Sensitivity Analysis of Outgoing Radiation at the Top of the Atmosphere in the NCEP/MRF Model

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Abstract

The outgoing radiation at the top of the atmosphere (TOA) determines the net radiative heating of the earth's surface and atmosphere system. Variabilities of the TOA outgoing radiation depend on features of the earth's surface and atmosphere. The fundamental goal of the present study is to quantify the sensitivity of the outgoing radiation to those features in a sophisticated radiation model used in the National Center of Environment Prediction (NCEP) medium range forecasting system. The technique of adjoint sensitivity analysis is employed.

The results obtained show that the TOA outgoing radiation is highly sensitive to spatial structures and temporal variations of water vapor, atmospheric and earth's temperature, and cloud cover. The outgoing short wave radiation variabilities are much more sensitive to the water vapor in the middle troposphere at high latitudes in the summertime hemisphere than at the tropical and subtropical areas. The outgoing long wave radiation variabilities are one order of magnitude more sensitive to specific humidity in the upper troposphere at mid- and lower altitudes than to that in other areas. In the tropics, more than 80% of a perturbation increase in the earth's surface emission is trapped by the clear sky column atmosphere, while only about 60 to 70% is trapped at middle and high latitudes; a rapid decrease band of trapping is found within the subtropics. For the outgoing short wave radiation, the lower clouds are more sensitive than the middle and high clouds. The most sensitive lower clouds tend to be located at mid-latitudes rather than in the tropical regions. Deep clouds and thin clouds display different impacts. For the outgoing long wave radiation, as expected, the high clouds are the most sensitive, but do not differ significantly from the middle clouds in the deep cloud case. The net radiation (shortwave and longwave radiation) is most sensitive to lower clouds at mid-latitudes in the summertime hemisphere with a cooling effect on the atmosphere. Implications of the obtained results for understanding supper green house effect over tropical oceans and the influence of cloud feedback on the climate change are explored. Our analyses reveal some deficiencies in the model when the calculated results are compared with relevant observations.

1 Introduction

Interaction between radiation and atmospheric processes creates one of the main challenges in understanding the climate change and developing climate models. This interaction comprises two aspects. The variations in temperature and atmospheric constituents, particularly clouds, water vapor and aerosols directly change the radiative flow of energy through the atmosphere and across the earth's surface. The modified radiative flow of energy, in turn, impacts on the atmospheric structure, atmospheric constituent distributions and the optical properties, as well as on the underlying surface properties through cooling and heating the atmosphere and earth's surface. This study aims to contribute to a better understanding of the first aspect of the interaction using adjoint sensitivity analysis.

Outgoing radiation at the top of the atmosphere (TOA) is of fundamental importance for the radiative flow of energy, since it determines the energy budget of the earth's surface and atmosphere system. Its importance also derives from the fact that it can be observed with satellite-based instruments. The Earth Radiation Budget Experiment (ERBE) [Barkstrom, 1984; Barkstrom and Smith, 1986] attempted to measure the energy balance of the earth from space, with emphasis on understanding the role of clouds in the climate system. Based on the concept of cloud radiative forcing (CRF) [Cess and Potter, 1987], many a study contributed to the quantitative description of cloud impacts on the climate and climate change [Ramanathan et al., 1989; Harrison et al., 1990; Hartmann and Doelling, 1991; Kiehl, 1994; and others]. The clear sky atmospheric greenhouse effects can also be understood in a much more quantitative way, in particular the dependence on the water vapor variability associated with the deep convection over the tropical oceans (the supper greenhouse effect) has been estimated [Raval and Ramanathan, 1989; Ramanathan and Collins, 1991; Stephens and Greenwald, 1992; Kiehl and Briegleb, 1992; Inamdar and Ramanathan, 1994; Hallberg and Inamdar, 1993, and others]. Such studies have provided a great improvement to our knowledge of the energy balance of the climate system, and have also laid an observational basis for the understanding of climate warming related to the increase in greenhouse trace gases.

While observations reveal more and more facts about the TOA outgoing radiation, some studies have shifted their focus on a detailed explanation of observed phenomena [e.g. Inamdar and Ramanathan, 1994; Kiehl, 1994]. Which factors control spatial features and temporal variations of the outgoing radiation? It is fair to say that we still possess only a partial understanding of this problem, and systematical and specific studies have become a pressing requirement. In this study, we will concentrate on how spatial distributions of atmospheric constituents, such as water vapor and clouds, affect the outgoing radiation. The importance of spatial distributions of atmospheric constituents for radiation variations has been recognized in a few numerical and observational studies. Kiehl and Briegleb [1992] showed that the longwave (LW) radiation (terrestrial radiation) clear sky fluxes are quite sensitive to upper tropospheric moisture. Zhang et al. [1995] argued that for the net LW fluxes the relative importance of clouds and water vapor significantly changes with latitudes. Understanding of such effects is necessary for the analysis of climate change. For example, climate warming may cause a decrease in cloud amount [Cess et al., 1990]. The feedback of clouds may be considerably different when the decrease in cloud amount occurs at different spatial locations, if the radiation response to cloud cover has strong spatial dependence.

The basic atmospheric state is characterized by strong meridional (in the latitudinal direction) and vertical variations. Associated with the land-sea contrast, there are also significant zonal variations, and the permanent and semi-permanent troughs (low atmospheric pressure regions) and ridges (high atmospheric pressure regions) dominate the zonal variations [Palmén and Newton, 1967]. The radiative properties of the earth's surface also display strong regional variations. Such spatial heterogeneities can exert profound impacts on the radiation response to variations in the atmosphere and the surface, and lead to sizable spatial difference in the behavior of the radiation response. The spatial dependence of radiation response roots in the nature of the atmosphere and the surface, and it is difficult to understand variations in the radiative flow of energy without understanding such dependence.

