The Effect of Changes in Cloud Amount on the Net Radiation at the Top of the Atmosphere

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(Manuscript received 19 June 1979, in final form 12 September 1979)

ABSTRACT

Due to the opposing albedo and greenhouse effects of clouds, the possibility exists that the net radiation at the top of the earth-atmosphere system is, in the mean, insensitive to changes in cloud amount. If so, this would have important implications for climate studies. This question is examined with the use of data on the components of the radiation budget at the top of the atmosphere obtained from the processing of 45 months of scanning radiometer observations of the NOAA satellites. Year-to-year changes in monthly mean values of outgoing longwave radiation and albedo are analyzed at a sample of geographic and climatic areas of the earth. By using the albedo changes as a measure of changes in cloud amount, it is possible to determined the sensitivity of the outgoing longwave radiation and the net radiation to changes in cloud amount. For each geographic/climatic area, the results indicate that the net radiation at the top of the atmosphere *is* sensitive to cloud amount changes and the sensitivity is such that the albedo effect of the clouds predominates over their greenhouse effect. Thus, for the earth as a whole the net radiation at the top of the atmosphere *is* sensitive to changes in cloud amount. Estimates of the numerical values of the global mean sensitivity of net radiation and outgoing longwave radiation to changes in cloud amount are presented and compared with previous findings.

1. Introduction

Clouds have two important effects on the radiation budget of the earth–atmosphere system. As a result of their scattering properties at solar radiation wavelengths, they act to increase the albedo of the system, and an increased albedo means a reduction in the amount of absorbed solar radiation. As a result of their absorption properties at terrestrial radiation wavelengths, they act to decrease the longwave radiation loss to space. Cess (1976) has recently suggested that the latter effect (the greenhouse effect) may completely cancel the former (the albedo effect). In this case, the net radiation at the top of the earth-atmosphere system would be insensitive to changes in cloud cover (amount). If true, this finding has important implications for global climate studies. For some simple climate models, it means that the global mean surface temperature is insensitive to changes in cloud amount. There are also implications for global climate models of the general circulation type. Much of the climate sensitivity work being performed with general circulation models is done with prescribed rather than predicted cloudiness (Manabe and Wetherald, 1975; Wetherald and Manabe, 1975), which is certainly a model deficiency. If the net radiation of the system is insensitive to cloud amount, this deficiency of the models may not be that critical, and it may be possible to obtain approximate estimates of climate sensitivity even with models that use prescribed cloudiness (or even with models that predict cloudiness, but do so erroneously). It is therefore of some importance to determine whether the opposite effects of clouds on the radiation budget at the top of the atmosphere do indeed cancel each other. In the present paper we use satellite observations to examine this question.

2. Previous work

Schneider (1972) has introduced a cloud sensitivity parameter δ for determining the effect of a change in cloud amount on the net radiation at the top of the earth-atmosphere system:

$$\delta \equiv \left[\frac{\partial(\mathrm{Net})}{\partial A_c}\right] = \left(\frac{\partial Q}{\partial A_c} - \frac{\partial F}{\partial A_c}\right), \tag{1}$$

where δ is the change in net radiation, Net, due to a change in the amount of clouds A_c; Q is the solar radiation absorbed by the system; and F is the longwave radiation lost to space by the system. Q may be written as

$$Q = Q_0(1 - \alpha), \tag{2}$$

where Q_0 is the available solar radiation and α the planetary albedo; α may be written as

$$\alpha = \alpha_c A_c + \alpha_s(1 - A_c), \tag{3}$$

where α_c is the cloud albedo and α_s is the clear-sky albedo as seen from space.

Assuming that α_c and α_s are not functions of cloud amount, one can obtain

$$\left(\frac{\partial \alpha}{\partial A_c}\right) = (\alpha_c - \alpha_s) \tag{4}$$

and, therefore,

$$\delta = -Q_0(\alpha_c - \alpha_s) - \left(\frac{\partial F}{\partial A_c}\right). \tag{5}$$

A negative value of δ implies that the albedo effect predominates; a positive value implies that the greenhouse effect predominates. Eq. (5) may be applied to global average conditions, in which case Q_0 is simply (SC/4) where SC is the solar constant, or to zonal or regional conditions.

The magnitude of the albedo effect depends on Q_0, which for regional or zonal conditions depends on latitude and time of year, and on the difference between the cloud albedo and the clear-sky albedo. This latter difference is greatest over low-latitude oceanic areas, where α_s is small, and smallest over high-latitude snow/ice covered regions, where α_s is large. The magnitude of the greenhouse effect, $(\partial F/\partial A_c)$, depends mainly on the height of the cloud top and the tropospheric lapse rate. The greater the height of the cloud and the greater the lapse rate, the greater the magnitude of $(\partial F/\partial A_c)$, other things being equal. A more detailed discussion of the albedo and greenhouse effects within the context of the cloud sensitivity parameter δ is presented by Schneider (1972).

