Interannual Variability of the Global Radiation Budget

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ABSTRACT

Interannual variability of the global radiation budget, regions that contribute to its variability, and what limits albedo variability are investigated using Clouds and the Earth’s Radiant Energy System (CERES) data taken from March 2000 through February 2004. Area-weighted mean top-of-atmosphere (TOA) reflected shortwave, longwave, and net irradiance standard deviations computed from monthly anomalies over a 1° x 1° region are 9.6, 7.6, and 7.6 W m\(^{-2}\), respectively. When standard deviations are computed from global monthly anomalies, they drop to 0.5, 0.4, and 0.4 W m\(^{-2}\), respectively. Clouds are mostly responsible for the variation. Regions with a large standard deviation of TOA shortwave and longwave irradiance at TOA are the tropical western and central Pacific, which is caused by shifting from La Niña to El Niño during this period. However, a larger standard deviation of 300–1000-hPa thickness anomalies occurs in the polar region instead of the tropics. The correlation coefficient between atmospheric net irradiance anomalies and 300–1000-hPa thickness anomalies is negative. These indicate that temperature anomalies in the atmosphere are mostly a result of anomalies in longwave and dynamical processes that transport energy poleward, instead of albedo anomalies by clouds directly affecting temperature anomalies in the atmosphere. With simple zonal-mean thermodynamic energy equations it is demonstrated that temperature anomalies decay exponentially with time by longwave emission and by dynamical processes. As a result, the mean meridional temperature gradient is maintained. Therefore, mean meridional circulations are not greatly altered by albedo anomalies on an annual time scale, which in turn provides small interannual variability of the global mean albedo.

1. Introduction

The interannual variability of top-of-atmosphere (TOA) irradiances observed by Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996) instruments is remarkably small. Four years of CERES data from March 2000 through February 2004 indicate that the annual and global mean TOA reflected shortwave irradiance is 97.0 W m\(^{-2}\), with maximum and minimum values of 97.2 and 96.8 W m\(^{-2}\), respectively. The difference between the maximum and minimum values is 0.4% of the mean value. Similarly, the annual mean TOA longwave irradiance is 239 W m\(^{-2}\) and the difference between the maximum and minimum values is 0.3 W m\(^{-2}\), which is only 0.1% of the mean value. Somewhat larger variability of TOA reflected shortwave, longwave, and net irradiance is also reported by Duvel et al. (2001), who used Earth Radiation Budget Experiment (ERBE), Scanner for Radiation Budget (ScaRaB)-Meteor Satellite, and ScaRaB data, but it is still small compared with the mean values. In addition, according to Loeb et al. (2007b), shortwave variability from the International Satellite Cloud Climatology Project (ISCCP; Schiffer and Rossow 1983) is consistent to 40% with that from CERES. To understand possible reasons for the small interannual variability of global and monthly mean TOA shortwave and longwave irradiances at TOA, we need to understand which regions contribute and the ways in which the variation is correlated with the variation of atmospheric and surface properties. Smith et al. (1990) and Bess et al. (1992) analyzed the interannual variability of absorbed shortwave and longwave irradiances at TOA, respectively, observed by Nimbus-6 and -7 Earth Radiation Budget Instruments. They found that El Niño–Southern Oscillation (ENSO) causes the largest interannual variability when seasonal cycles are removed. Work by Loeb et al. (2007a) also suggests that most albedo variations are in the tropics and they are highly correlated with cloud cover.

In this paper, the interannual variability of global radiation budget is further investigated using CERES...
data taken from the *Terra* platform. Specifically, the magnitude of the variation, where larger variations happen, and what limits the variability are investigated. Section 2 describes data used in this study. Section 3 discusses the standard deviation of anomalies computed in different ways, quantifies the variability, and identifies processes cause large variability. In section 4, a simple model based on zonal mean thermodynamic equation is used to qualitatively show that temperature anomalies decay exponentially with time.

2. Method

The 48 months of reflected shortwave and longwave irradiances at TOA estimated from observed CERES broadband radiances from March 2000 through February 2004 were used in this study. TOA reflected shortwave and longwave irradiances were from a CERES data product, the Single Scanner Footprint (SSF) edition 2B (Ed2B). The rev1 correction (Matthews et al. 2005) was applied to account for a small drift of shortwave CERES instrument calibrations. Surface shortwave and longwave irradiances used in this study to estimate atmospheric irradiances were from a CERES data product, Clouds and Radiative Swath (CRS) Ed2B. Forty-eight months of surface irradiances starting from March 2000 were also used. The instantaneous TOA and surface shortwave irradiances were converted to a daily value with the assumption that meteorology does not change over the course of a day. The conversion process is described in Kato et al. (2008). The daily value of the longwave irradiance is simply the mean of daytime and nighttime values weighted by the day–night fraction of a given day. Daily mean shortwave and longwave irradiances were averaged over a $1^\circ \times 1^\circ$ grid box and over 1 month. The monthly mean value computed from 4 months of $F(x)$ was then subtracted from each month value $F(t, x)$ to obtain a deseasonalized anomaly (hereafter anomaly) $\Delta F$,

