The tropical precipitation response to Andes topography and ocean heat fluxes in an aquaplanet model

ELIZABETH A. MAROON, * DARGAN M. W. FRIERSON, and DAVID S. BATTISTI

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

*Corresponding author address: Elizabeth A. Maroon, Department of Atmospheric Sciences, University of Washington, Box 351640, Seattle, WA 98195-1640.
E-mail: emaroon@uw.edu
This aquaplanet modeling study using AM2.1 examines how ocean energy transport and topography influence the location of tropical precipitation. Adding realistic Andes topography regionally displaces tropical rainfall from the equator into the northern hemisphere, even when the wind-evaporation feedback is disabled. The relative importance of the Andes compared to the asymmetric hemispheric heating of the atmosphere by ocean transport is examined by including idealized and realistic zonally-averaged surface heat fluxes (also known as q-fluxes) in the slab ocean. A hemispherically asymmetric q-flux displaces the tropical rainfall toward the hemisphere receiving the greatest heating by the ocean. The zonal mean displacement of rainfall is greater in simulations with a realistic q-flux than with realistic Andes topography. Simulations that add both a q-flux and topography displace rainfall farther to the north in the region 120 degrees to the west of the Andes than in simulations that only have a q-flux. Cloud and clear sky radiative feedbacks in the tropics and subtropics of this model both act to amplify the energy flux and the precipitation response to a given hemispheric asymmetry in oceanic forcing.
1. Introduction

More precipitation falls in the Northern Hemisphere (NH) tropics than in the Southern Hemisphere (SH) tropics annually. The zonally averaged tropical precipitation is tied to the ascending branch of the Hadley circulation, also known as the Intertropical Convergence Zone (ITCZ) (Hadley 1735; Dima and Wallace 2003). However, the variation of precipitation from the zonal mean is more complex. Tropical rain over land follows the seasonal cycle of insolation, but the precipitation patterns over the world’s oceans differ from basin to basin (Mitchell and Wallace 1992; Waliser and Gautier 1993; Schumacher and Houze 2003). For the majority of the year, there is greater precipitation in the NH of the Atlantic and Pacific Ocean basins, while there is greater precipitation in the SH of the Indian Ocean (Adler et al. 2003; Huffman et al. 2007). Only during the months of March and April is there an equivalent or greater amount of rain in the SH of the Pacific basin. Sometimes there is a double ITCZ, a feature that current general circulation models (GCMs) struggle to capture accurately (Mechoso et al. 1995; Zhang 2012; Hwang and Frierson 2013).

This greater NH precipitation has often been explained through processes and mechanisms local to the tropics. Philander et al. (1996) suggest that the NH Pacific and Atlantic ITCZ is a result of tropical coastal configuration preferring cold water coastal upwelling in the eastern side of the basin, south of the equator. This cold water, in turn, drives a wind-evaporation-sea surface temperature (WES) feedback that further pushes more precipitation to the NH tropics (Xie and Philander 1994). If waters are colder in the SH, then a cross-equatorial pressure gradient is induced that drives a southerly wind across the equator. In the SH, the anomalous higher pressure results in anticyclonic motion that enhances the Trade winds to the south of the equator, and in the NH, anomalous lower pressure results in cyclonic motion that decreases Trade winds near the equator. Just south of the equator, the stronger Trade winds enhance evaporation, which further cools the surface near the equator in the SH, and the opposite occurs in the NH. Thus, changes in the Trade winds on both sides of the equator reinforce the changes in SST. Southerly winds across the equator advect...
moisture to the NH, affecting the location of the ITCZ. The atmosphere-only model of Philander et al. (1996) was coupled to both a fixed-SST boundary and a fully dynamic ocean, did not include topography, and required stratus cloud feedbacks to see a large movement of the ITCZ to the NH. Xie (1996) explains how such a WES mechanism can be propagated westward, and argues that the structure of the eastern coast is key.

Xu et al. (2004) and Takahashi and Battisti (2007a) have argued that the presence of the Andes in models can be an important asymmetry that keeps the Pacific ITCZ north of the equator in the annual mean. The mountains block midlatitude westerlies and air is forced to follow isentropic surfaces to the north and south (Terra and Mechoso 2003). Drier air subsides as it moves toward the tropics, where it turns westward, enhances the SH Trade winds and dries the SH tropics (Rodwell and Hoskins 2001). Takahashi and Battisti (2007a) tested the influence of the Andes on tropical precipitation relative to the influence of other mountain ranges: they found that the presence of the Andes was more important to the location of the Pacific ITCZ than the Rockies or the Tibetan Plateau were. This topographic theory cannot, however, explain the greater NH precipitation in the Atlantic basin, as there is no equivalent “Andes” range in western Africa. They presented simulations that added an Andes mountain range with both a mixed-layer ocean (similar simulations as we are about to show results for) and fixed SST and found that coupling the model to a simple mixed-layer ocean was essential for increasing tropical precipitation in the NH: changing SSTs through ocean-atmosphere surface flux feedbacks are important for the location of precipitation.

In contrast to these local tropical mechanisms, modeling studies in the past decade have noted a connection between tropical precipitation and extratropical forcing (Chiang et al. 2003; Chiang and Bitz 2005; Zhang and Delworth 2005; Broccoli et al. 2006; Kang et al. 2008, and others). Paleoclimate evidence also shows evidence of the ITCZ moving during climatic shifts (Black et al. 1999; Thompson et al. 2000; Koutavas and Lynch-Stieglitz 2003, for example). A study by Fučkar et al. (2013) using an idealized coupled model (Farneti and Vallis 2009) has shown that the location of the zonally averaged ITCZ is determined by
the hemisphere with the greater upward heat flux from the extratropical ocean. By opening
a channel to create a circumpolar ocean in one hemisphere, deep water production in that
hemisphere is inhibited, pushing the sinking branch of the ocean meridional overturning
circulation (MOC) to the opposite hemisphere, which causes the zonally averaged ITCZ to
move into the NH.

With greater heat release to the atmosphere in the hemisphere with the sinking branch,
an energetic argument proposed by Kang et al. (2009) can be used to explain the tropical
precipitation response: greater heat release into the extratropics of one hemisphere means
weaker quasi-diffusive energy transport into the extratropics of that hemisphere. Conversely,
less heat release from the ocean to atmosphere results in stronger energy transport poleward
in that hemisphere. A hemispherically-asymmetric poleward transport of energy eventually
affects the tropics and the Hadley circulation: a weaker meridional temperature gradient
results in an a weaker, or summer-like, Hadley circulation in that hemisphere; a stronger
gradient strengthens the Hadley circulation. Thus, the zonally averaged ITCZ location -
which demarks the equatorward limit of the two Hadley cells - shifts into the hemisphere
with deep water production. As a consequence, there is a net cross-equatorial atmospheric
energy transport (AET) into the colder hemisphere (i.e. the hemisphere without deep water
production). Frierson et al. (2013) find observational evidence for this energetic constraint
between the tropics and extratropics in satellite data and reanalysis. They point to the
Atlantic MOC as the key zonal asymmetry that can explain greater zonally-averaged tropical
rain in the NH. Kang et al. (2014b) find that the Atlantic MOC is responsible for making
the NH warmer than the SH, which also would lead to greater tropical precipitation in the
NH.

