Tropospheric Temperature Changes and Their Relation to Increasing Greenhouse Gases and Sea Surface Temperatures

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ABSTRACT

In this Note, a set of analyses using general circulation model simulations shows that historical trends in both the chemical composition of the atmosphere – primarily related to increasing greenhouse gases (GHGs) and aerosols – and in the sea-surface temperatures can have very similar influences upon global-scale mid- and upper-tropospheric heights/temperatures. However the change in outgoing longwave radiation estimates at the top of the atmosphere resulting from large-scale sea-surface temperature forcing is very different from that associated with increasing GHG concentrations. The results suggest that accurate measurements of outgoing longwave and net radiation at the top of the atmosphere are a key “fingerprint” which can differentiate between boundary condition climate forcing and climate forcing related to the chemical composition of the atmosphere.
1. Introduction

Many recent studies have focused upon the detection and attribution of systematic surface and tropospheric warming over the last 35+ years. For a survey the reader is directed to IDAG (2005). In general, these studies have used rigorous statistical and empirical methodologies to analyze historical changes in surface and upper-air temperatures, ocean and atmospheric heat content, and regional variations in climate parameters. One forcing agent not considered in many of these investigations however is the historic evolution of the underlying sea-surface temperatures (SSTs). It is well known that regional, hemispheric and global-scale climate variations are related to interannual to decadal changes in the SST fields in the eastern equatorial Pacific (Alexander et al., 2002) as well as to changes in the SST fields across the North Atlantic (Rodwell et al., 2004). In addition, variations in tropical SSTs may have initiated a global-scale climate regime shift in 1976-77 leading to subsequent tropospheric warming (Bratcher and Giese, 2002; Giese et al., 2002; Levitus et al., 2005) and large-scale variations in atmospheric circulation patterns (Hurrell et al., 2004; Bracco et al., 2004; Deser and Phillips, 2006). In this study we are interested in using global climate models forced by various surface boundary conditions and atmospheric chemical constituents to address the issue of whether there is a particular “fingerprint” metric that can effectively differentiate between long-term climate variations that result from changes in the internal energy balances of the atmosphere (as could arise from changes in radiatively-active gas concentrations such as carbon dioxide), compared with forcing from surface boundary conditions (as could arise from large-scale changes in land use or SSTs). If so, observed values of this metric may serve to discriminate between climate variations associated with natural variability and those related to anthropogenic emissions over the last 35+ years.
2. Data

To characterize the atmospheric changes associated with various climate forcings, this study will focus upon the 500hPa temperature field and 250hPa height field as proxies for the integrated temperature of the troposphere. The former is approximately the radiative height derived from the MSU Channel 2 satellite retrieval (Santer et al., 2003a). The latter is used as a surrogate for the height of the tropopause, a useful “fingerprinting” signature for climate change (Santer et al., 2003a; Santer et al., 2003b). While modeling results have shown that the tropopause height responds both to tropospheric warming as well as stratospheric cooling, we will show here that the 250hPa height field, which requires less pre-processing than is needed to derive the tropopause heights, follows the 500hPa temperature evolution closely.

Coupled atmosphere/ocean/land-surface model estimates of the 500hPa and 250hPa fields are produced from NCAR’s Coupled Community System Model (CCSM3; Collins et al., 2005) using two basic emission scenarios. In the first, T85-resolution (approximately 75km) simulations of the CCSM3 are forced by a combination of historical greenhouse gas (GHG) concentrations, sulfate aerosols, volcanic particulates, stratospheric and tropospheric ozone, and solar activity for the period 1870-1999 (termed the Forced simulation). Here we examine the period from 1950-1999 in order to match the integration period for additional atmosphere-only model simulations (see below). In the second emissions scenario, the CCSM3 is run with constant GHG concentrations (set to 1990 levels) for a 400+ year integration period (termed the Control simulation). To compare this simulation with the Forced simulation, three 50-year segments (or “realizations”) of the Control simulation are archived for ensemble averaging similar to the three Forced CCSM3 ensemble simulations.
In addition to the coupled-climate simulation estimates, dynamic atmosphere-only model estimates are produced from the Community Atmosphere Model (CAM3.1 – Collins et al., 2004), which is the atmospheric model core used within the CCSM3. Again, two scenarios are considered. In the first, five T85-resolution simulations of the CAM3.1 are forced only by historical changes in global SSTs (termed the AMIP simulation). In the second scenario, 5 simulations are forced by historical changes in SSTs, GHG concentrations, sulfate aerosols, volcanic particulates, stratospheric and tropospheric ozone, and solar activity (termed the AMIP-ATM simulation). Table 1 provides a summary description of the coupled and atmosphere-only model simulations used in the study.

