Low-Latitude Cloudiness and Climate Feedback: Comparative Estimates from Satellite Data

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ABSTRACT

At low latitudes the seasonal variation in the radiation budget of the earth-atmosphere system is due largely to seasonal variability in cloudiness. Making use of this, we have estimated, from three different sets of satellite data, the relative albedo versus infrared modifications associated with cloudiness variability at low latitudes. Employing satellite data sets due to Ellis and Vonder Haar (1976) and Campbell and Vonder Haar (1980), we find that the albedo modification is somewhat less than that of the infrared. But when use is made of radiation budget data derived from NOAA-NESS scanning radiometer measurements, the albedo modification dominates over that of the infrared by nearly a factor of 2. This obviously suggests that estimates of climate feedback associated with changes in cloudiness are highly dependent on the satellite data set which is employed. It is further suggested that these differences might be in part attributable to the NOAA-NESS data being derived from narrow spectral measurements.

1. Introduction

Cloud amount, as a terrestrial climate feedback mechanism, has stirred considerable controversy and debate. Roads (1978) summarizes several predictions for the response of cloud amount to climatic change, and recent general circulation model results are presented by Schneider *et al.* (1978) and by Wetherald and Manabe (1980).

But the response of cloud amount to climatic change comprises only one aspect of the problem. If cloud amount is, in fact, significantly altered as a consequence of climatic change, then both the albedo and the infrared opacity of the earth-atmosphere system are correspondingly modified. Thus, even if one knows the manner in which cloud cover is altered, in order to appraise cloud-amount feedback one additionally requires knowledge as to how the albedo and the infrared opacity of the earth-atmosphere system are related to a change in cloud cover.

As discussed by Schneider (1972), a convenient parameter for illustrating the relative importance of these separate albedo and infrared modifications may be defined as

$$\delta = \frac{\partial Q_a}{\partial A_c} - \frac{\partial F}{\partial A_c}, \tag{1}$$

with Q_a denoting the absorbed solar radiation, F the outgoing infrared radiation and A_c the cloud cover fraction. For $\delta < 0$ the albedo modification dominates, whereas if $\delta = 0$ the albedo and infrared feedback mechanisms compensate.

A variety of negative values for δ have been suggested from model studies. But as discussed by Cess and Ramanathan (1978), these possibly stem from inadequate modeling of $\partial F/\partial A_c$, since this quantity cannot be evaluated from a model calculation unless the model has the capability of predicting how the amounts of individual cloud layers change in relation to a change in total cloud cover. Cess and Ramanathan further point out that a similar caveat applies to Ellis' (1977) empirical estimate of $\partial F/\partial A_c$.

In contrast to model studies, there have been three empirical investigations pertaining to the relative albedo versus infrared components of cloud amount feedback (Cess, 1976; Ohring and Clapp, 1980; Hartmann and Short, 1980). In the following section we review these three investigations, after which we present yet a fourth approach, one which employs seasonal variability in low-latitude cloud amount. We primarily utilize this fourth method to illustrate intercomparisons and uncertainties in such empirical

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predictions resulting from the satellite data set which is employed.

2. Empirical cloud feedback studies

Employing zonal annual climatological data for the Northern Hemisphere, Cess (1976) has suggested that $\partial Q_a/\partial A_c = -88\ W\ m^{-2}$ and $\partial F/\partial A_c = -91\ W\ m^{-2}$, implying that cloud amount is not a significant climate feedback mechanism, since $\delta \simeq 0$. A similar result was obtained for the Southern Hemisphere, and this suggestion was further consistent with an interpretation of seasonal climatological data for a limited number of midlatitude zones (Cess, 1976).

Nevertheless these empirical estimates for $\partial Q_a/\partial A_c$ and $\partial F/\partial A_c$ are highly uncertain, since they employ ground-based observations of cloud-cover fraction which may be subject to substantial error. Indeed, it is by no means clear that a unique definition for A_c even exists; the effective cloud cover fraction for incoming solar radiation might differ significantly from that for outgoing infrared radiation.

