: shoaling of the thermocline in the eastern IO and deepening it in the western IO respectively, strengthening of equatorial and coastal upwelling, and anomalous westward equatorial currents. These changes in the ocean, particularly the shoaling of the thermocline, would act to cool the eastern equatorial Indian Ocean, while the deepening and westward zonal currents would act to warm the western IO. The resulting SST gradient along the equator (colder east, warmer west) would drive an additional easterly wind anomaly, effectively amplifying the initial *uncoupled* wind response.

Contrasting the coupled simulations with the uncoupled ones, we find evidence for this positive feedback loop. Shelf exposure leads to a strengthened SST gradient along the equatorial IO in all of these simulations (Figs. 7c, 11b, 9d), as well as consistent changes in thermocline depth, upwelling, and ocean currents (Fig 8). These simulations exhibit changes in surface winds that are much stronger than in the uncoupled simulation (21kaShelves vs. 21kaShelvesAtm, Fig. 11b vs. Fig. 12a), indicating that the SST gradient contributes to the final wind response, and more importantly that coupled processes amplify the initial effect of shelf exposure. Last, this ocean response is critical because it leads to remote SST changes which drive large-scale circulation and rainfall changes reaching as far as eastern equatorial Africa.

This positive feedback loop was first proposed by *Bjerknes* [1969] to explain the growth of El Niño events. Modeling and theoretical studies have argued that the climatology of the Pacific is established by the same mechanism [*Dijkstra and Neelin*, 1995]. Several studies have speculated that it also plays a role amplifying long–term climate changes in the *Pacific* ocean [e.g. *Clement et al.*, 1996; *Koutavas et al.*, 2002], while others have questioned it [e.g. *DiNezio et al.*, 2009, 2011]. However, none of these studies have presented a rigorous proof of an active Bjerknes feedback amplifying externally–forced climate changes. In contrast, our results and a previous model diagnostics study by [*Xie et al.*, 2013] provide more firm mechanistic evidence that the Bjerknes feedback could play a key role in the response to of the Indian Ocean to external forcings.

It is unclear what causes this different behavior between the Indian and the Pacific oceans. Unlike the modern Pacific, where the equatorial thermocline exhibits a strong east–west tilt, the depth of the thermocline is rather uniform along the modern equatorial IO. Therefore anomalous easterly winds blowing over eastern IO, such as those driven by shelf exposure, would make the thermocline shoal there. Moreover, the depth of the thermocline is about 100 m, which is sufficiently shallow for the wind–driven shoaling to influence the surface. In contrast, the wind changes over the Pacific occur in the western and central part of the basin, where the thermocline is deep and less effective at influencing SSTs.

The Bjerknes feedback appears to be more effectively excited by the exposure of the Sahul and NW Australian shelves than of the exposure of the Sunda shelf. SSTs are more sensitive to changes in thermocline depth where climatological upwelling is strongest [*Li et al.*, 2003]. Therefore the location and seasonality of climatological upwelling is key to understanding differences between the response to each shelf. The region off the coast of Java and southern Sumatra exhibits strong upwelling during boreal spring and summer [*Susanto et al.*, 2001; *Potemra and Lukas*, 1999]; namely, the season when the coupled response develops. CESM1 simulates stronger coastal upwelling during June-July-August (JJA) in agreement with ocean reanalysis data (Fig. 19, top), suggesting that this seasonal upwelling could play a role in the response to Sahul exposure.

Our uncoupled simulations show that exposure of the Sahul shelf drives anomalous easterly winds over the SE IO during MAM and JJA (Fig. 14). These winds would act to shoal the thermocline in the eastern IO during the season when coastal upwelling is strongest, thus enhancing the cooling effect on SSTs. In contrast, the uncoupled wind response to exposure of the Sunda shelf is located over the NE IO (Fig. 12b), a region where coastal upwelling is not as strong during these seasons. This suggests that the distinct North–South asymmetry in the strength of coastal upwelling could explain the different effect of Sahul vs. Sunda exposure. Seasonality of the wind response could also be important. The uncoupled wind response to Sahul exposure emerges during MAM and becomes strongest during JJA (Fig. 14), the season when climatological upwelling is strongest, whereas the wind response to Sunda is strongest in DJF (not shown), when climatological upwelling is weaker.