Estimates of the average value of $(\partial F/\partial A_c)$ based on model calculations require assumptions on cloud-amount and cloud-height distributions, and temperature and water vapor profiles. The range of values in the literature is from -33 W m^{-2} (Ohring and Adler, 1978) to -75 W m^{-2} (Schneider, 1972). Most of this range is probably attributable to different assumptions on the above parameters, although some variability is also introduced as a result of different radiation codes and differences in the way the average value of $(\partial F/\partial A_c)$ is computed [e.g., from a global average model (Wang and Domoto, 1974), from a zonally averaged hemispheric model (Ohring and Adler, 1978), or from calculations at 260 representative points on the earth (Budyko, 1974)]. Even though the range of estimates of $(\partial F/\partial A_c)$ is quite large, when combined with reasonable estimates of the value of $(\alpha_c - \alpha_s)$, they all yield negative values of δ.

Cess (1976) has determined $(\partial F/\partial A_c)$ from a regression relationship between zonal, annual averages of outgoing longwave radiation based on satellite observations (Ellis and Vander Haar, 1976), and zonal, annual averages of cloudiness and surface temperature based on conventional sources. For the Northern Hemisphere, he obtains a value of -91 W m^{-2} for $(\partial F/\partial A_c)$. When combined with his estimate of 0.26 for $(\alpha_c - \alpha_s)$, this value of $(\partial F/\partial A_c)$ leads to $\delta = -2.6$ W m^{-2}. Using similar analysis techniques for the Southern Hemisphere, Cess obtains $\delta = 0.6$ W m^{-2}. These values of δ suggest almost complete cancellation of the albedo effect of clouds by their greenhouse effect.

To explain why the value of $(\partial F/\partial A_c)$ deduced from the satellite observations is so much larger than that obtained in model calculations, Cess and Ramanathan (1978) suggest that with an increase in total cloud amount, there is typically an increase in the high-cloud fraction and a decrease in the low-cloud fraction visible from space. Since higher clouds have a greater greenhouse effect, the value of $(\partial F/\partial A_c)$ for the real atmosphere is larger than is obtained in model calculations, in which it is generally assumed that the percentage change in cloud amount is the same for all cloud layers and is equal to the percentage change in total cloud amount, or in the notation of Cess and Ramanathan (1978) that $(dA_{ci}/A_{ci}) = (dA_c/A_c)$, where the subscript $_i$ refer to a cloud layer at a particular height. Combining the seasonal cloud-cover fractions of London (1957) into hemispheric averages of high, middle and low clouds, Cess and Ramanathan do indeed find that with an increase in total cloud amount the high-cloud fraction increases and the low-cloud fraction decreases. Using this information to calculate a hemispheric average value of $(\partial F/\partial A_c)$, they obtain a value of -104 W m^{-2} (-83 W m^{-2} for non-black cirrus clouds) as opposed to a value of -49 W m^{-2} (-41 W m^{-2} for non-black cirrus) for a calculation in which the percentage change in cloud amount is the same for all cloud layers. These results tend to confirm Cess's (1976) original conclusion that the albedo effect of a change in cloud amount is essentially completely cancelled by the greenhouse effect of a change in cloud amount, i.e., $\delta \approx 0$. They also point to the desirability of arriving at estimates of δ, and particularly $(\partial F/\partial A_c)$, from observations. In the next section we describe a method for obtaining such estimates from satellite observations.

3. Methods and data

The net radiation at the top of the atmosphere can be written as

$$\text{Net} = Q_0(1 - \alpha) - F. \tag{6}$$

Partial differentiation with respect to albedo yields

$$[\partial(\mathrm{Net})/\partial\alpha] = -Q_0 - (\partial F/\partial\alpha). \tag{7}$$

Using individual monthly mean data from the scanning radiometer observations of the NOAA satellites, we have examined the changes in longwave radiation and albedo for the same month in successive years at selected geographical locations. We interpret the albedo change to be due to a change in mean cloudiness characteristics and, in particular, to a change in mean cloud amount for the month [see Eq. (4)]. Unless the selected geographical area is susceptible to albedo variations due to snowcover variations, an interpretation in terms of a change in cloudiness characteristics appears to be valid. For oceanic areas, this is almost certainly true, and even for land areas this seems like a reasonable assumption, since there is no reason to expect the clear-sky albedo to vary significantly from one month of one year to the same month the following year. Thus, we may associate an increase (decrease) in albedo with an increase (decrease) in cloud amount. That is, the change in albedo is used as an index of the change in cloud amount. Ideally, it would be desirable to have satellite observations of the monthly mean cloud amount itself; unfortunately, such information is not currently available. By plotting the change in longwave radiation against the change in albedo, we can examine the relationship between changes in longwave radiation and cloud amount.

