What determines meridional heat transport in climate models?

Aaron Donohoe* and David S Battisti*

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

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Corresponding author address: Aaron Donohoe, University of Washington, 408 ATG building, box 351640, Seattle, WA, 98195
Email: aaron@atmos.washington.edu

1. Introduction

The total meridional heat transport (*MHT*) in a steady state climate system is equal to the net radiative surplus integrated over the tropics or, equivalently, the net radiative deficit integrated over the extratropics (Vonder Haar and Oort 1973). In this regard, the *MHT* is equal to the equator-to-pole contrast of absorbed solar radiation (*ASR*) minus the equator-to-pole contrast of outgoing longwave radiation (*OLR*). Therefore, any change in *MHT* must be accompanied by a change in the equator-to-pole contrast of *ASR* or *OLR* without compensating changes in the other quantity. The magnitude of the *MHT* in the climate system varies widely between state of the art coupled climate models (Lucirani and Ragone, 2010). In this paper we demonstrate that this inter-model spread in *MHT* in the models used for the IPCC's fourth assessment (IPCC 2007) is due to inter-model differences in the equator-to-pole contrast of ASR. We then explore the processes that control the equator to pole contrast of ASR, its variability amongst climate models, and its impact on *MHT*.

In a seminal paper, Stone (1978) calculated that approximately two thirds of the observed equator-to-pole gradient in ASR is due to the Earth-Sun geometry and the resulting meridional distribution of incident solar radiation at the top of the atmosphere (TOA) and the remaining one third is due to the equator-to-pole contrast in planetary albedo. Stone emphasized that the latter component was nearly energetically balanced by the equator-to-pole gradient in outgoing longwave radiation (*OLR*) such that the equator-to-pole gradient in net radiation was equal to *ASR* gradient associated with the Earth-Sun geometry. Subsequent work by Enderton and Marshall (2009) demonstrated that this result is not supported by modern observations or in climate model simulations. Enderton and Marshall (2009) found that approximately 35% of the observed equator-to-pole gradient of *ASR* in the Northern Hemisphere and 40% in the Southern Hemisphere is due to the equator-to-pole gradient in planetary albedo and that climate states with altered meridional gradients in planetary albedo exhibit very different strengths of atmospheric and oceanic circulation (i.e. different *MHT*).

Partitioning of the *ASR* gradient into components associated with orbital geometry and planetary albedo is useful because, while the former is externally forced, the latter is a strong function of the climate state and thus may provide important feedbacks when external forcing changes. More importantly, while the equator-to-pole gradient in solar insolation varies by approximately 5% over the entire obliquity cycle, there is little *a priori* constraint on the possible range of the meridional gradient in planetary albedo. Thus, a small perturbation in the external forcing may produce a disproportionately large change in the meridional gradient in ASR via changes in the meridional gradient of planetary albedo (e.g., changes in cloud or snow/ice cover) associated with the *response* of the climate system. Hence, an assessment of the sources that contribute to the meridional gradient in planetary albedo is essential for understanding how and why the atmospheric and oceanic circulation (the *MHT*) will respond to external forcing.

The Earth has a pronounced equator-to-pole gradient in surface albedo due to latitudinal gradients in the fraction of area covered by ocean and land, the latitudinal gradients in land vegetation, and the spatial distribution of land and sea ice (Robock 1980). The contribution of the equator-to-pole gradient in surface albedo to the equator-to-pole gradient in planetary albedo is still an unresolved question in climate dynamics, however, because there is considerable attenuation of the surface albedo by the atmosphere. While simplified energy balance models (EBMs) have often assumed that the local planetary albedo is a function of surface albedo only (i.e., Budyko 1969 and North 1975), this assumption is unwarranted due to the atmosphere's influence on planetary albedo; the step function change of planetary albedo at the ice-edge specified by EBMs is inconsistent with the observed meridional structure of planetary albedo (Warren and Schneider 1979) and more recent parameterizations of planetary albedo in EBMs have suggested that the atmosphere damps the influence of surface albedo on the top of atmosphere (TOA) radiative budget (Graves et al. 1993). Recent work by Donohoe and Battisti (2011) has demonstrated that the vast majority of the global average planetary albedo is due to atmospheric as opposed to surface processes; this result suggests that the meridional gradient of planetary albedo and hence the MHT in the climate system may also be strongly dictated by atmospheric processes (i.e. cloud properties).

This paper is organized as follows. In Section 2, we present the inter-model spread of *MHT* in coupled climate models and how the equator-to-pole contrast of *ASR* and *OLR* relate to the *MHT* spread. We find that the equator-to-pole contrast of *ASR* is highly correlated with the inter-model spread of *MHT*. In Section 3, we analyze the processes that determine the observational and inter-model spread of the equator-to-pole contrast of ASR; specifically, we find that model differences in the equator-to-pole contrast in ASR are dictated by the meridional gradient of planetary albedo which, in turn, is primarily due to cloud reflection. In Section 4, we look at the processes that control the inter-model spread in *OLR* and how these processes relate to equator-to-pole contrast of net radiation. A conclusion follows.

2. Meridional heat transport and the equator-to-pole contrast of absorbed solar radiation

In this section, we analyze the MHT in climate models and observations in terms of the equator-to-pole contrast of ASR and OLR. We demonstrate that the inter-model spread in peak MHT is largely determined by the equator-to-pole contrast of ASR.

a. Model runs and datasets used

We use data from the World Climate Research Programme's (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset: a suite of standardized coupled simulations from 25 global climate models that were included in the International Panel on Climate Change's Fourth Assessment Report (https://esgcet.llnl.gov:8443/index.jsp). The set of model simulations is commonly referred to as the WCRP's CMIP3 multi-model dataset (Meehl et al. 2007). We use the pre-industrial (PI) simulations from the 15 coupled models that provided the output fields required for the analysis presented in this study (Table 2). Each PI simulation is forced with temporally invariant external forcing (CO_2 is set to 280 ppm) and, in principle, represents an equilibrium climate that is in energy balance. In practice, both the global average and the local energy budgets are not balanced in the simulated climatologies (Lucarini and Ragone 2010); hence, we make corrections to balance the global annual mean radiative budget by adding a spatially and temporally invariant constant to the OLR field, prior to performing the analysis[1]. All calculations reported here are based on solar weighted annual average fields.

The observational analysis uses the TOA and surface shortwave radiation data products from the Clouds and Earth's Radiant Energy System (CERES) experiment (Wielicki et al. 1996). We use Fasullo and Trenberth's (2008a,b) long term climatologies of the CERES TOA data that correct for missing data and global average energy imbalances. For the surface shortwave fluxes we use the CERES "AVG" fields which are derived by assimilating the satellite observations into a radiative transfer model to infer the surface fluxes (Rutan et al. 2001). All calculations are preformed separately for each of the four CERES instruments (FM1 and FM2 on Terra from 2000 -2005 and FM3 and FM4 on AQUA from 2002 – 2005). We then average the results over the four instruments. Our calculations are performed on the annual average (solar-weighted) data.

b. Methodology for MHT calculation and definitions of ASR* and OLR*

We determine the total (atmosphere plus ocean) zonally averaged MHT to the extratropics of each hemisphere by noting that, in an equilibrium climate, the net radiative deficit spatially integrated from latitude Θ to the pole is exactly balanced by MHT into the region poleward of Θ (Trenberth and Caron 2001, Fasullo and Trenberth 2008b, and Vonder Haar and Oort 1973):

[1] The only calculated field discussed here that is affected by this correction is the MHT; this correction ensures the global mean heat transport divergence is zero and the resulting MHT is independent of whether the heat transport divergence is integrated from the South Pole to the North Pole or vice versa.