$$\Delta F(t, x) = F(t, x) - \overline{F(t, x)},$$

where $t$ and $x$ indicate that the irradiance is a function of time (month) and location ($1^\circ \times 1^\circ$ grid), respectively. The standard deviation of $\Delta F(t)$ for given $x$ was computed as the measure of the interannual variability of the shortwave or longwave irradiances for the region. In the following section, we investigate the standard deviation of these deseasonalized anomalies. Note that the uncertainty in the absolute calibration of CERES instruments (Loeb et al. 2009) does not affect the variability estimate unless the calibration shifts with time. A study by Loeb et al. (2007b) indicates that the stability of CERES shortwave instruments on *Terra* is approximately within 0.3 W m$^{-2}$ decade$^{-1}$. The uncertainty associated with the conversion of instantaneous irradiance to daily mean with a constant meteorology assumption is discussed in Kato et al. (2008). Briefly, if the uncertainty in the accuracy is defined as the difference between modeled and observed irradiances, the uncertainty in the daily mean downward shortwave irradiance at the surface is 7.8% and the downward longwave irradiance at the surface is 1.1%. The uncertainty in the net atmospheric shortwave irradiance is 16.1%, the net atmospheric longwave irradiance is 4.4%, and the net atmospheric irradiance is 10.8%. Similar to TOA irradiances, however, the stability of the surface and atmospheric irradiances is expected to be better than these uncertainties.

3. Result

a. Global TOA irradiance variability

The area-weighted global annual mean TOA all-sky albedo (the global annual mean TOA upward shortwave irradiance divided by the global annual mean downward shortwave irradiance) is 0.283, while the clear-sky albedo counterpart is 0.171 (Fig. 1). These correspond to the TOA reflected shortwave irradiance of 96.8 W m$^{-2}$ for all-sky conditions and 58.5 W m$^{-2}$ for clear-sky conditions, respectively. Similarly, the area-weighted global annual mean TOA all-sky longwave irradiance is 239 W m$^{-2}$ and the clear-sky value is 267 W m$^{-2}$. The all-sky TOA reflected shortwave (longwave) irradiance is significantly larger (smaller) than the clear-sky value, and it is 36 W m$^{-2}$ larger (23 W m$^{-2}$ smaller) than the clear-sky maximum TOA reflected shortwave (minimum longwave) irradiance of 60.8 (262) W m$^{-2}$ during the period. To show the variability difference under all-sky and clear-sky conditions, the area-weighted mean standard deviations from global $1^\circ \times 1^\circ$ TOA reflected shortwave, longwave, and net (TOA net shortwave minus TOA longwave) irradiance anomalies are plotted as a function of month (Fig. 2). The standard deviation for a given month is computed by

$$\sigma_{ts} = \sqrt{\frac{1}{n_m m} \sum_{j=1}^{n_m} \sum_{i=1}^{m} w_i \Delta F(t, x)^2},$$

where $w_i$ is the area weight, $n_m$ is the number of months used in the study for a given month of a year, and $m$ is the number of $1^\circ \times 1^\circ$ regions over the globe. Figure 2 shows, therefore, a typical $1^\circ \times 1^\circ$ variation of the TOA reflected shortwave, longwave, and net monthly anomalies as a
function of month. Note that the TOA net irradiances are defined as the TOA downward shortwave irradiance minus the TOA reflected shortwave minus the TOA longwave irradiance. Note also that computed clear-sky irradiances with all-sky atmospheric conditions were used in Fig. 2 to minimize the effect of atmospheric properties other than clouds and make the number of clear-sky samples equal to the number of all-sky samples. When the TOA reflected shortwave, longwave, and net irradiance standard deviations from $1^\circ \times 1^\circ$ regions are averaged over a year, they are 9.6, 7.6, and 7.6 W m$^{-2}$, respectively. The standard deviation of the TOA reflected shortwave, longwave, and net irradiance under clear-sky conditions is 3.4, 3.2, and 4.6 W m$^{-2}$, respectively. Clouds, therefore, increase the standard deviation of TOA reflected shortwave and longwave anomalies by more than a factor of 2. Because there is a significant anticorrelation between TOA reflected shortwave and longwave cloud radiative effects (Kiehl 1994), an increase of the TOA net irradiance standard deviation by clouds from the clear-sky value is smaller, $\approx 65\%$, when it is averaged over a year.