There are two ways to investigate this asymmetry in tropical precipitation, as a zonal
mean change or as the sum of precipitation changes in individual basins. Both will ultimately
yield the same answer, but will lead to differing causal interpretations. Here, we favor the
zonal mean approach to examine changes in the closed energetic budget. We investigate
the roles of local and remote forcing on tropical precipitation in an atmospheric general circulation model (GCM). We use the Andes mountain range as our tropical forcing, as well as a variety of surface heat fluxes (also called “q-fluxes”) to test the relative effectiveness of topography and oceanic heat release in moving tropical precipitation. In Section 2, we describe the model we used. Section 3 presents changes in the hydrologic cycle and TOA radiation in the simulations that only add an Andes mountain range. Section 4 discusses the results of Andes-only simulations and examines the importance of the WES feedback. The results from simulations with both a mountain range and surface heat fluxes are presented in Section 5, and conclusions are presented in Section 6.

2. Model Description

The model used here is an aquaplanet version of the AM2.1 atmospheric GCM with a finite volume dynamical core and a slab ocean (Anderson et al. 2004; Delworth et al. 2006). The simulations of Kang et al. (2008) use a similar configuration. An atmospheric GCM coupled to a slab ocean conserves energy (unlike a fixed SST boundary), which allows us to examine the energetic budget. All fields that would produce an asymmetric radiative forcing (e.g., ozone) have been symmetrized about the equator. Annual mean insolation is used. The radiative impact of aerosols is not included. The horizontal resolution is 2.5° × 2° and there are 24 vertical hybrid pressure-sigma coordinate levels. The heat capacity of the slab ocean is 1 × 10^7 JK^{-1}m^{-2} (2.4m). The shallow mixed layer depth decreases integration time and has little effect on the mean climate in comparison to deeper depths (Kang et al. 2008). A relaxed Arakawa-Schubert convection scheme is used for moist convection.

Aquaplanet simulations were completed with realistic Andes topography (T), with zonally symmetric idealized (IQ) and real-world (RQ) surface heat fluxes (also known as q-fluxes), and with combinations of both topography and surface heat fluxes (T+RQ, T+IQ10, T+IQ30). Figure 1 shows the geometry of the topography and the q-fluxes used. All simula-
tions that add topography or q-flux forcing are compared against a completely flat aquaplanet simulation that does not include q-fluxes (the “A” simulation).

The T simulation uses real-world Andes topography interpolated with a spline fit to the model’s grid, and everywhere outside of the Andes topography is flat. The mountains used are “water mountains” where only the height of the surface, not the surface type or roughness, is changed. These simulations have no land. There are also two simulations presented that test the importance of the WES feedback to the response of tropical precipitation. One simulation is flat (NoWES) and one has real-world Andes topography (T+NoWES).

The idealized q-flux described in Kang et al. (2008) is used in simulations with amplitudes of 10 and 30 W/m² (IQ10 and IQ30). The real-world zonally-averaged q-flux (RQ) is also used: it is derived from the CERES TOA energetic budget and the ERA-Interim atmospheric reanalysis (Wielicki et al. 1996; Dee et al. 2011). To insure that the land does not introduce a spurious energy transport into this zonally-averaged q-flux, the surface flux values over land from these datasets are set to zero, and then this surface flux is zonally averaged and becomes the RQ q-flux. The implied ocean energy transport (OET) of these q-fluxes indicates that heat is moved southward at all latitudes in IQ10 and IQ30, while in the RQ simulations the implied heat transport is generally poleward, but is northward at the equator (Figure 1c). In simulations that have both Andes topography and q-fluxes (i.e. T+RQ, T+IQ10, T+IQ30), the q-flux over the water mountains is the same q-flux as all other points at that latitude.

Table 1 lists all simulations that will be discussed in the following sections. Simulations were spun up for five years and five additional years beyond the spin-up are used for analysis. All results discussed in this paper are robustly defined using these averaging periods.
3. Addition of a single southern hemisphere mountain range

a. Changes in the mean state due to the addition of topography

In this section, we describe the impact of adding a realistic Andes mountain range to the model. The results using idealized Andes topography are similar to realistic topography experiments (Maroon 2013) and so only the latter are described here. Adding a mountain range creates subsidence and equatorward meridional wind to the west of the mountains as in Takahashi and Battisti (2007a). When westerlies hit the mountain barrier in the subtropics (see Figure 2a), they are deflected poleward and equatorward as required by Sverdrup balance (Rodwell and Hoskins 2001). Increased subsidence inhibits precipitation south of the equator, and tropical rainfall is displaced northward on the west side of the Andes (Figure 3). The decrease in rain in the SH is greater than the increase of rain in the NH; the decrease in SH SST near the Andes is also greater than the very slight increase in NH SST. There is a decrease in the maximum of the zonal averaged tropical precipitation when adding topography. Evaporation does not change as much as precipitation when adding topography in equilibrium (Figure 3c).

Increasing SH subtropical subsidence increases the low cloud amount and decreases high clouds, which both cool the surface (Figure 2a&b) in the added topography experiment compared to the aquaplanet experiment. Terra and Mechoso (2003) also found an increase in stratocumulus clouds due to the presence of the Andes. As discussed in Takahashi and Battisti (2007a), the addition of Andes topography results in greater subsidence of dry air into the boundary layer, which temporarily increases evaporation and starts cooling. Stratus clouds then act as a feedback to increase the cooling, resulting in a decrease of evaporation once steady state is reached. The decrease in SST is located in the same region as an increase in low clouds and a decrease in high clouds (Figure 2a&b). The increase in stratus clouds reduces the insolation reaching the surface, while the decrease in high cloud promotes
radiation escape to space: this reduces SST and further reinforces subsidence. The changes in evaporation are also linked with the changes in low cloud amount.

The westward extent of the northward precipitation displacement is much greater than the Rossby radius of deformation; this suggests that westward propagation is through some mechanism other than flow deformation due to the topography. Enhanced cross-equatorial winds near the equator would potentially indicate a WES feedback is responsible. However, subsidence and associated stratus cloud feedbacks inhibit convection in the SH, thereby preferring a NH ITCZ, without necessarily needing to invoke a WES feedback. In section 4, we will perform an experiment to examine the importance of the WES feedback for the response of tropical precipitation to topography.

Examining the change in the moisture budget from adding topography shows that the majority of the change in the precipitation is balanced by the convergence term (Figure 3). The change in evaporation opposes the change in precipitation, but it is almost an order of magnitude smaller than the precipitation change. Time mean and zonal mean moisture flux convergence can be further decomposed into (see, e.g., Peixoto and Oort (1992)):

\[- \nabla \cdot \left[ vq \right] = - \nabla \cdot \left[ \bar{v} \bar{q} \right] - \nabla \cdot \left[ v^* q^* \right] - \nabla \cdot \left[ v' q' \right]. \tag{1} \]