In summary, our simulations show that: 1) the Bjerknes feedback is the key mechanism amplifying the climate response, and 2) the initiating mechanism is caused by land–atmosphere interactions over the shelves. The increased albedo of the exposed land compared to open ocean is the critical initiating physical process because it cools the surface of the shelves leading to reduced atmospheric convection and reduced rainfall. This response is activated under a wide range of vegetation types, suggesting that, in the context of CESM1, differences in vegetation types are not crucial in activating the coupled response. In other words, the change from a darker ocean surface to a relatively brighter vegetated land surface is sufficient to generate the cooling and associated convective response required to activate the coupled response. The reduction in rainfall drives anomalous descending motion over the shelf resulting in anomalous easterlies to the east. This anomalous circulation is strikingly similar to the response of the model of *Gill* [1980] to equatorially *asymmetric* forcing. In this idealized case, Gill's model simulates an anticyclone (vectors) to the southeast of the negative diabatic heating anomaly (negative rainfall change), just as seen in Fig. 12c.

Our results are not strictly comparable with a previous study looking at the effect of sea level on tropical climate [*Bush and Fairbanks*, 2003] because that study focused on the effect of the Sunda shelf exposure alone. However, *Bush and Fairbanks* [2003] found stronger convection over the shelf, whereas all of our simulations show weaker convection, either over the Sunda or Sahul shelves (Fig 12). In contrast to our simulations, their model simulated warmer shelf surface, explaining the increased convection over the Sunda shelf [*A. Bush, pers. comm.*]. Together these results highlight the importance of the changes in surface temperature over the exposed land. The agreement of the CESM1 simulations with the proxy–derived patterns of hydroclimate changes suggests that cooling of the shelves could be a more plausible initiating mechanism.

Our simulations also show that the climate response to exposure of the Sahul Shelf has a distinct seasonal character. The response is initiated in MAM and JJA by land–atmosphere processes and the full response develops during SON amplified by coupled ocean–atmosphere processes. Our experiments do not allow us to rigorously pin down the main cause of the seasonality, but we can speculate about the critical mechanisms. The uncoupled simulations suggest that the seasonality is not caused by coupled dynamics and instead is caused by shelf exposure, since the easterly winds first occur during MAM and JJA (Fig. 14b and 14d). Moreover, our "wetland" simulation indicates that this seasonality is not caused by the lower thermal inertia of land relative to ocean. When set to "wetland", the shelves cool down year round, yet the response has the same seasonal evolution, including the drying during MAM and JJA. Instead, we propose that the seasonality of the uncoupled response is caused by the seasonal migration of the Inter Tropical Convergence Zone (ITCZ). The anomalous drying occurs during the dry season, when the ITCZ is in the northern hemisphere.

The annual-mean changes capture virtually all the spatial features of the seasonal changes. Therefore we expect that even seasonally biased proxies would capture a response similar to the model's annual-mean changes. The one exception is N. Australia, where CESM1 simulates drier conditions during MAM and wetter conditions during SON. As a result the annual-mean response does not show changes in rainfall. This is a region where one proxy record shows wetter conditions for the LGM, while nearby proxies show drier conditions (Fig. 3a). It is possible that this particular record is capturing the seasonally wetter conditions seen in our simulations.

According to our simulations, the change in the east–west SST gradient is about 1 K, with the west warming about 0.3 degree and the east cooling 0.6 degree. Paleoclimate evidence for an altered SST gradient would provide further support for this mechanism. A multi-proxy reconstruction of SST at the LGM, however, does not show evidence for changes in the zonal gradient [*Waelbroeck et al.*, 2009]. Over the IO, this reconstruction relies mainly on foraminiferal assemblages, which could be prone to uncertainties, in particular the lack of modern analogues [*Mix et al.*, 1999] and possible depth dependencies. However, newer estimates based on Mg/Ca paleothermometry indicate a 3K decrease in SST off the coast of Sumatra at the LGM [*Mohtadi et al.*, 2010a, 2014]. In contrast, alkenone data from the western Indian Ocean suggest a more modest cooling of 1–2K

[*Sonzogni et al.*, 1998; *Dahl and Oppo*, 2006]. This could be indicative of a stronger gradient; however, it should be noted that these changes are within error of most proxy measurements and are subject to proxy-specific uncertainties, such as seasonal biases [*Timmermann et al.*, 2014]; and therefore require further investigation to confirm.