To demonstrate that monthly anomalies from $1^\circ \times 1^\circ$ grid boxes partially cancel when they are averaged over a larger region, we computed the global mean standard
deviation of TOA shortwave, longwave, and net irradiances at TOA in several different ways. When anomalies are averaged over a 1° latitude zone and the area-weighted mean standard deviation is then computed from 48 × 180 zonal monthly anomalies, the standard deviations are reduced, respectively, to 2.4, 2.1, and 2.0 W m⁻². When anomalies are averaged over the globe and the standard deviation is then computed from 48 global monthly anomalies, the standard deviations of TOA reflected shortwave, longwave, and net irradiance anomalies are further reduced, respectively, to 0.5, 0.4, and 0.4 W m⁻².

b. Regional shortwave and longwave variations

The standard deviation of 48 anomalies over a 1° × 1° region x was computed by

$$\sigma_t = \sqrt{\frac{1}{n}\sum_{j=1}^{n} \Delta F(t_j, x)^2},$$

where n (=48) is the number of months. The map of $\sigma_t$ is shown in Fig. 3. Larger standard deviations of 1° × 1° TOA monthly reflected shortwave irradiance anomalies occur over the tropical western and central Pacific, the eastern Pacific, the tropical Indian Ocean, Europe, eastern Asia, and parts of North and South America regions (Fig. 3a). While standard deviations over storm-track regions—the northwest Pacific and the northwest Atlantic Ocean—are smaller than that in the tropical western Pacific, the mean TOA reflected shortwave irradiance over the northern Pacific storm-track region is larger than that over the tropical western Pacific; the annual mean TOA reflected shortwave irradiance over the northern Pacific is approximately 140 W m⁻² compared to the approximately 100 W m⁻² over the tropical western Pacific. Significant standard deviations of TOA reflected shortwave anomalies are caused by disturbances that last more than 1 month and occupy a larger part of a 1° × 1° grid box. Such anomalies involve an ocean system because a typical midlatitude frontal system moves of the order of 10 m s⁻¹, which takes a few days to move 1000 km. The variability resulting from the synoptic system, therefore, mostly cancels when the TOA reflected shortwave irradiance is averaged both over 1 month and a 1° × 1° region. This is apparent in Fig. 3; large standard deviations occur predominantly over the oceans. Anomalies resulting from ENSO persist for more than 6 month or longer (Trenberth 1997). The Multivariate ENSO Index (Wolter and Timlin 1998) changed from negative to positive during the period from March 2000 through February 2004, indicating that the tropical Pacific Ocean was changed from La Niña to El Niño. Therefore, ENSO is responsible for large standard deviations over the tropical western Pacific during this period.

The standard deviation of TOA longwave irradiance anomalies shows a similar pattern as the TOA reflected shortwave irradiance, but the contrast between the tropics and midlatitudes is more pronounced (Fig. 3, middle plot). This is probably because the frequency of

![Fig. 2. Standard deviation of the TOA (a) shortwave, (b) longwave, and (c) net (net shortwave minus longwave) irradiances as a function of month. All sky (closed circles) and clear sky (open circles). The standard deviation $\sigma_t$ given by (2) is computed from global 1° × 1° monthly deseasonalized anomalies over the period from March 2000 through February 2004 and averaged over the globe weighting by the area.](image-url)
Fig. 3. Standard deviation of the TOA reflected (top) shortwave, (middle) longwave, and (bottom) net irradiances. The standard deviation $\sigma_i$ given by (3) is computed over $1^\circ \times 1^\circ$ regions based on monthly deseasonalized anomalies over 4 yr and averaged over $5^\circ \times 5^\circ$ regions for plotting purposes. The TOA net irradiance is defined as the TOA net shortwave irradiance minus TOA longwave irradiance.
occurrence of thick convective clouds and their cloud-
top height decrease with increasing latitude, which re-
duces the longwave anomaly standard deviation. When
the TOA reflected shortwave and longwave irradiances
are combined for the TOA net irradiance, regions
that have large TOA reflected shortwave and longwave
variabilities are no longer prominent (Fig. 3, bottom
plot). Because a large positive TOA reflected shortwave
anomaly resulting from thick convective clouds also
causes a large negative longwave anomaly (Kiehl 1994;
Cess et al. 2001), such an anomaly has a smaller impact
on the TOA net irradiance. Large standard deviations of
the TOA net irradiance are also over the oceans. Re-
gions with a larger standard deviation are eastern Pacific
subtropical regions, where low-level clouds are present,
and the region off the coast of the Antarctic Peninsula.
The top and bottom plots of Fig. 3 suggest that large
TOA net irradiance standard deviations over these re-
gions are caused by large TOA reflected shortwave
standard deviations. Low-level cloud property changes
predominately affect the TOA reflected shortwave irradiance if
clouds are absent.