Additional data-driven estimates of 500hPa temperatures and 250hPa heights will be taken from the NCEP Reanalysis-I data product (Kalnay et al., 1996; Kistler et al., 2001) and the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-40 product (Kallberg et al., 2004).

To characterize changes in the overall earth/ocean/atmosphere energy balance, this study analyzes top of atmosphere (TOA) outgoing longwave radiation (OLR) and net solar radiation (incoming minus outgoing) estimates from both the CCSM3 and CAM3.1 simulations. In addition we archive the observed TOA OLR taken from the NOAA interpolated OLR dataset (Liebmann and Smith, 1996– termed the ‘NOAA’ dataset) and the Lucas et al. OLR dataset, which corrects for changes in equatorial crossing times and satellite transitions (Lucas, et al., 2005 – termed the ‘ECT-corrected’ dataset). Monthly-mean data are available for the period 1975-2004 and 1975-1999 respectively, with the exception of 1978 when there was a sensor failure during part of the year. We also include the newest version of the ERBE Wide Field of View (WFOV) nonscanner data (Wielicki et al., 2002; Wong et al. 2006), although the quality-
controlled, 72-day mean values are only available from 60S-60N. This version of the data (Edition 3, Revision 1) has been corrected for a change in the orbit altitude as well as instrument drift (Wong et al. 2006). Mean values are available for every 72-day period from 1985-1997 except during 1994 when there was a sensor failure during part of the year (see: http://eosweb.larc.nasa.gov/PRODOCS/erbe/quality_summaries/s10n_wfov/erbe_s10n_wfov_nf_sf_erbs_edition3.html).

3. Results

Figure 1 shows the globally-averaged 250hPa heights and 500hPa temperatures for the ensemble-mean Forced and Control CCSM3 simulations, as well as the ensemble-mean CAM3.1 AMIP and AMIP-ATM simulations. For all four datasets a 6-year (72 month) Bartlett (or Triangle) filter has been applied to the monthly mean values and the climatological means over the full time-periods have been removed. The anomaly fields are then area-weighted based upon latitude and averaged over the whole globe. Results indicate that SST-induced changes in globally-averaged upper-tropospheric heights, derived from the CAM3.1 AMIP simulations, have similar magnitude to the GHG-induced changes found in the Forced CCSM3 simulations. However, the inclusion of atmospheric chemical constituents into the CAM3.1 simulations produces only slight changes in the overall evolution of 250hPa heights compared with the SST-only simulations. The Control simulation shows no global-scale variations comparable to the trends in the other 3 datasets (approximately 20m). We provide error bars in this panel (designating the 95% confidence interval based upon the standard error of the mean); here we use them to quantify the spread of the ensemble members about the mean value in order to show that the mean values are representative of the ensemble members themselves and are not simply
residuals between very disparate values. We do not show error bars in other figures because all datasets indicate similar agreement between individual ensemble members and the ensemble mean.

The time-evolution of 500hPa temperatures is almost identical to that of the upper-tropospheric height fields (Figure 1b). If we assume the 500hPa temperature changes are constant through the air column and use the hydrostatic equation to derive an equivalent height change at 250hPa, we find, as an example, that the anomalous 500hPa temperature value in the last year of the CAM3.1 AMIP simulation (which has no contribution from changes in ozone concentrations) would produce approximately a 24.5m change in globally-averaged 250hPa heights, equivalent to the 25.5m anomaly seen in Figure 1a. As such, although the upper-tropospheric height fields may also be responding to changes in the stratospheric temperatures, we feel justified using the height fields as a proxy for the integrated tropospheric temperature structure.

Overall, these results indicate that increases in the 500hPa temperatures and 250hPa heights can be produced either by forcing a coupled-climate model with historical radiatively-active chemical constituents or forcing an atmosphere-only model with historical SSTs. In this sense observed changes in the integrated temperature fields may be “consistent” (Barnett et al., 2001) with those produced by changes in the chemical composition of the atmosphere (Santer et al., 2003b; IDAG, 2005), however they could also be considered consistent with forcing provided by global-scale variations in the sea-surface state.

