Characterizing the Hadley Circulation Response through Regional Climate Feedbacks

N. FELDL AND S. BORDONI

California Institute of Technology, Pasadena, California

(Manuscript received 17 June 2015, in final form 9 October 2015)

ABSTRACT

The robust weakening of the tropical atmospheric circulation in projections of anthropogenic warming is associated with substantial changes in regional and global climate. The present study focuses on understanding the response of the annual-mean Hadley circulation from a perspective of interactions between climate feedbacks and tropical circulation. Simulations from an ensemble of coupled ocean–atmosphere models are used to quantify changes in Hadley cell strength in terms of feedbacks, radiative forcing, ocean heat uptake, atmospheric eddies, and gross moist stability. Climate feedbacks are calculated for the model integrations from phase 5 of CMIP (CMIP5) using radiative kernels. Tropical mean circulation is found to be reduced by up to 2.6% K^{-1} for an abrupt quadrupling of carbon dioxide concentration. The weakening is characterized by an increase in gross moist stability, by an increase in eddy heat flux, and by positive extratropical feedbacks, such as those associated with lapse rate and sea ice response. Understanding the impact of radiative feedbacks on the large-scale circulation provides a framework for constraining uncertainty in the dynamic climate response, including the hydrological cycle.

1. Introduction

The weakening of the Hadley cell is one of the robust responses of global climate models to a warming Earth (Held and Soden 2006; Lu et al. 2007; Vecchi and Soden 2007). It is often attributed to an increase in atmospheric stability (Knutson and Manabe 1995) or follows as a consequence of the water vapor budget (Held and Soden 2006; Schneider et al. 2010). Since the strength of the Hadley cell is constrained by the angular momentum balance, large-scale atmospheric eddies may also modify the circulation response (Levine and Schneider 2011). While the literature agrees on the sign of the change, the magnitude of the slowdown is not well constrained, ranging from 0%–4% K^{-1} of global-mean surface temperature change (Lu et al. 2007). This spread leads to uncertainty in projections of regional climate change and extremes, which depend on the large-scale circulation.

The general circulation transports heat, moisture, and momentum and as such defines the large-scale atmospheric state. In the annual mean, atmospheric heat transport, including contributions from the Hadley circulation and stationary and transient eddies, is dictated by the top-of-atmosphere (TOA) radiative flux, and in particular by its meridional gradient; a surplus of absorbed solar radiation in the tropics and a deficit at high latitudes leads to a poleward horizontal flux of energy. Perturbations to the energy balance caused directly and indirectly by increasing greenhouse gas concentrations affect the heat flux between latitudes, which in the tropics manifests as changes in the intensity and structure of the Hadley cell (Kang et al. 2008). In other words, smaller-scale processes embedded in the circulation, such as clouds, produce heating and cooling via phase changes, interactions with solar and infrared radiation, and turbulence that in turn drive changes in the circulation itself. We propose that the formal framework of climate feedbacks may offer insights into quantifying changes in circulation strength due to these processes.

A guiding principle in climate dynamics is that the response of the climate system to an external forcing can be partitioned into individual climate feedbacks—physical processes that act to amplify or damp the system response. Such feedbacks have long been invoked to explain past climate change, such as the Pleistocene glaciations (Arrhenius 1896), and, more recently, anthropogenic climate change (e.g., Charney et al. 1979; Hansen et al. 1984; Schlesinger 1985). Intermodel

Corresponding author address: N. Feldl, MC 131-24, 1200 E. California Blvd., California Institute of Technology, Pasadena, CA 91125.
E-mail: feldl@gps.caltech.edu

DOI: 10.1175/JCLI-D-15-0424.1

spread in climate sensitivity[1] is a major factor contributing to the range in projections of anthropogenic climate change, along with uncertainties in future emissions and rates of ocean heat uptake. In particular, feedbacks associated with marine boundary layer clouds are the largest contributor to the uncertainty in sensitivity (Sherwood et al. 2014; Zhao 2014). The importance of these clouds, controlled in part by mixing of the lower troposphere by shallow convection, to the global response hints at the interconnected nature of feedbacks and circulation. It also emphasizes an important point: while climate sensitivity is a global quantity, the processes that produce it are regional, as are the societal impacts.

The limitations of the global framework and a practical push for regional predictability have led to an emerging emphasis on the spatial patterns of climate feedbacks. The net effect of regional feedbacks tends to be more negative in the extratropics relative to the tropics. Following the global paradigm, one might then naively expect the greatest warming in regions of strong positive feedbacks (i.e., tropical amplification), which is not observed. Rather, feedbacks drive both a local and remote response, via atmospheric heat transport, and are a valuable approach for understanding robust features of climate change, such as polar amplification of surface warming (Feldl and Roe 2013b; Roe et al. 2015).