Terms with asterisks are the anomalies from the zonal mean, and terms with primes are the departure from the time mean. The first term on the right hand side of Equation 1 represents the portion of the moisture flux convergence that is attributed to the mean meridional circulation (MMC). The second term is the part from the stationary eddies and the third term is the part from all transient circulations. The majority of the change in moisture flux convergence is due to changes in the MMC term (the Hadley circulation) (Figure 4a). Although one might first think that adding mountains would result in a substantial response in the stationary eddy term, this is not the case; the zonal anomalies of specific humidity and meridional wind do not covary. Figure 2a shows that the change in streamlines from the addition of the Andes causes an anomalous cross-equatorial Hadley cell that transports more moisture near the surface into the NH. Moisture has decreased in the SH
tropics and subtropics and increased in the mid-troposphere of the NH tropics coincident with the anomalous Hadley cells’ ascending branch (Figure 2c). We conducted two additional simulations that doubled and quadrupled the width of an idealized Andes range and found very little change to the magnitude of stationary eddy and MMC terms (not shown). The location of tropical precipitation also did not change from doubling and quadrupling the range’s width. Similarly, Takahashi and Battisti (2007b) mention a simulation where they doubled the width of their Andes and found little change in the zonal extent of the Pacific dry zone that they were studying.

b. Examining the energetic changes in the AM2 simulation with added topography

Adding topography results in an anomalous transport of energy from the NH to SH (Figure 5). This southward transport of energy is accomplished by an increase in the transport of energy in the upper branch of an anomalous Hadley cell (recall Figure 2c), and is consistent with a northward shift in the ITCZ. In the absence of storage by the ocean, the cross-equatorial energy transport can be examined through the top-of-atmosphere (TOA) radiative imbalance between the two hemispheres:

$$\Delta OET + \Delta AET \equiv \Delta TET,$$

where,

$$\Delta TET \equiv \int_0^{2\pi} \int_0^{\pi/2} 2\pi a^2 \cos(\phi)(\Delta SW - \Delta OLR)d\phi d\lambda$$

$$\Delta TET \equiv - \int_0^{2\pi} \int_0^{\pi/2} 2\pi a^2 \cos(\phi)(\Delta SW - \Delta OLR)d\phi d\lambda,$$

and where $\phi$ is latitude, $\lambda$ is longitude, $a$ is the radius of the Earth, $\Delta OET$ is the change in ocean energy transport at the equator, $\Delta AET$ is the change in the atmospheric energy transport at the equator, and $\Delta TET$ is the total meridional energy transport at the equator, $\Delta SW$ is the change of net shortwave radiation (with positive defined as into the atmosphere) and $\Delta OLR$ is the change in outgoing longwave radiation (with positive defined as leaving the atmosphere). In the limits of the integral, the equator is 0 and the north and south poles...
are $\pi/2$ and $-\pi/2$, respectively. SW and OLR fluxes are evaluated at model TOA and are integrated over the respective hemispheres.

The change in total meridional energy transport across the equator is exactly balanced by the change in the net energy that enters the climate system (atmosphere and ocean) integrated over either hemisphere. In simulations where $OET = 0$, all changes in the TOA budget stem from changes in atmospheric energy transport. To illuminate the regions that are important for the change in the energy transport, we rearrange Equation 2 and 3 to read:

$$\int_0^{2\pi} \left( \int_0^{\pi} 2\pi a^2 \cos(\phi) (\Delta SW - \Delta OLR) d\phi - \int_{-\pi}^{0} 2\pi a^2 \cos(\phi) (\Delta SW - \Delta OLR) d\phi \right) d\lambda = -2\Delta TET. \tag{4}$$

The imbalance in net radiation between the hemispheres is related to the cross-equatorial energy transport. The poleward limit of the integrals is the poles, so Equation 4 describes the change in the cross-equatorial energy transport in terms of the change in the net hemispheric TOA energy imbalance. The schematic in Figure 6 translates how a change in the hemispheric imbalance of TOA radiation is related to the cross-equatorial heat transport. A positive change of TOA radiation is defined as an increase in energy absorbed in the atmosphere and a negative change in TOA radiation is a decrease in absorbed energy (an increase in energy emitted by the atmosphere). The hemispheric imbalance is the NH minus SH net (ie, LW+SW) TOA radiation – hereafter called NH-SH $\Delta TOA$ – where positive indicates that there is more radiation being absorbed into the NH than into the SH. If the average NH-SH $\Delta TOA$ is positive, then more energy has been is absorbed in the NH atmosphere than the SH; hence, the change in energy transport across the equator is negative (southward). Conversely, if the average NH-SH $\Delta TOA$ is negative, then there is more of an increase in energy absorbed in the SH atmosphere than in the NH, and there is an accompanying northward change in energy transport across the equator.

The average NH-SH $\Delta TOA$ is positive due to the addition of Andes topography, indicating that there is greater radiation absorbed in the NH than in the SH. Thus, the change
of energy transport across the equator in the atmosphere is southward (Figure 5). Figure 7 shows how each latitude ($\phi$) contributes to the hemispheric average energy imbalance; that is, the change in hemispheric differences in each of the radiative terms $\Delta SW(\phi) - \Delta SW(-\phi)$, etc. Compared to the aquaplanet experiment, in the tropics adding topography causes more OLR to be emitted in the SH than in the NH and more SW to be absorbed in the SH than in the NH. The change in the absorbed SW compensates for most of the change in OLR. The change in the hemispheric imbalance of OLR emitted is greater than the change in the hemispheric imbalance in SW absorbed; that is, the change in the imbalance of OLR contributes slightly more to the anomalous southward heat transport. Within 7 degrees of the equator, the change in SW radiation and OLR is due to the ITCZ shift. The shift of high, thick clouds into the NH leads to a decrease in SW radiation and an increase in OLR within the NH (and vice versa in the SH). The change in Hadley circulation also decreases the clear sky OLR in the NH. The sum of these terms leads to a small but positive total NH-SH $\Delta$TOA near the equator (Figure 7a). NH-SH $\Delta$TOA is largest between 5 and 20 degrees latitude. OLR is contributes more to the total $\Delta$TOA on the equatorward side of this region, while SW contributes more of the poleward side.

If the change in TOA net radiation is instead split into clear and cloud sky components, then we see that net cloud and clear sky radiation contribute to the southward change in the cross-equatorial energy transport, though in different regions (Figure 7b). At the equator, the cloud-only components of SW and LW (red and green lines, respectively), instead show that the radiative effect of clouds there is to decrease NH-SH $\Delta$TOA in agreement with previous studies (Kang et al. 2008, 2014a; Seo et al. 2014). The positive clear-sky radiation changes are important for maintaining the interhemispheric asymmetry in $\Delta$TOA; decreased humidity in the SH deep tropics occurs where there is anomalous subsidence from the shifted ITCZ. SW+LW cloud radiation is positive from 5-20° (black line) where the contribution from clear sky component (blue line) is smaller; at these latitudes in the SH, adding a mountain range has increased the cloud fraction in the low stratus deck, which reduces
the absorbed SW insolation. Figure 2a shows that high clouds in the SH (NH) have also decreased (increased) which results in more (less) OLR. This decrease in high cloud is a result of changes in the Hadley circulation at those longitudes. In the NH (SH), there is less (more) subsidence outside the deep tropics due to the weaker (stronger) overturning caused by pushing the ITCZ off the equator (Lindzen and Hou 1988); high clouds are then either anomalously enhanced or inhibited depending on the anomalous subsidence. Together, outside the deep tropics, increased SW reflection from SH low cloud along with increased (decreased) SH (NH) OLR from changes in high clouds decrease the energy absorbed in the SH relative to the NH.