Indeed, the variations in the sea-surface state itself, as found in the CAM3.1 and Forced simulations, are similar (Figure 2). Here, we plot the zonally-averaged SST anomalies used to force the historical CAM3.1 simulations (Hurrell et al., 2005), the ensemble-mean SST
anomalies from the CCSM3 Forced simulations, and the ensemble-mean SST anomalies from the CCSM3 Control simulation. While the CAM3.1 variations show more regional structure than either of the CCSM3 simulations (particularly in the tropics), in both the Forced CCSM3 simulations and the CAM3.1 simulations there is a systematic increase in SSTs over the 45-year period that results in comparable global-scale changes at all latitudes. No similar systematic variations are seen in the CCSM3 Control simulations. These results indicate that the SST evolution in the GHG-forced coupled climate simulations is similar to that of the observed system (which is used to force the CAM3.1 simulations). While more rigorous optimal detection methodologies suggest that the observed SST evolution is consistent with changing GHG concentrations and is not an independent forcing in of itself (Barnett et al., 2001), these detection methodologies are based upon the “natural” variability of the simulated ocean state, which is known to be lower than observed (IDAG, 2005; Levitus et al., 2005). As such, similar to the tropospheric temperature/height fields, observed increases in the SST fields may be consistent with increased GHG concentrations in the atmosphere, however they may also be consistent with independent natural variations in the ocean state itself.

Instead, we argue that to differentiate between climate variations related to possible global-scale SST forcing and those related to changes in the chemical composition of the atmosphere, it is appropriate to think of the SST field as an “external” forcing of the atmosphere. In this scenario, on a globally-averaged basis any excess thermal energy put into the atmosphere from the ocean will subsequently be removed from the system via changes in the top-of-atmosphere (TOA) radiation (Trenberth and Solomon, 1994):

\[
\frac{\partial E_{\text{atmos}}}{\partial T} = R_{\text{net}}^{\text{TOA}} + F_s
\]
Here, $E_{atmos}$ is the vertically-integrated, globally-averaged total energy within the atmosphere, $R_{net}^{TOA}$ is net incoming radiation through the top of the atmosphere, and $F_s$ are energy fluxes into the atmosphere from the underlying surface (either land surface or ocean surface). If we assume the time-rate of change of the integrated atmospheric energy is small compared with the other two terms (Trenberth et al., 2002), we find:

$$R_{net}^{TOA} = -F_s$$

Hence, while many processes within the atmosphere itself (for instance related to changes in cloud cover and upper-tropospheric humidity – Cess et al., 1990) can modify the magnitude of the various TOA energy flux terms, ultimately any input of energy by changing surface conditions at the bottom boundary of the atmosphere should be balanced by negative anomalies in globally-integrated, net-incoming radiation at the top of the atmosphere (equivalently, net outgoing radiation anomalies should be positive). In contrast, for a scenario in which changes in the integrated temperatures of the atmosphere are related to changes in the GHG concentrations of the atmosphere, there will be a decrease in the TOA outgoing longwave radiation term (due to enhanced trapping of longwave radiation by intervening GHGs). If not compensated for by a decrease in net incoming solar radiation (or a subsequent increase in OLR related to other atmospheric processes), this decrease in OLR will translate into an increase in the energy transfer from the atmosphere to the surface, producing a relative warming of the ocean (as well as land) surfaces.

While other studies have implicitly used this balance as a means of attributing the historical increase of oceanic heat content (and by extension SSTs) across the globe to changing GHG concentrations in the atmosphere (Hansen et al., 2005), these studies assumed that simulated decreases in outgoing TOA radiation fluxes are representative of the observed system. However,
without analogous estimates of the observed changes in the planetary radiative balance, observed long time-scale (decadal and longer) oceanic heat content variations could be attributable to either natural and/or anthropogenic forcing (Levitus et al., 2005). Indeed, it is important to note that in both climate-change scenarios presented here (i.e. tropospheric warming due to increased GHG concentrations vs. increased global-scale SSTs), the SST signal is the same (i.e. the SSTs are increasing in both scenarios). However, in one case (related to increasing GHG concentrations), the net outgoing radiation anomaly should be negative; in the other (related to increasing SSTs), the net outgoing radiation anomaly should be positive. The above argument, therefore, suggests that a defining set of metrics for differentiating between SST-induced climate variability and climate variability related to changes in the chemical composition of the atmosphere is the historical evolution of the globally-integrated TOA radiation fields, and OLR in particular.