In addition, *De Deckker and Gingele* [2002] observe that a giant species of diatom, *Ethmodiscus rex*, is abundant during the last glacial period in the southeast Indian Ocean. *E. rex* requires a high nutrient supply [*Villareal et al.*, 1999], and so its presence may indicate more seasonal upwelling, although it can likewise be explained by a relative absence of the monsoonally-induced low-salinity 'cap' and subsequently more entrainment of deep-water nutrients [*De Deckker and Gingele*, 2002]. Last, we note that low $\delta^{18}O_{sw}$ values over the SCS and western tropical Pacific [*Lea et al.*, 2000; *Xu et al.*, 2010] suggest much fresher conditions during the LGM , in agreement with a reduction in freshwater export by the ITF as simulated by CESM1.

Our simulations show that the active Bjerknes feedback is associated with a shoaling of the thermocline in the eastern IO (Fig. 8a). However, paleoproxy evidence suggests a *warmer*, and therefore deeper, thermocline during the LGM [*Mohtadi et al.*, 2010b]. A deeper thermocline in the east would lead to warmer SSTs and increased rainfall there, a climate response that is at odds with the drying inferred from proxies [DNT13]. An alternate explanation may be a shoaling of *N. dutertrei* habitat, perhaps due to a generally shallower thermocline. Reconciling these conflicting lines of evidence is critical to achieve a dynamically consistent picture of glacial-interglacial climate changes in the IO.

Given the limitations of the paleoclimate record and the qualitative nature of our proxy-model evaluation, we cannot determine whether the magnitude of the simulated hydroclimate changes are realistic. The simulated SSS changes over the Arabian Sea could provide some guidance on the magnitude of the rainfall response. Our simulation shows changes of about -0.2 psu there, in contrast to the proxy data, which suggest a much larger freshening based on inferred $\delta^{18}O_{sw}$ [*Dahl and Oppo*, 2006]. Even if the $\delta^{18}O_{sw}$/SSS slope changed during the LGM, the observed $\delta^{18}O_{sw}$ changes of about -0.5 ‰ would lead to a SSS reduction between 0.5 and 1 psu. Thus CESM's hydroclimate response, while correct in sign, may be a lower bound of the LGM response. A rigorous quantitative estimate of the climate response to shelf exposure will not be possible until we have reliable estimates of the zonal SST gradient, subsurface temperature, or conversely a simulation including oxygen isotopes for direct comparison with observed $\delta^{18}O$.

In DNT13 we hypothesized that exposure of the *Sunda* Shelf was the key driver of the hydroclimate changes simulated by HadCM3 in response to full LGM forcings. However, this type of simulation did not allow us to isolate the specific driver. Our new CESM1 simulations show that the exposure of the *Sahul* Shelf and the corresponding albedo changes are the key initiating process, and that exposure of the Sunda Shelf plays a secondary role. We hypothesize that this difference in the responses depends on whether the Bjerknes feedback is activated or not. Both shelves drive uncoupled responses with anomalous easterlies over the equatorial IO and they are not much different in magnitude (Fig. 12). The response to Sahul exposure, however, has an easterly wind anomaly flowing along the coast of Sumatra (Fig. 12c). This location is particularly important because it is a region of strong ocean–atmosphere coupling [*Susanto et al.*, 2001; *Li et al.*, 2003]. For instance, this strong coupling gives rise to the Indian Ocean Dipole [*Webster et al.*, 1999; *Saji et al.*, 1999], a mode of climate variability that strongly resembles the climate response to shelf exposure. The uncoupled response to Sunda exposure, in contrast, does not exhibit along-shore winds over this region (Fig. 12b). This difference in the location of the shelves could be crucial to excite the Bjerknes feedback.

Our conclusions appear to be specific to the HadCM3 and CESM1 simulations. We contend that these models' responses are indicative of a robust mechanism because of their agreement with the proxies and because the mechanism involves rather simple and well understood physical and dynamical processes. However, it remains unclear why other models are unable to simulate these mechanisms. Our study, however, presents a framework that could be used to further explore this issue. It is possible that unlike the other models, HadCM3 and CESM1 simulate an (uncoupled) atmospheric response that is strong enough to excite the Bjerknes feedback. Conversely, CESM1 and HadCM3 could have a stronger Bjerknes feedback in the IO, or with the right seasonality so that it is effectively excited by the atmospheric response to shelf exposure. If this is the case, then simulating a realistic climatology and variability of the Indian Ocean maybe critical to simulating the correct response to LGM sea level.