Our motivation is to investigate whether the spatial pattern of climate feedbacks likewise affects atmospheric heat flux in the tropics. We specifically focus on the mean meridional circulation in the tropics, or Hadley cell. Recently, Voigt and Shaw (2015) demonstrated that changes in clouds (especially high, icy clouds in the deep tropics) and water vapor are key to the regional response of tropical circulation in aquaplanet simulations. In the present study, we characterize the contributions to changes in Hadley circulation strength in coupled ocean–atmosphere models using the framework of climate feedbacks. Untangling cause and consequence of the circulation response—and its relationship to the radiative demands imposed by feedbacks—is of fundamental importance to our understanding of climate dynamics and regional impacts and has been identified as one of the grand challenges of climate science by Bony et al. (2015). Even small changes in the Hadley circulation may have a profound impact on the position and variability of the tropical rain belt, Asian monsoon, and subtropical deserts. Furthermore, the width of the subsiding branches of the circulation determines the prevalence of the low clouds so critical to global sensitivity and patterns of temperature change.

[1] Climate sensitivity is defined as the equilibrium global-mean surface temperature change following a doubling of atmospheric CO_2 concentration.

2. Methods

We leverage phase 5 of the Coupled Model Intercomparison Project (CMIP5) to investigate feedback–circulation interactions in a multimodel ensemble of coupled ocean–atmosphere GCMs. Four experiments are selected for the decomposition: 1) a coupled ocean–atmosphere simulation in which the CO_2 concentration is abruptly quadrupled then held fixed (abrupt4$\times$$CO_2$); 2) an atmospheric simulation with SSTs and sea ice prescribed from the preindustrial simulation (sstClim); 3) the same experiment as sstClim, but for quadrupled CO_2 (sstClim4$\times$$CO_2$); and 4) the preindustrial control simulation (piControl) (Taylor et al. 2012). Monthly climatologies are computed from the 30-yr periods of sstClim and sstClim4$\times$$CO_2$, from the last decade of the 150-yr abrupt4$\times$$CO_2$ integration, and from the last 30 years of the 500-yr piControl integration.

Climate anomalies used to derive feedbacks λ_i are calculated as the difference between monthly climatologies of experiments 1 and 3. In sstClim4$\times$$CO_2$, sea surface temperatures are held fixed to preindustrial values, but in abrupt4$\times$$CO_2$, sea surface temperatures evolve and activate feedbacks. In other words, the indirect response to CO_2 is captured, even though the CO_2 concentration is the same in both experiments. The difference between experiments 3 and 2 provides an estimate of the troposphere-adjusted radiative forcing R_f (i.e., the direct effect of enhanced CO_2) (Taylor et al. 2012; Andrews et al. 2012; Feldl and Roe 2013b). Remaining quantities, such as the change in ocean heat uptake, are calculated as the full difference between experiments 1 and 4. The results are qualitatively the same when only the first three experiments are used, with piControl being replaced in the analysis by sstClim. Since CO_2 is quadrupled, rather than doubled, equilibrium climate sensitivity is estimated as $-R_f/2\Sigma_i\lambda_i$ [for a more comprehensive estimate, see Vial et al. (2013)]. Only the 13 models with sufficient output for all four experiments are included in the analysis.

Feedbacks are calculated using the radiative kernel technique (Shell et al. 2008; Soden et al. 2008). Individual feedback parameters (Fig. 1) are the product of the radiative kernel for the relevant variable $K_i = \partial R/\partial x_i$ and the climate change anomaly Δx_i, as described above, normalized by the zonal-mean, seasonal-mean surface air temperature response ΔT_s to give units of watts per square meter per kelvin:

