ABSTRACT

Previous studies have shown that increases in poleward ocean heat transport (OHT) do not strongly affect tropical SST. The goal of this paper is to explain this observation. To do so, the authors force two atmospheric global climate models (GCMs) in aquaplanet configuration with a variety of prescribed OHTs. It is found that increased OHT weakens the Hadley circulation, which decreases equatorial cloud cover and shortwave reflection, as well as reduces surface winds and evaporation, which both limit changes in tropical SST. The authors also modify one of the GCMs by alternatively setting the radiative effect of clouds to zero and disabling wind-driven evaporation changes to show that the cloud feedback is more important than the wind-evaporation feedback for maintaining constant equatorial SST as OHT changes. This work highlights the fact that OHT can reduce the meridional SST gradient without affecting tropical SST and could therefore serve as an additional degree of freedom for explaining past warm climates.

1. Introduction

Climate records of the past 100 million years show that sea surface temperature (SST) has varied much less in the tropics (Pearson et al. 2001; Tripati et al. 2003) than temperatures at higher latitudes (Sluijs et al. 2006; Littler et al. 2011). Many researchers have considered mechanisms that may limit changes in tropical SST in response to changes in the earth’s radiative forcing: that is, changes in greenhouse gas concentration or solar insolation (Newell 1979; Ramanathan and Collins 1991; Clement et al. 1996; Miller 1997; Lindzen et al. 2001; Williams et al. 2009). Nevertheless, other factors, such as changes in poleward ocean heat transport (OHT), are also likely to have affected tropical SSTs.

One might wonder how OHT could deviate much from its present value, especially at low latitudes. The total planetary heat transport (PHT) is roughly fixed as long as absorbed solar radiation and outgoing longwave radiation (OLR) are locally coupled (Stone 1978), and it is difficult to change the atmosphere–ocean partitioning of PHT at low latitudes because the Hadley and wind-driven ocean overturning circulations are closely linked through surface wind stresses (Held 2001). However, this reasoning can break down for climates in the distant past. Changes in continental configuration and especially a circumglobal ocean passageway at low latitudes would have increased OHT in the Cretaceous and perhaps the early Eocene. Coupled global climate models (GCMs) show that an open passageway increases OHT by ~1 PW (Hotinski and Toggweiler 2003; Enderton and Marshall 2009). This occurs because an open passageway prevents a geostrophically balanced equatorward ocean flow at depth, which leads to equatorial upwelling of deeper and colder water (Hotinski and Toggweiler 2003) and weakens equatorial gyres that transport heat equatorward (Enderton and Marshall 2009). OHT could also have changed via tropical cyclones, which affect OHT by mixing warm surface water through the thermocline (Emanuel 2001). The strength of cyclones and mixing increases with SST and suggests that warm climates could be tied to high OHT regimes (Emanuel 2002). Coupled GCM simulations confirm that imposed ocean mixing is capable of significantly increasing OHT (Korty et al. 2008; Manucharyan et al. 2011), although this depends on the location, strength, and timing of the imposed mixing (Sriver and Huber 2007; Sriver et al. 2008; Jansen and Ferrari 2009; Jansen et al. 2010). If the SST–cyclone mixing–OHT feedback (Emanuel 2002) is confirmed it could help explain several observations, including the low temperature gradients in the Pacific
during the Pliocene (Brierley et al. 2009; Fedorov et al. 2010, 2013) and the low meridional SST gradient during the Eocene (Korty et al. 2008).

This motivates consideration of how the climate system responds to changes in OHT. One might think that increased OHT affects SST by raising PHT, which would warm the extratropics and cool the tropics (e.g., Covey and Thompson 1989). However, several recent GCM studies have shown that, when OHT is increased, atmospheric heat transport (AHT) reduces to maintain a roughly constant PHT (Winton 2003; Herweijer et al. 2005; Vallis and Farneti 2009; Barreiro et al. 2011). The degree of compensation varies and depends on model parameterizations (Kang et al. 2008), but in some cases PHT even decreases as OHT is increased (Rose and Ferreira 2012). Despite PHT staying nearly constant in GCMs when OHT is varied, higher OHT robustly causes extratropical warming (Winton 2003; Herweijer et al. 2005; Rose and Ferreira 2012). The warming is driven by positive feedbacks: increased OHT leads to more evaporation in the subtropics and midlatitudes, which triggers convection, raises the amount of water vapor in the upper troposphere, and thus increases the extratropical greenhouse effect (Herweijer et al. 2005; Rose and Ferreira 2012). In the storm tracks, convection also leads to a shift from reflective low stratus cloud cover toward high convective clouds, which causes additional warming (Rose and Ferreira 2012).