An analysis of the changes from adding topography in the MMC, stationary eddy, and transient circulation terms of the MSE transport can be done, similar to the analysis of moisture flux convergence in the previous section. In calculating the stationary wave component of MSE transport ($\int \nu^* M \text{SSE}^e \, d\tau$), however, we find that the major contribution is directly over the range itself. Calculating the stationary wave component without including the region in the immediate vicinity of the mountain range would give a better estimate of the change to the stationary wave energy transport elsewhere, despite removing a region that is essential for the energy budget to remain closed. Calculating $\int \nu^* M \text{SSE}^e \, d\tau$, while ignoring the contribution over the mountain range itself, shows that the contributions to the stationary eddy transport is small in comparison to changes in energy transport by the MMC, similar to the results for moisture flux convergence.

Adding topography decreases the global mean surface temperature when compared with the equivalent flat run (Table 2, Figure 2b). This effect occurs even when discounting the small amount of surface that is elevated by the topography. The increase in stratus clouds is one possible explanation, and the decrease in SH high clouds also could be involved. Most of the cooling occurs between 30S and 0 north, and where there is also the largest increase (decrease) in low (high) clouds.

In simulations that add topography on top of a defined q-flux, there is also a decrease
in the global mean temperature, with the exception of the T+IQ30 experiment. In these experiments, there is a large surface cooling co-located with an increase in low clouds in SH subtropics to the west of the added topography, but other far-field regions of the globe have also significantly cooled due to the addition of topography.

4. Removal of wind-evaporation feedback in a simulation with a SH mountain range

Hemispheric asymmetry in the boundary conditions can cause one hemisphere to have greater tropical precipitation than the other. In the previous section, this asymmetry is due to the SH mountain range, which generates a regional circulation that extends westward, far beyond the equatorial radius of deformation. Often the wind-evaporation-SST (WES) mechanism is cited as the mechanism for the westward extension of a circulation perturbation. Here, we test if a WES feedback is important for the large scale response in these model simulations by artificially excluding the impact of wind changes on evaporation. We are testing what are the effects of having a wind feedback in the surface energy budget. If there is no change in the radiative terms, then in steady state, evaporation need not change either (Xie 1996).

The expression for evaporation used in the model ($E$) is,

$$E = C_q |U|(q_{surf} - q_{atm})$$

where $C_q$ is the drag coefficient of moisture determined by Monin-Obukhov drag theory, $|U|$ is the absolute magnitude of the wind in the lowest sigma layer of the atmosphere, $q_{surf}$ is the saturated specific humidity at the surface temperature, and $q_{atm}$ is the specific humidity in the first layer of atmosphere. $|U|$ is calculated using the surface wind magnitude and includes a gustiness parameter to account for unresolved wind speed variability; in all simulations here this gustiness parameter is zero. In simulations with the WES mechanism
removed (the NoWES experiments), $|U|$ is prescribed as a function of latitude in Equation 5 and is taken from the flat aquaplanet simulation and is symmetrized about the equator. This prescribed $|U|$ is not used in the sensible heat or surface momentum flux equations. The response of precipitation to removing the dependence of evaporation on wind is small in a flat aquaplanet. Results comparing the flat NoWES simulations with the aquaplanet simulation with WES are discussed in the Appendix.

Adding topography shifts the ITCZ northward with or without the wind-evaporation feedback. The difference between the precipitation change from adding topography in a NoWES framework and the precipitation change from adding topography “with WES” is shown in Figure 8a ($(P_{T+NoWES} - P_{NoWES}) - (P_T - P_A)$). This quantity is the difference in the impact of adding mountains because of mountain induced changes in the WES feedback. If WES is important the difference will be large; if turning off the wind-evaporation feedback is completely unimportant to the regional precipitation then the differences in Figure 8a will be near zero. Figure 8a shows that this difference is small in comparison to the precipitation change from adding topography with the normal evaporation parameterization (compare Figure 8a with Figure 3a). In fact, without changes in the WES feedback, the mountains cause the precipitation to shift slightly more into the NH, although the difference is very small.

The change in the SST pattern (Figure 8b) that is due to WES feedback is also small compared to changes due to the addition of topography (compare Figures 8b and 2b). Similarly the cloud changes due to the orography with WES feedbacks are very similar to the cloud changes without WES feedback. Hence, it is not surprising that the changes in SST due to the presence of mountains is not sensitive to whether or not there are WES feedbacks. With or without WES feedbacks allowed, the addition of mountains shifts the ITCZ northwards. Similar results are found in Kang and Held (2012) and Kang et al. (2014a) where the WES feedback is also suppressed in AM2 simulations that look at the impact of various q-fluxes; in these studies, both the zonal-mean structure (Kang and Held 2012)
and the zonal structure (Kang et al. 2014a) of tropical precipitation are also found to be largely independent of the WES feedback. Other studies (e.g. Mahajan et al. (2011); Li and Xie (2014)), however, find a large role for the WES feedback. In particular, Mahajan et al. (2011) turn off the WES feedback and find that it is important for amplifying the tropical SST anomaly that results from a high latitude cooling.

5. The climate response to the addition of both topography and surface heat fluxes

In the experiments described in the previous sections, there were no prescribed q-fluxes. In this section, we evaluate the sensitivity of the response to the mountains by modifying the q-fluxes to be (a) the zonal average of the real-world q-fluxes (RQ), found to shift the ITCZ northward in Frierson et al. (2013), and (b) the idealized midlatitude q-fluxes (IQ10 and IQ30) used in Kang et al. (2008) and Kang et al. (2009), which shifts the ITCZ southward.

a. Zonal precipitation response in all simulations

The zonally-averaged precipitation for all simulations is presented in Figure 9a. The ITCZ location is computed by interpolating the zonal averaged precipitation field fit using a cubic spline algorithm and finding the maximum.

The ITCZ location over all q-flux and topography simulations varies greatly, swinging from 16°S to 9°N depending mainly on the q-flux. AM2’s response is larger than that seen in the real world in the RQ simulation, as evidenced by the greater displacement of the zonally averaged ITCZ in the RQ experiment (8°N) compared to that observed (4°N, Adler et al. (2003)). In Frierson et al. (2013), it is argued that this is due to the lack of continents, specifically that of the Sahara desert which cools the NH and shifts rainfall southward.

The energetic constraint described in Kang et al. (2009) is valid in these simulations,
even when adding a mountain range. In both models, tropical precipitation varies with
the magnitude of the energy transport across the equator (accomplished by the Hadley
circulation) (Figure 9b). With a hemispheric imbalance in net absorbed radiation, the
Hadley circulation moves energy to the hemisphere with less energy, and the lower branch
of the Hadley circulation responds in turn, moving moisture in the opposite direction. The
ITCZ in AM2 moves approximately 5 degrees per PW of cross-equatorial energy transport
\( R^2 = 0.98 \).

Mountains are expected to affect the zonally-averaged precipitation less since they have
a limited regional response, and the results in Figure 9 confirm this expectation. In general,
q-fluxes used in this study induce a zonal mean response of precipitation that is substantially
larger than the mountain-induced response. A “Pacific” ITCZ location is calculated using
the zonal average within 120° longitude to the west of the mountain range (Figure 9c). In
the region to the west of the mountain range (where the Pacific ocean would be), the ITCZ
is only slightly farther north than the global mean ITCZ location of the same simulation.

b. An energetic analysis of the addition of q-fluxes in simulations with the same topography

Here we define atmospheric compensation as the ratio of cross-equatorial transport of
energy by the atmosphere to the cross-equatorial transport of energy by the ocean (into the
opposite hemisphere),
\[
\frac{AET_{no-Q} - AET_Q}{OET_Q - OET_{no-Q}} = \frac{-\Delta AET}{\Delta OET},
\]
where \( Q \) denotes an experiment with a q-flux and \( no - Q \) denotes the corresponding experi-
ment with no added q-flux and the same topography as the \( Q \) simulation. \( \Delta \) then represents
the change in the q-flux added experiment from the otherwise identical experiment without
the q-flux. The ratio in Equation 6 is plotted in Figure 10a for each of the experiments with
a q-flux. The best fit line for the experiments is in red; overcompensation is indicated by a
slope that is steeper than the one-to-one line (black line), while a shallower slope indicates
that the atmosphere undercompensates.