$$MHT(\theta) = -2\pi R^2 \int_{x=\sin(\theta)}^{1} \left[ASR(x) - OLR(x) \right] dx \quad . \qquad (1)$$

We gain insight into the processes that determine the *MHT* by decomposing the *ASR(x)* and *OLR(x)* into global averages (denoted by overbars) and spatial anomalies (denoted by primes) and by setting the limit of integration to $x_m = sin(\Theta_m)$, where Θ_m is the latitude where the zonally averaged *ASR* and *OLR* are equal. Then Equation 1 yields the maximum zonally averaged meridional heat transport (MHT_{MAX}):

$$\begin{aligned} MHT_{MAX} &= MHT(x_m) = -2\pi R^2 \int_{x=x_m}^{1} \left[ASR'(x) + \overline{ASR} - (OLR'(x) - \overline{OLR}) \right] dx \\ &= -2\pi R^2 \int_{x(ASR'=OLR')}^{1} [ASR'(x) - OLR'(x)] dx \end{aligned} \quad . \qquad (2)$$

Reduction to the last equation relies on the fact that a steady climate system achieves global average radiative equilibrium:

$$\overline{ASR} = \overline{OLR} \quad . \qquad (3)$$

Equation (2) can be rewritten as

$$MHT_{MAX} \cong -2\pi R^2 \left[\int_{x(ASR'=0)}^{1} ASR' \ dx \ - \int_{x(OLR'=0)}^{1} OLR' \ dx \right] = ASR^* - OLR^* \quad , \qquad (4)$$

where

$$ASR^* = -2\pi R^2 \int_{x(ASR'=0)}^{1} ASR' \ dx \qquad (5)$$

and

$$OLR^* = -2\pi R^2 \int_{x(OLR'=0)}^{1} OLR'(x) dx \quad . \qquad (6)$$

The near equality in Equation 4 holds exactly if the meridional nodes of the *OLR'* and *ASR'* are co-located; in all calculations presented here the near equality holds to within 1% of the MHT_{MAX} (the average error in the approximation is 0.3%). Figure 1 presents graphical representation for calculating MHT_{MAX} from equation 1 (panel A) and from equation 4 by application of the definitions of *ASR** and *OLR** (panels B and C respectively); the difference of panels 2B and 2C recompose the area representing $\mathrm{MHT}_{\mathrm{MAX}}$ in panel A.

The negative sign in equations 5 and 6 are chosen so that the *ASR* and *OLR* deficit over the extratropics defines *ASR** and *OLR** as a positive numbers. Equation (5) and (6) are the ASR and OLR deficit (*ASR** and *OLR**) in the Northern Hemisphere (NH) extratropics; a similar expression with modified limits of integration holds for the Southern Hemisphere (SH). By definition, the sum of *ASR** (or *OLR*)* in the two hemispheres is equal to the *ASR* (*OLR*) surplus (relative to the global average) integrated over the tropics. Therefore, this quantity represents the difference between energy absorbed (emitted) in the tropics and in the extratropics. In a steady climate system, the equator-to-pole contrast of *ASR* (*ASR*)* must either be balanced radiatively by an equator-to-pole contrast of *OLR* (*OLR*)* or by a dynamic heat transport from the tropics to the extratropics (MHT_{MAX}). In this regard, *ASR** represents the equator to pole scale forcing of the climate system and *OLR** and peak MHT_{MAX} are the radiative and dynamic responses to the forcing[2].

As a quantative example, we calculate from the CERES data that ASR* is 8.2 (9.0) PW in the NH (SH) and that this deficit in *ASR* over the extratropics is balanced by

[2] We will demonstrate that *ASR** is not externally forced, but is itself a function of the climate system.

a 2.4 (3.2) PW OLR anomaly (OLR*) and 5.8 (5.8) PW of heat import via MHT_{MAX} (Table 3).

c. Results

The CMIP3 models and the observations all have similar meridional structures of *MHT* (Fig. 2a) with a peak heat transport around 36° in each hemisphere. The inter-model average *MHT* structure and magnitude is in close agreement with the observational estimates of *MHT*[3] in the Northern Hemisphere (NH) and has a peak value (MHT_{MAX}) of 5.6 PW (Table 3). In the Southern Hemisphere (SH), the inter-model average MHT_{MAX} of 5.2 PW is one standard deviation (σ) below the observed MHT_{MAX}. MHT_{MAX} varies widely between models (Fig. 2b-c); the inter-model spread (defined throughout as two standard deviations -- 2σ) is 0.8 PW in the NH and 1.1 PW in the SH. In the SH, the model with the largest MHT_{MAX} has approximately 50% more heat transport than the model with the smallest MHT_{MAX}.

The inter-model spread (2σ) in *ASR** is 0.9 PW in the NH and 1.2 PW in the SH and approximately twice the inter-model spread in *OLR** (0.5 PW in the NH and 0.6 PW in the SH). The inter-model spread in MHT_{MAX} is well correlated with the inter-model spread in *ASR** (Fig. 3a) with and R^2 value of 0.57 in the NH and 0.85 in the SH (Table 4), both significant at the 99% confidence interval. In contrast, the inter-model spread in MHT_{MAX} is not correlated with *OLR** (Fig. 3b).

We understand this result as follows. The inter-model spread in MHT_{MAX} can be diagnosed from Eq. (4) and the statistics of *OLR** and *ASR**:

[3] The latter has uncertainties of approximately 20% at the latitude of peak heat transport (Wunsch 2005).

$$\sqrt{\left[MHT_{MAX}{}'^2\right]} = \sqrt{\left[ASR^{*}{}'^2\right] + \left[OLR^{*}{}'^2\right] - 2\left[OLR^{*}{}' \bullet ASR^{*}{}'\right]} \qquad (7)$$

where primes indicate the departure of the quantity from the inter-model average and the brackets are averages over all the models. Eq. (7) demonstrates that the inter-model spread in MHT_{MAX} is a consequence of the spread in ASR^*, the spread in OLR^*, and the covariance between ASR^* and OLR^*. For example, in the limit that OLR^* and ASR^* are linearly independent, then the spread in MHT_{MAX} is the quadrature sum of the spread in ASR^* and OLR^*. In contrast, in the limit of perfect correlation between OLR^* and ASR^*, with a regression coefficient of unity, there would be no spread in MHT_{MAX}, independent of how much either OLR^* and ASR^* varied between the different models. These two limits correspond to what we will call the dynamic and radiative limits of the extratropical energy budget. In the dynamic limit, the inter-model spread in either longwave or shortwave radiative deficit integrated over the extratropics is balanced by a *MHT* anomaly (from the inter-model average). In the radiative limit, the inter-model spread in shortwave and longwave deficits over the extratropic balance each other and there is no inter-model spread in MHT_{MAX}.

The square root of the inter-model covariance of OLR^* and ASR^* is approximately the same magnitude as the spread in OLR^* and is significantly smaller than the spread in ASR^* (Table 4) suggesting that the CMIP3 models are closer to the dynamic limit than the radiative limit; inter-model anomalies in ASR^* and OLR^* only partially balance each other leading to a MHT_{MAX} spread that is comparable in magnitude to the ASR^* spread. We can understand the correlation of ASR^* and MHT_{MAX} and the lack of correlation between OLR^* and MHT_{MAX} from the statistics of ASR^* and OLR^*.