5. Conclusion

Our simulations show that shelf exposure and corresponding surface albedo changes are the main mechanism whereby lowered glacial sea level influences IPWP hydroclimate. The climate response is initiated by changes driven by land–atmosphere interactions over the Sahul shelf. This response is then amplified by ocean–atmosphere interactions over the Indian Ocean leading to the full coupled response. We isolated the following physical and dynamical processes which are essential for the response. Lowered glacial sea level exposes the Sahul and NW Australian shelves, which become colder because of the higher albedo of land relative to seawater. The surface cooling increases the static stability of the overlying atmosphere causing atmospheric convection and rainfall to weaken over the shelves. The reduction in atmospheric convection drives a divergent circulation with anomalous easterly winds blowing over the eastern IO, particularly off the coast of Sumatra.

The ocean dynamical adjustment to these wind changes is a critical element of the full climate response. The anomalous easterlies shoal the thermocline in the eastern IO, strengthen equatorial upwelling, and drive anomalous westward equatorial currents. These changes make the eastern IO cooler and the western IO warmer. The strengthened east–west SST gradient further amplifies the initial easterly wind anomaly, resulting in a positive feedback loop between the ocean and the atmosphere akin to the Bjerknes feedback. The higher SSTs over the western IO are key for the simulation of wetter conditions there. Conversely, lower SSTs over the eastern equatorial IO enhance the initial drying caused by shelf exposure. The essential processes of this "Sahul – Indian Ocean Bjerknes" mechanism are summarized in a schematic diagram (Fig. 20).

Exposure of Sahul Shelf plays a key role because it drives anomalous winds off the coast of Sumatra, a location where the ocean is more sensitive to wind changes and therefore is more effective at exciting the positive Bjerknes feedback. Exposure of the Sunda shelf does not excite this positive feedback leading to a much weaker and localized response.

North–south asymmetries in the climate of the IO could be the cause of this differential response mainly because climatological upwelling is stronger in the southeastern IO than in the northeastern IO. We did not find an equally prominent role for other processes, such as changes in the ITF or tidal mixing, other than the surface freshening of the SCS and western Pacific due to changes in the routing of the ITF.

We did not test the effect of glacial greenhouse gases, orbital, or ice sheet boundary conditions. However, it appears that the response to sea level could explain the hydroclimate proxies over the IO and N. Australia. Other glacial boundary conditions may need to be invoked to explain the drying over SE Asian and India seen in the proxies. Exploration of the interplay of the different glacial BCs with the sea level response is left for future work. However, our simulations provide a picture of IPWP climate during the LGM which could be further tested using other proxies.

Specifically, our mechanism requires a strengthened equatorial SST gradient to produce the proxy-inferred dipole of wetter and drier conditions across the IO. However, this SST pattern is not consistently evident in marine paleoreconstructions. Further tests of this key prediction of our mechanism will require SST reconstructions with the spatial extent and accuracy to capture the magnitude of the gradient. Subsurface data could be key to test our hypothesis, since the wind–driven thermocline changes lead to large subsurface temperature signals. Continuous records of these parameters spanning several glacial cycles will be key to test the proposed "Sahul – Indian Ocean Bjerknes" mechanism. Last, this mechanism supports previous studies showing that Indian climate may be more sensitive to external perturbations than the Pacific [*Xie et al.*, 2013; *Tierney et al.*, 2013].

This issue deserves further attention due to the potential impacts of future climate change over the heavily populated Indian Ocean rim.

Acknowledgments. This research was funded by NSF P2C2 program through grants # ATM–1204011 and # OCN–1304910. We wish to acknowledge members of NCAR's Climate Modeling Section, CESM Software Engineering Group (CSEG), and Computation and Information Systems Laboratory (CISL) for their contributions to the development of CESM. PDN gratefully acknowledges CGD for supporting a long-term visit to NCAR. RFI was supported by a NERC Independent Research Fellowship #NE/K008536/1. The source code for the model used in this study, the CESM1, is freely available at `http://www.cesm.ucar.edu/models/cesm1.0/`. Both the data and input files necessary to reproduce the experiments with CESM1 are available from the authors upon request (pdn@ig.utexas.edu). The model output is archived at the University of Texas Institute for Geophysics (UTIG).