Expanding upon this definition of compensation illuminates how radiative feedbacks in these AM2 simulations allow so much overcompensation. Total energy transport at the equator must equal the sum of the atmospheric and oceanic energy transport (OET) at the equator (Equation 2). If the atmospheric response perfectly compensates the applied q-flux, then $\Delta OET = -\Delta AET$ and $\Delta TET = 0$. If the atmosphere overcompensates (as in AM2), then $\Delta OET < -\Delta AET$ and $\Delta TET > 0$. That is, the atmospheric response moves more heat out of the hemisphere than the ocean imports into that hemisphere; the difference is due to an increase in the net incoming energy from changes in the TOA energy fluxes.

Figure 10b shows the change in the symmetry of the TOA radiation for all simulations that include a q-flux. The difference in the integrands of Equation 4 are shown (e.g. $\Delta (SW(\phi) - SW(-\phi))$, where $\Delta$ now refers to the change in an experiment with a q-flux from an otherwise identical experiment without the q-flux. The lines in Figure 10b-g have been scaled by the cross-equatorial OET so that the different q-fluxes are compared equally; this scaling also changes the sign of simulations so that they all correspond to a 1 PW energy transport northward across the equator that heats the atmosphere in the NH and cools in the SH. The dashed curves show a scaled $\Delta OET$ for each q-flux. If the local atmospheric response to the imposed q-flux was a local radiative response, then the change in the TOA energy flux would lay exactly on these dashed lines. If atmospheric energy transport compensated for the ocean energy transport locally, then the change in the TOA energetic terms should be zero everywhere (a purely dynamic response).

The integrals of the curves in Figure 10b is proportional to the $-\Delta TET$ in each experiment. If the integral is negative, then the atmosphere is undercompensating; if positive, the atmosphere is overcompensating. It is notable that the TOA radiation imbalance in the extratropics in isolation would lead toward an undercompensating atmosphere: somewhere between a pure radiative and a pure dynamic response. The atmosphere there responds locally to the surface heating by transporting some heat away and by increasing (decreasing)
its OLR in the NH (SH).

Overall however, the change in the NH-SH TOA energy imbalance is positive: more energy is exported from the hemisphere than is imported by the ocean. Water vapor and cloud feedbacks are responsible for the overcompensation. The atmosphere is transporting more energy into the SH than the ocean delivered to the NH, and the tropics and subtropics are the important regions. The surface heatings are scaled here such that the q-fluxes add (remove) heat to the NH (SH) atmosphere from the surface. However, a positive change in the NH-SH net TOA radiative imbalance indicates that the atmosphere transports even more energy from the NH into the SH. More (less) clear sky OLR in the SH (NH) and more (less) LW cloud warming due to changes in the Hadley circulation are part of this energetic response (see Figure 10c-g). With an even greater energy imbalance between the hemispheres, the atmosphere works even harder to transport heat across the equator; hence, the ITCZ shifts farther into the NH.

Figures 10c&d decompose the TOA radiation imbalance into the hemispheric asymmetry in net TOA SW and OLR respectively; Figures 10e&f shows the changes in the hemispheric asymmetry in total TOA radiation (OLR + net TOA SW) in clear and cloud skies respectively. Both net clear sky radiation changes (mostly due to changes in LW water vapor absorption) and LW cloud radiation changes play a role in the overcompensation, but in different latitude bands. In the deep tropics, SW cloud forcing works against positive cloud sky OLR, and the result is that net cloud forcing works against the atmospheric energy transport’s overcompensation; this result is also found in Kang et al. (2008) and Seo et al. (2014). However, clear sky OLR in the tropics acts to amplify the response, unlike total cloud forcing. The response of net clear sky TOA radiation peaks closer to the equator, while changes in TOA cloud sky radiation peaks farther into the subtropics. In the extratropics, a northward ocean heat transport causes an increase in the energy lost to space in clear sky regions in the NH relative to the SH - a loss that is partially compensated for by a gain in energy in cloudy regions.
The different effects of clear and cloud sky radiation on the change in TET sometimes act together and sometimes against each other. A q-flux is added that warms the NH and cools the SH extratropics; this heating is eventually felt in the tropics and convection shifts into the northern hemisphere. Changes in circulation increase the humidity in the NH tropics and decreases it in the SH; the increase in the NH increases the amount of energy trapped in the atmosphere there (the SH clear sky OLR is not as important as the NH clear sky OLR). Less (more) cloud sky OLR is emitted to space in the NH (SH) deep tropics because of greater (fewer) deep clouds. In region with the ITCZ shift, more cloud sky OLR is emitted to space in the SH than in the NH due to lower cloud tops and drier air columns. Here though, SW cloud sky radiation due to reflection more than compensates for the changes in cloud OLR. In the NH subtropics, increased high clouds reduce cloud sky OLR there, while decreased high clouds increase cloud sky OLR in the SH subtropics. The warmer atmosphere in the NH subtropics emits less clear sky OLR because of enhanced water vapor. Changes in stratus decks is minimal in these simulations (in the added Andes simulations, stratus clouds play a larger role). Examining all these changes together in a NH minus SH context shows that there is more absorbed radiation in both cloud and clear sky conditions in the NH tropics than in the SH tropics, which acts to increase the southward transport of energy across the equator by the atmosphere. The tropical feedbacks are sufficiently large that the southward atmospheric energy transport is even larger than the northward oceanic transport that instigates the asymmetry in the atmosphere. The larger cross-equatorial AET then leads to a larger ITCZ shift.

In Figure 10 each simulation has been scaled by the OET forcing, so the atmosphere responded linearly to the magnitude of the q-flux, then all simulations would show the same response. In the extratropics, all simulations do seem to be reacting similarly, even in the RQ simulations (red lines) which have a significantly different meridional q-flux structure. The tropical response is not linear however. The dark purple and magenta lines indicate simulations with the strongest q-flux (IQ30), and yet, when scaled, the peak of the response
is less per PW than simulations with smaller q-fluxes. The width of the ITCZ becomes broader the farther the ITCZ moves from the equator, and the differences in the structure and amplitude of the ITCZ shift clearly contribute to the radiative differences in the cases with the largest forcing.

c. An energetic analysis of the addition of topography in simulations with the same surface heating

Figure 11a shows the difference in zonally-averaged AET of each topography simulation from its flat equivalent versus the difference in the zonally averaged ITCZ latitude. Although all five experiments show a southward AET and a northward displaced ITCZ, only three of the experiments show the same sensitivity of ITCZ location to energy transport that was found in the q-flux experiments (-0.19 PW/degree latitude). The other two experiments, and particularly for the case of real heat transport, suggest that the details of the distribution of ocean heat transport affects the atmospheric response to topography. Panels 11b-f show the same quantities as in Figure 10, but for the impact of adding topography. For example, if the topography simulation also includes a q-flux, then its climate response is subtracted with the flat simulation that has the same q-flux, leaving only the effect of the added topography. The differences shown in Figure 11 are $T+\text{IQ30}$ minus $\text{IQ30}$ (purple), $T+\text{IQ10}$ minus $\text{IQ10}$ (blue), $T+\text{NoWES}$ minus $\text{NoWES}$ (yellow), $T$ minus $A$ (orange), $T+\text{RQ}$ minus $\text{RQ}$ (dark red).