Fig. 1 shows an example of such a graph. Each point represents the change in longwave radiation ΔF and the corresponding change in albedo $\Delta\alpha$ from one month of one year to the same month the following year at the particular geographic location. The slope (b) of the least-square regression line

$$\Delta F = a + b\Delta\alpha \tag{8}$$

through the points will give the best estimate of the average value of $(\Delta F/\Delta\alpha)$ for each geographical area, which we interpret as $(\partial F/\partial\alpha)$.

We have interpreted changes in monthly mean albedo as being due to changes in monthly mean cloud amount. There is also the possibility that there may be random variations in monthly mean cloud albedo that would contribute to the changes in albedo. However, unless there were a correlation between the changes in monthly mean values of cloud amount and cloud albedo, which does not seem likely, they would only add noise to scatter diagrams, such as that in Fig. 1.

We have also interpreted changes in monthly mean longwave radiation as being due to changes in mean cloud amount. Here, the possibility exists that interannual variations of monthly mean surface temperature, the other major determinant of outgoing radiation, would also influence the longwave

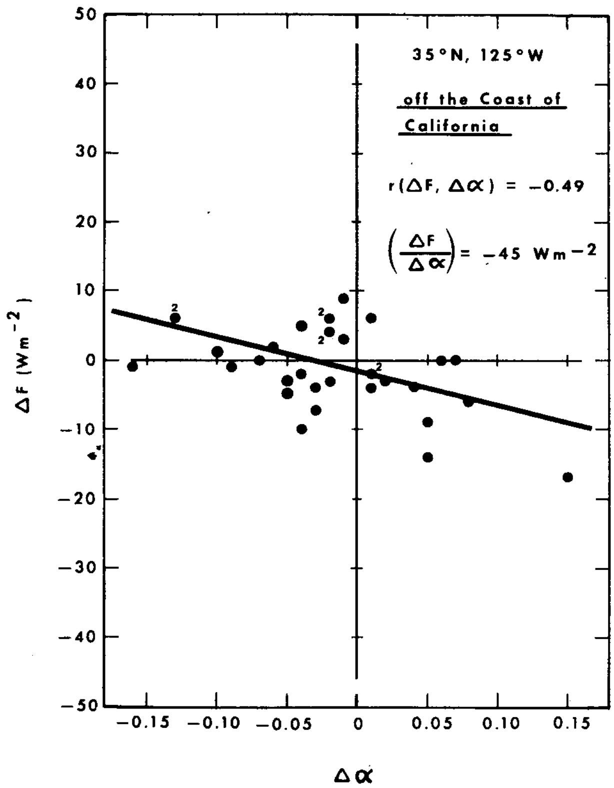

FIG. 1. Scatter diagram showing interannual changes in monthly means of longwave radiation ΔF plotted against interannual changes in monthly means of albedo $\Delta\alpha$ for area 3, centered at 35°N, 125°W. Two data points at same location are indicated by the number 2. The sloping line is the least-squares regression line through the points. $r(\Delta F, \Delta\alpha)$ is the correlation coefficient between ΔF and $\Delta\alpha$. The slope of the regression line gives the best estimate of the average value of $(\Delta F/\Delta\alpha)$, which is interpreted as $(\partial F/\partial\alpha)$.

radiation. If the interannual variations of monthly mean surface temperature were random, their direct effect on the longwave radiation would only add noise to the scatter diagrams. However, if a correlation exists between monthly mean surface temperatures and cloud amounts, the possibility arises of a systematic effect that would either increase or decrease the average value of $(\Delta F/\Delta\alpha)$, depending on the sign of the correlation between cloud amount and surface temperatures. As with the cloud amount-cloud albedo correlation, we believe that such a correlation, even if it does exist, is probably small, and would not significantly affect our results.

The procedure for determining the average values of $(\Delta F/\Delta\alpha)$ from the slopes (b) of the regression lines suppresses the effects of long-period trends in longwave radiation due to possible trends in surface temperature. It also suppresses the effects of possible erroneous trends in the observations of longwave radiation and albedo caused, for example, by degradation of satellite sensors. (See discussion of data below.)

This suppression of trends results from the fact that they enter only into the intercept of the regression line, which is

$$a = \overline{\Delta F} - b\,\overline{\Delta\alpha}, \tag{9}$$

where $\overline{\Delta F}$ and $\overline{\Delta\alpha}$ are the long-period trends in F or α.

