

The dependence of clear-sky outgoing long-wave radiation on surface temperature and relative humidity

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SUMMARY

A simulation of the earth's clear-sky long-wave radiation budget is used to examine the dependence of clear-sky outgoing long-wave radiation (OLR) on surface temperature and relative humidity. The simulation uses the European Centre for Medium-Range Weather Forecasts global reanalysed fields to calculate clear-sky OLR over the period from January 1979 to December 1993, thus allowing the seasonal and interannual time-scales to be resolved. The clear-sky OLR is shown to be primarily dependent on temperature changes at high latitudes and on changes in relative humidity at lower latitudes. Regions exhibiting a 'super-greenhouse' effect are identified and are explained by considering the changes in the convective regime associated with the Hadley circulation over the seasonal cycle, and with the Walker circulation over the interannual time-scale. The sensitivity of clear-sky OLR to changes in relative humidity diminishes with increasing relative humidity. This is explained by the increasing saturation of the water-vapour absorption bands with increased moisture. By allowing the relative humidity to vary in specified vertical slabs of the troposphere over an interannual time-scale it is shown that changes in humidity in the mid troposphere (400 to 700 hPa) are of most importance in explaining clear-sky OLR variations. Relative humidity variations do not appear to affect the positive thermodynamic water-vapour feedback significantly in response to surface temperature changes.

KEYWORDS: Clear-sky long-wave radiation Super-greenhouse effect Water-vapour feedback

1. INTRODUCTION

Outgoing long-wave radiation to space (OLR) is the fundamental cooling mechanism of the earth–atmosphere system. Knowledge of its dependence on surface and atmospheric parameters is therefore required for a quantitative understanding of the climate and of climate change. The response of the climate system to a change in the earth's radiation budget is made more complex by feedbacks, one of which involves changes in water-vapour amount and distribution in response to a change in surface temperature, T_s . Water vapour is generally regarded as providing a positive feedback; climate models imply an amplification of surface warming of about 1.6 (IPCC 1990).

Raval and Ramanathan (1989) used satellite observations of clear-sky OLR from the Earth Radiation Budget Experiment (ERBE) to argue in favour of a positive water-vapour feedback by showing an increase in normalized atmospheric greenhouse trapping with spatially increasing T_s . However, spatial variations cannot be used reliably to infer water-vapour changes in a warming climate; Spencer and Braswell (1997) noted that a positive feedback will always apply spatially between the ascending and descending branches of tropical circulation systems. Rind *et al.* (1991) showed satellite-derived water-vapour amounts to increase with the greater convection associated with hemispheric warming over the seasonal cycle. They concluded that the quantitative agreement with climate-model predictions strengthens the argument in favour of a positive water-vapour feedback. Ramanathan and Collins (1991) inferred a positive feedback spatially and over short time-scales, such as the seasonal cycle and the El Niño/Southern Oscillation (ENSO). However, Lau *et al.* (1996) asserted that it is not prudent to assume that feedbacks operating over these time-scales are relevant for climate change on decadal time-scales unless the entire circulation systems are accounted for.

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Bony *et al.* (1995) calculated the sensitivity of ERBE clear-sky OLR to T_s using monthly means of data from 1987 and 1988. The difference between the sensitivity of the clear-sky OLR to T_s over different time-scales in the ERBE and climate-model warming experiments was attributed to changes in temperature lapse rate and relative humidity (RH). Raval *et al.* (1994) also highlighted the strong seasonal correlation between ERBE OLR and column-mean RH for both clear and cloudy skies over the tropical oceans. The relationship between upper-tropospheric RH, deep-convection frequency and the normalized clear-sky atmospheric greenhouse trapping was examined by Soden and Fu (1995). From the correlation they inferred that variations in upper-tropospheric RH explain about half of the deviation of normalized greenhouse trapping from its climatological mean, with the remainder being tentatively explained by variations in the temperature lapse rate and moisture at lower levels. Sinha and Harries (1997) concluded that, while clear-sky OLR is most sensitive to upper-tropospheric RH, changes in the RH in the middle troposphere are largest. The prime motivation of the present study is to extend this earlier work by using a simulation of clear-sky OLR based on 15 years of data. A longer period of analyses than earlier studies allows a more robust characterization of the variability, and the use of a model to simulate the OLR allows the causes of the variability to be more clearly identified than is possible using statistical analyses for which relationships between variables may not adequately imply a direct causal link.

Another focus for this study is to examine the occurrence of the ‘super-greenhouse’ effect (SGE) (see, for example, Hallberg and Inamdar 1993) which describes the apparently unstable clear-sky long-wave radiative feedback operating at high sea surface temperatures whereby clear-sky OLR is anti-correlated with T_s . An additional feedback must be present to stabilize the tropical sea surface temperatures; suggested mechanisms are short-wave cloud radiative forcing (Ramanathan and Collins 1991), evaporative feedbacks (Hartmann and Michelsen 1993) and heat export to the extratropics (Pierrehumbert 1995). Bony *et al.* (1997) explained the SGE occurrence by considering the changes in the convective regime associated with the large-scale circulations over the seasonal and interannual time-scale.

The present study uses a simulation of the earth’s clear-sky long-wave radiation budget, described in section 2, to assess the impact of changes in temperature and humidity throughout the atmosphere on clear-sky OLR. Slingo *et al.* (1998) showed the quality of the simulation by comparing clear-sky OLR with that from ERBE. The present study benefits from a sufficiently long dataset to allow climate fluctuations to be resolved on a longer time-scale than the ERBE period, and it includes several ENSO events. Land and oceans are sampled, thus ensuring global coverage of clear-sky irradiance and sampling of the entire atmospheric circulation system. Thus, the sensitivity of clear-sky OLR to surface and atmospheric parameters may be calculated in a global-mean sense. The framework of interpretation of the present analysis is set out in section 3 while results are described in sections 4 and 5. Discussion with reference to the SGE and the water-vapour feedback are presented in the final section.

2. SIMULATION

A comprehensive reanalysis of atmospheric variables for a 15-year period has been performed by the European Centre for Medium-Range Weather Forecasts ReAnalysis project (ERA). Slingo *et al.* (1998) used the ERA analyses to perform simulations of clear-sky long-wave fluxes and heating rates, with the acronym CLERA (Clear-sky

Long-wave from ERA). Radiative calculations were performed with the 9-spectral-band version of the Edwards and Slingo (1996) radiation code every 6 hours from January 1979 to December 1993. Surface emissivity was set to unity for consistency with the ERA model. The ERA data were used on a 2.5° by 2.5° horizontal grid, with 31 σ -levels* in the vertical. The well-mixed greenhouse-gas concentrations (CO_2 , CH_4 , N_2O , CFC11, CFC12) were updated annually in accordance with IPCC (1995). Ozone was derived from a monthly-mean climatology (Cullen 1993) and was assumed to be fixed throughout the analysis period (interannual variations in ozone are unlikely to affect the OLR variability significantly). Clear-sky profiles were assumed throughout, even in cloudy regions, thus ensuring that information on clear-sky irradiance was sampled globally. This definition is consistent with the way in which clear-sky fluxes are calculated in climate models, but it differs from those used by ERBE studies which only class regions as clear-sky if the statistical likelihood of cloud is sufficiently low. Studies using ERBE data are, thus, prone to cloud contamination of clear-sky regions (Slingo and Webb 1992) and a tendency towards the sampling of drier profiles. The present study used monthly means of the simulated clear-sky OLR together with the ERA data.

Unique to the CLERA system are the length and time resolution of the dataset and its global coverage (land and ocean). Because the ERA combines a wide range of atmospheric and surface data using the same analysis system throughout the data period, it constitutes one of the most robust representations of global conditions and variability (including that of RH) over a 15-year period. However, because most of the relevant RH data is assimilated, independent validation of this parameter is problematic. The CLERA simulation calculates RH with respect to water. A more accurate definition is not required because RH is used in this study as a diagnostic parameter, rather than as a prognostic variable to determine the degree of saturation of water vapour in the atmosphere. Errors in the ERA fields of T_s and surface emissivity, while affecting the absolute calculation of clear-sky OLR, are unlikely to be significant with regard to monthly-mean variability.

3. FRAMEWORK

To assess the dependence of clear-sky OLR on atmospheric structure it is useful to examine the radiative transfer equation. OLR for clear-sky conditions may be expressed as:

$$\text{OLR} = \int_{\nu=0}^{\nu=\infty} \left[\{B_\nu(0)Tr_\nu(p_s, 0)\} + \int_{p_s}^0 B_\nu(p) \frac{dTr_\nu(p, 0)}{dp} dp \right] d\nu, \quad (1)$$

where $B_\nu(p)$ denotes the black-body emission at pressure, p , and $Tr_\nu(p_s, 0)$ denotes the transmittance between the surface pressure, p_s , and the top of the atmosphere. The subscript ν denotes wave number, and both B and Tr include implicit integration over the zenith angle to convert radiance to irradiance. Equation (1) may be simplified to

$$\text{OLR} = sT_s^4 Tr + \text{OLR}_a(T(p), q(p)), \quad (2)$$

where sT_s^4 is the wavelength-integrated (black-body) surface emission, Tr is the *effective* atmospheric transmittance and OLR_a is the atmospheric component of the OLR and s is the Stefan–Boltzman constant ($5.67 \times 10^{-8} \text{ W m}^{-2}\text{K}^{-4}$). Tr is an effective

* $\sigma = p/p_s$, where p is the pressure and p_s is the pressure at the surface.

transmittance because in reality the transmission varies with wavelength and, therefore, the Planck function becomes more important in the relatively transparent window regions of the spectrum. The second term on the right-hand side of Eq. (2) represents the atmospheric emission which is dependent primarily on temperature, T , and specific humidity, q , throughout the profiles, as well as on the concentrations of CO₂ and other greenhouse gases.