Let R be the TOA outgoing shortwave (SW) radiation (solar radiation), long wave radiation or SW and LW radiation as a whole (net radiation), and x the value of an atmospheric state or constituent variable at a given spatial position, such as cloud amount, water vapor mixed ratios, and etc. We define the ratio $\triangle R/\triangle x$ as the sensitivity of R to x at a given model grid point, where $\triangle R$ is the radiation response to the perturbation $\triangle x$. We can simply obtain sensitivities by directly integrating a radiation model; this is the idea of the commonly used sensitivity analysis, called direct sensitivity analysis. To achieve our goal, we are required to calculate sensitivities for every model grid points and for various model variables, and then by comparison we can identify the sensitivity structure (the pattern showing the sensitivity's spatial variations) for a specified variable and estimate the relative sensitivity among the variables. Due to the dimensionality of our model, direct sensitivity analysis becomes computationally prohibitive for our purposes. However, with the help of the adjoint technique, we can obtain the gradient of R with respect to x, i.e. $lim_{\triangle x \to 0} \triangle R/\triangle x$, for all model grid points and all model variables by carrying out one calculation of the adjoint model corresponding to the original radiation model [see Cacuci, 1981, Zou et al., 1993]. As can be seen in section 3, the gradients well approximate the sensitivities defined by the ratio $\triangle R/\triangle x$ for a suitable range of sizes of $\triangle x$. Thus we can just use the gradients in the sensitivity analysis, and shall refer to the gradients as sensitivities in the following discussion.

The outline of this paper is as follows. In Section 2, the method of the adjoint sensitivity analysis is described, and the limitations of the method are discussed. The salient details of relevant parameterizations in the radiation model are given in Section 3, along with a discussion of the problems related to derivation of the linearized and adjoint models used in the adjoint sensitivity analysis, as well as validity of the linearization. Section 4 presents in detail the results of the adjoint sensitivity analysis. Finally, Section 5 summarizes the results of this study, along with a discussion describing the implications of the results to understanding the supper green house effect over tropic oceans as well as the influence of cloud feedback on the climate change.

2 Description of Adjoint Sensitivity Analysis

We can write a radiation model as

$$f = F(x), \tag{1}$$

where f consists of radiation fluxes and radiative heating rates, and quantities describing the atmospheric and earth's surface states constitute x.

Define a response function

$$R = H(f). \tag{2}$$

R may have various physical meanings depending on a suitable choice of the function H. In the following sections, we will choose R such that it represents outgoing shortwave radiation, outgoing longwave radiation, or outgoing shortwave radiation and outgoing longwave radiation as a whole.

The sensitivity of the response function to variations in the model variables is mathematically defined as [see Cacuci, 1981]

$$\delta R = H(f + \delta f) - H(f). \tag{3}$$

where δR represents a measurement of the sensitivity of the response R to the perturbation δf.

We now introduce a linearization of Eq. (2)

$$\delta f = L\delta x. \tag{4}$$

Substituting (4) into (3), we obtain

$$\delta R =< H'(f), L\delta x > +O(\delta x^2), \tag{5}$$

where $<,>$ denotes an Euclidean inner product. Following Eq. (5), we then have the gradient of R and the approximation of the defined sensitivity using the first-order Taylor expansion:

$$\nabla R_x = L^T H'(f); \tag{6}$$

$$\delta R \approx < L^T H'(f), \delta x > . \tag{7}$$

where L^T is the transpose of L. L^T is also called the adjoint of the linear operator L.

By definition, a change δx in the atmospheric and earth's surface states will lead to a change in the radiation response δR. From Eq. (7), we can immediately conclude that $\mid \delta R \mid$

attains its maximum for all δx with a prescribed norm when δx is parallel to ∇R_x. In other words, in regions where the gradient ∇R_x is large, a change in the atmospheric and earth's surface states would give rise to a large change in the radiation response; in regions where the gradient ∇R_x is small, a change in the atmospheric and earth's surface states would affect the radiation response very little. Consequently, the pattern of the gradient ∇R_x defines the desired sensitivity structure of perturbations of the atmospheric and earth's surface states. To conform to the discussion in the introduction, we refer to the components of ∇R_x as sensitivities.

Equation (6) gives the representation of the gradient ∇R_x, and it also provides a powerful method for evaluating the gradient ∇R_x. Equation (6) states that only one calculation of the adjoint model is required to obtain the gradients of R with respect to all the components of x. It is noteworthy that the evaluated gradients of R are accurate, and their accuracy is not affected by the linearization approximation though Eq. (4) is employed in the calculation.

The accuracy of the calculated gradient is not limited by the linearization, but the practical applicability of sensitivity analysis relies on the range of validity of the linearization. It is necessary to ensure that the approximation of Eq. (7) is valid for a suitable tolerance over a reasonable range of sizes of δx. For models representing complex atmospheric physical processes like radiation, the validity of (7) is quite questionable. Lack of validity may be due to strong nonlinearity or to approximate representations of relationships between model variables. Radiation transfer models for practical use in numerical weather prediction and climate models are generally developed based on parameterizations. Use of discontinuous functions is inevitable in such parameterizations, which may cause a lack of validity in Eq.(7). In the following, however, we shall show that Eq.(7) is valid over a reasonable range when the discontinuities in the radiation model are properly handled.

3 Radiation Model

3.1 Nonlinear Radiation Model

The radiation codes that are employed in the following calculations are taken from the NCEP medium-range weather forecasting model [Campana, 1990]. The effects of the major radiatively active atmospheric constituents are taken into account in the modeling of both shortwave (SW) and longwave (LW) processes, including water vapor (H_2O), carbon dioxide (CO_2), ozone (O_3), and clouds. The model outputs forecast model layer radiative heating rates, outgoing SW and LW fluxes, as well as downward LW and net SW fluxes at the earth's surface. The model has an horizontal resolution of 2.8125 degrees with 28 vertical layers. The salient details of relevant parameterizations in the radiation models are described below.

The cloud parameterization is a diagnostic scheme based on the work of Slingo [1987]. The main difference between the present scheme and Slingo's scheme is that marine stratus clouds are computed [Campana, 1990]. The model employs both stratiform and convection cloud-types. The model clouds are classified into three types with high clouds above 350 hPa, middle clouds between 642 and 350 hPa, and low clouds below 642 hPa. The parameterization outputs cloud top, bottom and amount for lower, middle, and high clouds respectively, and accommodates thick clouds (thicker than one model layer) or thin clouds (at one model layer).