Eq. (7) indicates that with the value of $(\partial F/\partial\alpha)$ determined from the slope of the regression line and with the appropriate value of Q_0, which is the average annual available solar radiation (a function only of the latitude of the area), we can obtain an estimate of the average value of $\partial(\mathrm{Net})/\partial\alpha$ for the area. Since α is being used as an index of cloud amount, a positive value of $\partial(\mathrm{Net})/\partial\alpha$ implies that the greenhouse effect is greater than the albedo effect, a negative value implies that the albedo effect is greater than the greenhouse effect, and a value close to zero implies that the greenhouse effect approximately cancels the albedo effect.

The basic data are from the visible (0.5–0.7 μm) and longwave window (10.5–12.5 μm) channels of the scanning radiometers on the polar orbiting NOAA satellites. These satellites were in sun synchronous polar orbits with a 0900 LT southbound equator crossing and a 2100 LT northbound equator crossing. The albedo is determined from the visible channel, assuming an isotropic angular model and no diurnal variation of albedo. The longwave radiation to space is obtained from the longwave window observations by means of a regression relationship derived from simulations. Further details may be found in Gruber and Winston (1978). The original observations have horizontal resolutions of 4 km for the visible channel and 8 km for the longwave channel at the satellite subpoint. The data are archived in the form of daily values for a global 2.5° latitude by 2.5° longitude grid. Forty-five consecutive months of data are available.

For another purpose, investigation of the annual variation of the components of the radiation budget at the top of the atmosphere, a set of 2.5° latitude by 2.5° longitude areas had been selected for processing into monthly means at NOAA's Meteorological Satellite Laboratory. Most of these areas had been selected to correspond to locations for which Budyko (1974) had determined the annual variation of the components of the heat balance at the surface of the earth; the 2.5° grid area closest to each such location had been chosen for processing. Some additional grid areas had also been included. These areas represent a wide variety of climatic and geographic regions and seemed appropriate for the type of analysis discussed above. We did not include the very high-latitude locations in order to avoid the possibility of contamination of the results by snow-cover variations. For each grid area, the values of ΔF and $\Delta\alpha$ were determined from the changes in F and α for the same month in successive years. ΔF was correlated with $\Delta\alpha$ and the slopes of the regression lines were computed. Thus, for each selected grid area, the correlation between ΔF and $\Delta\alpha$ is based on 33 pairs of data.

All of the selected locations are north of 20°S latitude. Time series plots of the 45 monthly mean values of albedo for the entire Northern Hemisphere and for the entire latitude belt 20°N–20°S indicate a trend in the data that may be instrumental rather than real. The monthly mean albedos at the end of the series are ~0.03 lower than at the beginning of the series, with ~0.01 of this difference being accounted for by a drop in albedo between November and December of 1975. The longwave radiation series appears to be much more stable. However, the monthly mean values of longwave radiation are ~2 W m^{-2} lower at the end of the series than at the beginning. Most of this shift appears to have occurred between September and October 1976 and may be due to a change in satellite, NOAA 5 having taken over from NOAA 4 on 15 September 1976. Aside from the fact that the trends in the data are small compared to the interannual variability of longwave radiation and albedo, the procedure used to determine $(\partial F/\partial\alpha)$ supresses their effect on the results.

4. Results and discussion

The results are summarized in Table 1 where, for each selected grid area, information is presented on its location and type of climate; the correlation coefficients between ΔF and $\Delta\alpha$, $r(\Delta F,\Delta\alpha)$; the average values of $(\Delta F/\Delta\alpha)$, as determined from the slopes of the regression lines; the values of Q_0; and the average values of $(\Delta\mathrm{Net}/\Delta\alpha)$, as determined from Eq. (7). The correlation coefficients $r(\Delta F,\Delta\alpha)$ are all negative, as expected, ranging in value from −0.27 to −0.93; the average values of $(\Delta F/\Delta\alpha)$ range from −31 to −254 W m^{-2}. For all locations, the average value of $(\Delta\mathrm{Net}/\Delta\alpha)$ is negative, ranging from −157 to −356 W m^{-2}. Except for the deliberate elimination of high-latitude areas subject to snow/ice cover variations, the selected sample is representative of a broad range of climatic, geographic and surface type conditions. [Actually, one high-latitude location subject to such snowcover variations was also included (grid area 11 at 52.5°N, 82.5°E).] Since $(\Delta\mathrm{Net}/\Delta\alpha)$ is negative for all of these locations, it would appear logical to conclude that for the globe as a whole $(\Delta\mathrm{Net}/\Delta\alpha)$ is negative and, hence, the greenhouse effect of a change in cloud amount does not balance the albedo effect.