Increases in T in Eq. (2) act to increase the long-wave emission to space from the atmosphere, while increases in q act to decrease the surface component of the OLR and also to decrease OLR_a by shifting the emission to a higher, generally colder, layer. Thus, it is useful to explain the clear-sky OLR in terms of surface and atmospheric temperature and water-vapour amounts. Shine and Sinha (1991) show that clear-sky OLR is sensitive to relative changes in q throughout the atmosphere, but that clear-sky OLR is far more sensitive to absolute changes in q in the upper troposphere than at other levels. Therefore in developing an atmospheric humidity parameter that describes adequately the clear-sky OLR variability, it is more useful to consider the RH rather than the total column moisture. A further reason for using RH is that T and q have opposing effects on the clear-sky OLR that almost cancel in the water-vapour bands if the relative humidity is held constant (Slingo and Webb 1997). However, the cancellation is not complete because water vapour does not absorb strongly over the entire spectrum. Because atmospheric temperature, T_a , is coupled to a certain degree with T_s , a change in clear-sky OLR may be expressed as a function of a change in T_s and a change in tropospheric relative humidity, TRH

$$\Delta \text{OLR} = \left[4sT_s^3 \overline{Tr} + \left\{ sT_s^4 \frac{\partial Tr}{\partial T_a} + \frac{\partial(\text{OLR}_a)}{\partial T_a} \right\}_{(\overline{\text{RH}})} \frac{dT_a}{dT_s} \right] \Delta T_s + \left\{ sT_s^4 \frac{\partial Tr}{\partial(\text{TRH})} + \frac{\partial(\text{OLR}_a)}{\partial(\text{TRH})} \right\} \Delta(\text{TRH}), \quad (3)$$

from which it follows that

$$\Delta \text{OLR} = (4sT_s^3 \overline{Tr} + \alpha) \Delta T_s + \frac{\partial(\text{OLR})}{\partial(\text{TRH})} \Delta(\text{TRH}). \quad (4)$$

The TRH used here is defined in section 4(b); it approximates to the column-average relative humidity between about 100 and 900 hPa with a maximum weight at 450 hPa. Equation (4) shows that there is a dependence of OLR variability on T_s and TRH; it is a similar model to that developed by Thompson and Warren (1982). The parameter, α ($\text{W m}^{-2} \text{K}^{-1}$), denotes an amplification of the clear-sky OLR response to ΔT_s to changes in atmospheric temperature with constant RH.

The analysis framework that has been used to infer sensitivities and correlation is that of linear least-squares fit (LSQF) (e.g. Freund 1976). This removes the potential for misinterpretation by considering only particular months, as pointed out by Soden (1997). The sensitivity and the degree of linear correlation of clear-sky OLR to T_s and TRH may be calculated over the 12-month seasonal cycle of a particular year, and the 15-year interannual time-scale for a particular month. This differs from the technique used by Bony *et al.* (1995, 1997) who calculated a *spatial* correlation between sea surface temperature and ERBE clear-sky OLR differences between seasons and between years for all grid points in a particular temperature bin. A correlation coefficient, r , of greater than 0.75 (or less than -0.75) is required for a correlation at the 99.5% significance level for the smallest sample size used here (12 months).

4. REGRESSION STUDIES

(a) *Surface temperature*

Figure 1(a) shows the LSQF regression correlation for clear-sky OLR and T_s calculated over the seasonal cycle for 1985. There is a strong positive correlation between clear-sky OLR and T_s for latitudes polewards of about $\pm 30^\circ$. In the tropics the correlation tends to be negative, showing that changes in atmospheric structure act to counteract the direct ΔT_s contribution to changing clear-sky OLR in Eq. (4). In these regions the simulated clear-sky OLR decreases with increasing T_s , constituting a potentially unstable feedback. This has been termed the super-greenhouse effect (SGE) (see, for example, Ramanathan and Collins (1991)). Over high-latitude oceans, changes in atmospheric temperature act to amplify the sensitivity of clear-sky OLR to T_s (Webb *et al.* 1993); clear-sky OLR changes are greater than changes in surface emission due to large increases in OLR_a with increased T_s . In this situation the parameter α in Eq. (4) is large because atmospheric temperatures are changing at a greater rate than T_s . The relationship between clear-sky OLR and T_s over the seasonal cycle of other years is similar to that of 1985.

Over the interannual time-scale for July (Fig. 1(b)) an SGE at high values of T_s is still present. These regions are confined to the tropical oceans and are less coherent than for the seasonal cycle. Strong positive correlation is confined mainly to land areas of latitude greater than about 45° where changes in T_s are greatest and convection strongly couples T_a and T_s , thus increasing α . The simulated sensitivities are between about 1 and $3 \text{ W m}^{-2} \text{ K}^{-1}$ away from the tropical oceans and are negative over the tropical oceans; this is consistent with Bony *et al.* (1995) who used ERBE data over the period 1987–1989, together with data from a climate model applied to seasonal and interannual time-scales. Applying the LSQF analysis to interannual time-scales for months other than July results in similar sensitivities of clear-sky OLR to T_s , although regions exhibiting an SGE vary in latitude.

The regions exhibiting an SGE are consistent with the results of Bony *et al.* (1997) who used ERBE clear-sky OLR observations over the period 1985–1989; this may be explained by considering the changes in the large-scale circulation. Over the seasonal cycle (Fig. 1(a)), regions exhibiting an SGE are generally situated within the 5 to 15° latitudinal zones either side of the equator. The change in convective regime, related to the north–south shift of the Hadley Cell over the seasonal cycle, is larger in these regions than at the equator, and forces large changes in tropospheric RH that decouple the clear-sky OLR from the surface emission. Over the interannual time-scale the shift in the Hadley circulation is small, although a residual SGE signal is present over the Atlantic and Indian Oceans. The strongest SGE signal is evident in the tropical central Pacific, due to shifts in the Walker circulation associated with ENSO events.

(b) *Tropospheric relative humidity*

Increasing RH acts to decrease the surface component of the clear-sky OLR by decreasing Tr in Eq. (2) and effectively shifts the height of atmospheric long-wave emission to space up to a colder level, thus decreasing OLR_a . As absorption bands are radiatively less saturated with decreasing q , the sensitivity of the clear-sky OLR to RH is most negative for drier profiles and also for changes in RH in the relatively dry upper troposphere. This is complicated by the absorption properties of the water-vapour continuum (see, for example, Clough *et al.* (1992)) which, in the moister profiles, acts to decrease the surface component of the clear-sky OLR with increased RH in the lower troposphere.

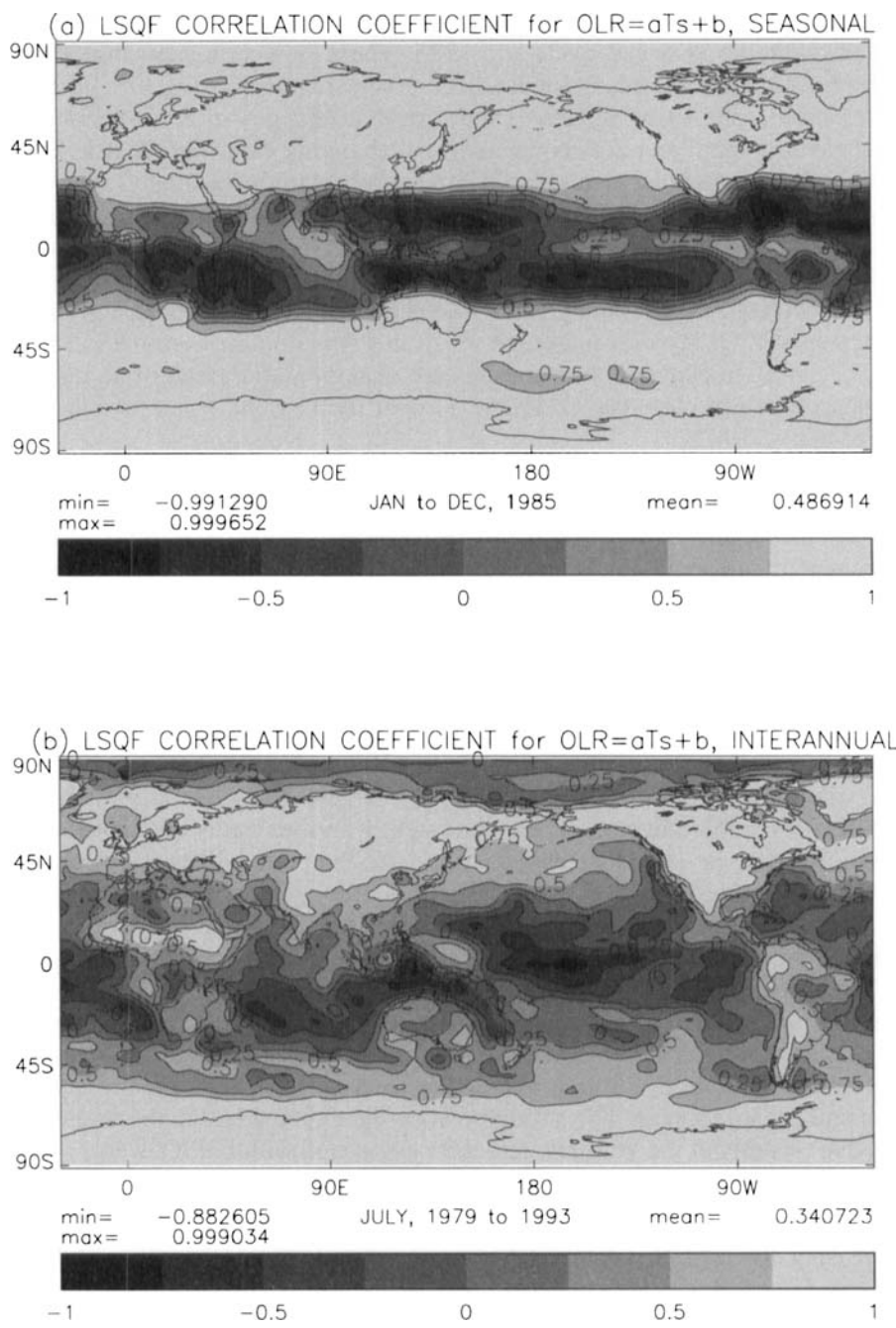


Figure 1. Correlation coefficient for the least-squares fit of the outgoing long-wave radiation (OLR) and the surface temperature (T_s) to $OLR = aT_s + b$ using data from the CLERA simulation (see text) over (a) the seasonal cycle for 1985 and (b) the July interannual variations over the period 1979–1993. The contour interval is 0.25.