Alternate approaches to estimating the albedo versus infrared components of cloud-amount feedback, which do not explicitly utilize cloud-cover climatologies, have recently been presented by Ohring and Clapp (1980) and by Hartmann and Short (1980). The results of both studies may be expressed in terms of the quantity

$$\epsilon = \frac{\partial F}{\partial Q_a}, \tag{2}$$

with the partial derivative implying that changes in F and Q_a are due solely to variations in cloud cover. Thus $\epsilon = 1$ refers to albedo-infrared compensation with changing cloud amount ($\delta = 0$), while for $\epsilon < 1$ the albedo modification dominates ($\delta < 0$).

Employing interannual variability in regional monthly-mean satellite radiation data, and attributing this variability as being the sole result of interannual variability in cloud cover, Ohring and Clapp (1980) estimate that globally $\epsilon \approx 0.4$. In the related study, Hartmann and Short (1980) employed regional day-to-day variability in satellite radiation data and assumed, as in Ohring and Clapp's interannual study, that this variability is due solely to day-to-day cloud cover variability. They estimate that globally $\epsilon \leqslant 0.5$. Thus the estimates by Ohring and Clapp (1980), and by Hartmann and Short (1980), are in effect mutually consistent, suggesting that the albedo component of cloud amount feedback dominates over the infrared component by roughly a factor of 2 or more, as opposed to Cess' (1976) suggestion of compensation.

At this point it is well to note a problem in all three of these studies. Hartmann and Short (1980) have correctly emphasized that Cess' (1976) $\partial Q_a/\partial A_c$ and $\partial F/\partial A_c$ values do not represent true partial derivatives, since his approach certainly does not eliminate all latitudinal variations in F and Q_a, except those induced by cloud cover changes, when estimating these derivatives. But by the same token this criticism is equally applicable to Ohring and Clapp, and to Hartmann and Short, since $\partial F/\partial Q_a$ inferred by them is also not a true partial derivative, in the sense that changes in F and Q_a are due only to cloud cover variability. Certainly, interannual and day-to-day regional radiation budget variations will be the result not only of changes in cloud cover, but will also contain effects due to variations in other quantities, such as atmospheric water vapor content, lapse rate, etc. One would, in fact, expect some type of correlation between cloud cover and atmospheric water vapor content, and perhaps an estimate of $\partial F/\partial Q_a$ which includes such an effect would prove to be more useful than one which does not.

A further difficulty pertaining to all three empirical studies is that they employ linear regressions of climatological data, such that they utilize regression analyses in which all variables are subject to error. For example, both Ohring and Clapp (1980) and Hartmann and Short (1980) estimate $\epsilon = \partial F/\partial Q_a$ by evaluating $\partial F/\partial \alpha$, where α denotes the albedo, as the slope of the least-squares regression of F on α. But this actually constitutes a lower-bound regression estimate of $\partial F/\partial \alpha$ and hence ϵ, since α is not a true independent variable. The upper-bound estimate corresponds to the reciprocal of the slope of the least-squares regression of α on F (e.g., Johnston, 1963), with the ratio of the two bounds being equal to the square of the correlation coefficient. Thus the estimates of ϵ by Ohring and Clapp, and by Hartmann and Short, should be interpreted as lower-bound regression estimates rather than actual regression estimates. This same caveat applies to the study by Cess (1976), since his estimate that $\delta \approx 0$ from (1) may be rephrased as $\epsilon = (\partial F/\partial A_c)/(\partial Q_a/\partial A_c) \approx 1$, and he estimated $\partial F/\partial A_c$ from a regression analysis.

There is yet another point concerning these empirical cloud feedback studies. Cess utilized the satellite radiation budget data compilation of Ellis and Vonder Haar (1976), whereas Ohring and Clapp, and Hartmann and Short, employed radiation budget data derived from NOAA–NESS scanning radiometer measurements. In the following we suggest yet a fourth "climate experiment" pertaining to cloud amount feedback, for which we employ both sets of satellite radiation budget data, in addition to yet a third recent data set, for the purpose of comparison with the three existing empirical interpretations of cloud feedback.