Plots of ΔF versus $\Delta\alpha$ are shown for two interesting areas in Figs. 1 and 2. Areas 3 (Fig. 1) and 6 (Fig. 2) are both oceanic, with area 3 at 35°N and area 6 at the equator. For area 3, $(\Delta F/\Delta\alpha) = -45$ W m^{-2} for area 6 $(\Delta F/\Delta\alpha) = -254$ W m^{-2}. The explanation for the large difference in $(\Delta F/\Delta\alpha)$ values for the two locations most likely lies in the

TABLE 1. Relationship of changes in outgoing longwave radiation and net radiation to changes in albedo at selected geographic climatic areas. $r(\Delta F,\Delta\alpha)$ is the correlation coefficient between ΔF and $\Delta\alpha$, and Q_0 is the mean annual available solar radiation.

Area	Geographic/climatic type	Latitude	Longitude	$r(\Delta F,\Delta\alpha)$	$(\Delta F/\Delta\alpha)$ (W m^{-2})	Q_0 (W m^{-2})	$(\Delta\mathrm{Net}/\Delta\alpha)$ (W m^{-2})
1	Equatorial, continental	0	67.5°W	−0.79	−192	413	−221
2	Equatorial, monsoon	10°N	107.5°E	−0.85	−249	408	−159
3	Subtropical, eastern side oceanic high	35°N	125°E	−0.49	−45	345	−300
4	Subtropical, continental, semi-arid	35°N	105°W	−0.84	−129	345	−216
5	Subtropical, continental, monsoon	25°N	75°E	−0.86	−221	378	−157
6	Equatorial, oceanic	0	180°W	−0.93	−254	413	−159
7	Equatorial, oceanic	0	110°W	−0.46	−57	413	−356
8	Tropical, western side oceanic high	17.5°N	87.5°W	−0.88	−178	396	−218
9	Tropical, eastern side oceanic high	15°S	12.5°E	−0.48	−95	401	−306
10	Subtropical, monsoon	32.5°N	130°E	−0.88	−134	354	−220
11	Midlatitude, continental	52.5°N	82.5°E	−0.68	−95	270	−175
12	Subtropical, continental, arid	40°N	55°E	−0.71	−132	326	−194
13	Equatorial, oceanic	0	30°W	−0.92	−219	413	−194
14	Equatorial, oceanic, monsoon	15°N	70°E	−0.92	−219	400	−181
15	Tropical, western side oceanic high	20°S	40°E	−0.82	−185	391	−206
16	Tropical, eastern side oceanic high	20°S	10°E	−0.69	−45	391	−346
17	Midlatitude, warm current	60°N	10°W	−0.27	−31	234	−203
18	Midlatitude, oceanic, monsoon	45°N	160°E	−0.62	−79	304	−225

different effect of the cloud regimes at the two locations on the longwave radiation. Area 6 represents a region in the equatorial Pacific that is subject to incursions of convective cloudiness from the south and west, while area 3 is representative of the stratiform cloud regime off the western coast of California. The low clouds off the California coast have only a small effect on the longwave radiation to space while the thick convective type clouds in the equatorial Pacific location, with their higher cloud top heights, have a much greater effect.

Our highest latitude location (area 17 at 60°N, 10°W) is an oceanic location, which, as a result of the presence of a warm current, is not subject to sea ice conditions. Because of its dependence on Q_0, the value of $(\Delta\mathrm{Net}/\Delta\alpha)$ might be expected to be close to zero at high latitudes in winter, when the average value of Q_0 is low. This is indeed the case at area 17, where $\Delta\mathrm{Net}/\Delta\alpha$ is -31 W m^{-2} for the winter half of the year compared to a value of -382 W m^{-2} for the summer half of the year.

Thus far we have shown that, for our sample of representative climatic regimes, $(\Delta\mathrm{Net}/\Delta\alpha)$ is clearly negative, indicating that the albedo effect of a change in cloudiness is not cancelled by the greenhouse effect. It would be desirable to obtain estimates of just how much greater is the albedo effect, in other words, estimates of the numerical value of δ. Such estimates can be obtained from our values of $(\Delta\mathrm{Net}/\Delta\alpha)$ by writing

$$\delta \approx \left(\frac{\Delta\ \mathrm{Net}}{\Delta A_c}\right) = \left(\frac{\Delta\ \mathrm{Net}}{\Delta\alpha}\right)\left(\frac{\Delta\alpha}{\Delta A_c}\right) = (\alpha_c - \alpha_s)\left(\frac{\Delta\ \mathrm{Net}}{\Delta\alpha}\right) \quad (10)$$

and estimating typical values of $(\alpha_c - \alpha_s)$ for the various areas. In a similar manner, estimates of $(\partial F/\partial A_c) \approx (\Delta F/\Delta A_c)$ can be obtained from $(\Delta F/\Delta\alpha)$. We have attempted to obtain such estimates. Values of the surface albedo (α_{s0}) for the different areas were obtained from the surface albedo maps of Posey and Clapp (1964), and are listed in Table 2. In this analysis, those areas for which it was difficult to obtain a representative surface albedo value were