We further separate the atmosphere from the surface
and analyze the contribution of clouds to the variability
of the net atmospheric irradiance to understand the
cloud radiative effect on the variability of atmospheric
energy deposition. The atmospheric net irradiance
$F_{\text{atm,net}}$ is the TOA net irradiance minus surface net
irradiance,

\[ F_{\text{atm,net}} = \left( F_{\text{dn,sw}}^{\text{toa,sw}} - F_{\text{up,sw}}^{\text{toa,sw}} - F_{\text{up,lw}}^{\text{toa,lw}} \right) 
- \left( F_{\text{dn,sfc,sw}} - F_{\text{up,sfc,sw}} + F_{\text{dn,sfc,lw}} - F_{\text{up,sfc,lw}} \right), \]  

where subscripts toa, sfc, sw, and lw indicate, respec-
tively, the top of atmosphere, surface, shortwave irra-
diance, and longwave irradiance, and superscripts up and
dn indicate, respectively, upward and downward. The atmospheric net irradiance $F_{\text{atm,net}}$ is also the sum of the atmospheric shortwave irradiance $F_{\text{atm,sw}}$ and atmospheric longwave irradiance $F_{\text{atm,lw}}$, where

\[ F_{\text{atm,sw}} = \left( F_{\text{dn,sw}}^{\text{toa,sw}} - F_{\text{up,sw}}^{\text{toa,sw}} \right) - \left( F_{\text{dn,sfc,sw}} - F_{\text{up,sfc,sw}} \right), \]  

and

\[ F_{\text{atm,lw}} = - F_{\text{up,sw}}^{\text{toa,lw}} - \left( F_{\text{dn,sfc,lw}} - F_{\text{up,sfc,lw}} \right). \]  

Figure 4 shows the standard deviation of TOA irradi-
ances as a function of latitude, and Fig. 5 shows the
standard deviation of atmospheric net irradiances as a
function of latitude for all-sky and clear-sky conditions.
The standard deviation is computed in two different
ways. The first value, which is shown by solid lines in
Figs. 4 and 5, was computed from $1^\circ \times 1^\circ$ anomalies for a
given latitudinal zone and averaged over $n$ months, such that

\[ \sigma_{\text{tx}} = \sqrt{\frac{1}{nm^2} \sum_{j=1}^{m} \sum_{i=1}^{n} \Delta F(t_j, x_i)^2}, \]
where \( m_z \) is the number of \( 1^\circ \times 1^\circ \) regions for a given \( 1^\circ \) zone. This standard deviation provides a mean standard deviation of \( 1^\circ \times 1^\circ \) region within a latitudinal zone. The second value, which is shown by dashed lines in Figs. 4 and 5, was computed from 4 yr of monthly zonal anomalies such that,

\[
\sigma_{tx} = \sqrt{\frac{1}{n} \sum_{j=1}^{n} \left[ \frac{1}{m_z} \sum_{j=1}^{m_z} \Delta F(t_j, x_j) \right]^2},
\]

which provides the temporal variation of zonal anomalies. A smaller \( \sigma_{tx} \) than \( \sigma_{tx} \) indicates that \( 1^\circ \times 1^\circ \) anomalies move along the longitude or \( 1^\circ \times 1^\circ \) anomalies are partially canceled when anomalies are averaged along the longitude. Figure 5 also indicates that clouds increase the standard deviation of the atmospheric net irradiance in a latitudinal zone at all latitudes for all seasons. The implication of this increasing variability to atmospheric energy transport is following. Because eddy potential energy is generated by increasing the standard deviation of atmospheric temperature for a given latitudinal zone (Lorenz 1955; Peixoto and Oort 1992), a larger standard deviation of \( 1^\circ \times 1^\circ \) anomalies in a given latitudinal zone under all-sky conditions than that under clear-sky conditions suggests that clouds increase the generation of eddy potential energy.

The above results indicate that clouds increase the variability of the reflected shortwave, longwave, and net irradiances at TOA, as well as the variability of the atmospheric net irradiance. To understand the
contribution of cloud cover change to the variation of these irradiances, the regional correlation coefficients between TOA reflected shortwave and cloud cover anomalies and between TOA longwave and cloud cover anomalies are shown in Fig. 6. Over most of the tropics and a part of the midlatitudes, the correlation coefficient between TOA reflected shortwave irradiance and cloud cover anomalies is greater than 0.8. The correlation coefficient between TOA longwave irradiance and cloud cover is less than −0.8 over most of the tropics and the Northern Hemisphere midlatitudes, except for regions where predominately low-level clouds are present.