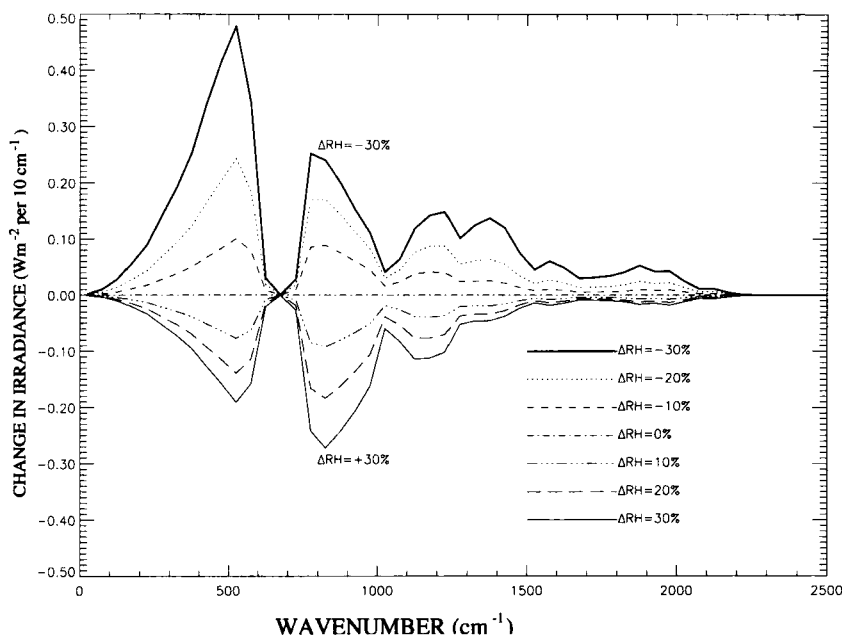


Figure 2. Spectral changes of the clear-sky outgoing long-wave radiation (calculated from a narrow-band model) displayed as a function of wave number (cm^{-1}), assuming uniform relative humidity changes (over the atmospheric layer between 100 and 1000 hPa) to a tropical standard profile having relative humidities set to 40% initially.

To illustrate this point a narrow-band radiative-transfer scheme (Forster and Shine 1997) was applied to a tropical standard profile (McClatchey *et al.* 1972) which had RH set to 40% between 100 and 1000 hPa (Fig. 2). In the major water-vapour bands (approximately wave numbers less than 600 cm^{-1} and greater than 1200 cm^{-1}), absolute changes in OLR are greater for a given decrease in RH than for an *increase* in RH of the same magnitude, confirming the nonlinear increase in OLR with decreasing RH. In the window region of the spectrum (approximately 800 to 1200 cm^{-1}) there is an approximately linear negative dependence of clear-sky OLR on RH. The water-vapour bands vary in their degree of saturation, and so the altitude of maximum contribution to OLR varies throughout the upper and middle troposphere (i.e. about 100 to 800 hPa) depending on wave number. Because continuum absorption in the window region mainly acts in the lower troposphere (i.e. pressures greater than about 600 hPa), RH variations throughout the atmosphere are significant in determining the clear-sky OLR. This is consistent with the findings of Shine and Sinha (1991).

Considering Eq. (4), a tropospheric RH parameter that adequately describes variations in clear-sky OLR must account for the changes in RH in the regions of the atmosphere that are most important in determining clear-sky OLR variability. The dependence of clear-sky OLR on changes of RH ($\Delta\text{OLR}_{\Delta\text{RH}}$) may be written as a function of its sensitivity to RH at pressure, p , and the change in RH(p):

$$\Delta\text{OLR}_{\Delta\text{RH}} = \frac{\partial(\text{OLR})}{\partial(\text{RH}(p))} \Delta\text{RH}(p). \quad (5)$$

Figure 3(a) characterizes the RH variability in the ERA by calculating its standard deviation, SD (%), at each σ -level for the July monthly means (1979–1993) between

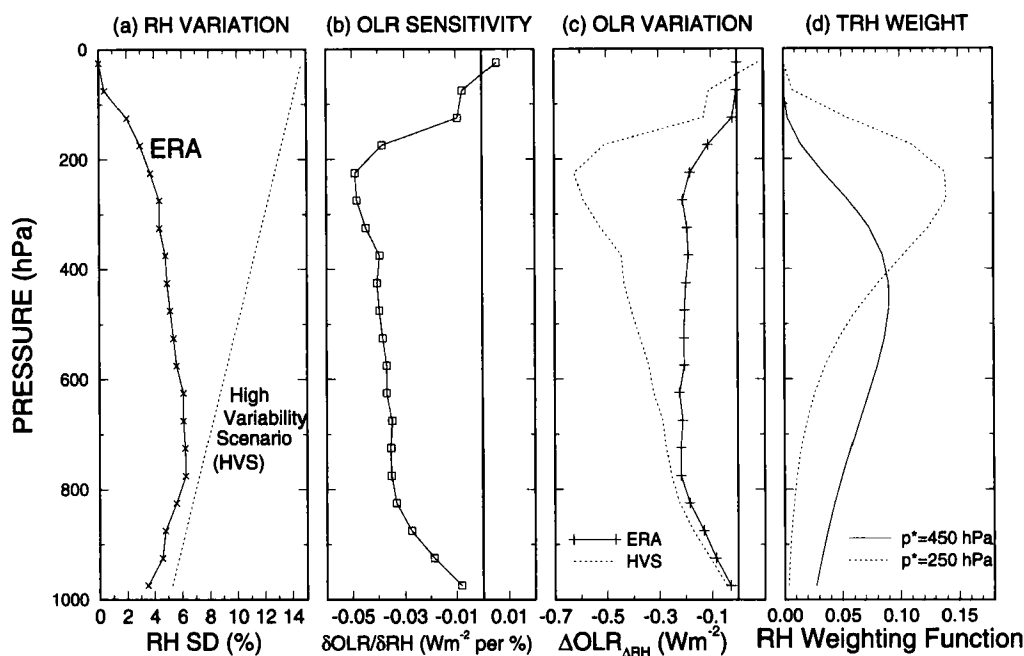


Figure 3. (a) Profiles of the standard deviation of relative humidity (RH SD) obtained from the 15 monthly-mean Julys of the ECMWF reanalysis dataset (full line) over the latitude range 40°N – 40°S , and from a high-variability scenario (dotted line) (see text). (b) Profile of the sensitivity ($\partial\text{OLR}/\partial\text{RH}$) of the clear-sky outgoing long-wave radiation to height-dependent changes (in 50 hPa slabs) of the relative humidity in a tropical standard profile, calculated from a narrow-band model. (c) Inferred variations of clear-sky outgoing long-wave radiation due to relative humidity variations ($\Delta\text{OLR}_{\Delta\text{RH}}$) in the 15 monthly-mean Julys of the ECMWF reanalysis dataset (full line), and in data from a high-variability scenario (dotted line) (see text). (d) Weighting function applied to the relative humidity (for evenly spaced 50 hPa slabs) used to calculate the tropospheric relative humidity (TRH) in Eq. (6) with maximum weights at $p^* = 450$ hPa (full line) and $p^* = 250$ hPa (dotted line).

40°N and 40°S . The $\partial(\text{OLR})/\partial(\text{RH}(p))$ term is provided by the narrow-band scheme applied to a tropical standard atmosphere for 1% increases in RH (Fig. 3(b)). While clear-sky OLR is most sensitive to RH in the upper troposphere (about 250 hPa), variations in the ERA RH are greatest in the middle and lower troposphere (about 600 to 800 hPa). Therefore the inferred OLR variation due to the ERA $\Delta\text{RH}(p)$ (from Eq. (5)) is greatest between about 200 to 850 hPa, with contributions from the entire troposphere (Fig. 3(c)). The calculated OLR variation assumes that the ERA provides a good representation of the true RH variability. In reality, there is significant uncertainty attached to the humidity throughout the atmosphere, especially at higher altitudes (see, for example, McNally and Vesperini (1996)). Therefore estimates of the RH interannual variability must be received with caution. A further calculation was performed whereby the RH standard deviation was assumed to increase linearly from 5% at 1000 hPa to 15% at the top of the atmosphere (Fig. 3(a)). This ‘high-variability scenario’ (HVS) assumes the ERA humidity variability to be underestimated, with the largest underestimation at high altitudes. By calculating the resulting OLR variability (Fig. 3(c)), the potential impact of the uncertainty in RH variability at higher altitudes is illustrated. The maximum OLR variability due to RH variations is at about 250 hPa, but with significant contribution to the OLR variability from RH variations at lower altitudes. It is seen that the upper troposphere has the *potential* to affect the clear-sky OLR variability to a greater extent than lower levels if the ERA humidity variation in the

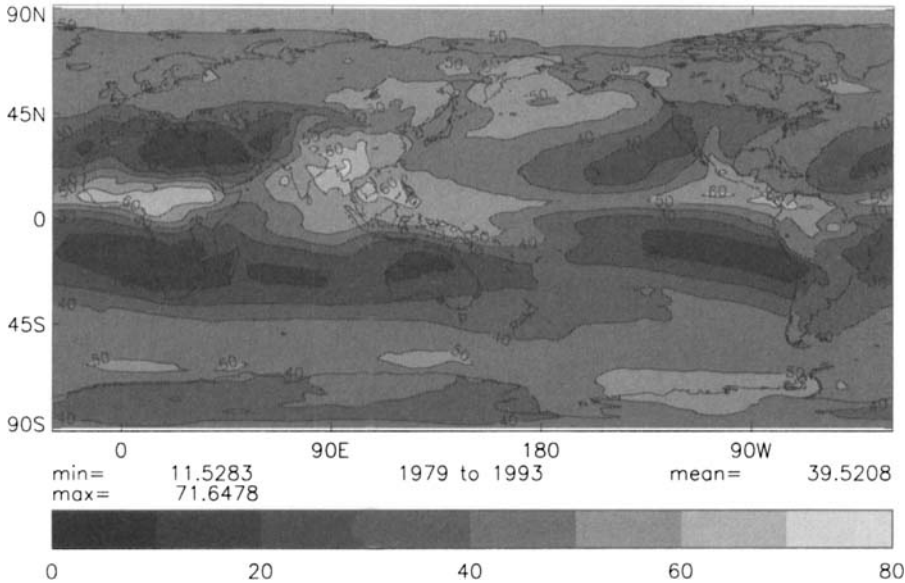


Figure 4. The mean July tropospheric relative humidity (%) over the period 1979–1993 calculated from Eq. (6) using the ECMWF reanalysis dataset. The contour interval is 10%.

upper troposphere is underestimated. This caveat should be considered when using the ERA humidity in studies such as the present one, thus highlighting the need for further work in addressing this uncertainty.