In the tropics, one would expect that increased OHT leads to cooling, which would be further amplified by the above feedbacks. However, Herweijer et al. (2005) found that tropical SST decreases by less than \(\sim 1\) K when OHT is increased from zero to its modern value in three GCMs. Similarly, Rose and Ferreira (2012) found that tropical SST is approximately constant across a huge range of imposed OHT regimes. Rose and Ferreira argued that this effect results from locally canceling shortwave and longwave changes in the top-of-atmosphere (TOA) energy budget. A complication for this argument is that the tropical surface energy budget is affected by PHT as well as TOA fluxes. Therefore, a change in the surface energy budget could potentially be balanced by a change in AHT without having a large impact on the TOA energy budget. Our goal is to consider this issue in detail and explain why tropical SST is insensitive to OHT changes.

Our approach differs from previous studies, which either used complex GCMs with modern-day boundary conditions (Winton 2003; Herweijer et al. 2005; Barreiro et al. 2011) or an “intermediate complexity” GCM in an aquaplanet configuration (Rose and Ferreira 2012). We use an aquaplanet configuration but compare two modern climate models with sophisticated physics parametrizations. This allows us to isolate the atmospheric response to OHT changes, including important cloud processes. Similar to previous studies, we find that SST in the inner tropics, where the OHT forcing is largest, remains roughly constant irrespective of OHT. We also identify the following mechanism: increased OHT reduces the strength of the Hadley circulation, which decreases tropical cloud cover and shortwave reflection in a way that compensates for the OHT forcing. The reduction in the strength of the Hadley circulation also decreases tropical surface winds, which reduces tropical evaporation and represents a smaller, secondary feedback [for other examples of stabilizing wind feedbacks, see Bates (1999) and Caballero (2001)].

## 2. Methods

We use two atmospheric GCMs: the Community Atmosphere Model versions 4 (CAM4) and 5 (CAM5). The changes between CAM4 and CAM5, particularly their cloud schemes, are substantial enough that they may be considered two distinct climate models (Neale et al. 2010, 2012). Our main results are robust across both models. We also consider three modified versions of CAM4 to explore how the SST response depends on cloud and evaporation feedbacks: in “no-CRF” we enforce permanent clear-sky radiation, thus disabling all cloud radiative forcing (CRF). In “\(U\) fixed” we fix the surface wind speed that is used to compute evaporation to its zonal-mean value from CAM4 with zero OHT. This does not affect the actual wind speeds in the model and only serves to make evaporation insensitive to changes in surface wind. In “no-CRF + \(U\) fixed” we combine the changes, but use wind speeds from no-CRF with zero OHT.

We run both models and all cases in aquaplanet configuration, coupled to a 50-m-deep slab ocean. In CAM4 we set the radiative effect of aerosols to zero. CAM5 requires some aerosol as cloud condensation nuclei, so we remove all aerosol species except sea salt. We set eccentricity to zero, retain a present-day obliquity (23.44°), and set CO\(_2\) to 100 ppm to balance the low surface albedo and lack of aerosol forcing. Because CAM4 has a global net cloud forcing of \(\sim -30\) W m\(^{-2}\), we reduce the solar constant by 120 W m\(^{-2}\) in no-CRF and no-CRF + \(U\) fixed. We do not include a sea ice model and allow the slab ocean temperature to fall below freezing. This setup eliminates the nonlinear ice–albedo feedback and emphasizes the atmospheric response to changes in OHT. The tropical SST in our runs with zero OHT is slightly higher than at present and lower than in Eocene reconstructions (Fig. 1; Huber and Caballero 2011).
We force the GCMs with a range of prescribed OHTs inside the slab ocean. The OHT shape is fixed and represents an idealized zonal-mean, annual-mean present-day OHT (Fig. 2a; cf. Trenberth and Caron 2001). Similar to present day, the OHT exports heat out of the inner tropics (equatorward of 17.5°) and converges heat in the subtropics and extratropics (17.5°–65°). Sensitivity tests with a different shape indicate that our results are qualitatively robust (not shown). We vary the forcing by multiplying the OHT shape with a constant and explore a regime that ranges from zero OHT to approximately twice the present-day OHT. In CAM4 we also explore the response to a reversed (equatorward) OHT for illustrative purposes. Finally, we double and halve CO₂ in CAM4, CAM5, and no-CRF with OHT set to zero.