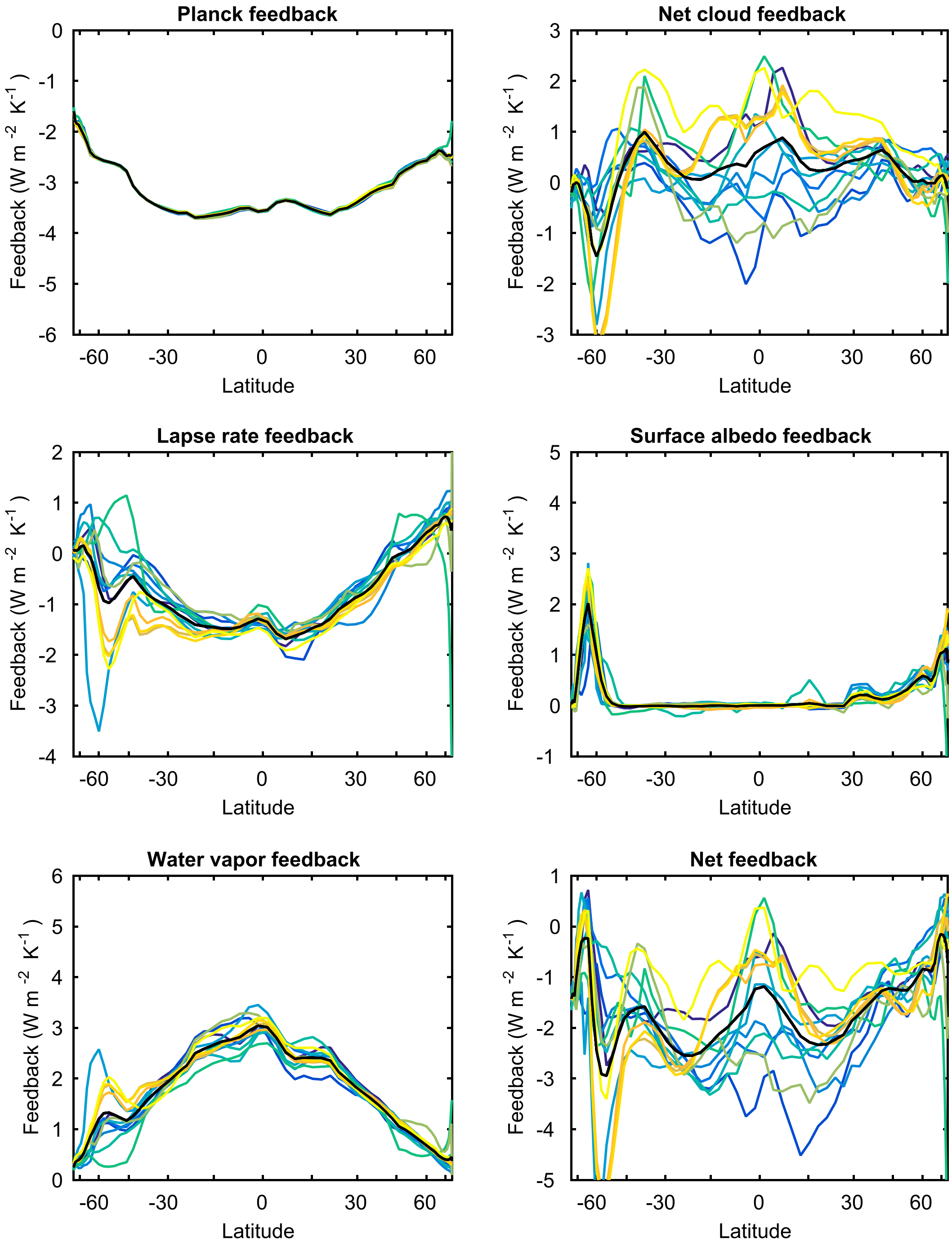

FIG. 1. Annual-mean, zonal-mean regional feedbacks for CMIP5. The multimodel mean is indicated by the black line. (Models are color-coded as in the legend of Fig. 4.)

$$\lambda_i = \frac{\partial R}{\partial x_i}\frac{\Delta x_i}{\Delta T_s}, \tag{1}$$

where R is the net radiative flux at the top of the atmosphere. Note that the conventional approach for global feedbacks is to normalize by the global-mean surface warming, as in Zelinka and Hartmann (2012); however, the regional feedbacks offer a number of advantages where spatial patterns of warming are of interest (Feldl and Roe 2013a; Armour et al. 2013). We consider the Planck, lapse rate, surface albedo, water vapor, and cloud feedbacks.

The Planck feedback is associated with a vertically uniform warming of the surface and troposphere ($x = T_s$), the lapse rate feedback with tropospheric warming that deviates from the vertically uniform profile ($x = T'$), and the surface albedo feedback with changes in surface albedo ($x = \alpha$). For the water vapor feedback, the specific humidity anomaly is given relative to the standard anomaly for a 1-K warming. In addition, since changes in absorptivity are proportional to changes in the logarithm of water vapor concentration (Raval and Ramanathan 1989; Huang and Bani Shahabadi 2014), the relevant variable to the TOA radiative flux is $\ln(q)$. The water vapor feedback becomes

$$\lambda_q = K_q \frac{\Delta \ln(q)}{\delta \ln(\mathcal{H} q_s)} \frac{1}{\Delta T_s} = K_q \frac{\Delta \ln(q)}{\delta \ln(q_s)} \frac{1}{\Delta T_s} \qquad (2)$$

assuming no change in relative humidity $\mathcal{H}$. Saturation specific humidity $q_s(T)$ is calculated from the control (i.e., sstClim4$\times CO_2$) atmospheric temperature field and its 1-K perturbation. Finally, we compute the cloud feedback from the change in cloud radiative effect ΔCRE, with corrections for cloud masking of noncloud feedbacks, following Soden et al. (2008):

$$\lambda_c = \frac{\Delta \mathrm{CRE}}{\Delta T_s} + \sum_i (\lambda_i^0 - \lambda_i), \qquad (3)$$

where clear-sky feedbacks are indicated as λ_i^0 and where ΔCRE is the difference between TOA radiative fluxes in all-sky (i.e., including clouds if present) and clear-sky conditions. The forcing term in Eq. (25) of Soden et al. (2008) is omitted here, as there is no change in CO_2 concentration between experiments 1 and 3. Together, the radiative forcing and the sum of the energy contributions from individual feedbacks balance the change in TOA radiative flux at equilibrium:

$$\Delta R = \sum_i \lambda_i \Delta T_s + R_f, \qquad (4)$$

where each term is a function of latitude.