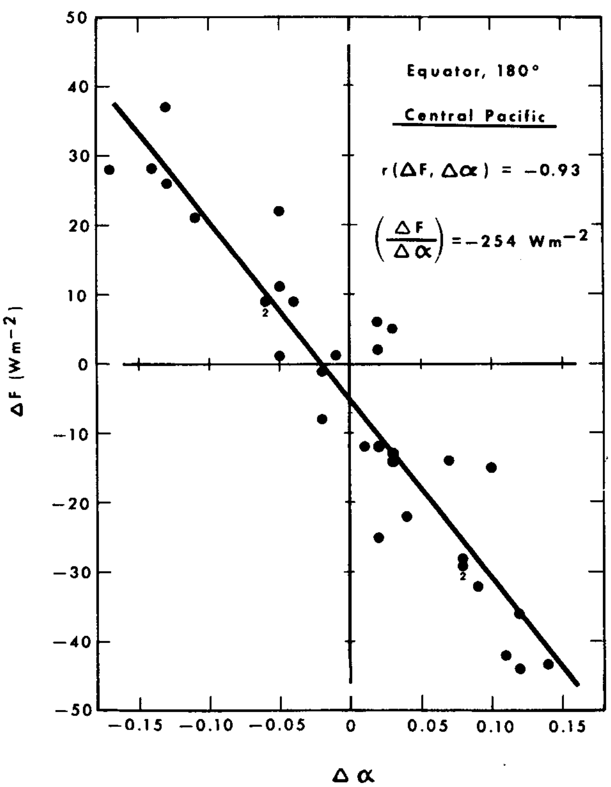

FIG. 2. As in Fig. 1 except for area 6.

TABLE 2. Estimates of the relationship of changes in outgoing longwave radiation and net radiation to changes in cloud amount at selected geographic/climatic areas.

Area	α_c (1)	α_c (2)	α_{s0}	α_s	$(\alpha_c - \alpha_s)$ (1)	$(\alpha_c - \alpha_s)$ (2)	$(\Delta F/\Delta A_c)$ (1) (W m^{-2})	$(\Delta F/\Delta A_c)$ (2) (W m^{-2})	$(\Delta\mathrm{Net}/\Delta A_c)$ (1) (W m^{-2})	$(\Delta\mathrm{Net}/\Delta A_c)$ (2) (W m^{-2})
1	0.35	0.41	0.07	0.12	0.23	0.29	−44	−56	−51	−64
3	0.46	0.48	0.07	0.13	0.33	0.35	−15	−16	−99	−105
4	0.46	0.48	0.15	0.18	0.28	0.30	−36	−39	−60	−65
5	0.42	0.46	0.10	0.14	0.28	0.32	−62	−71	−44	−50
6	0.35	0.41	0.06	0.11	0.24	0.30	−61	−76	−38	−48
7	0.35	0.41	0.06	0.11	0.24	0.30	−14	−17	−85	−107
12	0.46	0.49	0.20	0.22	0.24	0.27	−32	−36	−47	−52
13	0.35	0.41	0.06	0.11	0.24	0.30	−53	−66	−47	−58
14	0.38	0.43	0.06	0.12	0.26	0.31	−56	−68	−47	−56
15	0.40	0.44	0.06	0.12	0.28	0.32	−52	−59	−58	−66
16	0.40	0.44	0.06	0.12	0.28	0.32	−13	−14	−97	−111
17	0.52	0.57	0.12	0.20	0.32	0.37	−10	−11	−65	−75
18	0.46	0.51	0.08	0.15	0.31	0.36	−24	−28	−70	−81
Mean							−36	−43	−62	−72

(1) refers to use of α_c values of Cess (1976).
(2) refers to use of α_c values of Ohring and Adler (1978).

eliminated (e.g., coastal areas consisting of continental and oceanic segments or area 11, which is subject to winter snowcover variations). The surface albedo values were converted to clear-sky albedo values by using the computational results of Braslau and Dave (1973), which give α_s as a function of α_{s0} and solar zenith angle for a typical noncloudy atmosphere with average aerosol amounts (their C1 aerosol); these values of α_s are shown in Table 2. Two sets of estimates of typical cloud albedo values were used. One set is based on the meridional distribution of α_c used by Cess (1976), indicated by $\alpha_c(1)$ in Table 2; the other set is based on the meridional distribution of α_c used by Ohring and Adler (1978) in a zonally averaged climate model, indicated by $\alpha_c(2)$ in Table 2. The resulting estimates of δ and $(\partial F/\partial A_c)$ are shown in Table 2.