Multiplying Eq. 4 by *ASR**, averaging over all models, and diving by the standard deviation of *ASR** and MHT_{MAX} (from Eq. 7) gives

$$R_{MHT,ASR*} = \frac{[ASR^{*\prime} \bullet MHT_{MAX}{'}]}{\sqrt{[ASR^{*\prime 2}]}\sqrt{[MHT_{MAX}{'^2}]}} = \frac{[ASR^{*2}] - [ASR^{*\prime} \bullet OLR^{*\prime}]}{\sqrt{[ASR^{*\prime 2}]}\sqrt{[ASR^{*\prime 2}] + [OLR^{*\prime 2}] - 2[ASR^{*\prime} \bullet OLR^{*\prime}]}}$$

(8)

where $R_{MHT,ASR*}$ is the correlation coefficient between MHT_{MAX} and *ASR** across the models. A similar expression holds for $R_{MHT,OLR*}$. If *OLR** and *ASR** were uncorrelated, then the fraction of the MHT_{MAX} spread explained by *OLR** and *ASR** would be proportional to the spread in each variable and the fractional spread explained by each variable would sum to unity. *Ceteris paribus*, 70% (87%) of the inter-model variance of MHT_{MAX} in the NH (SH) would be explained by *ASR** spread and the remaining 13% (30%) would be explained by *OLR**. The positive correlation between *OLR** and *ASR** reduces the correlation coefficient between MHT_{MAX} and both *ASR** and *OLR** causing the fractional sum of the explained variance in MHT_{MAX} to decrease below unity. In our specific case, the covariance between *ASR** and *OLR** causes the MHT_{MAX} variance explained by *ASR** to decrease to 57% (85%) and the MHT_{MAX} spread explained by *OLR** to decrease to 0% (2%) in the SH (NH). The near zero correlation between MHT_{MAX} and *OLR** can be understood from the competing effects of the two terms in the numerator of Eq. 8. Averaged over the ensemble members, a one unit anomaly in *OLR** is accompanied by an approximately one unit anomaly in *ASR**, because the *OLR** spread and covariance between *ASR** and *OLR** are approximately equal (Table 4). Thus, the typical magnitude of an *ASR** anomaly associated with a given *OLR** anomaly nearly balances the *OLR** anomaly leading to no correlation between *OLR** and MHT_{MAX}.

In summary, the MHT_{MAX} spread in climate models is due to *ASR** differences between the models because the inter-model spread in *ASR** exceeds that in *OLR** and *ASR** and *OLR** are only weakly correlated. In the remainder of this paper, we will analyze the physical processes that determine *ASR**, *OLR**, their inter-model spread, and covariance.

3. The cause of inter-model spread in *ASR**

We now describe a method for partitioning *ASR** into components due to the Earth-Sun geometry and the meridional gradient of planetary albedo. We then further partition the planetary albedo contribution to *ASR** into components due to atmospheric and surface reflection and apply this methodology to the CMIP3 simulations.

a. ASR partitioning into incident and planetary albedo contribution*

1. METHODS

That *ASR** is non-zero is due to both the meridional gradient in incident solar radiation and to the meridional gradient in the planetary albedo. We can partition *ASR** into these two component contributions by writing the planetary albedo and incident solar radiation as the sum of a global mean and a spatial anomaly:

$$ASR(x) = a(x)\ S(x) = (\overline{a} + a'(x))\ (\overline{S} + S'(x)) = \overline{a}\overline{S} + \overline{a}S'(x) + a'(x)\overline{S} + a'(x)S'(x) \quad (9)$$

where *a(x)* is the co-albedo (one minus albedo), overbars denote a spatial average, and primes indicate spatial anomalies[4]. *ASR** can be calculated from Eq. 9 by subtracting the global average of each term and integrating over the extratropics:

$$ASR* = -2\pi R^2 \left[\bar{a} \int_{x(ASR'=0)}^{1} S'(x)\ dx \quad + \bar{S} \int_{x(ASR'=0)}^{1} a'(x)\ dx \quad + \int_{x(ASR'=0)}^{1} \left[S'(x)a'(x) - \frac{1}{2}\int_{-1}^{1} S'(x)a'(x)dx \right] dx \right] \quad (10)$$

The first term on the right hand side of Eq. 10 represents the equator-to-pole difference in incident solar radiation modified by the global average co-albedo and is primarily a function of the Earth-Sun geometry; it is the equator-to-pole contrast of *ASR* that would exist if there were no meridional variations in planetary albedo. The second term is the contribution of the meridional gradient in planetary albedo to *ASR** in the absence of

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[4] An alternative approach to dividing the fields into a global mean and spatial anomaly is to expand the variables in terms of even Legendre polynomials in each hemisphere, as was done in Stone (1978), North (1975), and Enderton and Marshall (2009). Our *ASR** and component contributions to *ASR** are proportional to the second Legendre coefficients provided that the spatial structure of ASR projects entirely onto the zeroth and second Legendre polynomials. The total ASR contrast calculated by these two methods agree to within 2%; the first order terms agree to within 5% of each other, and the second order term (the covariance) agrees to within 30%. The discrepancy is larger for the second order term because, even if the planetary albedo and incident solar radiation were fully captured by the first two Legendre polynomials, the covariance projects primarily on the 4^{th} Legendre polynomial and only secondarily onto the 2^{nd} polynomial (i.e. note the spatial structure of the covariance term in Figure 1B). In this regard, our index of the meridional difference is more accurate than that obtained by expansion in terms of Legendre polynomials truncated at the 2^{nd} order term , although the primary conclusions reached here are independent of the methodology employed.

spatial variations of incident solar radiation (Figure 4b). The last term is the covariance of the spatial anomalies in planetary albedo and incident radiation. The covariance contributes to a positive global average ASR because the high latitude regions have high albedos but receive a deficit of solar radiation such that the global average planetary co-albedo ($\overline{a}$) is smaller than the global average solar weighted planetary coalbedo. Similarly, the last term in Equation 10 makes a negative contribution to *ASR** because the high planetary albedo regions receive less incident radiation than the global average value that appears in the second term of Equation 3 such that the contribution to *ASR** due to the meridional gradient in planetary albedo is overestimated by the second term alone. Therefore, we can interpret the covariance term as a correction to the planetary albedo's contribution to *ASR**.

Equation 10 divides *ASR** into a geometric component that exists in the absence of any meridional gradient in planetary albedo (the first term, red line in Figure 4b,c) and a component that owes its existence to the meridional gradient in albedo (the sum of the second and third terms, Figure 4c). Hence, in the remainder of this study, we will define *the planetary albedo's contribution to ASR** to be the sum of the second and third terms[5].

2. RESULTS

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[5] It is equally valid to interpret Equation 3 as consisting of a component that exists in the absence of a meridional gradient in solar insolation (second term) and a component owing its existence to the meridional gradient of solar insolation (the sum of the first and third term). The interpretation is contingent on the phrasing of the question. In this regard, the grouping of the terms we adopt in this paper is a lower limit assessment of the planetary albedo gradient's contribution to *ASR**.

In the observational fields, spatial variations in planetary albedo contribute 2.9 PW to *ASR** in the NH via Eq. (10), representing 35% of the total *ASR** (8.2 PW – Table 3 and FIG. 4). In the SH, spatial variations in planetary albedo contribute 3.7 PW to *ASR** (41% of the total *ASR** of 9.0 PW). The inter-model average planetary albedo contribution to *ASR** in the NH is nearly identical to the observations (Table 3) whereas the models have a reduced equator-to-pole-contrast in planetary albedo in the SH resulting in small *ASR** values (by 0.5 PW on average) relative to the observations.