When the TOA reflected shortwave and longwave irradiance anomalies are combined with surface irradiance anomalies for the atmospheric net irradiance anomalies, tropical regions where both TOA reflected shortwave and longwave correlations are large maintain a large correlation with cloud cover (Fig. 6, bottom plot). However, the correlation in the midlatitudes is smaller even though the correlation both between TOA reflected shortwave irradiance and cloud cover anomalies and between TOA longwave irradiance and cloud cover anomalies is large. The correlation between the atmospheric net irradiance and cloud cover decreases with latitude and it is negative at high latitudes. The reason for the weak correlation in the midlatitudes is that the atmospheric net irradiance depends not only on cloud cover, but it also depends on either cloud height or cloud effective temperature. The sign changes around the 500-hPa pressure level (Fig. 7). Because high- and low-level clouds are both frequently present in midlatitudes and the sign of the atmospheric cloud radiative effect depends on the height, the correlation coefficient is reduced in the midlatitudes. In polar regions, where low-level clouds are dominant, the correlation between the net atmospheric irradiance and cloud cover is negative.

4. Discussion

The interannual variation of global mean reflected shortwave and longwave irradiances at TOA are less than 0.5% of the respective mean value during the 4-yr period from March 2000 through February 2004. Larger variabilities appear in the tropics, and are caused by clouds responding to ENSO, that is, the atmospheric response to regional variations of the sea surface temperature. Atmospheric processes are volatile compared with oceanic processes, but oceanic processes alone do not affect TOA reflected shortwave irradiance very much. Interactions between the ocean and the atmosphere provide persisting anomalies of both TOA reflected shortwave and longwave irradiance. For example, Norris and Klein (2000) investigated the variability of upward velocity at the 500-hPa level over the North Pacific and found that it is correlated with sea surface temperature variability. A large part of the energy input to the atmosphere in the tropical western Pacific comes from the ocean as the surface enthalpy flux (Trenberth et al. 2002). According to a bulk formula, increasing the temperature gradient between the sea surface and lower atmosphere increases the energy inputs to the atmosphere (Fairall et al. 1996).

The variation of the surface flux to the atmosphere in tropics is, therefore, predominantly due to anomalous ocean processes that cause the anomalous regional sea surface temperature. A significant part of El Niño can be modeled just shifting around a warm pool of seawater in the tropical Pacific (Enfield 1989). During El Niño, anomalous convection migrates eastward with the 29°C sea surface temperature isotherm (Enfield 1989). These suggest that most of variations in the TOA reflected shortwave and longwave irradiances over the tropics come from a shifting warm pool of seawater instead of local variations of either heating or cooling of the seawater by radiation. A smaller \( \sigma_{\text{rs}} \) than \( \sigma_{\text{r}} \) over the tropics, shown in Figs. 4 and 5, agrees that warm water predominantly moves longitudinally in the tropics. In addition, according to National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST version 2 (Reynolds et al. 2002), the mean sea surface temperature of tropical Pacific Ocean (30°N–30°S, 120°E–105°W) averaged over the 4-yr period (March 2000–February 2004) is 25.6°C and the standard deviation of monthly mean values is 0.16°C. Therefore, the standard deviation of TOA reflected shortwave and longwave irradiances over tropical Pacific is predominately due to shifts of warm seawater along the longitude.

There are few disturbances that persist in a large area, as demonstrated in Figs. 4 and 5, and for a long time (e.g., more than a year). This might be because large-scale dynamics is driven by the meridional temperature gradient that is determined by solar energy inputs and poleward energy transport. Stone (1978), using a simple model, concluded that the energy transport by dynamics are primarily controlled by the solar constant, size of the earth, the tilt of the earth’s axis, and hemispheric mean albedo. In other words, the solar energy input to the earth as a function of latitude controls poleward energy transport and the meridional temperature gradient. As long as the meridional temperature gradient is nearly constant, large-scale dynamics that transport energy...
FIG. 6. Correlation coefficient of TOA (top) reflected shortwave, (middle) longwave, and (bottom) net atmospheric irradiance with cloud cover. Monthly values averaged over a $5^\circ \times 5^\circ$ grid box are used to compute the correlation coefficient. The contour line increment is 0.2. Areas with a value greater than 0.8 for shortwave, less than $-0.8$ for longwave, and greater than 0.8 and less than $-0.8$ for the atmospheric net are shaded with a contour increment of 0.1.
poleward and the TOA longwave irradiance may be stable. Because clouds are generated by dynamical processes and the albedo of the earth or energy input to the earth largely depend on the cloud cover (Fig. 6), this implies that the interannual variability of cloud cover is also small. According to cloud cover retrieved from Moderate Resolution Imaging Spectroradiometer by the CERES cloud algorithm (Minnis et al. 2008), the interannual variability of cloud cover is indeed less than 1% of the mean cloud cover. However, why is the variability of cloud cover, which affects absorption of the shortwave irradiance by the earth system, so small compared with the mean cloud cover? In other words, what prevents the albedo perturbation by clouds to intensify with time? The compensation of shortwave anomalies by longwave anomalies in the tropics reduces the effect on the large-scale dynamics, but Fig. 5 shows that there is significant variability of the net atmospheric irradiance in the tropics.