The values of δ vary from −38 to −99 W m^{-2} for the $\alpha_c(1)$ set and from −48 to −111 W m^{-2} for the $\alpha_c(2)$ set. $(\partial F/\partial A_c)$ varies from −10 to −62 W m^{-2} for the $\alpha_c(1)$ set and from −11 to −76 W m^{-2} for the $\alpha_c(2)$ set. Particularly high values of $(\partial F/\partial A_c)$ are associated with monsoon type climatic regions and with the area in the central equatorial Pacific discussed earlier. The cloudiness in these regions extends to high tropospheric levels, thus having a large effect on the outgoing longwave radiation. The lowest values of $(\partial F/\partial A_c)$ occur in regions characterized by low-level cloudiness, e.g., areas on the eastern sides of the oceanic high pressure systems.

It is tempting to average the individual estimates of δ and $(\partial F/\partial A_c)$ to obtain estimates of the global means of these quantities; we have done this and the results are shown in the bottom line of Table 2. These estimates, of course, are only rough estimates of the global means and must be interpreted with care. The mean values of δ and $(\partial F/\partial A_c)$ are −62 and −36 W m^{-2} for the $\alpha_c(1)$ set and −72 and −43 W m^{-2} for the $\alpha_c(2)$ set. Due to the absence from our sample of high-latitude areas subject to snow/ice cover variations, the mean value of δ that we obtain is probably somewhat more negative than the actual mean value for the following reason. At high latitudes, the available solar radiation and, probably also, the difference $(\alpha_c - \alpha_s)$ are lower than at other latitudes. Thus, $(\partial Q/\partial A_c)$ is less negative at high latitudes. If $(\partial F/\partial A_c)$ remains the same at high latitudes, δ at high latitudes would be less negative than at other latitudes. (Due to the complete absence of solar radiation, δ is positive in the polar night region.) Thus, by omitting this region from our sample, we obtain a mean δ that is probably biased toward a more negative value.

It is of importance to try to explain why the present estimate of δ differs so much from Cess's (1976) estimate that $\delta \approx 0$. To correct for the bias discussed above, we shall add 5 W m^{-2} to each of our estimates of the mean value of δ, so that for the set $\alpha_c(1)$ the corrected average value of δ is −57 W m^{-2} and for the set $\alpha_c(2)$ it is −67 W m^{-2}. Thus, the difference between the present δ and Cess's δ is −57 W m^{-2} for $\alpha_C(1)$ and −67 W m^{-2} for $\alpha_c(2)$. The difference between the present $(\partial F/\partial A_c)$ and Cess's value of −91 W m^{-2} is 55 W m^{-2} for $\alpha_c(1)$ and 48 W m^{-2} for $\alpha_c(2)$. Thus, the difference in the $(\partial F/\partial A_c)$ values explains essentially all of the difference in δ for the $\alpha_c(1)$ set and a good part of the difference for the $\alpha_c(2)$ set.

It is not clear why there should be such a large difference in the values of $(\partial F/\partial A_c)$ between the present study and that of Cess (1976). Both studies make use of satellite observations of the outgoing radiation and thus require no assumptions on how a change in total cloud amount is distributed among the cloud layers at different heights, as do estimates of $(\partial F/\partial A_c)$ from model calculations. The only other satellite based estimates of $(\partial F/\partial A_c)$ are those of Ellis (1978), who obtains a value of ~ -40 W m^{-2}. However, this value is based on a comparison of the outgoing radiation for clear skies and for average cloudiness. As indicated by Cess and Ramanathan (1978), such a procedure for estimating $(\partial F/\partial A_c)$ contains the implicit assumption that $(dA_{ci}/A_{ci}) = (dA_c/A_c)$ and, hence, does not admit the possibility that changes of total cloud amount are not distributed among the various cloud layers according to the relative amounts of clouds prevailing in the various layers.