As discussed previously, examining the hemispheric imbalance of TOA radiation illuminates the latitudes that contribute most to the cross-equatorial energy transport; since the latitudinal shift in the ITCZ location due to topography is also proportional to cross-equatorial AET (Figure 9b), the decomposition of the terms that contribute to the changes in the TOA net energy flux can be used to interpret the causes of the ITCZ shift. Figure 11 shows this difference for the sets of simulations that add topography while holding q-fluxes constant. The TOA imbalance has a similar pattern for those simulations with small q-fluxes.
that do not shift the upwelling branch of Hadley circulation too far from the equator. With
the two largest q-fluxes (IQ30 and RQ), on the other hand, the TOA patterns induced by to-
pography are significantly different. The IQ30 q-flux changes the mean climate in such a way
that adding topography does not change the climate significantly. With the RQ q-flux, the
ITCZ was already well in the NH before topography was added, so adding topography did
not largely change the energy transport by the Hadley circulation across the equator. The
changes in SW and OLR compensate from adding topography with the RQ q-flux. A large
difference from adding topography to the RQ q-flux experiment is a larger cold subtropical
region to the west of the Andes; here, clear sky OLR has increased, but cloud sky radiation
compensates. For the smaller q-flux experiments, the locations of the radiative response to
the addition of topography are similar to the same response in sets of simulations that add
a q-flux and hold topography constant (Figure 10).

Note that the y-axis of Figure 10 has four times the range of those in Figure 11b-f,
but that zonal average TOA changes in Figure 10 are comparable to some of the regional
changes in TOA radiation as seen in Figure 5c. Mountains in AM2 cause a strong local TOA
response that is less noticeable in a zonal mean framework. The pattern of tropical clear
and cloudy sky changes due to the insertion of mountains is similar to those due to q-flux
forcing though weaker in amplitude. Compared to the impact of modest (IQ10) or more
realistic amplitude q-flux (RQ, IQ30) the impact of topography on the position of the zonal
ITCZ or the cross-equatorial energy transport is smaller.

6. Discussion and Conclusions

This study has examined the relative importance of cross-equatorial ocean heat flux and
Andes-like topography on the location of the zonally averaged atmospheric energy transport
and the attendant latitude of the ITCZ. The addition of a q-flux with a cross-equatorial
ocean energy transport results in large-scale changes to the atmospheric Hadley circulation.
Adjustment of the tropical atmosphere responds to amplify the effects of the original q-flux; this response produces an overcompensation of the atmosphere in this aquaplanet version of the AM2 GCM. This response to the q-flux occurs regardless of the exact structure of the q-flux itself, and with and without the addition of an Andes mountain range. This overcompensation is likely to be robust to the details of ocean circulation.

Adding an Andes-like mountain range results in a small northward shift in tropical precipitation. Although the extent of the displacement is sensitive to the details of the tropical ocean heat transport, in all cases, the changes in the precipitation are related to changes in the location of the ascending branch of the Hadley circulation. The result that tropical precipitation shifts northward is robust with or without wind-evaporation feedback included in the evaporation parameterization. The westward propagation of the northward precipitation displacement is not caused by the WES mechanism in this model.

Adding extra surface heat into one hemisphere shifts precipitation toward that hemisphere. With a mountain range and q-flux, the effects on the ITCZ are roughly additive. The addition of a realistic q-flux has a greater effect on the location of the ITCZ than does the addition of an Andes-like mountain range. Locally, the changes in the various components of the TOA radiation fluxes (in both cloudy and clear skies) due to an added Andes range are similar to the analogous TOA changes from an added q-flux albeit much smaller in amplitude. In the zonal mean however, adding a q-flux results in a greater shift in tropical precipitation than adding a SH mountain range does. Additional simulations that doubled and quadrupled the width of an idealized Andes did not result in a greater cross-equatorial energy transport or northward displacement of tropical precipitation. The changes in the MMC and stationary eddy components of the moisture and MSE transports also did not change much with increasing width. This result suggests that an Andes range affects the circulation of the tropics most through changing the Hadley circulation, rather than through the creation of stationary eddy circulations (as is seen in the extratropics).

The changes in the stratus cloud deck with an Andes mountain range reflects insolation
and cools the surface, making the SH less conducive for deep convection. In the energetic budget, we find that the increased SW reflection in the SH subtropics is partly responsible for the interhemispheric TOA radiative imbalance, which is in turn linked with the cross-equatorial AET and northward displacement of precipitation. In the simulations that add Andes topography, stratus clouds are created only by changing the pattern of subsidence due to the Andes, and without any representation of cold coastal upwelling or related ocean advection along the shore of South America. This result suggests a more secondary role for ocean advection in maintaining these low clouds. In the coupled ocean-atmosphere version of the AM2 GCM, this stratus cloud deck is not well reproduced and coupled feedbacks strongly amplify cloud deficiencies that already existed in the atmosphere-only GCM (Wittenberg et al. 2006). More work is necessary to determine the relative importance of ocean dynamics and large scale subsidence on south Pacific marine stratocumulus.

Changes in the high clouds and moisture due to the shift in the Hadley circulation are important in the simulations that add both topography and q-fluxes. The LW cloud and clear sky feedbacks are important components of the interhemispheric energy imbalance and corresponding shift of tropical precipitation. In the simulations with q-fluxes, cloud and clear sky feedbacks are responsible for the overcompensation of the AM2 atmosphere to a given q-flux.

Simulations with the same q-fluxes and topography described here have also been completed with the GFDL gray radiation model, GRaM (Frierson et al. 2006). GRaM lacks cloud and clear sky radiative feedbacks, and given the results here, it is not surprising that the atmosphere undercompensates for a given forcing (see also, Seo et al. (2014)). Aside from the smaller magnitude of climate response, the results were similar to those in the AM2 simulations here. Adding Andes topography shifts precipitation northward, and the precipitation shifts more for a large q-flux than for added topography. In these simulations, the dependence of evaporation on wind was also removed, and this removal did not significantly affect the change in precipitation (Maroon 2013). In the GRaM simulation with added
Andes, the northward shift of precipitation occurred within about 40° degrees longitude of
the Andes (compare to an extent of about 120° in the AM2 simulations); cloud and clear
sky radiative feedbacks, which are present in the AM2 model, may be responsible for the
westward extent.

The results from these aquaplanet simulations lend support to recent research showing
that the cross-equatorial ocean heat transport is important for the location of the zonally-
averaged ITCZ, and that this result is insensitive to the details of the ocean transport. Using
zonally averaged ocean heat flux derived from observations is more than sufficient to put the
ITCZ in the correct hemisphere. The addition of an Andes range creates zonal variation in
precipitation. The location of the ITCZ is modulated by both local and remote effects, and
neither should be neglected when working to understand its dynamics.