3. Results

a. Extratropics and subtropics

We find strong but not complete compensation between AHT and OHT in both models. In CAM4 the maximum PHT increases by about 0.26 PW per PW increase in maximum OHT (Fig. 2b). In CAM5 and the modified versions of CAM4 PHT increases between 0.18 and 0.31 PW per PW of OHT increase (not shown). The degree of atmospheric compensation does not correlate with the tropical SST response.

As in previous studies, we find that increased OHT leads to extratropical warming (Fig. 3). To understand the warming, we consider the surface and TOA energy budgets. At the surface and at all latitudes except near the equator, the forcing due to an increase in OHT is largely balanced by changes in latent heat flux: that is, evaporation (Fig. 4). Poleward of 17.5°, where OHT converges, evaporation increases. As in Herweijer et al. (2005), more evaporation increases the water vapor greenhouse effect in the subtropics. All three models show that higher OHT decreases the net longwave flux at the surface in the subtropics (Fig. 4). This occurs despite warmer SST and indicates that the greenhouse effect has increased. In the midlatitudes, the incoming shortwave flux increases in CAM4 and $U$ fixed (Figs. 4a,b). The increase in shortwave radiation is caused by cloud changes (cf. Fig. 4c) and explains why the
midlatitudes warm less in model runs without cloud radiative effects (cf. Figs. 3a,d). Similar to Rose and Ferreira (2012), we find that the mechanism for increased shortwave absorption in the midlatitudes is increases in extratropical convection, which reduce low cloud cover in the storm tracks (Fig. 6).

b. Inner tropics

We find that SST near the equator remains almost constant in both CAM4 and CAM5 (Fig. 5). When the OHT maximum is increased from 0 to 3 PW (= $75 \text{ W m}^{-2}$ local forcing), equatorial SST actually increases slightly ($<1 \text{ K}$) in both models. For reference, equatorial SST warms 1.3 K (CO$_2$ doubling)$^{-1}$. To understand why an OHT increase does not cool the tropics, we consider the CAM4 energy budgets. At the TOA, near the equator, increases in PHT are balanced by increases in absorbed shortwave, with some compensation from decreases in the greenhouse effect (Fig. 4a). Since there are negligible changes in clear-sky radiative fluxes, this points to an important role for clouds in regulating the net tropical response. Because the AHT changes, we also have to consider the surface energy budget to understand why SST remains nearly constant. At the surface equatorward of 17.5°, increases in absorbed shortwave and decreases in latent heat flux play a comparable role in balancing the negative OHT forcing (Fig. 4a). This suggests that an evaporative feedback, in addition to the cloud feedback, plays a role in regulating the equatorial SST as OHT changes. We note that the latent heat flux decreases despite slightly higher SST, so this change cannot be driven by local thermodynamics.

Both feedbacks result from changes in the Hadley circulation. As OHT increases, the meridional SST gradient weakens (Fig. 3), which reduces the strength of the Hadley circulation by about 20% per PW of OHT (Figs. 6a,b). This leads to a large decrease in cloud cover near the equator (Figs. 6a,b) and increases the absorbed shortwave. The reduction in the Hadley cell strength also reduces surface wind speed across the tropics (Fig. 6c) enough to explain why latent heat flux decreases despite increased SST (not shown).

Our simulations with variables held fixed demonstrate that the cloud feedback is more important than the wind–evaporation feedback. In the no-CRF case, in which there can be no cloud feedback, the equatorial SST decreases significantly (by 6.2 K) when the OHT maximum is increased from 0 to 3 PW (Fig. 5). The wind–evaporation feedback alone is therefore insufficient to explain why latent heat flux decreases despite increased SST (not shown).

FIG. 3. Zonal-mean SST change when OHT is varied. Line colors and numerical labels correspond to the forcings in Fig. 2a. Increased poleward OHT warms the extratropics in all models. In CAM4 and CAM5, equatorial SST is almost invariant to OHT changes. Compared to CAM4, there is some decrease in equatorial SST in the $U$ fixed case and much larger decreases in the no-CRF and no-CRF + $U$ fixed cases.
reduces slightly less when OHT is increased in $U$ fixed than in CAM4 (Figs. 4a,b). This is driven by an increase in near-surface relative humidity, which compensates for constant surface winds (see section 4).