To test this hypothesis, Figure 3 shows the globally-averaged TOA outgoing longwave radiation and net incoming solar radiation from 1) the Forced and Control CCSM3 simulations, and 2) the CAM3.1 AMIP and AMIP-ATM simulations. Trends in the outgoing longwave radiation, net incoming solar radiation, and total net radiation flux are given in Table 2. In the Forced CCSM3 there is an overall decrease in the OLR over the course of the simulation with modulations of the trend related to additional changes in the historical chemical composition of the atmosphere, principally associated with volcanic activity; these modulations are represented by dramatic drops in the net incoming solar energy term around 1964, 1982, and 1991. Given the historical sequence of volcanic activity, there is a negative trend in net incoming solar radiation during the simulation period (Table 2). The difference between the net incoming shortwave radiation and the outgoing longwave radiation gives the change in net TOA radiation
flux. During the periods following the volcanic eruptions, the temporary decrease in solar insolation is larger than the decrease in OLR, resulting in cooling temperatures and decreased geopotential heights during these periods (Figure 1). However, the negative trend in incoming solar radiation does not compensate for the decrease in outgoing longwave radiation, hence the net incoming radiation flux at the top of the atmosphere increases over the course of the simulation (Table 2).

For the atmosphere-only simulations, the CAM3.1 AMIP simulation shows a distinct rise in outgoing longwave radiation. In contrast, the CAM3.1 AMIP-ATM simulation has a decrease in OLR (Table 2) despite a very similar evolution in geopotential height and temperature fields (Figure 1). In addition, for the CAM3.1 AMIP simulation, the net incoming solar radiation experiences only a slight decrease over the course of the simulation (Table 2), resulting in an overall decrease in net incoming radiation flux, driven by the OLR trend, of about 1.10 W/m² over the course of the simulation (Table 2). In this model scenario, the radiative disequilibrium at the top of the atmosphere characterizes a disequilibrium between the oceanic evolution and the atmospheric evolution within the model such that the atmosphere is being forced by global-scale changes in the SSTs; given the model set-up, this result is to be expected.

For the CAM3.1 AMIP-ATM simulation, the net incoming solar radiation has large negative deviations due to the inclusion of volcanic aerosols (Figure 3b), resulting in an overall negative trend over the course of the simulation (Table 2). This evolution is almost identical to that in the Forced CCSM3 simulation, which is expected because they have the same prescribed changes in aerosols, volcanic particulates, and solar activity; the slight differences may be due to factors such as changing cloud cover or surface albedo. As with the Forced CCSM3 simulation, the short-term decreases in net incoming solar radiation tend to overwhelm the outgoing
longwave radiation, resulting in decreased tropospheric heights and temperatures during these periods (Figure 1). However, as with the CCSM3, the long-term trend in the net TOA radiation flux is driven by the decreasing OLR, resulting in an increase in the net incoming radiation flux through the top of the atmosphere, which stands in contrast to the CAM3.1 AMIP simulation (Table 2).

In the CAM3.1 AMIP-ATM scenario, it appears that excess energy input into the atmosphere by the underlying SSTs is not being removed via increased TOA OLR (as in the CAM3.1 AMIP simulation) but that the energy is ultimately being retained within the atmosphere via absorption by the increased GHG concentrations. In fact, the increase in net incoming TOA radiation flux in the CAM3.1 AMIP-ATM simulation indicates that there is a net flux of energy from the atmosphere to the earth’s surface (either the land surface or the ocean surface), not vice-versa as in the CAM3.1 AMIP simulation. The fact that this heat flux in the CAM3.1 AMIP-ATM is comparable to that needed to produce GHG-induced SST variations (as seen in the Forced CCSM3 simulations) equivalent to the SST variations in the CAM3.1 AMIP-ATM itself indicates that the global-scale SST evolution and GHG evolution in the CAM3.1 AMIP-ATM simulation are not independent forcings of the atmospheric temperature/height fields but are in quasi-equilibrium with one another.