The planetary albedo contribution to *ASR** varies widely between models (2σ = 0.9 PW in the NH and 1.2 PW in the SH). In contrast, the incident contribution to *ASR** varies by less than 1% among the different CMIP3 models. The small inter-model spread in spread in the incident contribution to ASR* is due to primarily to inter-model differences in global average planetary albedo and secondarily to small inter-model differences in the solar constant. The inter-model spread in the planetary albedo contribution to *ASR** explains 99% of the spread in ASR* in both hemispheres. Thus, the inter-model differences in *ASR** are a consequence of the planetary albedo differences between models.

b. Partitioning of planetary albedo into atmospheric and surface components

1. METHODOLOGY AND ADDITIONAL DATA SETS USED

We use the method of Donohoe and Battisti (2011) to partition the planetary albedo into a component due to reflection off of objects in the atmosphere and a component due to surface reflection. In short, their method builds a simplified radiative transfer model at each gridpoint that accounts for atmospheric absorption, atmospheric reflection, and surface reflection for an infinite number of passes through the atmosphere. By assuming that the atmosphere is isotropic to shortwave radiation, the simplified model provides analytical expressions for the upwelling and downwelling shortwave fluxes at both the surface and top of the atmosphere (4 equations) in terms of the incident radiation, the fractions of atmospheric reflection and absorption during each pass through the atmosphere, and the surface albedo (4 variables). The variables can be solved for given the radiative fluxes. The atmospheric contribution to the planetary albedo is equal to the fraction of radiation reflected during the first downaward pass through the atmosphere and will be denoted as $\alpha_{P,ATMOS}$. The surface contribution to planetary albedo is equal to the fraction of incident radiation that is reflected at the surface and eventually escapes to space and will be denoted as $\alpha_{P,\ SURF}$.

We calculate $\alpha_{P,ATMOS}$ and $\alpha_{P,SURF}$ for both the models and observations using annual average radiative fields. We have also performed the calculations on the climatological monthly mean data from the observations and then averaged the monthly values of $\alpha_{P,ATMOS}$ and $\alpha_{P,SURF}$ to obtain the annual average climatology. The zonal average $\alpha_{P,ATMOS}$ calculated from monthly data agree with those calculated directly from the annual average data to within 1% of $\alpha_{P,ATMOS}$ at each latitude.

2. Results

In both the models and observations, the vast majority (over 85%) of the global average planetary albedo is due to $\alpha_{P,ATMOS}$. $\alpha_{P,SURF}$ is approximately one third of the surface albedo because the atmosphere opacity attenuates the amount of incident solar radiation that reaches the surface and the amount of radiation that is reflected at the surface that escapes to space. These results are discussed at length in Donohoe and

Battisti (2011). Here, we focus on the implications of these results on the inter-model spread in $ASR*$ and MHT_{MAX}.

The contribution of $\alpha_{P,ATMOS}$ and $\alpha_{P,SURF}$ to $ASR*$ can be assessed by first dividing the planetary coalbedo (a) into separate atmospheric and surface components, each with a global average and spatial anomaly:

$$a(x) = (1-\alpha_{P,ATMS} - \alpha_{P,SURF}) = (1-\overline{\alpha}_{P,ATMOS} - \overline{\alpha}_{P,SURF} - \alpha'_{P,ATMOS} - \alpha'_{P,SURF}) \quad , \qquad (11)$$

Substituting Eq. (11) into the definition of $ASR*$ in Eq. (5) yields the contribution of $\alpha_{P,ATMOS}$ to $ASR*$:

$$ASR_{ATMOS}* = 2\pi R^2 \overline{S} \int_{x(ASR'=0)}^{1} \alpha'_{P,ATMS} dx \; + \; 2\pi R^2 \int_{x(ASR'=0)}^{1} \left[\alpha'_{P,ATMS} S' - \frac{1}{2} \int_{-1}^{1} \alpha'_{P,ATMS} S' dx \right] dx \quad , \qquad (12)$$

where we have again grouped the linear and covariance terms together to calculate the total contribution of spatial structure in $\alpha_{P,ATMOS}$ to $ASR*$ ($ASR*_{ATMOS}$). A similar expression is used to calculate $\alpha_{P,SURF}$'s contribution to $ASR*$ which we define as $ASR*_{SURF}$. In the observations $ASR*_{ATMOS}$ is found to contribute 2.5 PW to $ASR*$ while $ASR*_{SURF}$ is found to contribute 0.4 PW to $ASR*$ in the NH (Table 3). In the SH, $ASR*_{ATMOS}$ contributes 3.5 PW to $ASR*$ while $ASR*_{SURF}$ contributes 0.2 PW to $ASR*$. These results suggest that, even if the equator-to-pole gradient in surface albedo were to greatly diminish (e.g., in an ice-free world) the equator-to-pole scale gradient in ASR would decrease by less than 5% in each hemisphere, neglecting any changes in the atmospheric reflection.

In the NH, the breakdown of ASR^* into components associated with ASR^*_{ATMOS} and ASR^*_{SURF} is well reproduced by the inter-model average; the CMIP3 average $ASR^*_{,ATMOS}$ ($ASR^*_{,SURF}$) contribution to ASR^* is 2.4 PW (0.5 PW) while that observed is 2.5 PW (0.4 PW). In the SH, the CMIP3 ensemble average ASR^*_{ATMOS} contribution to ASR^* of 2.9 PW is one standard deviation smaller than the observed value of 3.5 PW while the ASR^*_{SURF} is small in magnitude and in close agreement with the observations. These results suggest that the model bias towards smaller than observed MHT_{MAX} in the SH (FIG.2) is a consequence of smaller than observed equator-to-pole gradient in shortwave cloud reflection (ASR^*_{ATMOS}).

Figure 5 shows a scatter plot of the total ASR^* against (a) ASR^*_{ATMOS} (b) ASR^*_{SURF} from the CMIP3 models (plus signs) in the northern (blue) and southern (red) hemispheres. There is a remarkably large range in the simulated ASR^* (2σ = 0.9PW and 1.2PW in NH and SH). Almost all of the inter-model spread in ASR^* is due to ASR^*_{ATMOS}; ASR^*_{ATMOS} (2 sigma = 1.2PW and 1.4PW in the NH and SH) is highly correlated with the total ASR^* ($R^2 = 0.94$), and the best-fit slope in each hemisphere is nearly unity. In comparison, the inter-model spread in ASR^*_{SURF} is small (2σ = 0.5PW and 0.4PW in the NH and SH, respectively) and not correlated with total ASR^*.

We take two limiting models for how the meridional structure of atmospheric and surface reflection contribute to ASR^*: "Model A" in which the surface albedo is spatially invariant so that ASR^* is determined entirely by the spatial structure of atmospheric reflection and "Model B" in which the atmosphere is transparent to shortwave radiation so that ASR^* is determined entirely by the surface albedo gradient. In the case of "Model A", ASR^* would equal the sum of the ASR^*_{ATMOS} and the incident (geometric) component of 5.2PW (black line, Figure 5a). “Model A” is an excellent fit to the inter-model spread in *ASR**. “Model A” slightly under predicts *ASR** in all cases because ASR^*_{SURF} is positive in all models (the vertical offset between the black line and the individual model results in Figure 5a). This suggests that, while surface processes do play a role in determining *ASR**, the majority of the inter-model spread in *ASR** (94%) is explained by differences in atmospheric reflection.