In the remaining part of this section, using zonal mean momentum and thermodynamic energy equations, we investigate whether or not the energy of the atmosphere explains a small interannual variability of the cloud cover. In the following, albedo variability is treated as variability in forcing to the system. In addition, a basic assumption is that large-scale dynamics that transports energy poleward as a response to the forcing depends on the meridional temperature gradient, which in turn controls the global mean cloud cover and albedo. The zonal mean momentum and thermodynamic energy

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**Fig. 7.** Contour of the (first row) daily mean cloud shortwave radiative effect, (second row) longwave radiative effect, and (third row) shortwave plus longwave radiative effect (net) to the atmosphere for Antarctic latitudes (60°–90°S), Southern Hemisphere midlatitudes (30°–60°N), the tropics (30°N–30°S), Northern Hemisphere midlatitudes (30°–60°N), and Arctic latitudes (60°–90°N) as a function of the cloud optical thickness (τ) and cloud-top height in the pressure coordinate estimated from July 2002 data. Only single-layer clouds are used. (fourth row) The logarithm (base 10) of the 2D normalized histogram of cloud occurrence. Shortwave effects are converted to daily values and daily mean longwave effects are computed by weighting daytime and nighttime longwave irradiances by the number of samples.
equations for quasigeostrophic motions on the midlatitude \(\beta\) plane are (Holton 1992, 369–371)

\[
\frac{\partial \mathbf{v}}{\partial t} - f_0 \mathbf{v} = \frac{\partial (\mathbf{u} \cdot \mathbf{v})}{\partial y} + \mathbf{X} \quad \text{and} \quad (9)
\]

\[
\frac{\partial T}{\partial t} - N^2 H W = -\nabla \cdot (\mathbf{u} T) - \frac{\partial (\mathbf{v} T)}{\partial y} + \frac{T}{c_p}, \quad (10)
\]

where \(u, v,\) and \(w\) are the zonal, meridional, and vertical components of wind, \(f_0\) is the Coriolis parameter, \(T\) is the temperature, \(N\) is the buoyancy frequency, \(H\) is the depth of the atmosphere, \(R\) is the gas constant, \(X\) is the drag force, \(J/c_p\) is the diabatic heating, and \(y\) increases toward the north. The overbar indicates the mean and the prime indicates the deviation from the mean along the longitude. The diabatic heating term is composed of

\[
\mathcal{J} = \frac{F_{\text{atm,sw}} + F_{\text{atm,lw}} + F_{\text{sh}} + F_{\text{precip}}}{\rho_0 H}, \quad (11)
\]

where \(F_{\text{atm,sw}}\) and \(F_{\text{atm,lw}}\) are atmospheric shortwave and longwave irradiance and \(F_{\text{sh}}\) is the sensible heat flux from the surface, and \(F_{\text{precip}}\) is the latent heat of vaporization multiplied by the mean precipitation rate.

We start with the zonal mean thermodynamic energy Eq. (10), and apply the equation to monthly mean values. We then separate each term into the climatological monthly mean and anomaly. We assume that climatological means and anomalies both satisfy (10) separately. We also assume that the vertical motion anomaly is proportional to the temperature anomaly,

\[
C \Delta T = \Delta w \frac{N^2 H}{R}, \quad (12)
\]

and the advection anomaly is proportional to the second derivative of temperature anomaly with respect to \(y\),

\[
-\Delta w \frac{\partial \Delta T}{\partial y} - \frac{\partial (\Delta v \Delta T)}{\partial y} = D \frac{\partial^2 \Delta T}{\partial y^2}. \quad (13)
\]

The sign of \(D\) is positive but \(C\) can be positive or negative depending on the correlation between \(\Delta T\) and \(\Delta w\). In addition, we separate the diabatic term into two parts—the portion that is linearly related to the temperature anomaly and the residual that is treated as a forcing. The longwave anomaly is related linearly to the temperature anomaly,

\[
\Delta F_{\text{lw}} = -4a \Delta T^3 \Delta T, \quad (14)
\]

where the negative sign indicates that the atmospheric longwave irradiance has a cooling effect. Atmospheric shortwave irradiance, latent heat, and sensible heat anomalies are treated as forcing so that the diabatic anomalies are expressed as

\[
\frac{\Delta T}{c_p} = a \Delta T + \frac{\Delta T'}{c_p}, \quad (15)
\]

where \(\Delta T'\) is the forcing term. The forcing term can also include diabatic heating that is not related to the temperature anomaly directly, such as the direct aerosol radiative forcing. The coefficient \(a\) is negative because the atmospheric longwave irradiance dominates in the first term on the right-hand side. The anomalies equation of (10) then becomes

\[
\frac{\partial \Delta T}{\partial t} = D \frac{\partial^2 \Delta T}{\partial y^2} + (a + C) \Delta T + \frac{\Delta T'}{c_p}. \quad (16)
\]