Soden and Fu (1995) used satellite retrievals of the upper-tropospheric RH (the RH over a layer corresponding approximately to 200–500 hPa) to show that the clear-sky normalized atmospheric greenhouse effect is strongly correlated with upper-tropospheric RH, although they note that the lapse rate and lower-tropospheric moisture may explain about half of this dependence. CLERA monthly-mean clear-sky OLR also shows a strong negative correlation with upper-tropospheric RH in the tropics (not shown). Considering the importance of RH variations over the entire troposphere shown in Fig. 3, some of the correlation between upper-tropospheric RH and clear-sky OLR may be due to a coupling between RH in the upper troposphere and RH at lower altitudes.

Taking the results of Fig. 3 into account, the parameter TRH is defined to account for RH throughout the troposphere, but with a weighting function having a maximum in the middle troposphere:

$$\text{TRH} = \frac{\sum_{i=1}^{31} \exp[-2\{\ln(p_i) - \ln(p^*)\}^2] \text{RH}(p_i) \Delta p_i}{\sum_{i=1}^{31} \exp[-2\{\ln(p_i) - \ln(p^*)\}^2] \Delta p_i}, \quad (6)$$

where i denotes σ -level and p^* is the pressure at which the maximum weight is prescribed to RH. The weighting function

$$\exp[-2\{\ln(p_i) - \ln(p^*)\}^2]$$

is plotted in Fig. 3(d) for evenly spaced 50 hPa thick slabs for $p^* = 450$ hPa and $p^* = 250$ hPa. A Gaussian form has been chosen so as not to exclude entirely the RH variations in regions of the atmosphere less important for the determination of clear-sky OLR. This is also more consistent with the form of the satellite-derived upper-tropospheric RH weighting function used by Soden and Fu (1995) than would be the

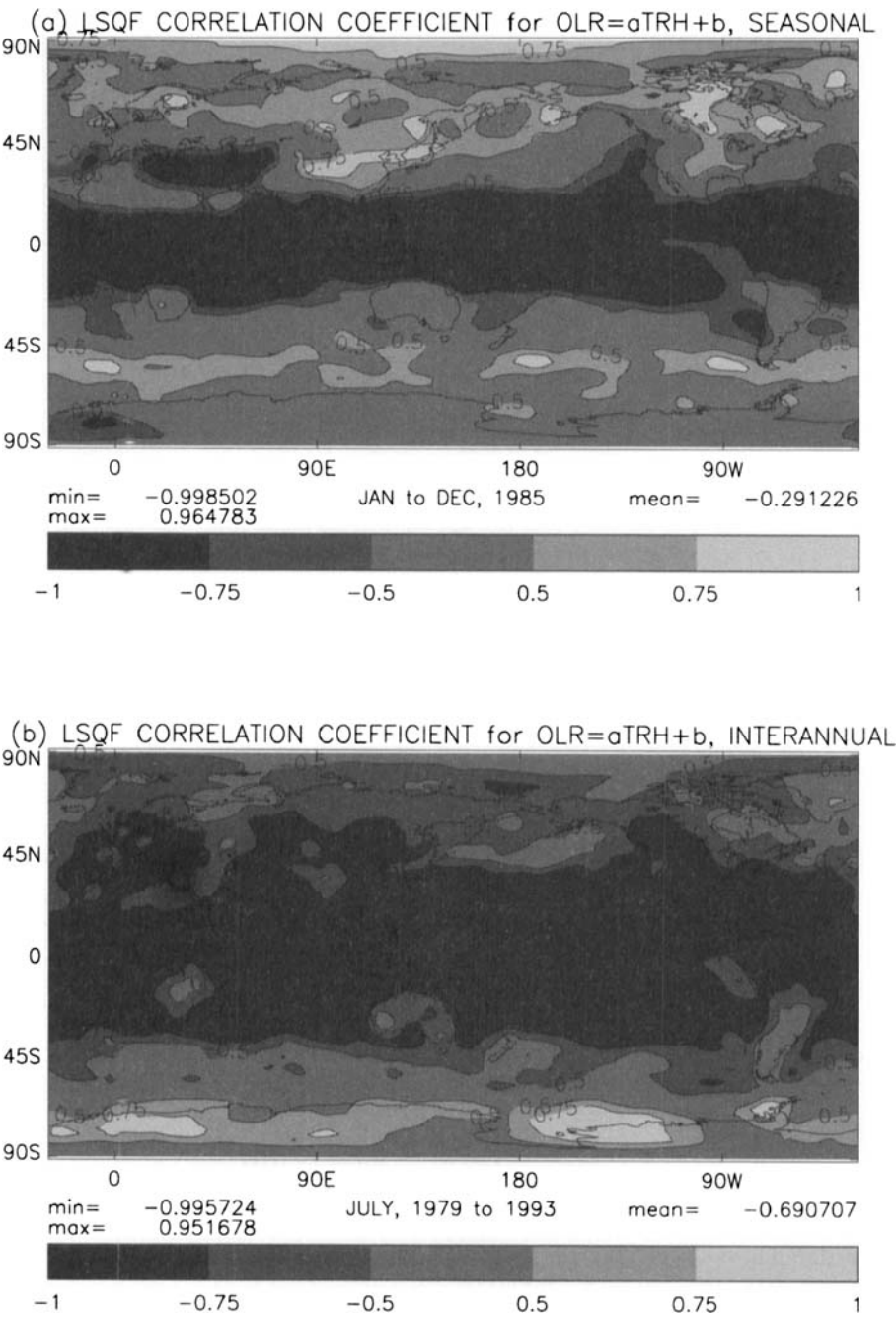


Figure 5. Correlation coefficient for the least-squares fit of the outgoing long-wave radiation (OLR) and the tropospheric relative humidity (TRH) to $OLR = aTRH + b$ using results from the CLERA simulations (see text) (a) over the seasonal cycle of 1985, and (b) for July over the period 1979–1993.

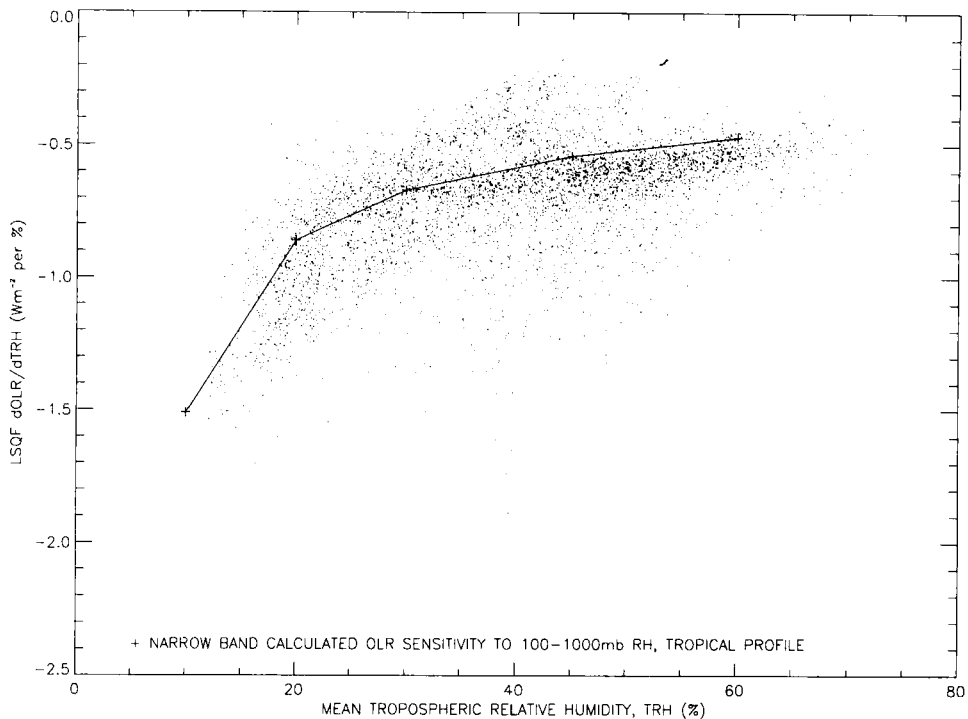


Figure 6. The sensitivity ($d(\text{OLR})/d(\text{TRH})$; units W m^{-2} per %) of the clear-sky outgoing long-wave radiation (OLR) to changes in tropospheric relative humidity (TRH) plotted against the 1979–1993 mean tropospheric relative humidity for July. Only grid points with correlations significant above the 99.5% level are plotted. The solid line denotes the corresponding narrow-band sensitivities when the relative humidity is varied within the 100–1000 hPa layer in a tropical standard profile.

case by taking a column-mean RH. It was decided to calculate TRH using $p^* = 450$ hPa so as not to exclude the contribution of RH variations in the lower troposphere to changes in clear-sky OLR variability. Although there is significant uncertainty with regard to RH variability in the upper troposphere and stratosphere, the low weight prescribed to pressures less than 150 hPa is still valid, given that clear-sky OLR is relatively insensitive to changes in RH at these levels (Fig. 3(b)). The mean July tropospheric RH over the period 1979–1993 calculated from Eq. (6) is plotted in Fig. 4. The dry descending regions of the subtropics are highlighted by a tropospheric RH of below about 30%, while the ascending regions of the tropical warm pool and the Asian monsoon, as well as over central Africa, are characterized by a high value of the tropospheric RH (above 60%).