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The response of removing wind-evaporation feedback in a flat aquaplanet

Removing the WES mechanism in an aquaplanet simulation does not change the mean climatology in a significant way (NoWESminusA). Comparing the flat simulations with and without WES (A and NoWES simulations, respectively), there is little change in the zonal-mean wind strength felt by evaporation in each simulation (Figure 12a - dark green line). The wind difference between the simulations with topography and WES (orange) shows that the wind felt by evaporation would otherwise vary much more than it is prescribed. When comparing how evaporation changes in simulations with topography and with or without WES (Figure 12b-c), we see that adding topography results in a similar change in evaporation from the flat simulation with WES regardless of whether WES is present or not. The changes in evaporation are almost an order of magnitude smaller than the scale of tropical precipitation changes.

The response of precipitation in a flat aquaplanet to turning off a WES feedback in these simulations is small (Figure 12b). There is a small southward shift of precipitation when turning off WES [The aquaplanet simulation with WES in AM2 contained a tiny amount of random asymmetry: the ITCZ peaked slightly in the NH tropics (see Figure 3), and symmetrizing the wind strength in the evaporation parameterization likely helped to symmetrize the climate state].
REFERENCES


Kang, S. M., I. M. Held, D. M. W. Frierson, and M. Zhao, 2008: The Response of the ITCZ


1 Description of experiments performed. 32
2 Global mean surface temperature (K) for each simulation. The means are expressed as deviations from the aquaplanet simulation for all other simulations. The mean temperatures for latitudes from 90S-30S, 30S-0, 0-30N, and 30N-90N are also included. The deviation from the mean of each simulation in these latitude bands is from the aquaplanet mean in those same latitudes. The standard deviation of the aquaplanet simulation is 0.17K in the extratropics and 0.14K in the tropics. Standard deviations of other simulations are similar; assuming a simulation has the same standard deviation with the same number of degrees of freedom as the aquaplanet simulation, then changes greater than 0.15K and 0.12K are significant at the 95% confidence level in the extratropics and tropics, respectively. Changes in the mean temperature from that of the aquaplanet are italicized if significant at the 95% confidence interval. Changes in the mean temperature of a topography simulation from its flat counterpart are bolded if significant at the 95% confidence interval. 33
Table 1. Description of experiments performed.

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tr>
<td>A</td>
<td>Aquaplanet only</td>
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<tr>
<td>T</td>
<td>Realistic Andes topography</td>
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<tr>
<td>NoWES</td>
<td>Flat aquaplanet, no WES feedback</td>
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<tr>
<td>T+NoWES</td>
<td>Realistic Andes topography, no WES feedback</td>
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<tr>
<td>IQ10</td>
<td>Kang et al. (2008) q-flux, amplitude of 10 W m(^{-2})</td>
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<tr>
<td>IQ30</td>
<td>Kang et al. (2008) q-flux, amplitude of 30 W m(^{-2})</td>
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<tr>
<td>RQ</td>
<td>Real-world zonally and annually averaged q-flux</td>
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<tr>
<td>T+IQ10</td>
<td>Realistic Andes topography + Kang et al. (2008), amplitude of 10 W m(^{-2})</td>
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<tr>
<td>T+IQ30</td>
<td>Realistic Andes topography + Kang et al. (2008) q-flux, amplitude of 30 W m(^{-2})</td>
</tr>
<tr>
<td>T+RQ</td>
<td>Realistic Andes topography + real zonal q-flux</td>
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Table 2. Global mean surface temperature (K) for each simulation. The means are expressed as deviations from the aquaplanet simulation for all other simulations. The mean temperatures for latitudes from 90S-30S, 30S-0, 0-30N, and 30N-90N are also included. The deviation from the mean of each simulation in these latitude bands is from the aquaplanet mean in those same latitudes. The standard deviation of the aquaplanet simulation is 0.17K in the extratropics and 0.14K in the tropics. Standard deviations of other simulations are similar; assuming a simulation has the same standard deviation with the same number of degrees of freedom as the aquaplanet simulation, then changes greater than 0.15K and 0.12K are significant at the 95% confidence level in the extratropics and tropics, respectively. Changes in the mean temperature from that of the aquaplanet are italicized if significant at the 95% confidence interval. Changes in the mean temperature of a topography simulation from its flat counterpart are bolded if significant at the 95% confidence interval.

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<td>+0.0 K</td>
<td>+3.0 K</td>
<td>+6.0 K</td>
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List of Figures

1 Topography and surface heat fluxes used.  

2 Changes in streamfunction, SST, clouds, and evaporation. Panel a shows changes in SST (shading), surface wind (vectors) and high cloud amount (gray contours) from adding mountains. Vectors represent the change in the winds of the 0.996 (bottom) sigma level. Panel b shows the change in evaporation (shading) and low cloud amount (contours). The gray solid (dashed) contours in panels a and b indicate an increase (decrease) in high and low cloud amount, respectively; contour intervals are every 10% and the zero contour is not shown. Black contours indicates topography higher than 1000m and the red lines show the 10 mm day$^{-1}$ contour of precipitation in the RT simulation and denotes the ITCZ location. Panel c shows the changes of streamfunction (contours) and relative humidity (shading) from adding mountains. Streamlines are displayed every $2 \times 10^{10}$ kg s$^{-1}$, starting at $1 \times 10^{10}$ kg s$^{-1}$. Solid contours indicate clockwise motion while dashed contours indicated counterclockwise motion. For comparison, the maximum Hadley cell streamfunction in the mean climate is approximately $2.5 \times 10^{11}$ kg s$^{-1}$.  

3 Change in precipitation from adding topography (panel a). Black contours indicate topography higher than 1000m. Zonally averaged precipitation (solid lines) and evaporation (dashed lines) are in the aquaplanet (green lines) and Andes topography (orange lines) simulations in panel b; the difference in precipitation and evaporation from adding topography is plotted in panel c. Note that the latitude axis of panels b and c are area-weighted (sine of latitude).
Decomposition of the moisture flux convergence from adding topography. The total moisture flux convergence (black) is decomposed into the part from the MMC (green), the part from the stationary eddies (red), and the part from the transient circulations (blue).

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Schematic showing how changes in TOA absorbed radiation relates to NH-SH changes in radiation and cross-equatorial energy transport.

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The range of tropical precipitation changes. Zonally averaged precipitation in all AM2 experiments is in panel a. Note that the x-axis is area-weighted. The latitudes of the zonally averaged ITCZ versus cross-equatorial atmospheric energy transport (AET) is shown in panel b. The best fit line has a slope of -0.19 PW/degree ($R^2 = 0.98$). The zonally averaged ITCZ location versus Pacific ITCZ location is shown in panel c; the one-to-one line is dashed and denotes if the location of the Pacific ITCZ is identical to the zonally-averaged ITCZ; flat simulations are not included in panel c.