To look into this matter somewhat more deeply we have performed an additional calculation. Cess's (1976) value of $(\partial F/\partial A_c) = -91$ W m^{-2} is based upon a regression equation between the outgoing radiation and the surface temperature and cloud amount for zonally averaged (10° latitude belts) annual values for the Northern Hemisphere. We have calculated a similar regression equation for zonally averaged seasonal values (actually winter, April, summer and October) using the same sources for F (Ellis and Vonder Haar, 1976) and A_c (London, 1957) as used by Cess for the annual values. Surface temperature data were taken from Oort and Rasmusson (1971) [with extrapolations from their highest latitude (75°N) to our highest latitude belt centered at 85°N being done graphically]. We obtained the regression equation

$$F[\mathrm{W\ m^{-2}}] = 240 + 1.8T_s - 60A_c, \qquad (11)$$

where T_s is in °C. The multiple correlation coefficient is 0.95. According to this equation $(\partial F/\partial A_c) = -60$. This value of $(\partial F/\partial A_c)$ falls between that obtained by Cess (1976) from the annual values and that obtained in the present study, although it is closer to the latter. It is also greater than the value of -71 W m^{-2} obtained by Budyko (1974) from simulations of monthly mean outgoing longwave radiation at 260 points, uniformly distributed around the globe. Its use in the expression for the global average value of δ yields $\delta = -28$ W m^{-2} using the value of $(\alpha_c - \alpha_s) = 0.26$ suggested by Cess (1976), and $\delta = -49$ W m^{-2} using $(\alpha_c - \alpha_s) = 0.32$ from Ohring and Adler (1978). In either case,it yields a negative value for δ, although not as large as that suggested by the present study.

5. Conclusions and recommendations

Based on our analysis of satellite observations of year-to-year changes in monthly mean values of outgoing longwave radiation and albedo of a sample of geographic and climatic areas of the earth, we conclude that the opposite effects of variations in cloud amount on the radiation budget at the top of the atmosphere do not cancel each other. In particular, the albedo effect is greater than the greenhouse effect of a change in cloud amount, i.e., the cloud sensitivity parameter $\delta < 0$. Thus, in climate sensitivity studies being performed with climate models, it would be desirable to be able to predict cloudiness (correctly) rather than to prescribe cloudiness, so that this cloud feedback effect can be properly simulated.

We obtain the following sets of estimates for δ and $(\partial F/\partial A_c)$ (based mostly on Northern Hemisphere data):

$$\delta = -57\ \mathrm{W\ m^{-2}}, \quad (\partial F/\partial A_c) = -36\ \mathrm{W\ m^{-2}}, \qquad (12)$$

$$\delta = -67\ \mathrm{W\ m^{-2}}, \quad (\partial F/\partial A_c) = -43\ \mathrm{W\ m^{-2}}. \qquad (13)$$

The first set is based on the use of the cloud albedo values of Cess (1969); the second set is based on the cloud albedo values of Ohring and Adler (1978).

Because of the nature of the satellite observations used in the present study, and because of the assumptions necessary to apply our method, the above conclusions should be considered as tentative and the above values as preliminary estimates only. The albedo and longwave radiation values used in the present study are from the visible and infrared window channels of the scanning radiometer of the NOAA satellites. The visible channel measures only part of the total solar radiation reflected to space and the infrared channel measures only part of the total longwave radiation to space. Although the processing of these observations into total albedo and total longwave radiation is based on very high correlations between the parts measured and the totals, some uncertainties remain. For example, the total longwave radiation is obtained from the infrared window radiation observations by means of a nonlinear regression relationship based on simulations for some 100 model atmospheres. About half of the simulations were for cloudy conditions, with the clouds assumed to be blackbodies. A reviewer has pointed out that because cirrus clouds are much more transparent in the infrared window region than at other parts of the longwave spectrum, the regression relationship used to convert from window radiation to total longwave radiation may not completely account for variations in outgoing longwave radiation due to variations in cirrus cloud amount. Because of uncertainties such as these, it would be highly desirable to apply the method used in the present study to observations from instruments designed to measure the total albedo and the total longwave radiation, when they become available from the earth radiation budget satellite systems.

In the present study, albedo changes were used as a measure of cloud amount changes. This was done because there were no cloud amount data. It would be highly desirable to conduct further studies of this type using satellite cloud amount data, when they become available.

The influence of a possible correlation between surface temperature and cloud amount on the results obtained would have to be checked with concurrent and collocated observations of albedo, outgoing longwave radiation, cloud amount and surface temperature. Except for oceanic areas, it might be difficult to obtain representative monthly mean surface temperatures for 2.5° × 2.5° areas.

And, finally, it would be desirable to extend analyses of the type performed here to obtain greater coverage of the geographical climatic regions of the globe.

Acknowledgments. Discussions with A. Krueger, S. Fritz, H. Jacobowitz and A. Gruber helped to clarify various aspects of the subject. T. I. Gray, Jr., was responsible for processing the satellite data into the form used in the present study. Sarah Rechavi and Benyimin Neeman programmed the correlation and regression analyses.

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