The shortwave spectrum is divided into 9 frequency bands [Lacis and Hanson, 1974]; O_3 and H_2O overlap is assigned to the first band, and CO_2 and H_2O overlap is assigned to the other eight bands. The absorption of O_3 and H_2O is parameterized following the work of Lacis and Hanson [1974], while the absorption of CO_2 is calculated using the formula of Sasamori [1972]. When clouds are present, clear sky transmission is weighted by clouds effects, which are functions of cloud fraction, cloud albedo, and cloud (water) absorption. The bulk reflectivities are prescribed with 0.21, 0.48, and 0.69 for high, middle, and lower clouds respectively. A weak cloud absorption is taken into account by setting absorptivities 0.005, 0.02, 0.035 for high, middle, and lower clouds for the last eight bands, whereas no cloud absorption occurs for the first band (O_3 band).

A mean cosine of the solar zenith angle, $\overline{\mu}$, is employed to calculate the SW fluxes at each model "grid point", and it is computed for each model latitude as in the following equation [Manabe and Strickler, 1964]

$$\overline{\mu} = \frac{\int_{daylight} \mu(t)dt}{\int_{daylight} dt}, \tag{8}$$

where $\mu(t)$ is the cosine of the solar zenith angle at time t. A mean cosine of the solar zenith angle is only a function of latitude.

The long wave radiation model [Fels and Schwarzkopf, 1975; Schwarzkopf and Fels, 1991] operates on seven frequency bands. This model computes the absorptivity of the dominant absorber in each region and then calculates the other through a series of highly parameterized approximate techniques. It includes an extended vibration-rotation water vapor continuum along with the other absorption lines. The code has p-type water vapor absorption with an extensive e-type water vapor continuum. Cloud LW radiation emissivity is included as input, but if it is not specified, it is prescribed inside the model. For lower and model clouds the default emissivity is 1. For high clouds, its value is set to 0.6 in the tropical regions and to 0.3 in polar regions.

3.2 Linearization and Adjoint

The linearized and adjoint codes of the radiation model have been derived. We select five variables to be perturbed in the linearized code, including atmospheric moisture q, temperature T, pressure p, cloud fraction C_i ($i = 1, 2, 3$ for high, model and lower clouds), and surface temperature T_s. These fives variables are model forecast ones, and strongly affect variations of outgoing radiation. The amounts of O_3 and CO_2 are not perturbed since these two variables are given climatological values with seasonal variations. Extension to these two variables, or even to other parameters, is straightforward. The method described by Navon et al. [1992], Zou, [1997], is employed in the derivation of the linearized and adjoint codes. This method is characterized by the fact that linearized and adjoint codes are derived directly from the original nonlinear codes, thus ensuring that the linearized and adjoint codes are completely consistent with the original nonlinear codes.

A concern here is the range of validity of the linearization. We know that radiation processes display strong nonlinearity. One may question whether linearized codes could be useful in practical applications. On the other hand, recent studies present evidence showing that linearizations of radiation models do work well. Chou and Neelin [1996] indicated that the linear scheme well reproduced the nonlinear results in simulation of longwave radiative fluxes associated with interannual variability, seasonal variations in intermediate tropical atmospheric models. Diagnostic results of radiative fluxes in several sophisticated radiation models by Zhang et al. [1994] also showed that nonlinear solutions display some linear properties. We shall numerically verify that the linearized and adjoint models are valid for a considerable range of sizes of perturbations.

We validated the linearized and adjoint calculations of the radiation model with the so-called gradient test. In the gradient test, a test value, t, is computed as the ratio between a perturbation of the response function, R, and its first-order Taylor expansion, i.e.

$$t = \frac{R(x+\delta x) - R(x)}{< \nabla R(x), \delta x >}. \tag{9}$$

$\nabla R(x)$ is evaluated using (6). The adjoint is fully consistent with the original nonlinear model if the value of t approaches 1 when the perturbation δx goes to zero. Comparing (9) with (7), we can find that t is a quantity which can be used to measure the accuracy of the linearization approximation.

Corresponding to sensitivity analyses in the following sections, we define two response functions:

$$R_s(x) = \sum_k (F^{\uparrow}_{stoa})_k, \tag{10}$$

$$R_l(x) = \sum_k (F^{\uparrow}_{ltoa})_k, \tag{11}$$

where k is defined on all horizontal model grids. $F^{\uparrow}_{stoa}$ denotes outgoing shortwave fluxes at the top of the atmosphere, and $F^{\uparrow}_{ltoa}$ outgoing longwave fluxes.

In all the following calculations, the NCEP reanalyses are used. Table 1 presents an example of the calculated results. In these calculations, the basic state consists of the reanalyses at 00GMT, 29 June 1989. The perturbations were produced by removing part of the largest spatial scale components for wind components, temperature and pressure. A scaling constant, α, serves to control the size of perturbations. Perturbed moisture and cloud fractions are obtained by simply multiplying the corresponding basic states with 3α, where 3α, rather than α, is used to account for large variabilities of moisture and cloud cover. This calculation verifies that the adjoint is fully consistent with the original nonlinear model as explained in Appendix A.

We can also conclude from Table 1 that the linearized and adjoint models are valid for a considerable range of perturbation sizes. For $\alpha = 0.1$, the largest temperature is around 4 K, wind speed $5m/s$, pressure $10hPa$, moisture mixed ratio 4 g/kg, and cloud fraction 0.3. In this case, the linearization approximation of (7) holds with a high accuracy for shortwave radiation, while the accuracy becomes relatively low for longwave radiation. We have performed comparisons between the linearized perturbation solution and the nonlinear one, and found that the linear solution reproduced well the spatial structure, and that the correlation between the two solutions was as high as 0.89. This suggests that the linearization is valid even for values of α as large as 0.1.

These results are very robust. A number of similar calculations to the above have been carried out, but for different basic states representative of different months and seasons. Consistent results were obtained. We also used other definitions of the response functions to carry out the gradient test, and again obtained very similar results.

It is worth noting that the above conclusions are true only when the discontinuities (in both functions and the first order derivatives) in the original radiation models are properly handled. We analyzed the impact of discontinuities on the validity of the linearization routine by routine, and found that the discontinuities associated with table look-ups and cloud calculations may cause a lack of validity of the linearization; we then removed these discontinuities by replacing table look-ups and functions with discontinuities with cubic spline interpolations and sine and cosine function expansions.