The total atmospheric energy flux is calculated as the zonal and meridional integral of the atmospheric energy balance (TOA minus surface fluxes) (Zelinka and Hartmann 2012):

$$F_a = \int_{-\pi/2}^{\phi} \int_0^{2\pi} (R - \nabla \cdot F_o) a^2 \cos\phi \, d\lambda \, d\phi, \qquad (5)$$

where ϕ is latitude, λ is longitude, and a is the radius of the earth. We estimate ocean heat uptake $\nabla \cdot F_o$ from the sum of surface radiative, latent, and sensible heat fluxes. The anomalous surface fluxes under quadrupling of CO_2 are negligible over land, and hence $\Delta(\nabla \cdot F_o)$ represents changes in divergence of ocean heat transport as well as ocean heat storage. In the tropics, energy flux by the mean meridional circulation depends on the Hadley cell mass flux $\psi_{\max}$ and the energy flux per unit mass flux, or gross moist stability, H (Fig. 2):

$$F_{\mathrm{HC}} = \psi_{\max} H. \qquad (6)$$

The gross moist stability can alternately be thought of as the effective energy stratification in the tropics (i.e., the difference between moist static energy aloft and near the surface). We calculate it from the vertical integral of energy flux over pressure p

$$H = \frac{2\pi a \cos\phi}{g} \frac{1}{\psi_{\max}} \int_0^{p_s} [\overline{v}][\overline{h}] \, dp, \qquad (7)$$

where g is the gravitational constant and moist static energy $h = C_p T + gz + L_v q$ and zonal-mean meridional wind $[\overline{v}]$ is adjusted to be in mass balance. Though the mean meridional circulation dominates the total atmospheric energy flux F_a in the tropics, stationary and transient eddies may influence the position of the climatological energy flux equator and contribute anomalous transports ΔF_e. For small perturbations, then, fractional changes in energy flux are given by

$$\frac{\Delta F_a - \Delta F_e}{F_{\mathrm{HC}}} = \frac{\Delta \psi_{\max}}{\psi_{\max}} + \frac{\Delta H}{H} \qquad (8)$$

[cf. derivations in Merlis et al. (2013) and Hill et al. (2015)].

Substituting Eqs. (4) and (5) into Eq. (8), we arrive at a characterization of fractional changes in circulation strength:

$$\frac{\Delta \psi_{\max}}{\psi_{\max}} = \frac{\iint \sum_i (\lambda_i \Delta T_s)' + \iint R_f' - \Delta F_o' - \Delta F_e}{F_{\mathrm{HC}}} - \frac{\Delta H}{H}, \qquad (9)$$

where integrals are as in Eq. (5). The primes denote deviations from the global mean; a uniform feedback or forcing does not alter tropical transports. Changes in Hadley cell strength are thus quantified in terms of contributions from feedbacks, radiative forcing, ocean heat uptake, atmospheric eddies, and gross moist stability. Contributions from eddy transport (stationary and transient) are calculated as the difference between ΔF_a and ΔF_{HC}. An advantage to considering fractional changes is that increases and decreases are clearly indicated (i.e., an increase is always positive), because the anomaly is relative to the climatology, which varies

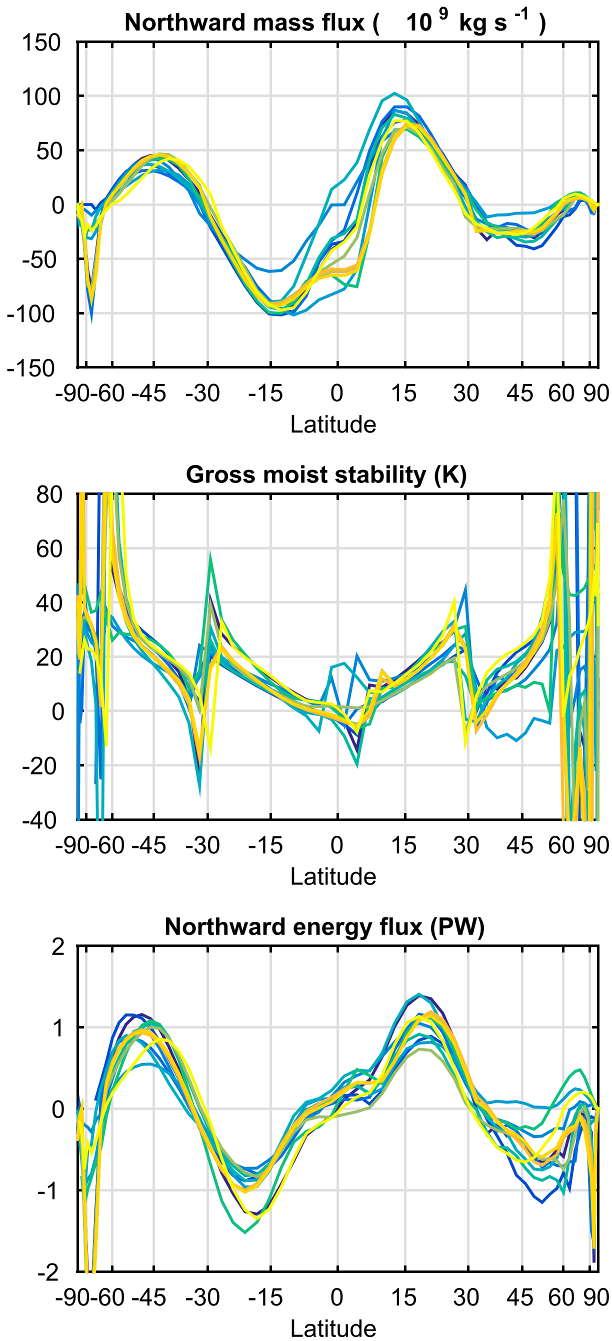

FIG. 2. Climatological mass flux, gross moist stability, and energy flux by the Hadley circulation for the piControl multimodel ensemble. Mass flux is calculated as the signed maximum magnitude of the meridional mass streamfunction. (Models are color-coded as in the legend of Fig. 4.)

among models. A limitation is that the solution is ill defined at the cell edges where ΔF_{HC} goes to 0.