4. Discussion

Above, we argue that observed trends in the TOA radiation terms, and OLR in particular, can serve as a “fingerprint” to differentiate between climate forcing associated with changes in the global-scale SST field compared with changes in the chemical composition of the atmosphere. To see how “best estimates” of the observed fields discussed here compare with
model simulated fields, Figure 4 shows the Reanalysis-based 250hPa height fields (taken from
the NCEP and ECMWF Reanalysis products) as well as the satellite-based OLR estimates (taken
from the NOAA-interpolated and ECT-corrected datasets, as well as the ERBE Wide Field of
View Nonscanner dataset). Both of these sets of data have limitations with regard to their
veracity. For instance much has been written about the Reanalysis data products for periods
prior to the “satellite” record (e.g. mid-1970’s – Kistler et al., 2001); in addition, the observed
and simulated OLR data products are known to have differences between one another,
particularly with regard to their interannual variances (Wielicki et al., 2002). However, a
qualitative comparison of the two reanalysis-based height fields suggests that, although they
show some differences in their short-term evolution, they both indicate the “observed” global
system has experienced a tropospheric warming similar to that found in the Forced coupled
model, as well as to that found in both the AMIP and AMIP-ATM atmosphere-only models. As
such, this observed historical warming, if accurate, could be consistent with changes in observed
GHG and aerosol concentrations (as in the Forced CCSM3 simulation) and/or changes in
observed SSTs (as in the CAM3.1 simulations).

We next examine the globally-averaged observed OLR data (Figure 4b). Trends in the
available observed radiative terms are provided in Table 2; given the short duration of these
observed fields, none of these trends are significant, however we present them here for
qualitative comparison both with the model data and with one another. Although the NOAA-
based OLR fields (NOAA and ECT-corrected OLR) show more interannual variability than the
simulations (Wielicki et al., 2002), trends in these two datasets better match the Forced CCSM3
and SST/GHG CAM3.1 simulations and do not show a signature consistent with boundary-
condition forcing as in the SST-only CAM3.1 simulation (Table 2). If correct, these results
suggest that the observed changes in the integrated temperature of the global atmosphere cannot be attributed to external forcing by the underlying SSTs alone.

In comparison, the ERBE-based WFOV OLR data show values similar to the NOAA-based OLR estimates up through 1992, when the nonscanner instrument went off-line. However, starting in 1994 there was a significant increase in the ERBE WFOV OLR over the final 4-year period (through 1997), resulting in an overall trend in the ERBE WFOV OLR data similar to that in the SST-only CAM3.1 simulations (Table 2). (All comparisons are quantitatively the same if the observed and simulated fields are averaged over 60N-60S to match the ERBE WFOV data.) If correct, these results suggest an increase in global-scale OLR. Further investigation of the ERBE TOA radiation data (Table 2) suggests that this increase in OLR is related to changes in top-of-atmosphere boundary-condition forcing associated with an increase in net incoming solar radiation, possibly related to a decrease in cloud cover and hence albedo during this period (T. Wong; Personal communication). The balance between these two produces an overall increase in the net incoming radiation flux. This increase in the net incoming radiation flux, which from above should be balanced by a net flux of energy from the atmosphere to the surface (Wong et al., 2006), suggests that the possible increase in the observed tropospheric temperatures (Figure 4a) is not associated with changing SST boundary-condition forcing as in the CAM3.1 AMIP simulation nor is it associated with increased GHG concentrations, but may be related to changes in solar radiation absorbed by the earth system.

5. Conclusions

Attribution of recent climate change to various influences has been systematically studied over the last two decades. Numerous results have shown that trends in the observed climate are
consistent with increasing radiatively-active atmospheric constituents resulting from increased anthropogenic emissions of greenhouse gases and aerosols (e.g. IDAG, 2005). At the same time to fully attribute the observed climate changes to anthropogenic emissions requires systematically removing other feasible forcing agents. While many potential forcing agents, including solar variability and volcanic activity, have been shown to produce behavior inconsistent with the observed global trends over the last 50+ years (Santer et al., 2003a,b), other forcing agents may still produce behavior similar to that observed and hence may also represent feasible contributors to recent climate variability.