4 Sensitivity Analysis

We shall present the sensitivity analysis results for 00GMT, June 29 and December 29, 1989, these results being representative of the northern summer and winter cases respectively. The following calculations are carried out for both clear sky and cloudy sky respectively. In cloud radiative forcing calculations, clear sky and cloudy sky are theoretically defined, but these definitions differ in practical applications [Cess and Potter, 1987; Ramanathan et al., 1989; Cess et al., 1990]. This difference leads to different computational schemes for cloud radiative forcing, and may cause a difference in the quantitative results [e. g. Harshvardhan et al, 1989; Zhang, 1994]. We adopt the same definition of clear sky as in so-called "method II" [Cess and Potter, 1987] of cloud radiative forcing computations; i. e. the clear sky atmosphere retains the states unchanged, but the clouds are set zero.

The radiation response functions for outgoing shortwave radiation (OLR) and outgoing longwave radiation (OSR) defined in Eqs. (10) and (11) respectively,are employed. Here we define another response function:

$$R_{sl} = R_s + R_l. \tag{12}$$

In R_{sl}, shortwave and longwave radiation are considered as a whole, i.e. the net radiation. As has been seen in Section 2, the gradients of R_s, R_l and R_{sl} define sensitivity structures of perturbations of the atmospheric and earth's surface states. In the present study, sensitivity structures of atmospheric moisture, temperature, clouds and the surface temperature will be depicted. Sensitivity structures of atmospheric pressure will not be discussed, since our analyses showed that the outgoing radiation was not sensitive to perturbed atmospheric pressure.

4.1 Clear Sky Case

4.1.1 Shortwave Radiation Sensitivity

Shortwave radiation can be absorbed by water vapor, and is independent of atmospheric temperature and the surface temperature [Lacis and Hansen, 1974]. An increase in water vapor will cause a reduction in OSR. Figure 1 presents the zonally averaged sensitivity structure of moisture. A remarkable feature is that OSR is much more sensitive to water vapor mixing ratios at high latitudes in the summertime hemisphere than at other latitudes. It is also worth noting that the most insensitive region is located at the tropics.

▷ [Figure 1 near here]

Such latitude dependent features are obviously linked to the surface albedo. Water vapor of an unit amount absorbs more shortwave radiation at the tropics than at other latitudes due to stronger incident solar radiation. The water vapor reduces more incident shortwave radiation to the surface and enhances the atmospheric heating rates. Since the tropical surface is characterized by oceans which have the largest darkness (near black bodies), the absorption by water vapor has little effect on OSR. The incident radiation to the atmosphere is almost fully absorbed by the atmosphere and surface system in the tropics irrespective of the absorption by water vapor, and thus the absorption by water vapor hardly causes a decrease in OSR. In contrast, the surface at the high latitudes is characterized by snow or ice cover which has the largest brightness, and the surface absorbs only a small part of the incident solar radiation. In this case, the absorption by water vapor is nearly equal to the decrease in OSR. In addition, the absorption of water vapor is further increased by relatively strong reflected solar radiation. The sensitivity structure at 600 hPa (Fig. 2), which corresponds to the maximum absorption areas, more clearly illustrates the predominant linkage to the surface albedo. This linkage is not obscured by spatial variations in the atmospheric structure.

▷ [Figure 2 near here]

We observed that the sensitivity structures are rather symmetric about the equator for the northern summer and the southern summer, though the sensitivity at the southern high latitudes in the southern summer is stronger than that at the northern high latitudes in the northern summer and the maximum sensitivity is slightly higher.

4.1.2 Longwave Radiation Sensitivity

Longwave radiation is strongly affected by water vapor, as well as by the atmospheric and earth's surface temperature. The emission and absorption of gases is complexly dependent on the vertical structure of their distributions and on the vertical structure of the atmospheric temperature. Such dependence exerts a significant effect on sensitivity structures.

Figure 3 illustrates the zonally averaged sensitivity structure of the moisture for OLR. A striking feature is the larger sensitivity found near the tropopause, with a maximum of 24 $Wm^{-2}g^{-1}kg$. The middle and lower troposphere have a magnitude of less than 2.5 Wm^{-2} $g^{-1}kg$. On the whole, OLR is one order of magnitude more sensitive to specific humidity in the upper troposphere than in the middle and lower troposphere. The high sensitivity of OLR to the water vapor in the upper troposphere has been noted in observational studies. Kiehl and Briegleb [1992] clearly indicated that the differences in upper tropospheric moisture were the major cause for the OLR difference during a two year observation period in most regions over the tropical oceans. This result is of particular importance in understanding the supper green house effect over tropical oceans, which will be discussed in the following section.

▷ [Figure 3 near here]

Some meridional features in Fig. 3 are also of interest. In the middle and lower troposphere, sensitivities have a small positive value at the high latitudes, while the positive value is a little larger in the wintertime hemisphere than in the summertime hemisphere. Near the tropopause, sensitivities are relatively large between the subtropics in the summertime hemisphere and the middle latitude in the wintertime hemisphere. It is interesting to note the existence of two weak sensitivity regions at the latitudes of around 30° in the summertime hemisphere and 40° in the wintertime hemisphere. These two regions are obviously linked to activities of polar fronts [Palmén and Newton, 1967].

Qualitatively, the above results for longwave radiation are easily understood from the radiation transfer equation. We may write the radiation transfer equation in a perturbed form as:

$$F'^{\uparrow}_{ltoa} = -\int_0^\infty \int_0^\infty \tau'_\nu(z,h) \frac{dB_\nu(h)}{dh} dh d\nu, \tag{13}$$

where ν is wavenumber, B_ν the Planck function, $F'^{\uparrow}_{ltoa}$ the perturbed OLR flux associated with water vapor perturbations, and $\tau'_\nu(z, h)$ the perturbed transmission function between level h and z. B_ν is a function of temperature T, and thus $dB_\nu(h)/dh$ depends on temperature lapse rates. A large temperature lapse rate leads to a large magnitude of perturbed OLR. The inversion enhances OLR, which is the leading cause to positive values of sensitivities at high latitudes.