Rather than calculating the implied northward transport by feedbacks and forcing, an alternative decomposition might have begun with the divergence of Eq. (6). However, as shown by Merlis et al. (2013), simplifying assumptions restrict the energy balance to the region of the streamfunction extremum. For a multimodel ensemble, in which the streamfunction maximizes at a range of different latitudes, this leads to a larger residual. In other words, the energy balance closes more neatly for its integral form. In the following section, we perform a detailed breakdown of changes in circulation strength for each model in the ensemble, following Eq. (9).

3. Annual-mean circulation changes

We characterize contributions to changes in the strength of the annual-mean Hadley circulation using the framework of climate feedbacks. The dynamical response to a warming world is consistent with the role of the atmosphere in reducing meridional energy gradients. An increase in atmospheric energy in the deep tropics, for instance, reinforces the existing equator-to-pole gradient of net radiation. This promotes an anomalous poleward transport, which may be accomplished by a strengthening of the tropical circulation. Likewise, an anomalous equatorward transport may be associated with a weakening of the circulation. Hence, it is the spatial distribution of anomalous energy, dictated by feedbacks, radiative forcing, and ocean heat uptake, that control the circulation response.

Figure 3 shows the anomalous northward atmospheric flux implied by the spatial patterns of net feedback, radiative forcing, ocean heat uptake, and atmospheric eddies. The pattern of radiative forcing from CO_2 emissions is determined by variations in cloudiness and surface temperature (Shine and Forster 1999). Forcing peaks in the tropics, resulting in anomalous divergence [i.e., a poleward flux in both hemispheres (as in Huang and Zhang 2014)]. The pattern of ocean heat uptake induces poleward fluxes in the extratropical atmosphere, compensating for downward fluxes into the subpolar oceans and small upward fluxes at low latitudes. However, this term also has a strong cross-equatorial component, associated with anomalous upward surface fluxes in the southern subtropics in some models. The eddy term, including both stationary and transient atmospheric eddies, is characterized by an increase in poleward eddy energy transport with a high degree of intermodel variability. The enhanced eddy transport is consistent with a robust increase in poleward total heat transport (Hwang and Frierson 2010) and the dominant role of eddies in the extratropics (Trenberth and Stepaniak 2003). Finally, the anomalous transport due to climate feedbacks may be either equatorward or poleward, which we return to investigate later in the section.

Fractional changes in annual-mean Hadley circulation strength are shown for each model in the leftmost