At the other end of the spectrum, if the atmosphere were indeed transparent to shortwave radiation (“Model B”), *ASR** would be equal to the incident (geometric) contribution plus the surface reflection contribution given by the global average solar insolation times the surface albedo anomaly integrated over the extratropics (plus a second order term):

$$SURF^* = \overline{S} \int_{x(ASR'=0)}^{1} \alpha' \; dx \; + \int_{x(ASR'=0)}^{1} \left[\alpha' S' - \frac{1}{2} \int_{-1}^{1} \alpha' S' \, dx \right] dx \quad , \tag{13}$$

where α' is the surface albedo anomaly from the global average. *SURF** is the contribution of the surface albedo to *ASR** if the atmosphere is transparent to shortwave radiation. The theoretical prediction of “Model B” is co-plotted with results from the CMIP3 PI simulations in Figure 5C; “Model B” is clearly a poor description of the CMIP3 ensemble. Surface albedo plays a negligible role in determining the inter-model differences in *ASR** because the surface albedo is strongly attenuated by the atmosphere and the inter-model spread in atmospheric reflection overwhelms the surface albedo contribution to planetary albedo spread.

Collectively, these results suggests that differences in atmospheric reflection are, by far, the primary reason for the remarkable spread in *ASR** in the CMIP3 ensemble of PI simulations.

4. Processes controlling the inter-model spread of OLR*

In the previous sections we concluded that the CMIP3 ensemble features large differences in *ASR** (due to cloud reflection differences) that are only weakly compensated by differences in *OLR** leading to large inter-model spread in MHT_{MAX}. This result is surprising because cloud longwave and shortwave radiative forcing are known to compensate for each other in the tropics (Keihl 1994 and Hartmann et al. 2001). In this section, we ask why the inter-model spread in ASR* and OLR* do NOT compensate for each other. We first analyze the processes that cause the inter-model spread in *OLR* (subsection A). We then project these results on to the inter-model spread in *OLR** (subsection B) and relate the results to the inter-model spread of *ASR** (subsection C).

A. Inter-model spread in OLR

OLR is a consequence of both clear sky processes (i.e. temperature and specific humidity) and cloud properties (i.e. cloud optical thickness and height). We partition the inter-model spread in *OLR* into cloud and clear sky contributions. We then further sub-partition the cloud contribution into cloud fraction and cloud structure components and the clear-sky contribution into surface temperature and specific humidity components in this subsection.

We diagnose the cloud contribution to *OLR* from the longwave cloud forcing (LWCF—Kiehl 1994):

$$LWCF = OLR_{CLEAR} - OLR \quad , \tag{14}$$

where OLR is the total-sky OLR and OLR_{CLEAR} is the clear-sky OLR. We decompose the inter-model spread in OLR into clear sky and cloud components as follows: the zonal average inter-model spread in OLR is regressed onto the inter-model spread in OLR_{CLEAR} (-$LWCF$) at each latitude and rescaled by the spread in OLR to define the clear-sky (cloud) contribution to OLR spread. By construction, the clear-sky and $LWCF$ contribution to the OLR spread add to the total-sky OLR spread (FIG. 8).

In the tropics, the inter-model spread in OLR is almost entirely due to differences in $LWCF$ (FIG 8A). In contrast, the inter-model spread in OLR in the polar regions is almost entirely due to differences in OLR_{CLEAR}. In the subtropics, $LWCF$ and OLR_{CLEAR} contribute nearly equally to the OLR spread. In the SH storm track region, $LWCF$ contributes more the OLR spread than OLR_{CLEAR} while the opposite is true in the NH storm track region.

We further divide the inter-model spread in $LWCF$ into components due inter-model differences in cloud fraction and cloud structure. The total-sky OLR can be written as the cloud fraction (C_{FRAC}) weighted sum of the OLR when the scene is clear (OLR_{CLEAR}) and the OLR when the scene is cloudy (OLR_{CLOUD}):

$$OLR = (1-C)\ OLR_{CLEAR} + C\ OLR_{CLOUD} = OLR_{CLEAR} + C(OLR_{CLOUD} - OLR_{CLEAR}) \ . \tag{15}$$

Eq. (15) can be rearranged into an expression for LWCF:

$$LWCF = C_{FRAC}(OLR_{CLEAR} - OLR_{CLOUD}) \equiv C_{FRAC}C_{STRUC} \quad . \tag{16}$$

Eq. (15) states that $LWCF$ is a consequence of how often the scene is cloudy and the optical properties/vertical structure of the cloud. For example, two models with the same C_{FRAC} could have very different $LWCF$ due to different cloud top heights (Hartmann et al.

1992). The inter-model spread in *LWCF* is divided into components due to inter-model differences in C_{FRAC} and C_{STRUC} by deconstructing C_{FRAC} and C_{STRUC} into the ensemble average and the perturbation from the average at each latitude:

$$LWCF = (\overline{C_{FRAC}} + C'_{FRAC})(\overline{C_{STRUC}} + C'_{STRUC}) = \overline{C_{FRAC}}\,\overline{C_{STRUC}} + \overline{C_{FRAC}}C'_{STRUC} + \overline{C_{STRUC}}C'_{FRAC} + C'_{STRUC}C'_{FRAC} \,.$$

(17)

The first term on the right hand side does not contribute to the inter-model spread in *LWCF*. The second and third terms correspond to the contribution of cloud structure ($LWCF_{STRUC}$) differences and cloud fraction ($LWCF_{FRAC}$) differences to the inter-model spread in *LWCF*. The last term is substantially smaller than the other terms at all latitudes (not shown).

Inter-model differences in C_{FRAC} are responsible for the majority of the inter-model *OLR* spread in the subtropics and SH storm track region and approximately 50% of the inter-model *OLR* spread in the NH storm track region (FIG. 8B). Differences in C_{STRUC} are responsible for the vast majority of the inter-model spread of *OLR* in the tropics. In the polar regions (poleward of 60˚), inter-model differences in *LWCF* are uncorrelated with the cloud fraction spread suggesting that cloud optical properties, as opposed to cloud amount determines *OLR* in this region (Curry and Ebert 1992).

The contribution of OLR_{CLEAR} to the *OLR* spread is subdivided into components that are linearly congruent with the inter-model spread in surface temperature and vertically integrated specific humidity as follows (Thompson and Solomon 2002): the correlation coefficient between inter-model differences in OLR_{CLEAR} and surface temperature (or the negated specific humidity) is multiplied by the OLR_{CLEAR} spread at each latitude. The inter-model differences in surface temperature explain the vast majority of the inter-model spread in OLR_{CLEAR} in the extratropics and make the largest contribution to the *OLR* spread in the polar regions (FIG. 8C). This spatial structure mimics the inter-model spread in surface temperature spread ($R^2 = 0.95$) which features values of approximately 4K in the polar regions (1σ) and below 1K equatorward of 40˚ (not shown). The regression coefficient between surface temperature and OLR_{CLEAR} for all gridpoints and ensemble members is 2.1 W m^{-2} K^{-1} which is consistent with other estimates of the linear parameterization of *OLR* with surface temperature (Warren and Schneider 1979). We understand these results as follows. Per unit perturbation of surface temperature, the *OLR* changes by approximately 2 W m^{-2} with some regional dependence[6]. Thus, the OLR_{CLEAR} spread scales as the surface temperature spread times approximately 2 W m^{-2} K with warmer temperatures corresponding to larger *OLR* values.

We also expect that the OLR_{CLEAR} to be negatively correlated with the water vapor content of the upper atmosphere due to the greenhouse effect. Indeed, inter-model variations in vertically integrated water vapor explain a portion of the OLR_{CLEAR} spread in the sub-tropics that was not previously explained by surface temperature variability (FIG. 8C) with higher vapor content corresponding to lower *OLR* values due to the raising of the effective emission level. In the high latitudes the opposite is true; high vapor content corresponds to more *OLR* due to the positive correlation between upper atmospheric water vapor and surface temperature (not shown) that is absent in the subtropics.