The sensitivity structure of atmospheric temperature displays a strong spatial variation (Fig. 4). In the vertical, the largest sensitivities are located between 600 and 500 hPa, and the globally averaged sensitivity at 600 hPa is 0.2 $Wm^{-2}K^{-1}$. The sensitivities rapidly decrease above and below the largest sensitivity regions. The sensitivity between 600 and 500 hPa is one order of magnitude larger than that near the tropopause, and three times stronger than that near the surface. As far as the latitudinal variation is concerned, the sensitivities are largest over tropical areas, and decrease toward the poles. Unlike water vapor, the sensitivity structure of the atmospheric temperature does not present a significant difference between the northern summer and southern summer. It is worth noting that a 1 K temperature perturbation in the middle troposphere produces an OLR perturbation which is equivalent to the one produced by a 0.1 g/kg water vapor perturbation in the middle and lower troposphere or by a perturbation of about 0.15 K in the surface temperatures (see the following discussion). Thus, temperature perturbations in the middle troposphere have a significant effect on OLR variabilities at the top of the atmosphere.

$\triangleright$ [Figure 4 near here]

Figure 5 shows the sensitivity structures of the surface temperature. A positive surface temperature perturbation results in an increase in the emission of longwave radiation, and thus in an increase in OLR when feedbacks are not considered (an increase in sea surface temperatures may lead to a decrease in OLR when the feedback of water vapor is involved [Inamdar and Ramanathan, 1994]. The globally averaged sensitivity is 1.4 $Wm^{-2}K^{-1}$. As expected, the tropical areas have the smallest sensitivities due to the strong longwave greenhouse trapping caused by abundant water vapor. On the average, the tropical sensitivities are about half as large as those in the middle and high latitudes. On the whole, the sensitivities decrease from the tropics to the pole regions. However, the magnitude of the gradients of sensitivities with respect to latitudes is not uniform, and a particularly large gradient is observed in the subtropical regions. Such a meridional variation is especially obvious in the sensitivity structure for the Northern Hemisphere on 29 December 1989. The regions of weak sensitivity over the tropical oceans are well consistent with regions of large greenhouse trapping observed in the Earth Radiation Budget Experiment (ERBE) [e.g., Raval and Ramanathan, 1989; Kiehl and Briegleb, 1992]. There are some other features of interest. The regions of high orography are characterized by large sensitivities, for instance the Tibetan Plateau and the Rocky mountains. North Africa and part of Australia are relatively sensitive regions, which is a consequence of their drier atmosphere. The northeastern regions of the Pacific and Atlantic oceans are relatively insensitive compared with other regions located at the same latitude in the northern winter, which are associated with the upward motion in the front of the troughs over the middle oceans [Palmén and Newton, 1967].

▷ [Figure 5 near here]

The sensitivity structure of the surface temperature displays to a large extent a great geographic variation in its greenhouse effects. To more accurately represent the greenhouse effects, we analyzed the sensitivity structure of the surface longwave emission. The sensitivity of the surface longwave emission is not a direct product of the adjoint model, but it can be calculated by the following equation:

$$\nabla_{es} R_l = \frac{\nabla_{Ts} R_l}{4\sigma T_s^3}, \tag{14}$$

where $\nabla_{Ts} R_l$ is the gradient of R_l with respect to the surface temperatures Ts, and σ the Stefan-Boltzmann constant. The magnitude of $\nabla_{es} R_l$ represents the factor of the perturbed OLR, which varies between 0 and 1. An increased perturbation emission at the surface is completely trapped by the column atmosphere when the gradient is 0, and escapes completely out of the top of the atmosphere when the gradient is 1. Figure 6 illustrates the structure of $\nabla_{es} R_l$. We can see that less than 20% of an increased perturbation emission at the surface goes out of the top of the atmosphere over the tropical areas, while only 10% goes out in some of the tropical areas such as over Southern Asia. In other words, the column tropical atmosphere traps more than 80% of an perturbation increase in the emission at the surface. The magnitude of $\nabla_{es} R_l$ at mid-latitude is in general larger than 30%, while at high latitudes it is larger than 40%. It has long been found and investigated in depth that the moistening effect of deep convection activity strongly enhances the atmospheric greenhouse effect [Raval and Ramanathan, 1989; Stephen and Greenwald, 1991; Kiehl and Briegleb, 1992, Hallberg and Inamdar, 1993; Inamdar and Ramanathan, 1994]. In our calculations, the regions with a greenhouse trapping larger 80% correlate well with deep convection activities. This is also the case for the region within the inner Eurasian continent on 29 June 1989. It is of interest to note that the deep upward motion in the front of the semi-permanent troughs over the middle Pacific and Atlantic oceans in the wintertime period displays effects similar to deep convective activities.

▷ [Figure 6 near here]

4.2 Cloudy Sky Case

The interaction between radiation and clouds has long presented a fundamental challenge. In atmospheric general circulations models, the representation of this interaction has undergone a considerable improvement during recent years, but substantial deficiencies are still present as indicated by model comparisons [Cess et al, 1996]. The interaction comprises two processes: clouds affect radiation by absorption and reflection; radiation affects clouds through changing the structure of the atmosphere by heating and cooling. The first aspect is generally thought to be more easily depicted in the model. Unfortunately, its depiction is by far unsatisfactory [Baer et al., 1996; Ellingson et al., 1991]. The following analysis intends to show the effects of variations in cloud amount on the outgoing radiation at the top of the atmosphere.

The clouds used as basic states are obtained by the diagnostic scheme described in Section 2. Figure 7 provides three zonal cross-sections of cloud distributions in the Northern Hemisphere on 29 December 1989. Associated with the deep convection near the equator, there are three main thick cloud regions: South Asia and the western Pacific, the central

Pacific and the western Atlantic. The top of most thick clouds reaches up to 200 hPa, and some of them attain up to 150 hPa. The convective activities become weaker northward, and correspondingly cloud regions shrink and the top of thick clouds is lower. There also exist thin clouds (present at a model layer). The amount of thick clouds is primarily less than 0.25. It seems that the patterns of clouds are quite reasonable, whereas the amount of clouds tends to be underestimated. Due to the above limitation in cloud estimates, we shall concentrate in the following discussion on the qualitative aspects.