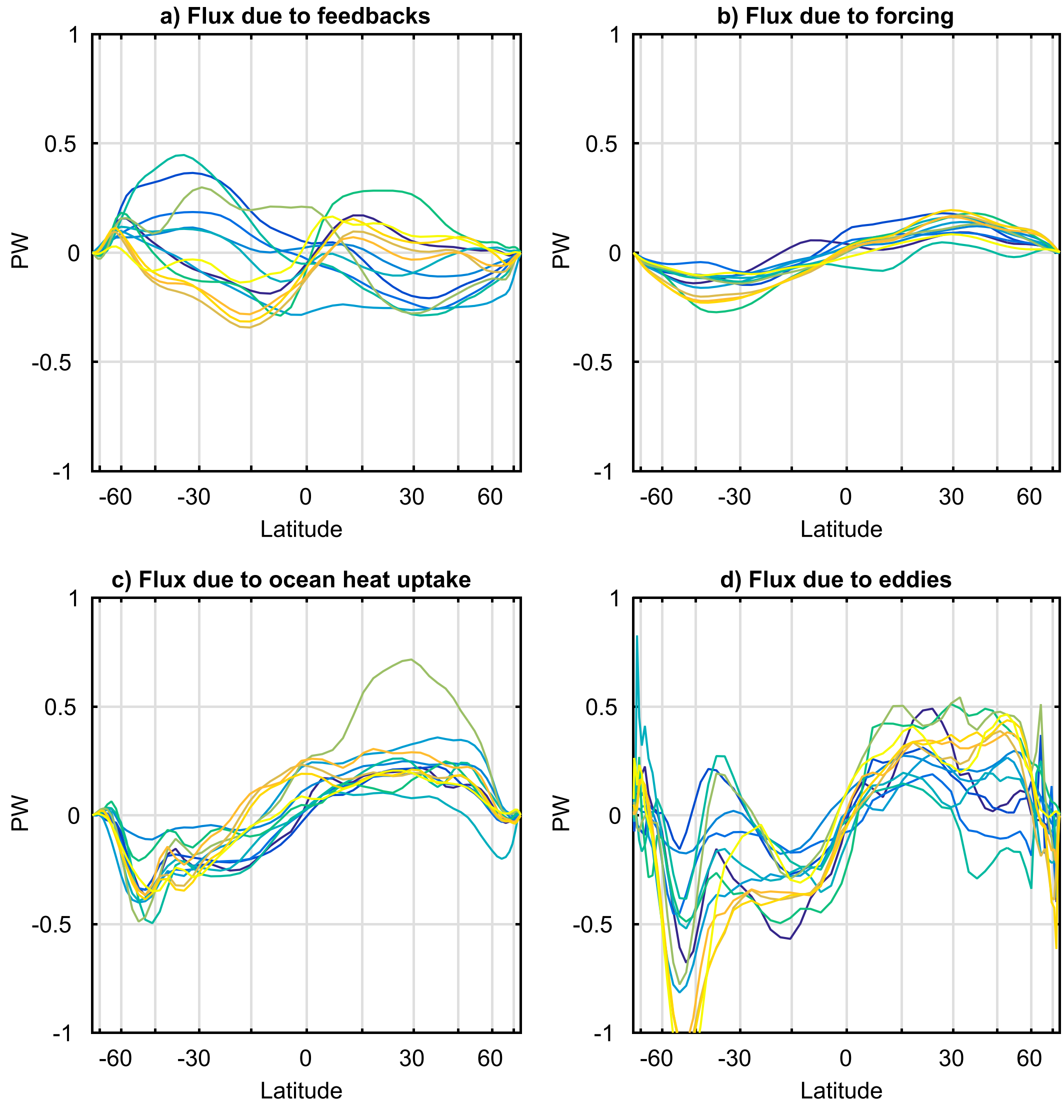

FIG. 3. Anomalous northward atmospheric energy fluxes [i.e., the numerators in Eq. (9) integrated from the South Pole to the North Pole]. The net feedback term is approximated as $\Delta R - R_f$ [Eq. (4)]. (Models are color-coded as in the legend of Fig. 4.)

column of Fig. 4; the other columns indicate the contributions to that circulation response following Eq. (9). Total changes in circulation strength range from -2.6% to $0.6\%\ \mathrm{K}^{-1}$ and are calculated as the change in maximum mass flux $\Delta\psi_{max}/\psi_{max}$, averaged between 15° and 25° latitude in both hemispheres. Models are color-coded such that yellow indicates the least weakening and purple indicates the most weakening. The sole model in which the Hadley cell strengthens is IPSL-CM5A-LR. FGOALS-s2 has outlier contributions, here from its ocean heat uptake term. The residual (i.e., the difference between the total circulation change and the sum of the individual contributions) is less than 2% for all models.

Feedbacks may contribute weakening or strengthening tendencies on the circulation response (Fig. 4), depending on the spatial structure of the net feedback. Radiative forcing is a relatively small contribution but robustly positive; the greater forcing in the tropics reinforces the climatological poleward energy transport and invigorates the Hadley cell. Ocean heat uptake is also a strengthening tendency for the Hadley circulation; the tropical atmosphere must export energy to the extratropics to compensate for anomalous fluxes into the ocean. However, in some models, hemispheric asymmetries in ocean heat uptake tend to strengthen the northern Hadley cell at the expense of the southern cell (not shown), a signature of a southward shift in the

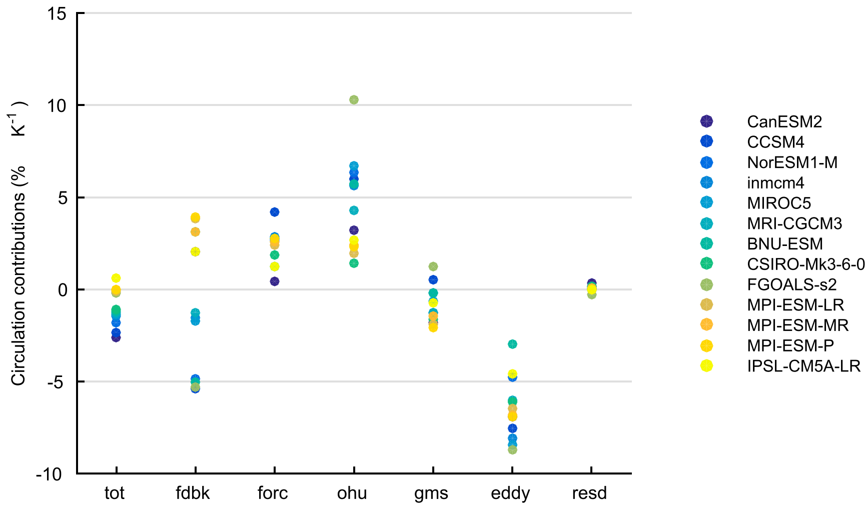

FIG. 4. Changes in circulation strength (% K^{-1}; global-mean surface temperature change) due to net feedback (fdbk), radiative forcing (forc), ocean heat uptake (ohu), gross moist stability (gms), and stationary and transient atmospheric eddies (eddy). The total change (tot) is in the leftmost column and the residual (resd) in the rightmost column. Unnormalized, total circulation changes range from −14.6% to 3.5%. (Expansions for model acronyms are available at http://www.ametsoc.org/PubsAcronymList.)

intertropical convergence zone (ITCZ). These same models have the greatest anomalous cross-equatorial heat flux in Fig. 3c. Weakening tendencies on the circulation are provided primarily by stationary and transient eddies and, in most models, by gross moist stability. In all models, the eddy energy flux out of the tropics increases, and, in isolation, this implies that the Hadley cell needs to transport less energy to compensate. Finally, the increase in gross moist stability with warming stabilizes the tropical atmosphere and hence opposes the circulation.