If we assume that the system is forced by

\[
\frac{\Delta T'}{c_p} = J e^{\lambda t} e^{(a + \alpha) y}, \quad (17)
\]

then \(\Delta T\) that satisfies (16) is

\[
\Delta T = J \rho_0 e^{\lambda t} e^{(a + \alpha) y} + \frac{1}{\lambda} \int_0^t e^{\lambda (t-\tau)} \left[ \frac{\alpha}{\lambda} \right] \frac{\Delta T}{c_p} d\tau + \frac{\Delta T'}{c_p}, \quad (18)
\]

where

\[
\cos \phi = \frac{Dk^2 - C - a}{((Dk^2 - C - a)^2 + \omega^2)^{1/2}}, \quad \text{and} \quad (19)
\]

\[
\sin \phi = \frac{\omega}{((Dk^2 - C - a)^2 + \omega^2)^{1/2}}; \quad (20)
\]

here, \(\Delta T_0\) is the temperature anomaly at \(t = 0\), and \(\lambda^2\) is a positive constant determined by boundary conditions. The forcing term can have more terms, composed with a Fourier series, but the result discussed below does not change by this simplification.

Because \(D > 0, \alpha > 0,\) and \(Dk^2 - C - a\) is likely to be positive (unless \(k\) is small and \(C + a\) is a large positive number), both \(\cos \phi\) and \(\sin \phi\) are positive. Therefore, we expect that \(\Delta T\) is positively correlated with the forcing term \(\Delta T'\). The phase shift increases and the correlation decreases as the ratio of \(\omega\) to \(Dk^2 - C - a\) increases. We also expect that the correlation between the temperature and longwave anomalies is negative because of (14).

Figure 8 shows the correlation coefficient between the anomalies of the geopotential height difference at
Fig. 8. Correlation coefficients between 300- and 1000-hPa thickness anomalies and atmospheric shortwave (SW), atmospheric longwave (LW), and atmospheric net [shortwave + longwave (NET)] irradiance anomalies. The coefficients are computed with 1° × 1° anomalies averaged over 1° latitudinal zone; 48 months of data from March 2000 through February 2004 are used.

The first two terms on the left-hand side of (21) are the change of the thermal wind relation with time. Clouds perturb the meridional gradient of diabatic heating in the atmosphere by warming in the tropics and cooling in polar regions (Stuhlmann and Smith 1988; Zhang and Rossow 1997; Kato et al. 2008). Therefore, the cloud radiative effect on the term \( R(Hc_p) \partial T/\partial y \) is negative. Because of warming in the tropics and cooling in polar
regions, clouds increases the meridional energy transport in the midlatitude and polar regions (Zhang and Rossow 1997; Kato et al. 2008). Meridional energy transport by eddies is a convex function of latitude (Peixoto and Oort 1992). If we assume that the cloud radiative effect on the meridional energy transport is also a convex function and has a maximum somewhere in the midlatitude or polar regions, \( \partial^2 \left( \bar{v} T^* \right)/\partial y^2 \) is negative. The atmospheric cloud radiative effect almost linearly decreases with latitude (Kato et al. 2008) so that the cloud radiative effect on \( R/H \partial/\partial y (w \partial \bar{T}/\partial y) \) is small. The cloud effect on \( N^2 \partial w/\partial y \) is also negative because increasing \( w \) by clouds decreases with latitude. Therefore, if we assume clouds predominantly affect temperature and upward velocity, (21) suggest that the cloud effect on the diabatic heating \( R(He_{atm}) \partial J/\partial y \) is compensated by two terms, \( \partial^2 \left( \bar{v} T^* \right)/\partial y^2 \) and \( N^2 \partial w/\partial y \). Therefore, if we assume that clouds predominantly affect the temperature and vertical velocity, the cloud radiative effect on the atmospheric temperature is partially canceled by meridional and vertical energy transport by dynamics. Recall that clouds increase the standard deviation of net irradiance in a latitudinal zone (Fig. 5), which can consequently increase meridional energy transport in the atmosphere by eddies, and clouds alter radiation field in a way that radiative anomalies can be compensated by energy transport by dynamics. In other words, the coefficient \( D_2 \) in (18) increases with the cloud radiative effect and the temperature anomalies generated by clouds decay exponentially. As a result, thermal wind balance is not altered on the annual time scale.