In summary, the inter-model spread in *OLR* is a consequence of nearly equal contributions from clear-sky and cloud processes with the cloud processes playing a dominant role in the lower latitudes and clear-sky processes dominating the extratropical

[6] The regression of surface temperature onto OLR_{CLEAR} at each latitude shows larger values in the dry subtropics and lower values in the high latitudes.

behavior. The cloud contribution is due to differences in both cloud fraction and cloud structure while the clear-sky contribution is primarily due to surface temperature differences with the exception of the subtropics where inter-model differences in water vapor also play a role.

*B. Inter-model spread in OLR**

The contributions to the *OLR* spread that were discussed in the previous subsection are projected onto the inter-model spread in *OLR** in this subsection. The spread in *OLR** is a consequence of the magnitude of spread in the component contributions to *OLR* (previously discussed) and the spatial decorrelation length scale of those processes. For instance, even though cloud fraction explains a large fraction of the *OLR* spread at each latitude, it would be poorly correlated with the spread in *OLR** if the cloud fraction anomalies were local (poorly correlated with anomalies at adjacent latitudes) as opposed to regional or global. Sliding one point correlation maps of the inter-model differences OLR_{CLEAR} and OLR_{CLOUD} suggest that inter-model differences in both fields are regional in scale (not shown); individual models tend to have OLR_{CLEAR} and OLR_{CLOUD} anomalies that extend over the entire tropical region, storm-track region, or polar regions with no significant correlation between anomalies in one region and the other region. The meridional de-correlation length scale (where the spatial auto-correlation is equal to e^{-1}) of the OLR_{CLEAR} anomalies is of order 15˚ in the extratropics (~30˚ in the tropics) and is slightly longer than that of OLR_{CLOUD}.

We define $OLR_{CLEAR}*$ and $OLR_{CLOUD}*$ for each model by substituting OLR_{CLEAR} and $\mathrm{OLR_{CLOUD}}$ into the integrand of Eq. (6) with the limits of integration defined from the total *OLR* field. The inter-model spread in $\mathrm{OLR_{CLEAR}}*$ is 0.26 PW (0.26 PW) and the inter-model spread in OLR_{CLOUD}* is 0.25 PW (0.24 PW) in the NH (SH – Table 3). The near equality of the clear-sky and cloud contribution to OLR* spread is consistent with the relative contributions of OLR_{CLEAR} and OLR_{CLOUD} to the OLR spread at each latitude (FIG. 8) and the fact that both inter-model differences OLR_{CLEAR} and OLR_{CLOUD} have similar decorrelation length scales. In the NH (SH), 44% (35%) of the inter-model variance in OLR* is due to differences in OLR_{CLEAR} and 40% (23%) is due to differences in OLR_{CLOUD} (Table 3).

We further subdivide OLR_{CLOUD}* into cloud fraction and cloud structure components by use of Eq. 17. Inter-model differences in cloud fraction and cloud structure make nearly equal contributions to the inter-model spread in OLR_{CLOUD}* (Table 3). This result is consistent with the previous conclusion that cloud structure and cloud fraction make comparable magnitude contributions to the spread in OLR_{CLOUD} with some regional dependence (FIG. 8) and that inter-model differences in cloud fraction and cloud structure are regional in scale (have similar decorrelation length scales – not shown).

The relationship between the equator-to-pole gradient in surface temperature and OLR_{CLEAR}* is analyzed by defining TS*, the surface temperature anomaly (from the global average) averaged over the extratropics:

$$TS^* = \frac{\int_{x(OLR'=0)}^{1} TS'(x)dx}{\int_{x(OLR'=0)}^{1} dx} \quad . \qquad (18)$$

Inter-model differences in TS* explain 81% (85%) of the inter-model spread in OLR_{CLEAR}* (Table 3). The regression coefficient between TS* and OLR* is 0.21 PW K^{-1} which corresponds to 2.0 W m^{-2} OLR_{CLEAR} anomaly per unit temperature anomaly averaged over the polar cap; this number is consistent with linear parameterizations of *OLR* from surface temperature (Warren and Schneider, 1979). A similar quantity for the equator-to-pole gradient in specific humidity, Q^*, can be defined by substituting the vertically integrated specific humidity into the integrand of equation 13. Q^* is not significantly correlated with OLR_{CLEAR}^* in either hemisphere (Table 3).

In summary the inter-model spread in OLR^* is a consequence of nearly equal magnitude contributions from clear-sky and cloud processes. Inter-model differences in both cloud structure and cloud fraction contribute to the spread in OLR_{CLOUD}^* and the vast majority of the OLR_{CLEAR}^* spread is due to inter-model differences in the surface temperature gradient.

C. Relationship between OLR and ASR**

We gain further insight into why inter-model differences in OLR^* and ASR^* do NOT compensate for each other by analyzing the meridional structure of *ASR* and *OLR* anomalies associated with a "typical" ASR^* anomaly from the ensemble average. We regress a normalized index of ASR^* onto the inter-model spread in zonal average *ASR*, *OLR*, OLR_{CLOUD} and OLR_{CLEAR} (FIG. 9). The resulting *ASR* curve shows the anticipated structure of an ASR^* anomaly with anomalously high values in the tropics and low values in the extratropics; both tropical and extratropical anomalies in $\alpha_{P,ATMOS}$ contribute to a "typical" ASR^* anomaly. In contrast, the *OLR* anomaly associated with an ASR^* anomaly only has appreciable magnitude in the tropics that is due to OLR_{CLOUD} anomalies of the same sign as the *ASR* anomalies. We interpret this result as the compensation between *LWCF* and shortwave cloud forcing in the tropics (Keihl 1994 and Hartmann et al. 2001); the same cloud properties that reflect anomalous shortwave radiation also reduce *OLR* by rising the effective longwave emission level. This compensation is not complete over the tropics for the inter-model spread (c.f. the magnitude of the *OLR* and *ASR* curves in tropics in FIG. 9). Over the extra-tropics, there is little compensation between *ASR* and *OLR* anomalies in a "typical" *ASR** anomaly because 1. the *OLR* spread is a consequence of both clear-sky and cloud properties in this region whereas the *ASR* spread is primarily due to cloud properties and 2. the cloud properties the control the inter-model spread in $\alpha_{P,ATMOS}$ are different from the cloud properties that dictate the *OLR* spread[7]. As a consequence, *ASR* and *OLR* anomalies are poorly correlated with eachother over the extratropics leading to *ASR** and *OLR** spread that is only partially compensating.

5. Summary/Discussion

The peak *MHT* in the climate system was diagnosed as the difference between the equator-to-pole gradient of *ASR* (*ASR**) and *OLR* (*OLR**). 65% (59%) of the observed *ASR** in the NH (SH) is a consequence of the meridional distribution of incident solar radiation at the TOA while the remaining 35% (41%) is due to the meridional distribution of planetary albedo. We have demonstrated that the vast majority (86% and 94% in the NH and SH) of the meridional gradient of planetary albedo is a consequence of atmospheric as opposed to surface reflection. These results suggest that surface albedo plays a significantly smaller role in setting equator-to-pole gradient in ASR than atmospheric reflection (e.g. cloud distribution).

The total equator-to-pole gradient in absorbed solar radiation, *ASR*,* and its partitioning into atmospheric and surface albedo components found in the observations is

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[7] The inter-model spread in $\alpha_{P,ATMOS}$ in the Southern Ocean is poorly correlated with cloud fraction whereas the spatial variations in $\alpha_{P,ATMOS}$ within a given model is well correlated with cloud fraction. This result suggests that inter-model variations in the parameterization of cloud albedo as opposed to cloud fraction differences are responsible for the $\alpha_{P,ATMOS}$ spread.

well replicated in the multi-model mean of the CMIP3 PI model simulations in the NH. However, in the SH, the ensemble average ASR^* is smaller than the observed value due to a smaller than observed equator-to-pole gradient in $\alpha_{P,ATMOS}$ (ASR^*_{ATMOS}). As a consequence, the ensemble average MHT_{MAX} is 0.6 PW smaller than the observed value in the SH.