In this study we look to systematically isolate a “fingerprinting” signature that can effectively evaluate whether the evolution of global-scale sea-surface temperatures (SSTs) is one such contributor. Here we investigate the evolution of the 500hPa temperatures and 250hPa height fields (which serve as a proxy for integrated tropospheric temperatures) taken from 1) coupled climate models forced by historical changes in the chemical composition of the atmosphere – primarily related to increasing greenhouse gases (GHGs) and aerosols – and 2) uncoupled atmosphere-only models forced with historical SSTs vs. historical SSTs combined with radiatively-active chemical constituents. We show that both GHG-only forcing and SST-only forcing can produce a very similar evolution of the global-scale tropospheric temperature and height fields. In addition, we find that the global-scale SST evolution in the GHG-forced coupled simulation is similar to that used to force the atmosphere-only simulations. As such, any comparable increases in the observed tropospheric temperatures could be consistent with either a forced response to changes in underlying SSTs or to changes in historical GHG concentrations.

However, we find that for SST-only forced atmospheric models, the excess energy applied to the atmosphere via increased SSTs is effectively removed via increased top-of-atmosphere
out-going longwave radiation, which is realized as an anomalous increase in globally-averaged TOA OLR and an anomalous decrease in the total net incoming radiation flux. In comparison, for coupled climate models, in which the climate forcing is applied principally via changes in the chemical composition of the atmosphere (i.e. GHGs and aerosols), we find that there is a long-term decrease in globally-averaged TOA OLR and an anomalous increase in the net incoming radiation flux, indicative of a trapping of longwave radiation by the intervening GHG concentrations. In addition, when atmosphere-only models are forced both with historical SSTs and GHG/aerosol concentrations, the anomalous OLR leaving the top of the atmosphere switches from positive to negative, indicating that the excess energy input to the atmospheric system via changing SSTs is in quasi-equilibrium with the amount of energy retained by the increasing GHG concentrations within the atmosphere.

As such, we argue that continuing and more precise measurements of the observed radiative terms at the top of the atmosphere, particularly OLR, should provide the “fingerprint” that can discriminate between GHG-driven energetics of the climate system and any potential surface-driven components. In addition, these measurements will be able to better discriminate between internally-generated SST variability vs. GHG-induced heating of those same surfaces. Combining these observations with a more complete budget analysis of the observed variations in the heat content of the ocean (and by extension the SST fields) will further enhance our understanding of natural variations in the state of the ocean system and its role as a potential independent driver of trends in the observed climate system.

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help acquiring and analyzing the ERBE WFOV data. Interpolated OLR data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA, from their Web site at http://www.cdc.noaa.gov/. ECT-corrected OLR data provided by the UCAR/NCAR Data Support Section, Boulder, Colorado, USA, from their Web site at http://dss.ucar.edu/datasets/ds684.1/. NCEP Reanalysis data provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA, from their Web site at: http://www.cdc.noaa.gov. ECMWF ERA-40 data used in this study/project have been provided by ECMWF and have been obtained from the ECMWF Data Server. CCSM3 data used in this study/project have been provided by UCAR/NCAR have been obtained from the Earth System Grid Data Server. ERBE WFOV OLR Edition3_Rev1 available from: http://earth-www.larc.nasa.gov/erbeweb/Edition3_Rev1/access_ed3_rev1_data.html
REFERENCES


http://www.ecmwf.int/publications/library/ecpublications/_pdf/era40/ERA_40_PRS17.pdf


TABLE LEGENDS

**Table 1** Name and characteristics of model simulations used in this analysis. “Coupled” models refer to fully-coupled ocean/atmosphere/land-surface models; “Atmosphere” models refer to atmosphere-only model simulations; “GHGs” refers to historic changes in greenhouse gas concentrations; “SSTs” refers to historic changes in sea-surface temperatures. The 1990-2140 duration for the CCSM3 Control simulation represents a 150-year simulation using constant greenhouse gas and sulfate aerosol concentrations set to 1990 levels.