▷ [Figure 7 near here]

4.2.1 Shortwave Radiation Sensitivity

Figure 8 presents the zonally averaged sensitivity structure of thick clouds for shortwave radiation. Since the cloud amount is perturbed when the cloud amount in the basic state is larger than zero, the perturbed cloud amount is thus zero without the presence of clouds in the basic state and as a consequence the sensitivity is assigned a value of zero. We should keep this in mind in the following analyses.

The cloud reflection of shortwave radiation becomes stronger with the increase in cloud amount, which is illustrated by the positive sensitivity. The sensitivity structure is roughly characterized by a direct proportion to the incident solar radiation, but some important features can not be accounted for by the difference in the incident solar radiation. The lower clouds are the most sensitive: this result may primarily be a consequence of assigning a larger reflectivity to the lower clouds. While we have no observations to serve as comparison, it remains a concern whether the result obtained is reasonable. The high latitudes in both the wintertime and summertime hemispheres are super insensitive regions. The high albedo of the surface in these regions is responsible for this super insensitivity, since an increase in cloud amount hardly causes an increase in OSR even for strong incident solar radiation when the surface's albedo is larger than that of clouds. The three types of clouds present somewhat different meridional variations. In order to illustrate the last point, we provide the latitudinal variations of sensitivity of the three types of clouds (same as defined in the model) (Fig. 9). In the wintertime hemispheres, the sensitivity of the three types of clouds consistently increases from zero at high latitudes to about 20 Wm^{-2} for a 0.1 increase in the cloud amounts at the equator, and the lower cloud has slightly larger sensitivity. While the high clouds are most sensitive at the latitude consistent with the location of the largest incident solar radiation, the lower clouds tend to have the largest sensitivity at mid-latitudes in the summertime hemisphere. Correspondingly, the difference between the high and lower clouds reaches its maximum in the summer middle latitudes, and the lower cloud sensitivity is in the average about 30 Wm^{-2} for a 0.1 increase in the cloud amounts in these latitudes. This sensitivity is twice as large as that of the high clouds which is about 15 Wm^{-2}. We have also calculated the sensitivity variation of the three types of the clouds with latitude for the thin clouds (Fig. 10). Compared with the thick clouds, the sensitivity of the lower clouds is more consistent with the strength of the incident solar radiation, and the high clouds exhibit a significant weaker sensitivity while the other two types of clouds show only a slight difference.

▷ [Figure 8 near here]

▷ [Figure 9 near here]

▷ [Figure 10 near here]

4.2.2 Longwave Radiation Sensitivity

Figure 11 shows the zonally averaged sensitivity structure of thick clouds for longwave radiation. The effects of clouds on OLR are similar to the effects of water vapor, and are intimately related to temperature lapse rates. The sensitivity structure between the thick clouds and water vapor (Fig. 2) shares some common features: the largest sensitivity is located near the tropopause and at the tropics with a shift of about 10° to the summertime hemispheres. Associated with the inversion at the polar latitude, the sensitivity has small positive values. The main difference is that the rates of vertical decrease in the sensitivities are largest between 900 and 700 hPa for the thick clouds while in the upper troposphere they are largest for the water vapor. Such a feature indicates that the OLR is more sensitive to the high and middle clouds than to lower clouds. To further illustrate this feature, we also depict the sensitivity variation of the three types of the clouds with latitudes (Fig.

12). The sensitivity of the high and lower clouds displays a consistent meridional variation, whereas the lower cloud presents a significantly smaller sensitivity apart from the high latitudes. At the tropics, the sensitivities of the middle and high clouds are on the average about 12 Wm^{-2}, and which is twice as large as that of the lower cloud. At high latitudes, the sensitivity of all the three type of clouds becomes smaller, and have even small positive values at polar latitudes. Such a feature can be seen in the observations by Yamanouchi and Charlock [1995]. For the thin clouds (Fig.13), the three types of clouds manifest a decrease in the sensitivity compared with that of the thick clouds. The high clouds have a similar tendency for meridional variations in the thick clouds with a relatively small decrease in sensitivity. The middle clouds present smaller meridional variations, and correspondingly the sensitivities at the lower and middle latitudes present a significant decrease to an average of about 2 Wm^{-2} for a 0.1 increase in the cloud amounts, i.e. a reduction by a factor of 5 compared with that of the corresponding thick cloud. The sensitivity of the lower cloud is rather weak, and has a value smaller than 1 Wm^{-2} for a 0.1 increase in the cloud amounts at all latitudes.

▷ [Figure 11 near here]

▷ [Figure 12 near here]

▷ [Figure 13 near here]

4.2.3 Net Radiation Sensitivity

The sensitivity of clouds to the total radiation for both longwave and shortwave radiation is related to the net forcing of clouds on the radiation responses. Figure 14 illustrates the zonally averaged sensitivity structure of the thick clouds. There are two striking features: A cancellation occurring between the longwave greenhouse effect and the shortwave reflection for middle and high clouds at lower latitudes and a strong positive sensitivity of the lower cloud at the summer mid-latitudes. These two features become evident when we carry out a comparison between the sensitivity structures of shortwave and longwave radiation in Figs.8 and 11. The middle and high clouds at lower latitudes strongly trap the longwave radiation, and simultaneously strongly reflect the shortwave radiation. Thus a large cancellation occurs in the middle and high cloud areas at lower latitudes. The lower clouds cause a small longwave trapping, but a larger shortwave reflection compared with the middle and high clouds. Such an opposite tendency creates a larger positive sensitivity of the lower clouds. We wish to emphasize the importance of the lower clouds at the summer middle latitude, i.e. the lower clouds display the strongest sensitivities for the shortwave radiation and a weak sensitivity for the longwave radiation, and thus become especially sensitive for the total radiation compared with that at other latitudes. This latitudinal dependence appears to be consistent with observed results that the net cooling effect on the earth is larger over the middle and high latitude oceans [e.g. Ramanathan et al., 1989]. The importance of the fact is clearly seen in the sensitivity variations of the three types of the clouds with the latitude (Fig. 15). It is thus expected that the lower clouds at mid-latitudes play an important role in the earth's radiation budget.