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Change in energy transport and energetic terms in all simulations that include topography. Panel a shows the change in atmospheric energy transport versus the change in zonally averaged ITCZ latitudes in simulations that include topography from their equivalent flat simulations. Differences shown are $T+IQ30$ minus IQ30 (purple), $T+IQ10$ minus IQ10 (blue), $T+NoWES$ minus NoWES (yellow), $T$ minus A (orange), and $T+RQ$ minus RQ (dark red). The NH-SH changes in TOA radiation budget for these simulations are shown in panels b-f. Panel b shows the change in the hemispheric asymmetry in the TOA radiation balance difference for all simulations that include a SH mountain range. Panels c-f are as in panel b, but for panel c absorbed SW (positive downward, energy increases in the atmosphere), panel d OLR (positive upward, energy leaving the atmosphere), panel e TOA clear sky radiation, and panel f TOA cloud sky radiation. Positive indicates that there is a greater change of energy at that latitude in the NH atmosphere, while negative indicates that there is greater energy at that latitude in the SH.

Zonal mean changes in wind, precipitation, and evaporation in experiments with and without wind-evaporation feedback. The changes in the wind fed into the evaporation parameterization are in panel a. Changes in precipitation are in panel b, and changes in evaporation are in panel c. Differences of NoWES minus A (dark green), $T+NoWES$ minus $T$ (light green), $T+NoWES$ minus NoWES (yellow), and $T$ minus A (orange) are shown.
Fig. 1. Topography and surface heat fluxes used in this study. a. Realistic Andes topography with contours every 1000m. b. Zonal q-fluxes used in this study. Blue and magenta lines show the surface heat flux described in Kang et al. (2008) with peak amplitudes of 10 and 30 W m$^{-2}$ at 65° latitude, respectively. The red dashed line is the zonally-averaged surface heat flux as derived from CERES TOA energy and ERA-Interim reanalysis. c. The implied northward ocean energy transport for the surface heat fluxes in panel b. Note that latitude axis of panels b and c is area-weighted (sine of latitude).
Fig. 2. Changes in streamfunction, SST, clouds, and evaporation. Panel a shows changes in SST (shading), surface wind (vectors) and high cloud amount (gray contours) from adding mountains. Vectors represent the change in the winds of the 0.996 (bottom) sigma level. Panel b shows the change in evaporation (shading) and low cloud amount (contours). The gray solid (dashed) contours in panels a and b indicate an increase (decrease) in high and low cloud amount, respectively; contour intervals are every 10% and the zero contour is not shown. Black contours indicates topography higher than 1000m and the red lines show the 10 mm day\(^{-1}\) contour of precipitation in the RT simulation and denotes the ITCZ location. Panel c shows the changes of streamfunction (contours) and relative humidity (shading) from adding mountains. Streamlines are displayed every \(2 \times 10^{10}\) kg s\(^{-1}\), starting at \(1 \times 10^{10}\) kg s\(^{-1}\). Solid contours indicate clockwise motion while dashed contours indicated counterclockwise motion. For comparison, the maximum Hadley cell streamfunction in the mean climate is approximately \(2.5 \times 10^{11}\) kg s\(^{-1}\).
Fig. 3. Change in precipitation from adding topography (panel a). Black contours indicate topography higher than 1000m. Zonally averaged precipitation (solid lines) and evaporation (dashed lines) are in the aquaplanet (green lines) and Andes topography (orange lines) simulations in panel b; the difference in precipitation and evaporation from adding topography is plotted in panel c. Note that the latitude axis of panels b and c are area-weighted (sine of latitude).
Fig. 4. Decomposition of the moisture flux convergence from adding topography. The total moisture flux convergence (black) is decomposed into the part from the MMC (green), the part from the stationary eddies (red), and the part from the transient circulations (blue).
Fig. 5. The change in energy transport from adding topography. Panel a shows the change in total energy transport, while panel b shows its division into LH and DSE transport. Panel c shows the change in TOA net radiation across the globe (positive downward).
More energy absorbed into the NH atmosphere
Less energy absorbed into the SH atmosphere
Area under curve is proportional to the cross-equatorial energy transport

Fig. 6. Schematic showing how changes in TOA absorbed radiation relates to NH-SH changes in radiation and cross-equatorial energy transport.
Fig. 7. NH-SH TOA energetic analysis. Panel a shows the change in the TOA radiation asymmetry at each latitude from the equator due to the addition of the Andes: net radiation (black), SW (red), and OLR (blue). Panel b shows the decomposition of the change from adding topography of the NH-SH net radiation difference into changes in SW+OLR clear sky TOA radiation (blue), SW+OLR cloud sky radiation (black), OLR cloud sky radiation (green), and SW cloud sky TOA radiation (red). Positive indicates that there is a greater input of energy at that latitude in the NH atmosphere than at the same latitude in the SH.
Fig. 8. The impact of WES in setting the response to realistic topography. Shown is the difference of the change from adding topography in a NoWES framework and the change from adding topography in a “with WES” framework. Panel a shows precipitation (shaded) \((P_{T+NoWES} - P_{NoWES}) - (P_T - P_A)\). Panel b shows SST (shading) and wind (vectors). Vectors are the winds in the 0.996 sigma level (near surface). Black contours indicates topography higher than 1000m.
Fig. 9. The range of tropical precipitation changes. Zonally averaged precipitation in all AM2 experiments is in panel a. Note that the x-axis is area-weighted. The latitudes of the zonally averaged ITCZ versus cross-equatorial atmospheric energy transport (AET) is shown in panel b. The best fit line has a slope of -0.19 PW/degree ($R^2 = 0.98$). The zonally averaged ITCZ location versus Pacific ITCZ location is shown in panel c; the one-to-one line is dashed and denotes if the location of the Pacific ITCZ is identical to the zonally-averaged ITCZ; flat simulations are not included in panel c.
Fig. 10. Compensation of atmospheric energy transport to the implied ocean energy transport at the equator (panel a) and the scaled changes in the NH-SH energetics (panels b-g; positive downward). In panel a, the black line shows where the simulations of a perfectly compensating model would lie, while the red line with a steeper slope indicates the overcompensation of atmosphere. Colors refer to the same simulations as in Figure 9. The symbols for the T+IQ30 and IQ30 cases are co-located, as are the T+IQ10/IQ10 and the T+RQ/RQ cases. All values in panels b-g have been scaled by the cross-equatorial ocean energy transport.
Fig. 11. Change in energy transport and energetic terms in all simulations that include topography. Panel a shows the change in atmospheric energy transport versus the change in zonally averaged ITCZ latitudes in simulations that include topography from their equivalent flat simulations. Differences shown are $T+IQ30$ minus $IQ30$ (purple), $T+IQ10$ minus $IQ10$ (blue), $T+NoWES$ minus $NoWES$ (yellow), $T$ minus $A$ (orange), and $T+RQ$ minus $RQ$ (dark red). The NH-SH changes in TOA radiation budget for these simulations are shown in panels b-f. Panel b shows the change in the hemispheric asymmetry in the TOA radiation balance difference for all simulations that include a SH mountain range. Panels c-f are as in panel b, but for panel c absorbed SW (positive downward, energy increases in the atmosphere), panel d OLR (positive upward, energy leaving the atmosphere), panel e TOA clear sky radiation, and panel f TOA cloud sky radiation. Positive indicates that there is a greater change of energy at that latitude in the NH atmosphere, while negative indicates that there is greater energy at that latitude in the SH.
Fig. 12. Zonal mean changes in wind, precipitation, and evaporation in experiments with and without wind-evaporation feedback. The changes in the wind fed into the evaporation parameterization are in panel a. Changes in precipitation are in panel b, and changes in evaporation are in panel c. Differences of NoWES minus A (dark green), T+NoWES minus T (light green), T+NoWES minus NoWES (yellow), and T minus A (orange) are shown.