Finally, one would expect the largest equatorial cooling when we disable both feedbacks. Indeed, when the OHT maximum is increased from 0 to 1.5 PW, SST decreases more in the no-CRF $U$ fixed case than all other simulations (Fig. 5). Note that this is not true when the OHT maximum is further increased to 3 PW, at which point no-CRF + $U$ fixed cools less at the equator than no-CRF. This occurs because, by holding $U$ fixed, the no-CRF + $U$ fixed case has an artificially increased source of tropospheric moisture compared to no-CRF. This increases the tropical greenhouse effect, warms the surface and outweighs the increased surface cooling from evaporation (not shown).

4. Discussion and conclusions

The goal of this paper was to explain why tropical SST is not strongly affected by changes in OHT. We find that higher OHT weakens the Hadley circulation, which reduces reflective cloud cover and, to a lesser extent, reduces the surface winds that sustain evaporation, allowing SST to remain almost constant over a large range of OHT states.

We note that we only discuss feedbacks that explain the tropical SST response in CAM4 and CAM5. However, because evaporation still responds to OHT changes
in the $U$ fixed simulations, one might wonder what would happen if evaporation were held completely fixed. We performed such simulations in CAM4 and found that the results depend extremely sensitively on the detailed way in which evaporation is fixed. This is because a runaway low-level cloud feedback similar to that observed by Stainforth et al. (2005) occurs in some cases. In the fixed-evaporation simulations in which this cloud feedback did not occur, clouds were able to adjust to maintain roughly constant tropical SST.

Our treatment of OHT is an idealization and ignores atmosphere–ocean interactions. That means some of the OHT states we consider might not be physically realizable. For example, we find that an OHT increase weakens surface winds, which, assuming Sverdrup balance, should in turn decrease the OHT. Nevertheless, coupled GCMs that resolve atmosphere–ocean interactions produce a large variety of possible OHT states (Enderton and Marshall 2009; Vallis and Farneti 2009; Ferreira et al. 2011), which suggests that OHT changes are plausible for past climates. We also use a meridionally symmetric OHT profile, which neglects the effect of cross-equatorial OHT on the position and strength of the ITCZ (Kang et al. 2008). Similarly, we use an aquaplanet configuration and do not capture zonally asymmetric features (e.g., the Walker circulation), the ice–albedo feedback, or the effect of continents. However, the insensitivity of tropical SST to changes in OHT has been demonstrated in models that include sea ice as well as present-day continents (Winton 2003; Herweijer et al. 2005). Finally, the fact that the cloud feedback operates in multiple climate models and is driven by large-scale dynamical changes suggests that our results should be qualitatively robust for different cloud parameterizations. For example, we find that the change in surface shortwave flux at the equator differs less than 3 W m$^{-2}$ between CAM4 and CAM5.

This work is relevant to the ongoing debate about past warm climates—for example, the Pliocene and Eocene.
which had a lower pole–equator temperature gradient than the modern climate. One can explain these climates to a large extent in terms of elevated CO₂ concentrations (Huber and Caballero 2011) combined with high-latitude feedbacks—for example, the ice–albedo effect (North 1975) and changes in polar convective clouds (Abbot and Tziperman 2008a,b). However, in the case of the Eocene, some high CO₂ scenarios might lead to tropical surface temperatures that surpass the heat tolerance of tropical plants. This should have led to large-scale extinctions (Huber 2008), for which there is currently no supporting evidence in the paleorecord (Jaramillo et al. 2010). Similarly, modern GCMs still have difficulty reproducing the observed temperature gradients of the early Pliocene (Fedorov et al. 2013). Our work is consistent with the idea that OHT could act as an additional lever for explaining past warm climates. This lever could be independent of CO₂ changes (e.g., changes in continental configuration) or act as a feedback on warming caused by CO₂ changes (e.g., tropical cyclone feedback). In this scenario, CO₂ would regulate tropical temperatures and cause changes in the pole–equator temperature gradient through high-latitude feedbacks, whereas changes in OHT would modify the pole–equator temperature gradient without significantly affecting tropical SST.

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