Figure 5(a) shows that there is a strong negative correlation between clear-sky OLR and the tropospheric RH over the seasonal cycle of 1985 within about 30° of the equator. Away from the tropics the correlation is smaller and tends to be positive, implying that changes in other parameters influence the clear-sky OLR changes. The region of strong negative correlation over the interannual time-scale for July extends up to 45° away from the equator and the global-average correlation coefficient is -0.7 (Fig. 5(b)). The TRH explains more of the variance in OLR over the interannual time-scale than over the seasonal cycle because temperature fluctuations are smaller from year to year than over a season. The correlation distribution is similar when setting p^* to 250 hPa (not shown). Comparing Figs. 1 and 5 it is apparent that the tropospheric RH explains most of the

clear-sky OLR variations at low latitudes, while T_s is important in explaining clear-sky OLR variations at high latitudes.

Figure 6 shows the sensitivity (defined by $d(\text{OLR})/d(\text{TRH})$ for the LSQF of $\text{OLR} = a\text{TRH} + b$ to the CLERA results) of the clear-sky OLR to changes in TRH during July. The scatter of points about the calculated narrow-band sensitivity are likely to be the result of OLR sensitivity to temperature variations, as described in Eq. (4). The TRH parameter also cannot capture the changes in RH throughout the atmosphere, introducing an additional source of error. Nevertheless, the simulation captures the tendency for $d(\text{OLR})/d(\text{TRH})$ to be most negative in the regions with the lowest mean TRH because the water-vapour absorption bands saturate with increasing RH (Fig. 2). The increased sensitivity of clear-sky OLR to TRH for the drier descending regions of low TRH that are highlighted in Fig. 4 are important with regard to the water-vapour feedback for two further reasons: (a) The dry descending regions contain a high proportion of clear skies and cover large areas of the tropics; and (b) radiative processes in these regions significantly affect the total heating.

5. HEIGHT-DEPENDENT VARIATIONS OF RH

(a) *Experimental design*

While section 4(b) highlights the strong negative correlation between tropospheric RH and clear-sky OLR, further experiments are necessary to quantify the effects of RH variation throughout the atmosphere. For this purpose, five additional experiments were undertaken whereby the monthly-mean clear-sky OLR was calculated by the CLERA system using the 15 monthly-mean July fields from the ERA. This differs from the reference version of the CLERA simulations used in section 4 which accumulated the monthly-mean clear-sky OLR from values calculated every six hours. To assess the effect of RH variations on the clear-sky OLR, a control experiment was run using the 15 monthly-mean ERA fields. Perturbation experiments were subsequently run in which the specific humidity was varied with the ERA temperature to conserve the ERA 15-year mean July RH at each grid point at specified σ -levels, as defined in Table 1. The experiments are described as follows:

CTRh—Reference version of the CLERA simulation which used six-hourly fields to construct the monthly-mean irradiance and allowed the RH to vary with time throughout the atmosphere;

CTRm—Control experiment in which the RH was allowed to vary with time throughout the atmosphere. The only difference from CTRh is that monthly-mean input was used;

CONRH—The RH was set equal to the 15-year July-mean value at each σ -level;

UTRH—The RH was allowed to vary with time in the upper troposphere only;

MTRH—The RH was allowed to vary with time in the middle troposphere only;

LTRH—The RH was allowed to vary with time in the lower troposphere only.

Note that the labels for the experiments are used for convenience and are most appropriate for the tropical troposphere; in the extratropics much of what is labelled as ‘upper troposphere’ lies in the lower stratosphere.

TABLE 1. DESCRIPTION OF THE FIVE EXPERIMENTS TO CALCULATE THE JULY MONTHLY-MEAN CLEAR-SKY OUTGOING LONG-WAVE RADIATION FOR THE 15 YEARS OF THE ECMWF REANALYSIS (ERA) DATASET, USING COMBINATIONS OF MONTHLY-MEAN PROFILES OF SPECIFIC HUMIDITY (q) OBTAINED FROM THE ERA DATA FOR EACH OF THE YEARS 1979–1993 AND PROFILES OF q COMPUTED FROM THE 15-YEAR MEAN JULY RELATIVE HUMIDITY VALUES AT EACH σ -LEVEL ($\overline{\text{RH}}(\sigma)$)

Experiment	σ -level number n	Specific humidity q	Regions having relative humidity varying with time
CTRM	$1 \leq n \leq 31$	ERA monthly-mean profiles	All levels
CONRH	$1 \leq n \leq 31$	$0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T)$	No levels
UTRH	$\left\{ \begin{array}{l} 1 \leq n < 5 \\ 5 \leq n \leq 14 \\ 14 < n \leq 31 \end{array} \right.$	$\left\{ \begin{array}{l} 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \\ \text{ERA monthly-mean profiles} \\ 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \end{array} \right.$	Upper troposphere (≈ 100 to 400 hPa)
MTRH	$\left\{ \begin{array}{l} 1 \leq n < 15 \\ 15 \leq n \leq 21 \\ 21 < n \leq 31 \end{array} \right.$	$\left\{ \begin{array}{l} 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \\ \text{ERA monthly-mean profiles} \\ 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \end{array} \right.$	Middle troposphere (≈ 400 to 700 hPa)
LTRH	$\left\{ \begin{array}{l} 1 \leq n < 22 \\ 22 \leq n \leq 29 \\ 29 < n \leq 31 \end{array} \right.$	$\left\{ \begin{array}{l} 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \\ \text{ERA monthly-mean profiles} \\ 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T) \end{array} \right.$	Lower troposphere (≈ 700 to 1000 hPa)

$q_{\text{sat}}(T)$ is the saturation specific humidity as a function of the July monthly-mean temperature profiles.

Where the RH was held constant, the specific humidity was calculated as:

$$q = 0.01\overline{\text{RH}}(\sigma)q_{\text{sat}}(T), \tag{7}$$

where σ is the height coordinate, $\overline{\text{RH}}(\sigma)$ is the 15-year July mean RH at a given level and $q_{\text{sat}}(T)$ denotes the saturation specific humidity for the ERA temperature T of each σ -level calculated using an empirical form of the Clausius–Clapeyron equation (Liou 1980). At levels where the RH was allowed to vary with time, the July ERA monthly-mean q for the given year was used. Because of the nonlinear dependence of clear-sky OLR on q and T , the global-mean clear-sky OLR from monthly-mean profiles (experiment CTRm) was 1.8 W m^{-2} less than the corresponding value for the reference CLERA simulation which used six-hourly calculations (experiment CTRh), as shown in Table 2. Thus, experiment CTRm was necessary to ensure consistency between the perturbation and control experiments.

(b) Global-mean clear-sky OLR variability

The variability of the simulated clear-sky OLR can be characterized by its standard deviation (SD_{OLR}) for the 15 years. Table 2 shows that the global-mean standard deviation from CTRm was almost identical to that from CTRh, indicating that using monthly-mean input does not affect the interannual variability. However, the global-mean standard deviation for CONRH (in which the RH is constant) was roughly half that of CTRm (in which the RH varies with time). Where the RH was allowed to vary with time in only the upper, middle or lower troposphere the standard deviation of the clear-sky OLR was larger than for experiment CONRH, showing that RH variations throughout the troposphere affect the global-mean clear-sky OLR variability. The OLR standard deviation was largest for experiment MTRH and was least for LTRH. Therefore the order of importance for global variations in clear-sky OLR due to tropospheric RH variations is the middle, the upper and then the lower troposphere.

TABLE 2. THE JULY MEAN (OVER THE PERIOD 1979–1993) OF THE GLOBAL-MEAN CLEAR-SKY OUTGOING LONG-WAVE RADIATION (OLR) AND THE STANDARD DEVIATION (SD_{OLR}) FOR THE RESULTS OF VARIOUS EXPERIMENTS, TOGETHER WITH THE SENSITIVITY OF THE OLR TO CHANGES IN THE GLOBAL-MEAN SURFACE TEMPERATURE (T_s), AND THE CORRELATION COEFFICIENT (r) FOR THE LEAST-SQUARES FIT OF THE RESULTS TO $OLR = aT_s + b$ (SEE TEXT)

Experiment	OLR ($W\ m^{-2}$)	SD_{OLR} ($W\ m^{-2}$)	Sensitivity ($W\ m^{-2}K^{-1}$)	Correlation coefficient r
CTRh	268.5	3.0	1.38	0.41
CTRm	266.7	2.9	1.42	0.42
CONRH	266.5	1.5	2.28	0.85**
UTRH	266.6	1.9	2.46	0.81**
MTRH	266.6	2.1	1.66	0.60*
LTRH	266.5	1.7	1.93	0.76**

The symbol ** denotes a statistically-significant correlation at greater than the 99.5% confidence level. The symbol * denotes a statistically-significant correlation at greater than the 95% confidence level.

The correlation between *global-mean* clear-sky OLR and T_s is not significant at the 99.5% level for experiments CTRh and CTRm. However, when the RH was not allowed to vary (CONRH), a strong positive correlation of 0.85 was calculated, with the global-mean clear-sky OLR increasing at the rate of $2.3\ W\ m^{-2}K^{-1}$. Thus, changes in RH act to mask the dependence of clear-sky OLR on T_s in a global-mean sense. When all 180 global monthly-mean clear-sky OLR values were used from CTRh, and the seasonal cycle was filtered out, a statistically significant correlation at the 99.9% confidence level was attained ($r = 0.49$) with $d(OLR)/dT_s$ taking a value of $2.1\ W\ m^{-2}K^{-1}$. Thus, changes in RH do not appear to affect greatly the dependence of clear-sky OLR on T_s , but merely affect the degree of linear correlation. The sensitivity of the global-mean clear-sky OLR to T_s is similar to that inferred from climate models, as discussed by Cess *et al.* (1990).