Figures 9a,b show \( \sigma_{\bar{r}} \) and \( \sigma_{\bar{r}} \) from \( \Delta F_{\text{atm,sw}} \), \( \Delta F_{\text{atm, lw}} \), and \( \Delta F_{\text{atm, net}} \). While \( \sigma_{\bar{r}} \) which is a mean 1\(^{st}\) \times 1\(^{st}\) standard deviation within a latitudinal zone, is large over the tropics (Fig. 9b), \( \sigma_{\bar{r}} \) which is a mean zonal standard deviation, is smaller than \( \sigma_{\bar{r}} \) and is nearly constant with latitude, except for the polar regions (Fig. 9a). We also computed \( \sigma_{\bar{r}} \) and \( \sigma_{\bar{r}} \) from 300- to 1000-hPa thickness anomalies for the same period. Unlike irradiances, \( \sigma_{\bar{r}} \) and \( \sigma_{\bar{r}} \) computed from 300- to 1000-hPa thickness anomalies are about the same magnitude. In addition, both \( \sigma_{\bar{r}} \) and \( \sigma_{\bar{r}} \) increase with latitude, which is different in shape from those computed with irradiance anomalies. Figure 9 indicates that atmospheric irradiance anomalies, especially atmospheric shortwave irradiance anomalies, are not direct driver of atmospheric temperature anomalies in the midlatitude and polar regions. Instead, based on the above results, it is postulated that temperature anomalies in the midlatitude and polar regions are the result of the variability in energy transport from tropics to polar regions.

In summary, the interannual variability of global and annual albedo is small for the following two reasons: 1) For most of the globe, the variability of atmospheric temperature is caused by the variability of dynamics that transport energy poleward and atmospheric longwave irradiance, which reduce the temperature anomaly in the atmosphere. 2) Albedo variability, which regulates the variability of solar irradiance input to the earth system, does not increase with time. To augment the albedo anomalies, temperature anomalies need to alter dynamical processes, which in turn can alter albedo by altering clouds. Temperature anomalies are, however, damped by dynamical processes that transport energy poleward and by longwave emission. As a result, the mean meridional temperature gradient is maintained and the mean meridional circulation is not altered by albedo anomalies on an annual time scale. If large-scale dynamics determines the global mean cloud cover, therefore, the interannual variability of the global and annual mean albedo is small.

While these results do not answer why the global and annual mean albedo is at the current value, they offer a qualitative explanation toward understanding why the interannual variation of global albedo is so small. The results also indicate that albedo variability affects atmospheric temperature variability indirectly and that the interannual variability of temperature is predominately caused by the variability of the atmospheric response of meridional and vertical energy transport to the solar forcing. These results, however, do not mean that the albedo has no trend over a long time period. As anthropogenic forcing increases, and if \( \tau \) depends on forcing, the time constant \( \tau \) might be altered significantly over time. When \( \tau \) becomes larger and temperature anomalies persist over a long time, temperature anomalies from different anomalous processes in the atmosphere can accumulate and consequently can alter the mean temperature.

5. Conclusions

Four years of CERES data from March 2000 through February 2004 show that the difference between the maximum and minimum annual mean TOA reflected shortwave and longwave irradiances is 0.4% and 0.1% of the respective annual mean value. Clouds are mostly responsible for these variations at TOA. The largest variation in the monthly mean 1\(^{st}\) \times 1\(^{st}\) TOA reflected shortwave and longwave irradiance occurs in the western and central tropical Pacific due to a shift from La Niña to El Niño during the period. Small global and interannual variability is a result of cancellation of larger regional anomalies when they are spatially and temporally averaged. Anomalies of 300–1000-hPa thicknesses are positively correlated with atmospheric shortwave
irradiance anomalies and negatively correlated with atmospheric longwave irradiance anomalies. The 300- and 1000-hPa thickness anomalies are negatively correlated with atmospheric net irradiance anomalies, which indicate that temperature anomalies are not directly driven by shortwave irradiance anomalies. In addition, the standard deviation of 300- and 1000-hPa thickness anomalies increases with latitude. Therefore, it is postulated that 300- and 1000-hPa thickness anomalies are caused by the variability of dynamical processes that transport energy poleward and by longwave emission. Because of these processes, temperature anomalies in the atmosphere decay exponentially with time and the mean meridional temperature gradient is maintained on an annual time scale. With an assumption that the global mean cloud cover depends predominantly on large-scale dynamics, the exponential decay of temperature anomalies leads to a small interannual variability of global mean cloud cover and albedo.

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FIG. 9. (a) Standard deviation $\sigma_x$ [given by (8)] of atmospheric SW, LW, and NET computed from 48 deseasonalized anomalies from March 2000 through February 2004. (b) As in (a), but for $\sigma_{\alpha}$ [given by (7)]. (c) The standard deviation $\sigma_x$ [given by (8)] of deseasonalized 300–1000-hPa thickness anomalies during the same period. (d) As in (c), but for $\sigma_{\alpha}$ [given by (7)].
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