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Simulation Name	Description	Climatology years
	pre-industrial control	
0ka	Fully coupled	301–500
0kaAtm	Atmosphere-only forced with climatological SSTs from 0ka	11–110
	LGM sea level	
21kaShelvesAtm	0kaAtm + exposed Sunda + Sahul land mask	11–110
21kaSundaAtm	0kaAtm + exposed Sunda land mask	11–60
21kaSahulAtm	0kaAtm + exposed Sahul land mask	11–60
21kaShelves	0ka + exposed Sunda + Sahul land mask	301–400
21kaSunda	0ka + exposed Sunda land mask	111–210
21kaSahul	0ka + exposed Sahul land mask	111–210
21kaSahulWet	0ka + exposed Sahul land mask set to wetland	11–60
21kaSLnoTM	21kaShelves + POP2 bathymetry with shallower ITF sills	301–400
21kaSL	21kaSLnoTM + LGM tidal mixing*	301–500

Table 1. Climate model simulations. Simulations performed with CEMS1.2 under different combinations of LGM sea level boundary conditions and ocean–atmosphere coupling. *LGM tidal mixing based on *Montenegro et al.* [2007].

Climate response	Simulation difference
Uncoupled	
Sunda and Sahul shelves exposed	21kaShelvesAtm – 0kaAtm
Sunda shelf exposed	21kaSundaAtm – 0kaAtm
Sahul shelf exposed	21kaSahulAtm – 0kaAtm
Coupled	
Sunda and Sahul shelves exposed	21kaShelves – 0ka
Sunda shelf exposed	21kaSunda – 0ka
Sahul shelf exposed	21kaSahul – 0ka
Sahul shelf exposed as wetland	21kaSahulWet – 0ka
Partial/full closure of key passages of the Indonesian Throughflow (ITF)	21kaSLnoTM – 21kaShelves
Change in tidal mixing (TM)	21kaSL – 21kaSLnoTM
Response solely due to ocean dynamics	21kaShelves – 0ka – (21kaShelvesAtm – 0kaAtm)
Linear Sunda and Sahul shelves exposed	21kaSunda + 21kaSahul – 2 × 0ka
Full LGM sea level	21kaSL – 0ka

Table 2. Climate responses. Approach used to isolate the climate response associated with each mechanism from the simulations listed in Table 1.

Transports			
	Volume (Sv)	**Heat (Sv)**	**Freshwater (Sv)**
CESM1 – 0ka	8.9	0.78	0.26
CESM1 – 21kaSL	7.4	0.64	0.22
Observed	10	0.58	0.11

Table 3. Indonesian throughflow transports. Volume, heat, and freshwater transports simulated by CESM1 in the PI and LGM sea level simulations. Values indicate transport in the southward direction, heat and freshwater gain for the Indian Ocean. **Heat and freshwater transports are computed relative to reference temperatures and salinity of 0 °C and 35 ppm respectively. Observed transport values are from *Gordon* [2005] (for volume and heat) and *Talley* [2013] (for freshwater).**

Figure 1. Topography and bathymetry of the Maritime Continent. Present-day (left) and Last Glacial Maximum (right). Light blue areas in the left panel indicate the Sunda and Sahul shelves, currently submerged between Sumatra and Borneo; and Australia and New Guinea, respectively. Red arrows indicate key passages of the the Indonesian Throughflow. The LGM topography and bathymetry is derived by applying a 120 m sea level drop to NOAA's ETOPO5 [*ETOPO5*, 1988]

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Figure 2. Changes in hydroclimate of the Indo-Pacific warm pool (IPWP) at the Last Glacial Maximum (LGM). Network of (a) terrestrial and (b) marine proxies showing changes in LGM hydroclimate compiled by [*DiNezio and Tierney*, 2013]. Dots show locations of proxies and colors indicate drier (brown), unchanged (white), or wetter (blue) conditions at the LGM for terrestrial records and saltier (red), unchanged (white), or fresher (blue) for marine records. Colored (black) triangles indicate locations where two or more proxies agree (disagree). In panel a), locations in the ocean denote marine cores in which terrestrial proxies were measured. Coastlines correspond to a 120 m drop in sea level.