In Fig. 4, we see that feedbacks dominate the intermodel spread in circulation response. That uncertainty is specifically attributed to the cloud feedback (Fig. 5). As in previous figures, FGOALS-s2 is an outlier in the contribution of individual feedbacks to the circulation response. The dominance of the cloud feedback variability is exacerbated by the evident compensation between the water vapor and temperature feedbacks such that their combined spread is reduced [as has been pointed out for the feedbacks themselves (e.g., Soden and Held 2006)]. Temperature and surface albedo feedbacks are weakening tendencies on the Hadley circulation. In particular, the surface albedo feedback maximizes at the sea ice edge and drives a small equatorward anomalous heat flux in both hemispheres, which acts against the Hadley cell. The water vapor feedback is a strengthening tendency on the circulation.

These tendencies can be understood by returning to Fig. 1. A positive tropical feedback (e.g., water vapor feedback) invigorates the Hadley cell by amplifying the atmospheric energy that must be exported to higher latitudes. On the other hand, a positive extratropical feedback (e.g., surface albedo feedback) or a negative tropical feedback (e.g., temperature feedback or cloud feedback in some models) weakens it by requiring less poleward heat transport. Notably, the cloud feedback imprints on the net feedback in the tropics: the model with the greatest circulation weakening (purple) exhibits the most negative tropical cloud feedback, and vice versa for circulation strengthening (yellow) and a positive tropical cloud feedback. An advantage to the decomposition presented here is the ability to quantify the magnitude of the effect and how uncertainty in feedbacks translates into uncertainty in circulation response.

4. Summary and discussion

Climate feedbacks have advanced understanding of prominent features of global change, such as polar amplification (Holland and Bitz 2003; Feldl and Roe 2013b; Pithan and Mauritsen 2014), enhanced poleward atmospheric heat flux (Hwang and Frierson 2010; Roe et al. 2015), and the importance of tropical clouds to global sensitivity (Bony and Dufresne 2005; Webb et al. 2013). Rather than focus on the surface temperature response, here we expand our investigation to how the energetic framework of feedbacks interacts with dynamical circulation response. We find that tropical mean circulation weakens by up to 2.6% K^{-1} (and 1% K^{-1} on

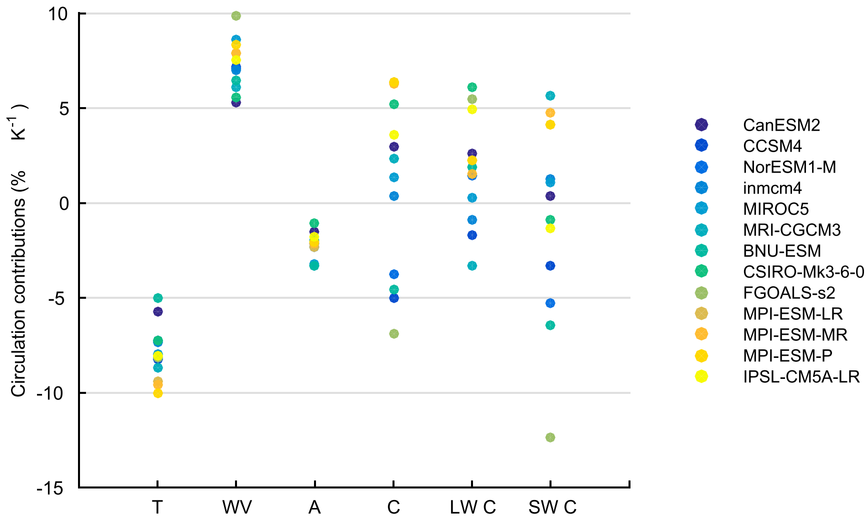

FIG. 5. Changes in circulation strength (% K^{-1}; global-mean surface temperature change) due to temperature feedback (T), water vapor feedback (WV), surface albedo feedback (A), net cloud feedback (C), and the longwave (LW C) and shortwave (SW C) components of the cloud feedback.

average) at quasi equilibrium following an abrupt quadrupling of CO_2 concentration. That weakening is characterized by 1) an increase in stationary and transient eddy heat flux, 2) an increase in gross moist stability, 3) temperature, and, to a lesser degree, surface albedo feedbacks—or more generally any feedback that becomes more positive toward the poles, thus reducing the climatological meridional energy gradient. The weakening is partially offset by ocean heat uptake, the radiative forcing itself, and the water vapor feedback (or, in general, any feedback that becomes more positive toward the equator, thus strengthening the meridional energy gradient). The intermodel spread in the feedback contribution to circulation change is dominated by the cloud feedback, consistent with Su et al. (2014), with correspondence between a strong negative tropical cloud feedback and greatest overall weakening.