▷ [Figure 14 near here]

▷ [Figure 15 near here]

At the tropics, observations [e. g. Ramanathan et al., 1989; Kiehl and Ramanathan, 1990] from ERBE showed that the longwave cloud forcing and the shortwave cloud forcing are nearly canceling each other in tropical convective regions. Our calculations (Fig. 14) indicate that even near the top of clouds at about 200 hPa the shortwave cloud forcing is larger than the longwave cloud forcing in magnitude, while the lower cloud produces a much stronger shortwave forcing than the longwave forcing. These calculated results are not consistent with the observations. The radiation model has been found to overestimate the cloud shortwave reflection [Campana, 1997, Personal Communication]. The inconsistency may be partly attributed to this overestimation. There are two other possible factors that may contribute to this inconsistency: first, the observations may overestimate the longwave cloud forcing [Inamdar and Ramanathan, 1994; Kiehl and Briegleb, 1992], and second the model cloud heights are not high enough, as showed by Kiehl [1994] who found that a cloud top height of about 16 Km is required for near cancellation.

5 Summary and Discussion

In this study we have used the adjoint technique to calculate the gradients of the outgoing radiation at the top of atmosphere with respect to the amount of the atmospheric constituents as well as the atmosphere and the earth's surface state variables. The gradient fields define the sensitivity structures, which present the spatial patterns of the sensitivity variations. The magnitudes of the gradients may also be used to estimate the relative sensitivity amongst various variables. The calculations revealed significant spatial variations in the sensitivities, and in particular the spatial variation patterns is of importance for understanding the variabilities of the outgoing radiation.

For the outgoing shortwave radiation (OSR), the most sensitive regions of water vapor mixing ratios are located at the summer high latitudes, while the tropics is an insensitive region. Such a meridional variation in the sensitivities is primarily determined by the albedo of the the surface. OSR is found to be more sensitive to the lower cloud than to the middle and high clouds, i.e. the lower clouds tend to be more sensitive at mid-latitudes. The albedo of the the surface and the variation in the incident solar radiation are mainly responsible for this meridional variation in sensitivities.

For the outgoing longwave radiation (OLR), the water vapor mixing ratios are found to be most sensitive near the tropopause, roughly about 30 times more sensitive than those in the lower and middle troposphere. The fastest vertically increasing band of sensitivities was found to be in the upper troposphere. The OLR is also sensitive to temperature variabilities in the middle troposphere as illustrated by the fact that the OLR variation produced by a 1 K temperature perturbation is equivalent to that produced by a 0.1 g/kg water vapor perturbation. OLR displays similar sensitivities to the middle and high clouds, but much weaker sensitivities to the lower clouds. The sensitivities of the middle and high clouds are on the average twice as large as those of the lower clouds at lower latitudes, while the sensitivity of the three types of clouds decreases rapidly at mid-latitudes.

The clear sky atmosphere displays the strongest longwave strong greenhouse trapping at the tropics, and more than 80% of a perturbation increase in the earth's surface emission is trapped. A rapidly decreasing band of the trapping was found within the subtropics, and about 60-70% of a perturbation increase in the earth's surface emission is trapped at middle and high latitudes. The rapid rise in the longwave trapping at the tropics compared with that at middle and high latitudes supports the suggestion by Hallberg and Inamdar [1993], Inamdar and Ramanathan [1994], Kiehl and Briegleb [1992], and others, that deep convection moistening effects strongly enhance greenhouse effects. Such deep convection effects were also observed at the summer northern mid-latitudes when convective activities prevail. The deep upward motion in the front of the semi-permanent troughs over the central Pacific and Atlantic oceans in the wintertime has a moistening effect similar to the deep convective activities, and enhances the longwave greenhouse trapping in these areas.

An increase in cloud amount leads to an increase in the net outgoing radiation in all areas with clouds, apart from the lower and middle troposphere at polar latitudes where an inversion exists. The net outgoing radiation is found to be more sensitive to the lower clouds. This is due to a stronger sensitivity for the shortwave reflection and a weak sensitivity for the longwave trapping, whereby the summer mid-latitude lower clouds become especially sensitive compared with middle and high clouds and lower clouds at other latitudes.

Over the tropical oceans, Raval and Ramanathan [1989] found that the increase in the atmospheric greenhouse trapping can exceed the increase in the longwave surface emissions associated with higher sea surface temperatures (SSTs) when the SST is higher than a certain critical temperature of about 298 K. Such a super greenhouse effect has been a concern for many climatologists, since this phenomenon evidences a strong feedback between surface temperature, water vapor and the greenhouse effect. Observational studies have tended to the following consistent conclusion [Stephens and Greenwald, 1991; Kiehl and Briegleb, 1992; Hallberg and Inamdar, 1993; Inamdar and Ramanathan, 1994]: in warm oceanic regions, the climatological effects of deep convection are to moisten the entire troposphere and enhance the greenhouse effect of the water vapor. The enhancement in the greenhouse effect is sufficiently strong so that both the surface and the surface-troposphere system lose their ability to radiate the excess energy to space. Inamdar and Ramanathan [1994] suggested that the increase in the temperature lapse rate within the lower troposphere can also be a contributing factor. However, there seems to exist a significant difference in the quantitative results. For example, Kiehl and Briegleb [1992] emphasized the importance of the increase in upper tropospheric moisture, while Inamdar and Ramanathan [1994] showed that the lower troposphere is more important. Our results are consistent with those suggested by Kiehl and Briegleb [1992]. OLR is about 10 to 30 times more sensitive to the water vapor mixing ratios in the upper troposphere and near the tropopause than in the lower troposphere, and thus an increase in water vapor in the upper troposphere will cause a significant increase in OLR trapping. The temperature sensitivities (Fig. 4) do not seem to support the importance of the increase in the temperature lapse rate within the lower troposphere. The deep convection can pump moisture up to the tropopause, and the related increase in water vapor in the upper troposphere and near the tropopause may be the most important factor for the occurrence of the super greenhouse effect.