Figure 3. Observed and simulated present-day hydroclimate of the IPWP. Annual-mean (top) rainfall and (bottom) sea-surface salinity (SSS) over the Indo-Pacific warm pool (IPWP) from observations (left) and simulated by CESM1 (right). Rainfall observations are from GPCPv2 [*Adler et al.*, 2003] and SSS observations are from the NOAA World Ocean Atlas [*Antonov et al.*, 2010]

Figure 4. Model boundary conditions representing the effect of sea level on IPWP climate. Atmosphere model land fraction used in simulations with (a) pre-industrial sea level and (b) lowered LGM sea level. Ocean model bathymetry used in simulations with (c) pre-industrial sea level and (d) lowered LGM sea level. Both LGM boundary conditions are defined based on a 120 m sea level drop with respect to pre-industrial.

Figure 5. Vegetation distribution over the Maritime Continent for simulations with LGM sea level. Coverage for plant functional types (PFTs) used over the Maritime Continent, including the exposed Sunda and Sahul shelves. Note that the exposed Sunda and Sahul shelves are covered mainly by a mix of (b) deciduous tropical trees and (d) C_4 grass.

Figure 6. Simulated ocean currents and mixing around the Maritime Continent. Surface ocean circulation in the (a) pre-industrial (0ka) and (b) LGM sea level (21kaSL) simulations. Velocity vectors are averages over the upper 50 m of the ocean model. Colors indicate current speed in cm s^{-1}. Vertical diffusivity due to tidal and background mixing at a depth of 100 m in the (c) pre-industrial (0ka) and (d) LGM sea level (21kaSL) simulations.

Figure 7. Climate response to LGM sea level. Annual mean change in (a) rainfall, (b) sea-surface salinity, and (c) sea-surface temperature (colors) and surface wind stress (arrows) simulated by CESM1 in response to changes in LGM sea level with respect to preindustrial. The changes are computed by differencing the output from simulations 21kaSL minus 0ka. The 21kaSL simulation includes changes to the following boundary conditions due to lowered LGM sea level: 1) exposure of shelves and associated vegetation changes, 2) closed seaways or raised sills in key passages of the Indonesian Throughflow, and 3) changes in tidal mixing. Refer to Tables 1 and 2 for details on the experimental design. Vectors show changes in ocean surface wind stress over the ocean. Hydroclimate reconstruction data is also shown for comparison in (a) and (b), as in Figure 2.

Figure 8. Climate response to LGM sea level – ocean currents. Annual mean change in (a) thermocline depth, (b) zonal surface currents, and (c) upwelling simulated by CESM1 in response to changes in LGM sea level with respect to preindustrial. Zonal currents are averaged over the upper 50 m layer. Thermocline depth is computed as the depth of the maximum vertical temperature gradient. Positive (negative) changes in zonal currents indicate increased westward (eastward) velocity. Upwelling is defined as the upward velocity averaged over the 50 m to 100 m depth range. Positive change indicates increased vertical velocity. The changes are computed as in Fig. 7.

Figure 9. Climate response to LGM sea level broken down by mechanism. Coupled response to different boundary conditions are show from top to bottom: exposure of the (a,b) Sunda and (c,d) Sahul shelves, (e,f) closure of key passages of the Indonesian Throughflow, and (g,h) changes in tidal mixing associated with exposure of shelves. Annual mean changes in rainfall are shown on the left and sea-surface temperature and winds on the right. Wind vectors show changes in wind stress over the ocean. Refer to Tables 1 and 2 for details on the experimental design. Hydroclimate reconstruction data is also shown for comparison with the rainfall changes as in Figure 2.

Figure 10. Ocean response to shelf exposure vs. closure of ITF passages. Changes in sea-surface salinity (SSS, left) and sea-surface height (SSH, right) simulated in response to exposure of Sunda and Sahul shelves (top) and partial closure of key passages of the Indonesian Throughflow (bottom). Refer to Tables 1 and 2 for details on the experimental design.

Figure 11. Climate response to shelf exposure broken down by process. (a,b) Coupled response to exposure of both the Sunda and Sahul shelves. (c,d) Uncoupled response to shelf exposure simulated by replacing the full dynamical ocean model with prescribed seasonally-varying sea-surface temperature from the 0ka control simulation. (e,f) Response due to coupled ocean-atmosphere interaction computed by differencing the coupled and uncoupled responses (a minus c, and b minus d, respectively). (g,h) Linear response to exposure of the Sunda and Sahul shelves computed as the sum of the response to Sunda exposure plus the response to Sahul exposure from individual simulations. Refer to Tables 1 and 2 for details on the experimental design. Annual mean changes in rainfall are shown on the left and sea-surface temperature and winds on the right. Winds vectors show changes in wind stress over the ocean.