Figure 7 shows the global-mean clear-sky OLR for each July plotted as a function of global-mean T_s for experiments CONRH and CTRm; it highlights the increase in correlation once the RH signal is removed. When considering the years before 1985, the clear-sky OLR for a given T_s is greater than for the years after 1985 for experiment CONRH. An additional run with the CLERA simulation was performed in which the well-mixed greenhouse gases were set at July 1979 levels, while all other parameters were as defined in CTRm for July 1993. The global-mean clear-sky OLR difference due to changes in greenhouse gases over the 15-year period were calculated as $0.5\ W\ m^{-2}$ for the July 1993 profiles. This is enough to explain the apparent decrease of clear-sky OLR at a given T_s in Fig. 7. When the clear-sky OLR was adjusted to account for the increasing concentrations of greenhouse gases, the correlation between the global-mean clear-sky OLR and T_s increased for all experiments.

When only the upper-tropospheric RH was allowed to vary with time (experiment UTRH) the sensitivity of the global-mean clear-sky OLR to T_s was greater than for experiment CONRH (Table 2). Conversely, when only the lower- or middle-tropospheric RH were allowed to vary with time $d(OLR)/dT_s$ was less than for experiment CONRH. This suggests that upper-tropospheric RH variations counteract the thermodynamic water-vapour feedback, while RH variations in the lower and middle troposphere enhance the thermodynamic water-vapour feedback. Using a *t*-test (Freund 1976), the values of $d(OLR)/dT_s$ for experiments CONRH and UTRH were found not to be significantly different at the 95% confidence level. Therefore a longer dataset is required

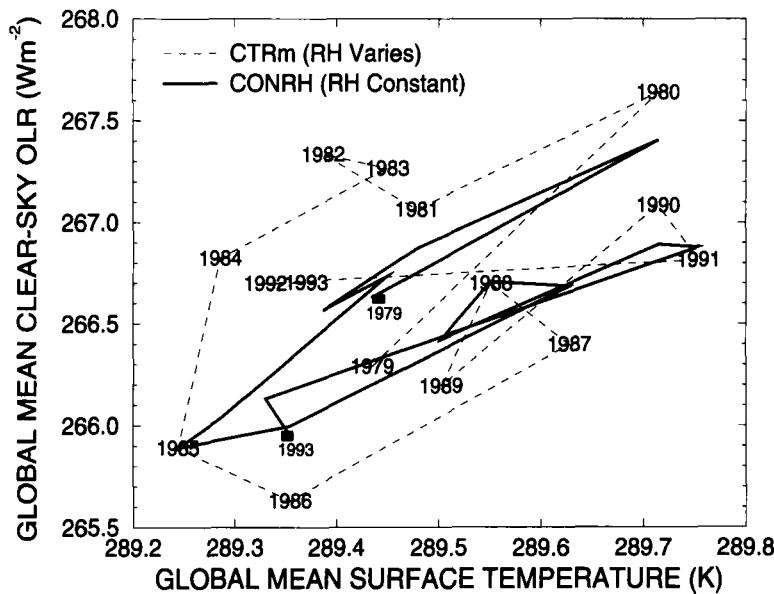


Figure 7. The global-mean clear-sky outgoing long-wave radiation (OLR: units W m^{-2}) plotted against the global-mean surface temperature (T_s) for the 15 monthly-mean Julys of the ECMWF reanalysis period (1979–1993) for experiments CTRm (dashed line) and CONRH (solid line). Numbers within the frame denote the year; experiment CTRm is annotated for individual years, but only end points are noted for experiment CONRH (see text and Table 1 for further information).

to identify a negative feedback due to upper-tropospheric RH. Thus, there is weak evidence that the negative water-vapour feedback mechanism proposed by Lindzen (1990) may operate, but only in a small portion of the atmosphere. The remainder of the atmosphere appears to counteract the upper-troposphere effect, causing the overall water-vapour feedback to be strongly positive.

The OLR standard deviation is plotted for experiments CTRm in Fig. 8(a) and CONRH in Fig. 8(b). Large values of SD_{OLR} (greater than 3 W m^{-2}) are present in both experiments for high-latitude regions (most especially over Antarctica) where the variation of clear-sky OLR is more dependent on changes in temperature than on changes in RH. The largest reduction in the SD_{OLR} occurs over ocean regions within about 30° of the equator, most especially over the tropical central Pacific and the region north of the Arabian Sea. This reinforces the conclusions of section 4(b), namely that changes in RH control the clear-sky OLR variation over middle and low latitudes while T_s appears to dominate the OLR variability over high-latitude land regions.

(c) Height-dependent RH variation and clear-sky OLR

Comparing standard deviations indicates the differences between the experiments in the simulated variabilities over the 15 months. However, a better measure of the fraction of clear-sky OLR variability due to changes in RH is to consider OLR-anomaly differences between the control and perturbation experiments. Therefore the fraction of clear-sky OLR anomaly, ΔOLR , due to changes in RH may be calculated as follows for

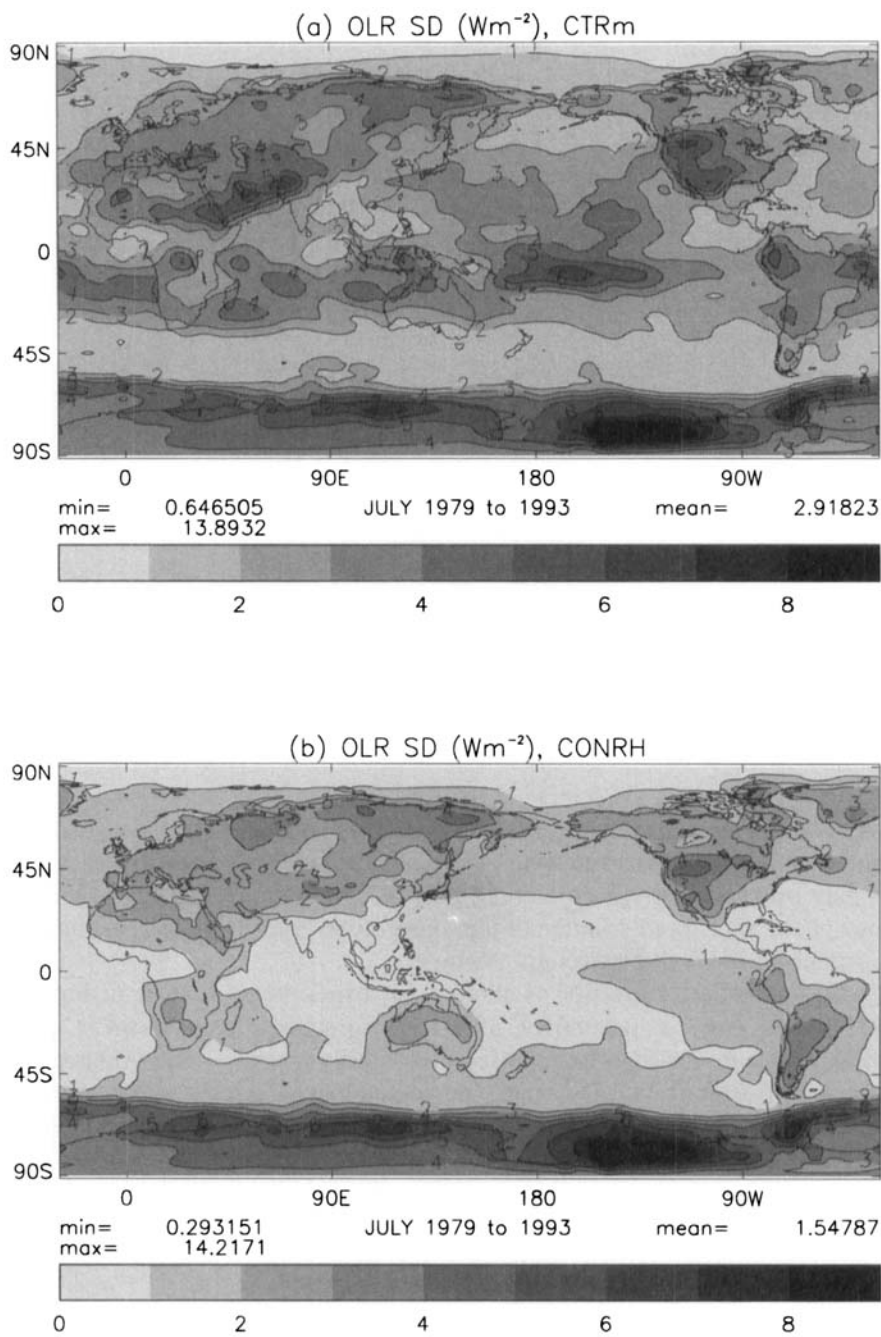


Figure 8. The standard deviation of the clear-sky outgoing long-wave radiation (W m^{-2}) for July over the period 1979–1993, plotted globally (a) where the relative humidity was allowed to vary with time at each level (experiment CTRm) and (b) where the relative humidity was held constant at each level (experiment CONRH). The contour interval is 1 W m^{-2} .