**Table 2** Trends in globally-averaged top-of-atmosphere radiative terms from model simulations and available observations: OLR – outgoing longwave radiation; Net SW – net incoming solar radiation; Net Rad – total net radiation calculated as the difference between Net SW and OLR. Units are in (W/m²)/decade. Sign convention such that positive values are represented by the direction of arrow. For all data except the ERBE Edition3-Rev1 fields, time-series are smoothed using a 6-year Bartlett filter before calculating trends. For the ERBE Edition3-Rev1 fields, a 3-year Bartlett filter is used due to the short duration of the time-series. Trends calculated using slope of best-fit linear interpolation for length of available record.
FIGURE CAPTIONS

Fig. 1 (a) Globally-averaged 250hPa geopotential height anomalies taken from Forced and Control CCSM3 simulations and the CAM3.1 AMIP and AMIP-ATM simulations – see text for details. Forced CCSM3 estimates derived from the average of 3 ensemble simulations. Control CCSM3 estimates derived from the average of the first, second, and third 50-year periods of the full 150-year time-series. CAM3.1 estimates derived from the average of 5 ensemble simulations. Geopotential heights are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval starting at 1953. The initial height anomalies from all the datasets are shifted such that they all start at 0. Error bars give the 95% confidence interval based upon the standard error of the mean value at each time-step. All values have units of meters. (b) same as Figure 1a except for 500hPa temperature anomalies. All values have units of degrees C.

Fig. 2 (a) Zonally-averaged observed sea-surface temperature anomalies used to force the CAM3.1 simulations for the period 1950-2000. Sea-surface temperatures are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. Anomalies calculated from climatology of the full time-series. Contour interval is 0.1K; minimum contour is +/-0.1K. (b) Same as Figure 2a except for the ensemble-mean SST evolution taken from the Forced CCSM3 simulations. (c) Same as Figure 2a except for the mean of the first, second, and third 50-year periods of the full 150-year time-series taken from the Control CCSM3 simulation.

Fig. 3 (a) Globally-averaged change in outgoing longwave radiation at the top of the atmosphere, taken from the Forced and Control CCSM3 simulations and the CAM3.1 AMIP and
AMIP-ATM simulations for the period 1950-2000. OLR values are smoothed using a 6-year Bartlett filter and initial values for all data plotted starting at 0. All values have units of W/m². (b) same as Figure 3b except for change in incoming net shortwave radiation.

Fig. 4 (a) Globally-averaged 250hPa geopotential height anomalies taken from the NCEP Reanalysis, the ERA40 Reanalysis, the 3 ensemble runs of the Forced and Control CCSM3 simulations, and 5 ensemble runs of the CAM3.1 AMIP and AMIP-ATM – see text for details. Geopotential heights are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. The initial height anomalies from all the datasets are shifted such that they start at 0. All values have units of meters. (b) Same as Figure 4a except for top-of-atmosphere outgoing longwave radiation. OLR estimates taken from the Equatorial Crossing Time (ECT) corrected data for the period 1975-1999 (with 1978 anomalies set to 0), the interpolated NOAA OLR for the period 1975-2004 (with 1978 anomalies set to 0), the ERBE WFOV Edition3_Rev1 OLR for the period 1985-1997 (with 1994 anomalies set to 0), the Forced CCSM3 simulations and the CAM3.1 AMIP and AMIP-ATM simulations. For all data except the ERBE WFOV OLR, time-series are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. For the ERBE WFOV OLR, a 3-year Bartlett filter is used due to the short duration of the time-series. Note that y-axis scale is the same for both the simulated and observed data, however the zero-line is offset to aid with visualization.
Table 1 Name and characteristics of model simulations used in this analysis. “Coupled” models refer to fully-coupled ocean/atmosphere/land-surface models; “Atmosphere” models refer to atmosphere-only model simulations; “GHGs” refers to historic changes in greenhouse gas concentrations; “SSTs” refers to historic changes in sea-surface temperatures. The 1990-2140 duration for the CCSM3 Control simulation represents a 150-year simulation using constant greenhouse gas and sulfate aerosol concentrations set to 1990 levels.

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<td>5</td>
</tr>
<tr>
<td>CAM3.1 AMIP-ATM</td>
<td>Atmosphere</td>
<td>SSTs</td>
<td>1950-2000</td>
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<td>Sulfate aerosols</td>
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<td>Volcanic particulates</td>
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<td>Ozone</td>
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<tr>
<td></td>
<td></td>
<td>Solar activity</td>
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</tbody>
</table>
TABLE 2 Trends in globally-averaged top-of-atmosphere radiative terms from model simulations and available observations: OLR – outgoing longwave radiation; Net SW - net incoming solar radiation; Net Rad - total net radiation calculated as the difference between Net SW and OLR. Units are in (W/m²)/decade. Sign convention such that positive values are represented by the direction of arrow. For all data except the ERBE Edition3-Rev1 fields, time-series are smoothed using a 6-year Bartlett filter before calculating trends. For the ERBE Edition3-Rev1 fields, a 3-year Bartlett filter is used due to the short duration of the time-series. Trends calculated using slope of best-fit linear interpolation for length of available record.