The CMIP3 simulations of the PI climate system exhibit a remarkably large spread (of order 1 PW or 20%) in MHT_{MAX} that exceeds the projected change under global warming by a factor of approximately five (Hwang and Frierson 2011). This spread is due to inter-model differences in the equator-to-pole gradient in *ASR* (ASR^*) and is uncorrelated with inter-model differences in the equator-to-pole gradient in *OLR* (OLR^*). The inter-model spread in ASR^* results from model differences in the meridional gradient of α_P that are primarily (94%) due to differences in cloud reflection ($\alpha_{P,ATMOS}$). As a consequence, total heat transport in the climate models is primarily determined by the optical properties of the atmosphere (FIG. 6). Our definition of MHT_{MAX} in terms of ASR^* and OLR^* is useful tool for analyzing the *MHT* and its inter-model spread because the meridional contrast of *ASR* and *OLR* are governed by different physical processes in the models; ASR^* is primarily controlled by cloud reflection where as cloud fraction, cloud structure, and surface temperature all contribute to OLR^*.

Our results indicate that, in the present climate, the peak MHT_{MAX} is mainly determined by the shortwave optical properties of the atmosphere (i.e., cloud distribution) and suggests that MHT_{MAX} is largely insensitive to subtleties in the model dynamics that contribute to the heat transport (Stone 1978). We can understand this result in the context of simplified energy balance models. In the annual mean, the extratropical deficit in *ASR*,

*ASR**, is balanced by the sum of *OLR* anomalies relative to the global mean (*OLR**) and meridional heat transport into the extratropics (MHT_{MAX}). If the heat transport is diffusive along the surface temperature gradient and the *OLR* anomaly is proportional to the surface temperature anomaly from the global mean (as in Budyko 1969 and Sellers 1969 amongst others) then both the extratropical *OLR* anomaly and MHT_{MAX} are proportional to the same equator-to-pole temperature gradient. The ratio between MHT_{MAX} and *OLR** is then dictated by the relative efficiencies of large scale heat diffusion and radiation to space which is commonly called δ in the literature (see Rose and Marshall 2009 for a review). If two climate models had different δ values yet the same *ASR**, MHT_{MAX} would differ between the models and the inter-model spread in *ASR** and *OLR** would be anti-correlated. For example, a more diffusive model (e.g. a model with more vigorous baroclinic eddies) would have more MHT_{MAX} and less *OLR** and vice versa. In contrast, if δ were nearly equal among climate models but *ASR** varied, then the MHT_{MAX} and *OLR** would be proportional to *ASR** with a regression coefficient dictated by the relative efficiency of dynamic and radiative heat exports (equal to $\frac{\delta}{\delta+1}$ and $\frac{1}{\delta+1}$ -- Donohoe 2011). The positive correlation between *ASR** and *OLR** (FIG. 3C) suggests that the CMIP3 suite of climate models all have a similar δ value such that MHT_{MAX} is dictated by *ASR** which in turn, we have demonstrated is controlled by the meridional distribution of clouds simulated. Furthermore, the relatively steep slope between MHT_{MAX} and *ASR** (a regression coefficient of 0.64 in the NH and 0.85 in the SH – FIG 3A) as compared to the relatively shallow slope between *OLR** and *ASR** (a regression coefficient of 0.36 in the NH and 0.15 in SH—FIG. 3C) suggests that δ is greater than unity; the dynamic export of heat out of the tropics (MHT_{MAX}) is a more efficient pathway towards achieving local energy balance than is the radiative export of energy anomalies (*OLR*). Thus, per unit *ASR** anomaly imposed by the modeled cloud distribution, the extratropical energy budget will be balanced primarily by a MHT_{MAX} anomaly and secondarily by an *OLR** anomaly.

We are currently applying the same diagnostics used in this manuscript toward simulations of anthropogenic climate change, paleoclimate states and idealized climate states. We also hope to more formally examine the inter-model, seasonal, and inter-climate state variability of δ in future work.

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Tables and table captions

Table 1. Variables used in this study.

Symbol	Meaning
MHT_{MAX}	Peak magnitude of meridional heat transport in each hemisphere
ASR^*	The equator-to-pole contrast in Absorbed Solar Radiation
OLR^*	The equator-to-pole contrast of Outgoing Longwave Radiation
X	Sine of latitude
A	Planetary co-albedo
S	Incident solar radiation
$\alpha_{P,ATMOS}$	Atmospheric contribution to planetary albedo
$\alpha_{P,SURF}$	Surface contribution to planetary albedo
ASR_{ATMOS}^*	The contribution of $\alpha_{P,ATMOS}$ to ASR*
ASR_{SURF}^*	The contribution of $\alpha_{P,SURF}$ to ASR*
$LWCF$	Longwave cloud forcing = OLR - OLR_{CLEAR}
C_{FRAC}	Cloud fraction
C_{STRUC}	The contribution of cloud structure to OLR when clouds are present
OLR_{CLEAR}^*	Clear-sky OLR contribution to OLR*
OLR_{CLOUD}^*	Contribution of LWCF to OLR*
$OLR_{CLOUD\ FRAC}^*$	The contribution of cloud fraction anomalies to the inter-model spread in OLR*
$OLR_{CLOUD\ STUC}^*$	The contribution of cloud structure anomalies to the inter-model spread in OLR*
TS^*	The equator-to-pole contrast of surface temperature
Q^*	The equator-to-pole contrast of vertically integrated specific humidity

Table 2. Models used in this study and their resolution. The horizontal resolution refers to the longitudinal and latitudinal grid-spacing or the spectral truncation. The vertical resolution is the number of vertical levels.

Abbreviation	Full Name	Horizontal Resolution	Vertical Resolution
BCCR-BCM2.0	Bjerknes Centre for Climate Research, University of Bergen Norway	T63	L31
CCCMA-CGCM3.1	Canadian Centre for Climate Modeling & Analysis, Canada	T47	L31
CNRM-CM3	Meteo-France/Centre National de Recherches Météorologique, France	T63	L45
CSIRO-MK3.0	Australian Commonwealth Scientific and Research Organization (CSIRO), Australia	T63	L18
GFDL-CM2.0	NOAA/Geophysical Fluid Dynamics Laboratory, USA	2.0˚ X 2.5˚	L24
GISS-ER	NASA/Goddard Institute for Space Studies, USA	4˚ X 5˚	L20
IAP-FGOALS	National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), China	T42	L26
MPI-ECHAM5	Max Planck Institute for Meteorology, Germany	T63	L31
INM-CM3.0	Institute for Numerical Mathematics, Russia	4 X 5	L21
IPSL-CM4.0	Institute Pierre Simon Laplace, France	2.5˚ X 3.75˚	L19
Micro3.2 (Hires)	National Institute for Environmental Studies, and Frontier Research Center for Global Change, Japan	T106	L56
MRI-CGCM2.3.2a	Meteorological Research Institute, Japan	T42	L30
NCAR-CCSM3.0	National Center for Atmospheric Research, USA	T85	L26
UKMO-HADCM3	Hadley Centre for Climate Prediction and Research/Met Office, UK	2.5˚ X 3.8˚	L19
MIUB-ECHOg	University of Bonn, Germany	T30	L19

Table 3. ***ASR****, **its partitioning into incident and planetary albedo components (2nd and 3rd columns) by application of Equation 3 and the subsequent partitioning of the planetary albedo component into atmospheric and surface contributions (4th and 5th columns) by application of Equation 7.** ***OLR**** **and the peak MHT are also shown. The observations and CMIP3 multi-model average and spread (2 standard deviations) are shown for each hemisphere. All entries are in PWs.**