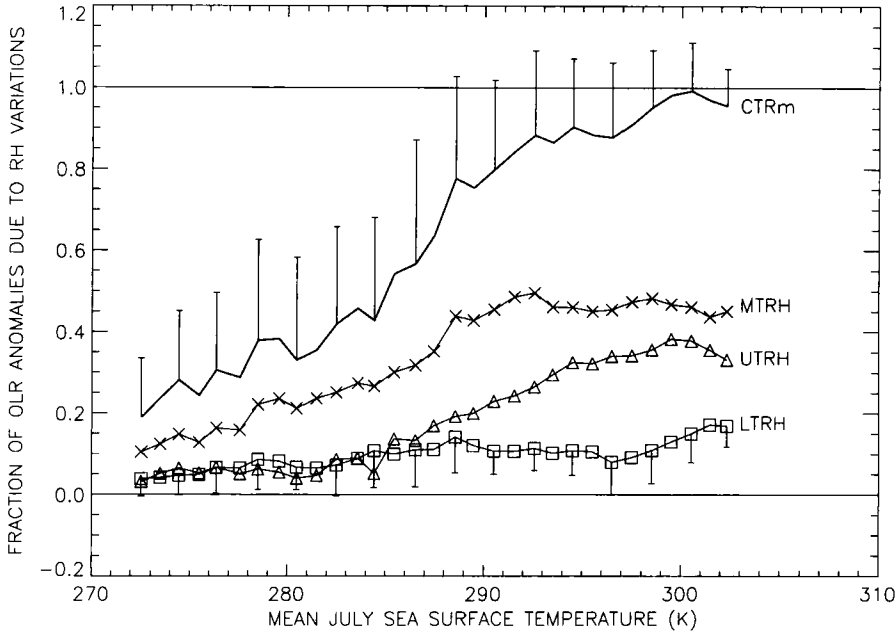


Figure 9. Fractional anomalies f (from the baseline experiment CONRH, see Eq. (8)) of the clear-sky outgoing long-wave radiation (OLR) due to changes in relative humidity for experiments CTRm, UTRH, MTRH and LTRH (see text and Table 1) plotted against the global-mean July surface temperature (averaged over 1 K bins for oceanic regions only). Only grid points having clear-sky OLR anomalies greater than 2 W m^{-2} have been considered. Vertical bars denote one standard deviation about the mean value of f in the temperature bin; for clarity, these are plotted every alternate temperature bin for experiments f_{CTRm} and f_{LTRH} .

each grid point:

$$f_j = \frac{1}{N} \sum_{i=1}^N \left[\frac{(\Delta \text{OLR}_{i,j} - \Delta \text{OLR}_{i,\text{CONRH}})}{\Delta \text{OLR}_{i,\text{CTRm}}} \right], \quad (8)$$

where i denotes the year, N is the number of years and j denotes UTRH, MTRH, LTRH or CTRm. Only grid points where $(\Delta \text{OLR}_{i,\text{CTRm}})^2$ is greater than $4 \text{ (W m}^{-2}\text{)}^2$ are considered, so as to increase the signal-to-noise ratio.

Figure 9 shows f calculated using Eq. (8) for experiments with time changes in RH throughout the troposphere (f_{CTRm}), and for experiments with time changes restricted to the upper, middle and lower troposphere. f_{CTRm} is about equal to the sum of f_{UTRH} , f_{MTRH} and f_{LTRH} . There is a rapid rise in f_{CTRm} with T_s , the RH changes explaining nearly all of the variation in clear-sky OLR at temperatures of about 300 K. For temperatures below about 290 K, much of this clear-sky OLR dependence on RH is due to middle-tropospheric RH variations. At temperatures greater than about 290 K, changes in the upper-tropospheric RH become increasingly important, as do changes in lower-tropospheric RH at temperatures above 295 K. f_{UTRH} decreases at temperatures above about 300 K (mainly over the tropical west Pacific regions). This may be a result of the smaller upper-tropospheric RH variability for the most moist profiles. While the middle-tropospheric RH appears to be the dominant control on clear-sky OLR away from high latitudes, the variations of RH *throughout* the troposphere contribute significantly to clear-sky OLR variability. When f is binned in land-surface temperature intervals (not shown), values are generally lower than over the ocean. However, the

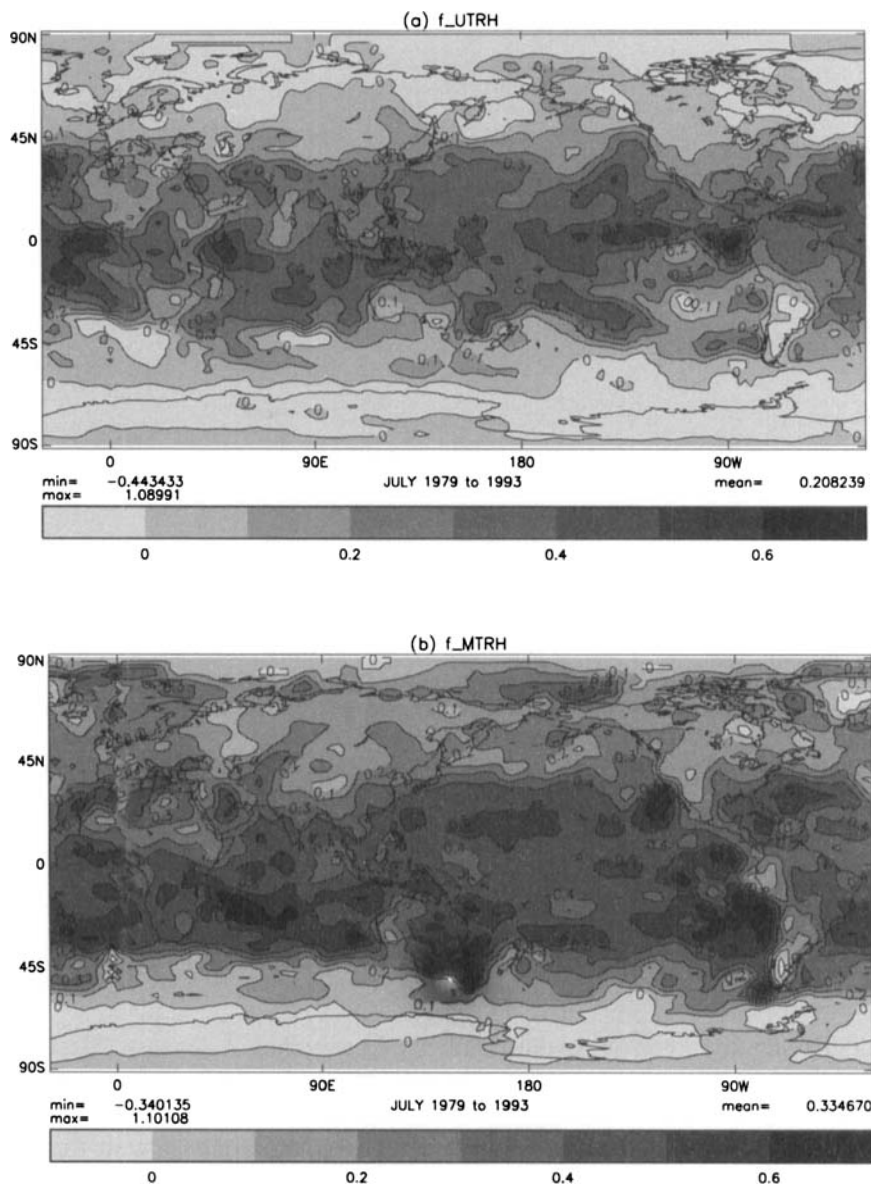


Figure 10. Fractional anomalies of clear-sky outgoing long-wave radiation (OLR) for July (see Eq. (8)) due to interannual changes in relative humidity in (a) the upper, (b) the middle and (c) the lower troposphere. Only grid points where clear-sky OLR anomalies are greater in magnitude than 2 W m^{-2} are considered. The contour interval is 0.1.

importance of middle-tropospheric RH and the rapid increase in f with T_s are still evident.

Figures 10(a)–(c) show the global distribution of f for experiments UTRH, MTRH and LTRH, respectively. Experiment MTRH explains a significant amount of clear-sky OLR variability throughout much of the globe, excluding Antarctica. This is especially so off the west coast of South America where middle-tropospheric RH variations appear to dominate clear-sky OLR variations. The global-mean f_{MTRH} is 0.33 which implies

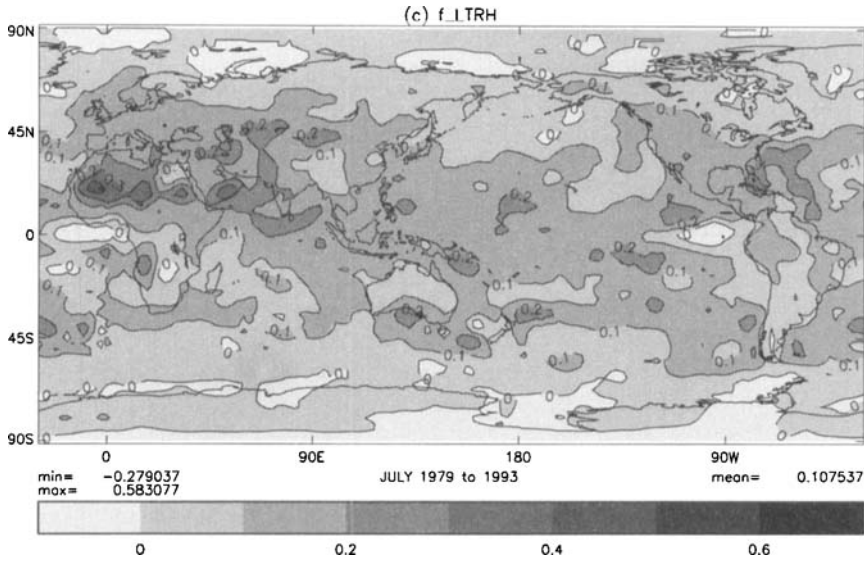


Figure 10. Continued.

that middle-tropospheric RH variations account for, on average, about a third of the clear-sky OLR variation. f_{UTRH} is generally above 0.4 within about 40° of the equator and accounts for about 21% of the variation in clear-sky OLR. The lower-tropospheric RH appears to contribute little to the clear-sky OLR variation (Fig. 10(c)). However, when averaged over the globe, about 11% of the clear-sky OLR variability is explained by changes in lower-tropospheric RH. Thus the conclusions of section 5(b) are also supported by this analysis: (a) the order of importance of regions in the troposphere within which RH variations influence OLR variations is the middle troposphere, the upper troposphere and then lower troposphere; and (b) RH variations throughout the troposphere contribute significantly to changes in OLR from year to year.