<table>
<thead>
<tr>
<th>Name</th>
<th>Duration</th>
<th>OLR(↑)</th>
<th>Net SW(↓)</th>
<th>Net Rad(↓)</th>
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<tr>
<td>CCSM3 Forced</td>
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<td>-0.21</td>
<td>-0.15</td>
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<td>-0.08</td>
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<td>-0.15</td>
<td>+0.07</td>
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<td>NOAA</td>
<td>1975-2004</td>
<td>-0.26</td>
<td>-</td>
<td>-</td>
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<td>ECT-Corrected</td>
<td>1975-1999</td>
<td>-0.23</td>
<td>-</td>
<td>-</td>
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<td>ERBE Ed3-Rev1</td>
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<td>+0.94</td>
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FIG. 1  (a) Globally-averaged 250hPa geopotential height anomalies taken from Forced and Control CCSM3 simulations and the CAM3.1 AMIP and AMIP-ATM simulations – see text for details. Forced CCSM3 estimates derived from the average of 3 ensemble simulations. Control CCSM3 estimates derived from the average of the first, second, and third 50-year periods of the full 150-year time-series. CAM3.1 estimates derived from the average of 5 ensemble simulations. Geopotential heights are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval starting at 1953. The initial height anomalies from all the datasets are shifted such that they start at 0. Error bars give the 95% confidence interval based upon the standard error of the mean value at each time-step. All values have units of meters. (b) same as Figure 1a except for 500hPa temperature anomalies. All values have units of degrees C.
Fig. 2 (a) Zonally-averaged observed sea-surface temperature anomalies used to force the CAM3.1 simulations for the period 1950-2000. Sea-surface temperatures are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. Anomalies calculated from climatology of the full time-series. Contour interval is 0.1K; minimum contour is +/-0.1K. (b) Same as Figure 2a except for the ensemble-mean SST evolution taken from the Forced CCSM3 simulations. (c) Same as Figure 2a except for the mean of the first, second, and third 50-year periods of the full 150-year time-series taken from the Control CCSM3 simulation.
FIG. 3 (a) Globally-averaged change in outgoing longwave radiation (OLR) at the top of the atmosphere, taken from the Forced and Control CCSM3 simulations and the CAM3.1 AMIP and AMIP-ATM simulations for the period 1950-2000. OLR values are smoothed using a 6-year Bartlett filter and initial values for all data plotted starting at 0. All values have units of W/m$^2$. (b) same as Figure 3b except for change in incoming net shortwave radiation.
FIG. 4  (a) Globally-averaged 250hPa geopotential height anomalies taken from the NCEP Reanalysis, the ERA40 Reanalysis, the 3 ensemble runs of the Forced and Control CCSM3 simulations, and 5 ensemble runs of the CAM3.1 AMIP and AMIP-ATM – see text for details. Geopotential heights are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. The initial height anomalies from all the datasets are shifted such that they start at 0. All values have units of meters. (b) Same as Figure 4a except for top-of-atmosphere outgoing longwave radiation. OLR estimates taken from the Equatorial Crossing Time (ECT) corrected data for the period 1975-1999 (with 1978 anomalies set to 0), the interpolated NOAA OLR for the period 1975-2004 (with 1978 anomalies set to 0), the ERBE WFOV Edition3_Rev1 OLR for the period 1985-1997 (with 1994 anomalies set to 0), the Forced CCSM3 simulations and the CAM3.1 AMIP and AMIP-ATM simulations. For all data except the ERBE WFOV OLR, time-series are smoothed using a 6-year Bartlett filter and plotted at the center of the averaging interval. For the ERBE WFOV OLR, a 3-year Bartlett filter is used due to the short duration of the time-series. Note that y-axis scale is the same for both the simulated and observed data, however the zero-line is offset to aid with visualization.