Our results have also implications for the understanding of climate change and cloud feedback. Climate models seems to predict the consistent tendency of a decrease in cloud amount and an increase in cloud heights when the climate warms [e.g. Manabe and Wetherald, 1988; Cess et al., 1990]. Although we recognize that climate models prediction may not always match the observed reality for global warming, it is still of interest to link some effects of the cloud feedback based on the cloud variabilities with a climate warming produced by climate models. Our results show that clouds have positive sensitivities, i.e. an increase in cloud amount causes an increase in the net outgoing radiation, except at polar latitudes, and clouds with lower heights have larger sensitivities. These results indicate that both a decrease in cloud amount and an increase in cloud heights lead to a decrease in the net outgoing radiation, and then to a positive feedback to climate warming in climate models. Other results of this sensitivity analysis suggest that the summer mid-latitudes lower clouds have the largest sensitivity, leading to the conclusion that the strongest feedback to climate warming is expected to occur at these latitudes.

We emphasize that our results depend on the model used, and we do not intend to use the results directly to explain the behavior of the real atmosphere. Our primary purpose is to emphasize the importance of the concept of sensitivity and especially the spatial patterns of sensitivities for an in-depth analysis of interactions between radiation and other atmospheric processes. In order for our conclusions to reliably explain the real atmospheric processes, our results must be compared with those obtained from other models as well as with those derived from observations such as data from the Atmospheric Radiation Measurement (ARM) Program [Stokes and Schwartz, 1994]. Such comparisons will be carried out in a follow-up paper and may be very helpful to identify model deficiencies.

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A Appendix

For a function $y(z)$ with continuous second derivatives, the Taylor expansion around z yields

$$y(z+\delta z) = y(z) + y'(z)\delta z + \frac{1}{2}y''(\xi)\delta z^2, \tag{A.1}$$

where $\xi \in (z, z+\delta z)$.

Equation (A.1) may be written as

$$\frac{y(z+\delta z) - y(z)}{y'(z)\delta z} = 1 + \epsilon \tag{A.2}$$

where

$$\epsilon = \frac{y''(\xi)\delta z^2}{2y'(z)\delta z} \tag{A.3}$$

For a double precision computer machine (15-digit precision), ϵ is determined by (A.3) when $\mid y''(\xi)\delta z^2/y(z) \mid$ is smaller than 10^{-15}. In such a case, ϵ is of the same order of magnitude as δz when $y''(\xi)$ and $y'(z)$ are of the same order of magnitude. Otherwise, ϵ is determined by $10^{-15}/y'(z)\delta z$. From the above discussions one draws the conclusion that ϵ attains its the smallest value with an order of magnitude between 10^{-7} and 10^{-8}. From this conclusion and Table 1, we obtain that the linearized and adjoint models are fully consistent with the original nonlinear model.

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Table Captions

Table 1. Results of the Gradient Test

Table 1. Results of the Gradient Test

α	test (t) shortwave	test(t) longwave
10^{-1}	0.9981116206659	0.5373190715952
10^{-2}	1.0001212323730	0.9875666899427
10^{-3}	1.0000835059890	0.9963953949818
10^{-7}	1.0000000340750	0.9999994083835
10^{-8}	0.9999946586258	0.9999999912610
10^{-9}	0.9999968097493	0.9999953800715

Figure Captions

Figure 1. The zonally averaged sensitivities of outgoing shortwave radiation at the top of the atmosphere to water vapor mixing ratios ($g \cdot kg^{-1}$) for 29 June 1989 (a) and 29 December 1989 (b). The contour interval is 0.2 $Wm^{-2}g^{-1}kg$. Dashed lines denote negative values, and indicate that an increase in water vapor mixing ratios leads to a decrease in outgoing shortwave radiation. Note that panel (a) is for 29 June 1989, panel (b) is for 29 December 1989, and dashed lines denote negative values in all of the following figures.

Figure 2. Sensitivities of outgoing shortwave radiation at the top of the atmosphere to water vapor mixing ratios at 600 hPa. The contour interval is 0.25 $Wm^{-2}g^{-1}kg$. Other descriptions are same as in Figure 1.

Figure 3. The zonally averaged sensitivities of outgoing longwave radiation at the top of the atmosphere to water vapor mixing ratios. The contour interval is 1.0 $Wm^{-2}g^{-1}kg$.

Figure 4. Same as in Figure 3 but for atmospheric temperature. The contour interval is 0.025 $Wm^{-2}K^{-1}$.

Figure 5. Sensitivities of outgoing longwave radiation to the earth's surface temperature. The contour interval is 0.25 $Wm^{-2}K^{-1}$. Stippled regions denote sensitivities smaller than 1.25 $Wm^{-2}K^{-1}$.

Figure 6. Sensitivities of outgoing longwave radiation to the earth's surface perturbation emission. These sensitivities are the ratios between an increase in the earth's surface emission and the corresponding increase in outgoing longwave radiation. The contour interval is 0.1. Stippled regions denote ratios larger than 0.2, or indicate that atmospheric columns trap more than 80% of a perturbation increase in the earth's surface emission.

Figure 7. Longitude-height cross sections of the clouds for the basic state of 29 December 1989. (a) is for 35° N, (b) for 60° N, and (c) for 1.5° N respectively. Heavy lines denote that the cloud amount is larger than 0.25.

Figure 8. The zonally averaged sensitivities of outgoing shortwave radiation at the top of the atmosphere to clouds. The contour interval is 50 Wm^{-2}, which is equivalent to an increase of 5 Wm^{-2} in the outgoing shortwave radiation due to an increase of 0.1 in the cloud amount.

Regions without presence of clouds in the basic states are stippled.

Figure 9. The zonally averaged sensitivities of outgoing shortwave radiation to three types of thick clouds. Solid curves are for high clouds, dashed curves for middle clouds and dash-dotted curves for lower clouds.

Figure 10. Same as in Figure 9 but for thin clouds.

Figure 11. Same as in Figure 8 but for outgoing longwave radiation and with a contour interval of 25 Wm^{-2}.

Figure 12. Same as in Figure 9 but for outgoing longwave radiation.

Figure 13. Same as in Figure 10 but for outgoing longwave radiation.

Figure 14. The zonally averaged sensitivities of outgoing shortwave and longwave radiation as a whole (net radiation) to clouds. Other descriptions are the same as in Figure 8.

Figure 15. Same as in Figure 9 but for net radiation.