(d) *The super-greenhouse effect*

Section 4(a) shows an SGE to act over tropical ocean regions for the interannual time-scale. This is explained in terms of the change in the convective regime due to the ENSO cycle that are associated with changes in T_s . When the RH is held constant in time, as in experiment CONRH, there are no regions exhibiting a statistically-significant SGE. This is consistent with Hallberg and Inamdar (1993) who showed that thermodynamics alone is not enough to explain the SGE. Figure 11 shows the sensitivities of the OLR to changes in T_s for experiments CTRm, CONRH, UTRH, MTRH and LTRH. The sensitivity indicated by experiment CONRH ranges from between about $1.0 \text{ W m}^{-2}\text{K}^{-1}$ in the tropics to about $1.9 \text{ W m}^{-2}\text{K}^{-1}$ elsewhere. An SGE is not apparent in any of the T_s bins for experiment CONRH, while an SGE applies at sea surface temperatures above about 296 K for experiment CTRm, albeit with a large scatter about the mean value for the bin. Allowing changes in RH for each layer results in a decreased $d(\text{OLR})/dT_s$ compared with the CONRH experiment. The sum

$$\sum_j \left[\left\{ \frac{d(\text{OLR})}{dT_s} \right\}_j - \left\{ \frac{d(\text{OLR})}{dT_s} \right\}_{\text{CONRH}} \right],$$

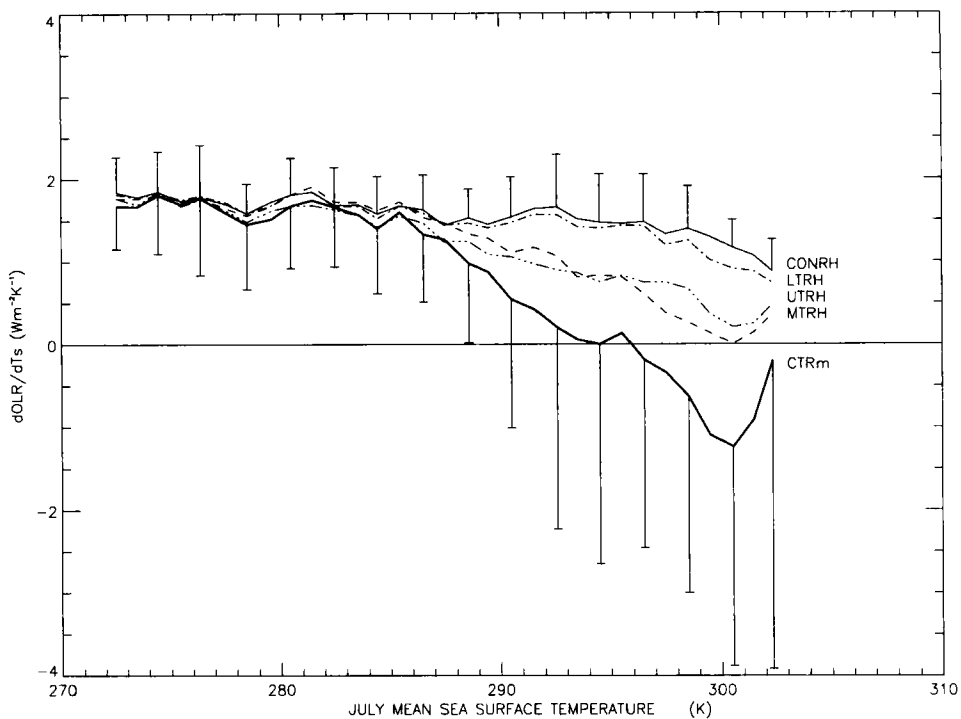


Figure 11. The sensitivity ($d(\text{OLR})/dT_s$) of the clear-sky outgoing long-wave radiation (OLR) to changes in T_s over the oceans ($\text{W m}^{-2}\text{K}^{-1}$) determined from the least-squares fit of the results to $\text{OLR} = aT_s + b$ (see text) plotted against the mean July sea surface temperature over the period 1979–1993 (calculated in 1 K bins) for experiments CTRm, CONRH, UTRH, MTRH and LTRH (see Table 1). Vertical bars denote one standard deviation about the mean value of $d(\text{OLR})/dT_s$; for clarity they are plotted for experiments CONRH and CTRm only.

where j denotes experiments UTRH, MTRH and LTRH, is approximately equal to

$$\left\{ \frac{d(\text{OLR})}{dT_s} \right\}_{\text{CTRm}} - \left\{ \frac{d(\text{OLR})}{dT_s} \right\}_{\text{CONRH}}.$$

A decrease in $d(\text{OLR})/dT_s$ at high values of T_s for experiment CONRH shows that thermodynamics contributes to producing SGE situations because, over the tropical oceans, the clear-sky OLR is essentially decoupled from T_s . This is due to the large column moisture and, therefore, a small Tr term in Eq. (4). Allowing RH to vary with time in the lower troposphere only in experiment LTRH results in a similar $d(\text{OLR})/dT_s$ to that of experiment CONRH. However, for experiments MTRH and UTRH, $d(\text{OLR})/dT_s$ is significantly lower at high sea surface temperatures. Thus it is mainly the changes in the upper- and middle-tropospheric RH that cause SGE situations to occur. Similar conclusions have been drawn from other studies (e.g. Hallberg and Inamdar 1993).

6. DISCUSSION

A simulation of the earth's clear-sky long-wave radiation budget is used to examine the dependence of clear-sky OLR on surface and atmospheric temperature and on relative humidity throughout the troposphere. Changes in T_s appear to dominate the

clear-sky OLR variation at high latitudes, especially over the land regions, while over middle and low latitudes changes in tropospheric RH appear to control clear-sky OLR variability. Globally, tropospheric RH explains a larger proportion of OLR variations over the interannual time-scale than over the seasonal cycle. This is because, for a given change in tropospheric RH, variations in T_s are smaller over the interannual time-scale at all latitudes except the tropics.

An SGE (defined as negative values of $d(\text{OLR})/dT_s$) is identified over both seasonal and interannual time-scales. When RH is held constant, or is allowed to vary with time in the lower troposphere only, no statistically-significant SGE is produced. However, when the upper- or mid-tropospheric RH is allowed to vary, an SGE is identified over the interannual time-scale. The necessary change in RH is provided only when a change in T_s is associated with a transition between strong and weak convection. This may only occur where ascent is imposed by the large-scale circulation (Bony *et al.* 1997). Over the seasonal cycle, the shift in the Hadley circulation imposes such a transition in the tropics (although away from the equator) which corresponds with the regions exhibiting an SGE. On the interannual time-scale, the shift in the Walker circulation associated with the ENSO is sufficient to alter the degree of convection in the tropical Pacific necessary to produce an SGE situation. Because the SGE constitutes an unstable radiative feedback to the surface (Inamdar and Ramanathan 1994) stabilizing mechanisms, such as short-wave cloud radiative feedbacks, evaporative feedbacks and heat export to the extratropics, are required to prevent a local runaway greenhouse effect from operating (Pierrehumbert 1995).

The sensitivity of clear-sky OLR to tropospheric RH is most negative in regions of low-tropospheric RH. This can be explained by the saturation of the water-vapour absorption bands with increasing RH. In all but the most moist regions clear-sky OLR is most sensitive to a given RH variation if that change is in the upper troposphere; Spencer and Braswell (1997) have used this result to argue that changes in the upper-tropospheric RH are very important for climate change. However, because RH variations in the upper troposphere are generally associated with larger changes in RH at lower levels, the importance of upper-tropospheric RH with regard to the clear-sky OLR may have been overstated. Because, in the ERA, the RH variations are greatest between about 600 and 800 hPa, the *middle*-tropospheric RH contributes most to clear-sky OLR variability over the interannual time-scale. This supports the conclusions of previous studies (e.g. Clough *et al.* 1992; Sinha and Harries 1997). By allowing RH to vary with time in only the upper, middle or lower troposphere, it is shown that the middle troposphere (about 400 to 700 hPa) contributes about one third of the clear-sky OLR variability. The upper and lower troposphere contribute about 21% and 11%, respectively. Therefore the RH *throughout* the troposphere contributes significantly to clear-sky OLR variation. Changes in surface and atmospheric temperature account for the remaining 35% of the variability. There is weak evidence in support of a negative water-vapour feedback as suggested, for example, by Lindzen (1990), although this is only the case for upper-tropospheric RH variations; changes in the RH at lower levels of the troposphere counteract this effect, resulting in an overall positive water-vapour feedback.

The results presented are relevant for interannual variations between Julys. Therefore the conclusions are relevant for the interannual variability due to changes in the large-scale circulation for this month. For example, changes in the Asian monsoonal circulation may be significant in dictating the variability of RH between Julys. The profound link between clouds and water vapour also render the present study incomplete with regard to the overall response of the climate system to radiative forcing. For the

experiments undertaken, changes in RH do not appear to significantly alter the relationship between global-mean clear-sky OLR and T_s , but merely to decrease their linear correlation.

While information concerning the sensitivity of the earth's radiation budget to observed changes in surface and atmospheric parameters yields valuable information as to the mechanisms acting in the climate system, climate-model experiments are required to perform relevant climate-sensitivity calculations. Unfortunately, these studies themselves are limited by the potential inability of models to parametrize accurately the moist convection so important in determining the water-vapour and cloud feedbacks. Therefore it should be a prerequisite that climate models should be able to reproduce the sensitivities and variability that are provided by observations and simulations such as the present study.

A major limitation of the CLERA simulation is the reliance on the numerical-model calculations used in the ERA project. This is especially true of the upper troposphere and the dry descending oceanic 'deserts' where observational data are limited. Hence, a large uncertainty is attached to the ERA humidity and its variability in these regions (e.g. Spencer and Braswell 1997). The inference of the present study, namely that clear-sky OLR variability is dependent mainly on RH variations in the middle troposphere rather than in the upper troposphere, is therefore dependent on the validation of the ERA humidity variability, most especially in these dry regions of the atmosphere. The dry descending regions of low-tropospheric RH are likely to be important with regard to the water-vapour feedback. Not only is clear-sky OLR most sensitive to tropospheric RH variations in these regions but, unlike the moist tropics, the cloud amount is small, thus increasing the importance of clear-sky feedbacks. Also, the radiative cooling is a dominant diabatic term in the descending portions of the Hadley circulation. Thus, in understanding the climate system and the water-vapour feedback, attaining robust observations of humidity in these regions must be paramount. By testing the ability of climate models to capture the variability of moisture and the associated variability in the clear-sky greenhouse effect over sufficiently long time-scales, confidence can be increased in the prediction of future climate change.

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