Figure 12. Uncoupled climate response to shelf exposure. Annual-mean changes in rainfall and wind stress in response to exposure of (a) both the Sunda and Sahul shelves (21kaAtmShelves), (b) Sunda shelf (21kaAtmSunda), and (c) Sahul shelf (21kaAtmSahul). These uncoupled responses are from simulations where the fully dynamical ocean model is replaced with prescribed seasonally-varying sea-surface temperature from the 0ka control simulation. Refer to Tables 1 and 2 for details on the experimental design. Annual mean changes in rainfall are shown on the left and sea-surface temperature and winds on the right. Winds vectors show changes in wind stress over the ocean.

Figure 13. Seasonality of the coupled climate response to exposure of the Sahul shelf. Seasonal changes in rainfall (left) and surface temperature and winds (right) in response to exposure of the Sahul shelf. Changes for the following seasons are shown from top to bottom: March-April-May (MAM), June-July-August (JJA), September-October-November (SON), and December-January-February (DJF). Vectors show changes in wind stress over the ocean. Refer to Tables 1 and 2 for details on the experimental design.

Figure 14. Seasonality of the uncoupled climate to exposure of the Sahul shelf. As in Fig. 13, but for a simulation with exposed Sahul shelf and prescribed seasonally-varying sea-surface temperature from the 0ka control simulation.

Figure 15. Sensitivity of surface relative humidity and albedo to land surface properties. Changes in surface relative humidity (left) and surface albedo (right) in simulations where the Sahul shelf is set as dry vegetated land (top) and wetland (bottom). Refer to Tables 1 and 2 for details on the experimental design.

Figure 16. Seasonality of the coupled climate response when the Sahul shelf is wetland. As in Fig. 13, but for a simulation with the exposed Sahul shelf set to wetland.

Figure 17. Impact of albedo and relative humidity on convective environment and rainfall. Changes in low-level entropy (left) and rainfall (right) in uncoupled simulations where the Sahul shelf is set as dry vegetated land (top) and wetland (bottom). The shelf surface has the same albedo in both simulations, however, with higher evaporation capacity and relative humidity in the wetland case. Changes are shown for the March-April-May (MAM) season when shelf exposure initiates the coupled response. Low level entropy is computed as the equivalent potential temperature on a terrain-following model level about 20 hPa above the surface.

Figure 18. Sensitivity of the coupled climate response to shelf coverage type. Rainfall changes during March-April-May (MAM, left) and September-October-November (SON, right) simulated when the exposed Sahul shelf set to bare soil (top) and C_4 grass, i.e. savanna conditions (bottom).

Figure 19. Seasonal controls on the coupled response. Upwelling during June-July-August (JJA, top) and sea surface temperature (SST) during September-October-November (SON, bottom) from observations (left) and simulated by CESM1 in the pre-industrial control (right). Observed SST data is from NOAA Optimum Interpolation (OI) product [*Reynolds and Smith*, 1995]. Upwelling data is from the ORAS-3 ocean reanalysis [*Balmaseda et al.*, 2008].

Figure 20. Sahul – Indian Ocean Bjerknes Mechanism. Schematic diagram illustrating the essential processes driving changes in Indian Ocean climate in response to exposure of the Sahul shelf. The response is initiated during the March–April–May season (top) when the surface of the Sahul shelf cools due to the increased albedo of land relative to ocean. The shelf is delimited by yellow lines between Australia and New Guinea. Drying and increased subsidence (brown vertical arrows) occur over the colder surface of the Sahul shelf driving a divergent surface circulation with anomalous easterly winds (black horizontal arrows) over the eastern Indian Ocean. The response peaks during the September–October–November season (bottom) when the the initial uncoupled easterly wind anomaly is amplified by the Bjerknes feedback resulting in warming (red) of the western Indian Ocean and cooling (blue) of the eastern Indian Ocean, particularly off the coast of Sumatra. Increased convection and wetter conditions (green vertical arrows) occur over the warmer ocean and increased subsidence and drier conditions (brown vertical arrows) occur over the colder ocean.