(PW)	Total *ASR**	Incident	Albedo	Atmospheric	Surface	*OLR**	MHT
NORTHERN HEMISPHERE							
Observations	8.2	5.3	2.9	2.5	0.4	2.4	5.8
Model Average	8.1	5.2	2.9	2.4	0.5	2.6	5.5
Model Spread (2σ)	0.9	0.1	0.9	1.2	0.5	0.6	0.8
SOUTHERN HEMISPHERE							
Observations	9.0	5.3	3.7	3.5	0.2	3.2	5.8
Model Average	8.4	5.2	3.2	2.9	0.3	3.2	5.2
Model Spread (2σ)	1.2	0.1	1.2	1.4	0.4	0.5	1.1

Table 4. (Top rows) The inter-model spread (2σ) and square root of co-variance of the terms in the extratropical energy budget (Eq. 4) among the CMIP3 PI simulations. All terms are in units of PW. (Bottom rows) Statistical relationships between the inter-model spread of the variables considered in this study. The squared correlation coefficients (R^2) and regression coefficients (listed in parenthesis when significant) are calculated separately in each hemisphere for the ensemble of 15 models listed in Table 2.

Inter-model Spread --2σ (PW)		
Field	NH	SH
$2\sqrt{\left[ASR^{*\prime 2}\right]}$	0.90	1.20
$2\sqrt{\left[OLR^{*\prime 2}\right]}$	0.58	0.46
$2\sqrt{\left[ASR^{*\prime}\bullet OLR^{*\prime}\right]}$	0.50	0.44
$2\sqrt{\left[MHT_{MAX}{}^{\prime 2}\right]}$	0.78	1.12
R^2 (and regression coefficients when significant) between variables		
Fields	NH	SH
MHT_{MAX} vs. ASR*	0.57 (0.64)	0.85 (0.85)
MHT_{MAX} vs. OLR*	0.02	0.00
OLR* vs. ASR*	0.28 (0.36)	0.15 (0.15)
MHT_{MAX} vs. ASR^*_{ATMOS}	0.63	0.84
ASR* vs. ASR^*_{ATMOS}	0.80 (0.88)	0.93 (0.82)
ASR* vs. ASR^*_{SURF}	0.09	0.21(-1.32)

Table 5. Division of OLR* spread into clear sky ($OLR*_{CLEAR}$) and cloud components ($OLR*_{CLOUD}$-- top rows) and the subsequent division of the cloud contribution into cloud fraction ($OLR*_{CLOUD\ FRAC}$)and cloud structure ($OLR*_{CLOUD\ STRUC}$) components (middle rows). The bottom rows show the correlation of the $OLR*_{CLEAR}$ spread with the equator-to-pole contrast of surface temperature (TS*) and specific humidity (Q*).

	NH		SH	
	Spread -- 2σ	R^2	Spread -- 2σ	R^2
Division of OLR* into clear and cloud components Correlation with OLR*				
$OLR_{CLEAR}*$	0.52 PW	0.44	0.52 PW	0.35
$OLR_{CLOUD}*$	0.50 PW	0.40	0.48 PW	0.23
Division of $OLR_{CLOUD}*$ into fraction and structure components Correlation with $OLR_{CLOUD}*$				
$OLR_{CLOUD\ FRAC}*$	0.44 PW	0.47	0.50 PW	0.30
$OLR_{CLOUD\ STRUC}*$	0.38 PW	0.39	0.52 PW	0.19
$OLR_{CLEAR}*$ correlation with TS* and Q*				
TS*	3.0 K	0.81	1.8 K	0.85
Q*	2.6 kG m^{-2}	0.12	1.6 kG m^{-2}	0.08

FIGURE CAPTIONS

FIG. 1. Graphical demonstration of the calculations of (A) MHT_{MAX}, (B) ASR^*, and (C) OLR^* from the CERES annual average data. The x axis is sine of latitude in all panels. (A) The zonal average *ASR* (red line) and *OLR* (green line). The blue (red) shaded area is the spatially integrated net radiative deficit (surplus) in the extra tropics (tropics) and equals the meridional heat import (export) from each region (MHT_{MAX}). (B) The zonal average *ASR* co-plotted with the global average *ASR*; the shaded area equals ASR^*. (C) As is (B) except for *OLR* and OLR^*. The black circles denote the latitude where $ASR' = 0$ in each hemisphere.

FIG. 2. (A) Meridional structure of meridional heat transport for the observations (thick-solid line) and each of the CMIP3 PI simulations (thin-dashed lines). (B) Histogram of maximum heat transport (MHT_{MAX}) in the Northern Hemisphere (NH). The observational value is shown by the dashed vertical line. (C) As in (B) except for the Southern Hemisphere (SH).

FIG. 3. (A) Maximum heat transport (MHT_{MAX}) versus ASR^* in each the NH and SH (blue and red plus signs) of the CMIP3 PI model ensemble and observations (filled squares). (B) As in (A) except for MHT_{MAX} versus OLR^*. (C) As in (A) except for OLR^* versus ASR^*. The blue and red lines are the linear best fits in the SH and NH and are only shown where significant.

FIG.4. (A) CERES Zonal average *ASR* anomalies from the global average (black) partitioned into incident (red), albedo (blue), and covariance (green) terms via Eq. 10.

(B) As in (A), except combining the albedo and covariance terms into a variable albedo term (blue) as discussed in the text. (C) The subdivision of the variable albedo term into atmospheric (magenta line) and surface reflection (cyan line) terms as discussed in section 3B. (D) The contribution of each of the terms to ASR^* in each hemisphere, calculated from the spatial integral of the curves over the extra-tropics (colors are the same as the curves in the previous panels).

FIG.5. (A) ASR^* versus atmospheric reflection contribution to ASR^*(ASR^*_{ATMOS}) in each the NH and SH (blue and red plus signs) of the CMIP3 PI model ensemble and observations (filled squares). The theoretical prediction of "Model A", as discussed in the text, is given by the black line. (B) As in (A) except plotted against surface albedo contribution to ASR^* (ASR^*_{SURF}). (C) As in (B) except for the surface albedo gradient ($SURF^*$). The theoretical prediction of "Model B", as discussed in the text, is given by the black line. The blue and red lines are the linear best fits in the SH and NH and are only shown where significant.

Figure 6. MHT_{MAX} versus atmospheric reflection contribution to ASR^* (ASR^*_{ATMOS}) in each the NH and SH (blue and red plus signs) of the CMIP3 PI model ensemble and observations (filled squares). The blue and red lines are the linear best fits in the SH and NH and are only shown where significant.

FIG. 7. (A) CMIP3 inter-model spread in OLR decomposed in cloud ($LWCF$) and clear sky (OLR_{CLEAR}) components as described in the text. (B) The LWCF contribution to the inter-model spread in OLR decomposed into cloud fraction ($LWCF_{FRAC}$) and cloud

structure (*$LWCF_{STRUC}$*) components. (C) The OLR_{CLEAR} contribution to the inter-model spread in *OLR* decomposed into components that are linearly congruent with the surface temperature spread ($OLR_{CLEAR\ TS}$) and the vertically integrated specific humidity spread ($OLR_{CLEAR\ Q}$).

FIG. 8. Regression of the normalized inter-model spread in ASR* on to the inter-model anomalies of *ASR* (black), *OLR* (green), OLR_{CLEAR} (red), and *LWCF* (blue). The resulting curves are the radiative anomalies associated with a one-standard deviation *ASR** anomaly.