The remote impact of extratropical feedbacks on tropical circulation is unsurprising in light of earlier work by Chiang and Bitz (2005), in which the position of the ITCZ was found to respond to imposed high-latitude land and sea ice via atmospheric teleconnections. In the present study, the anomalous atmospheric energy flux due to the surface albedo feedback is largely equatorward and maximizes at 50°–60° latitude in both hemispheres (not shown). We speculate that the increase in absorbed shortwave radiation due to the melting of sea ice contributes to polar amplified surface warming and a reduction in meridional temperature gradient; the reduced temperature gradient in turn produces a decrease in poleward dry static energy flux, which is consistent with the weakening Hadley circulation. More generally, Seo et al. (2014) demonstrated that the remarkable efficacy of high-latitude forcing on ITCZ shifts is a consequence of cloud feedbacks that amplify the response as it propagates equatorward. Beyond atmospheric feedbacks, compensation may also be expected between the atmospheric eddy and ocean heat uptake contributions, reflecting the general compensation between atmosphere and ocean heat transport. Our results, combined with previous studies, highlight the interacting nature of these myriad processes and justify the ongoing efforts required to unravel their contributions to circulation changes.

Recent studies emphasize a weakening of the tropical circulation as a direct response to CO_2. The hypothesis is that increased CO_2 concentration reduces net radiative cooling, which leads to a warming and slight stabilizing of the atmosphere and a decrease of ascent (Bony et al. 2013). This may appear at odds with the strengthening tendency due to radiative forcing shown in Fig. 4. The key to reconciling these apparently disparate results lies in the effect of cloud masking on radiative forcing. Deep convective clouds produce a minimum in radiative forcing by reducing the TOA impact of CO_2 at the equator (Feldl and Roe 2013b; Huang and Zhang 2014; Merlis 2015). The meridional gradient in forcing results in anomalous energy flux convergence along the equator in some models (Fig. 3b; i.e., a weakening of the ascending branch of the Hadley cell). This equatorial effect is likely more prominent in stratosphere-adjusted forcing estimates, as direct tropospheric adjustments to

CO_2 tend to smooth the forcing gradient. However, the dominant tendency across the broader tropics is one of divergence, or a strengthening of the Hadley cell. Since our approach focuses on the region of streamfunction maximum, the latter is emphasized. While somewhat conceptual, the separation of circulation response into a direct, forced component and a component that evolves with surface warming (on the same time scale as feedbacks) has consequences for the viability of solar radiation management to mitigate precipitation change.

The energy flux by the Hadley cell is shaped by the strength and structure of the circulation. Our discussion has highlighted the former; however, the analysis captures both. Shifts in the position of the ITCZ are suggested by increases in the strength of one cell at the expense of the other and are tied to ocean heat transport. As demonstrated by Clement (2006), the presence of tropical ocean heat flux divergence weakens the Hadley cell, reflecting the compensation between the atmosphere and ocean in transporting heat. Further, in some models, the pattern of ocean heat uptake tends to strengthen the northern Hadley cell and weaken the southern cell, which affects the magnitude of the cross-equatorial heat flux. These models display an increase in northward heat flux at the equator of up to 0.25 PW (Fig. 3c). Models in which both cells exhibit the same sign change, on the other hand, tend to have little to no change in cross-equatorial flux (i.e., the cells do not shift relative to the equator). Note that we consider ocean heat uptake anomalies 140–150 years after CO_2 quadrupling, and we anticipate the transient response to exhibit interesting differences. Our result supports other studies that have linked hemispheric asymmetries in ocean heat uptake to the climatological pattern of tropical precipitation (Frierson et al. 2013; Marshall et al. 2014), though here we emphasize the response under climate change.

While increases in atmospheric stability are linked to the weakening of the Hadley circulation in a warmer world, as anticipated, the full tropical circulation response is controlled by a number of local and remote processes in the atmosphere and ocean. The underlying principle does not depend on which processes are acting, only on the aggregate energetic requirements. However, by separating the climate response into its component feedbacks, the relative importance of particular processes to the tropical circulation can be deciphered. While it must be true that the radiative effects of water vapor and clouds are tightly coupled to the large-scale circulation, the details of that coupling and its response under climate change are not well understood. Future work will employ idealized simulations to assess the physical interactions in greater detail. Understanding the interplay between convection and the large-scale circulation is a necessary step for constraining uncertainty in regional climate change in the tropics and subtropics, where the projected environmental and societal impacts are substantial.

Acknowledgments. We thank three anonymous reviewers for their helpful comments on the manuscript. This work also benefited from discussions with Tim Merlis, Gerard Roe, and Yen-Ting Hwang. The authors were supported by the Caltech Engineering and Applied Science Discovery Fund, the Foster and Coco Stanback Postdoctoral Fellowship, and the National Science Foundation (AGS-1524569 and AGS-1462544).

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