3. Low-latitude seasonal variability

All satellite observations of the outgoing infrared flux at the top of the atmosphere indicate that, at low latitudes, the flux is a maximum in winter and a minimum during the summer. This constitutes an anticorrelation with seasonal cloud amount at these latitudes, indicating that at low latitudes the dominant seasonal influence upon the outgoing flux comprises seasonal variability in cloud cover. This suggests the following "climate experiment" for estimating, at least for low-latitude seasonal cloud variability, the relative roles of albedo and infrared opacity modifications due to changes in cloud amount.

Let F^* and Q_a^* denote F and Q_a for which seasonal variations due to factors other than seasonal changes in cloud amount have been removed. Thus (2) becomes

$$\epsilon = \frac{dF^*}{dQ_a^*}. \tag{3}$$

To convert satellite observed monthly mean values of F and Q_a to F^* and Q_a^*, we assume that to first order $F = F(A_c, T_s)$ and $Q_a = Q_a(A_c, \mu, Q)$, where $\mu = \cos(\text{solar zenith angle})$ and Q is the insolation at the top of the atmosphere. Correspondingly, letting an overbar refer to annual mean quantities, we have

$$F^*(A_c) = F(A_c, \bar{T}_s), \tag{4}$$

$$Q_a^*(A_c) = Q_a(A_c, \bar{\mu}, \bar{Q}). \tag{4b}$$

Thus to convert monthly means for $F(A_c, T_s)$ to $F^*(A_c)$, we use

$$F^*(A_c) = F(A_c, T_s) - \frac{\partial F}{\partial T_s}(T_s - \bar{T}_s), \tag{5}$$

with $\partial F/\partial T_s = 1.6$ W m^{-2} °C^{-1} (Cess, 1976). The actual choice of $\partial F/\partial T_s$ makes little difference, since the temperature correction is small. We have also employed $\partial F/\partial T_s = 3$ W m^{-2} °C^{-1}, as suggested by the general circulation model study of Wetherald and Manabe (1980) for tropical regions, and find that this increases our estimates of ϵ by only ~5–10%. In the same manner

$$Q^*(A_c) = \bar{Q}(1 - \alpha^*), \tag{6}$$

where

$$\alpha^*(A_c) = \alpha(A_c, \mu) - \frac{\partial \alpha}{\partial \mu}(\mu - \bar{\mu}). \tag{7}$$

The zenith-angle dependence of the albedo, $\partial\alpha/\partial\mu$, has been estimated as in Lian and Cess (1977), with monthly mean values for T_s being taken from Crutcher and Meserve (1970) for the Northern Hemisphere and from Taljaard *et al.* (1969) for the Southern Hemisphere, and monthly mean values of μ were evaluated from a standard orbital calculation (Sellers, 1965). Our use of $\partial\alpha/\partial\mu$, which stems from the separate zenith-angle dependences of both clear-sky and cloudy-sky albedos, is of course highly uncertain. We have taken $\partial\alpha/\partial\mu = -0.38$, independent of both latitude and season. But we have found that a ±25% variation in $\partial\alpha/\partial\mu$ produces a maximum ∓15% change in our estimates for ϵ. Since the present study is primarily concerned with comparative rather than absolute estimates for ϵ, we regard the obvious uncertainty in $\partial\alpha/\partial\mu$ as acceptable.

A further point concerns our earlier discussion of the difficulty in estimating true partial derivatives from climatological data. Realistically, as defined by (3), ϵ is not just a measure of the relative infrared versus albedo modifications associated with changes in cloud amount, but when estimated from climatological data it also will incorporate modifications due to changes in cloud altitude, cloud structure and cloud optical properties. These, of course, are all important ingredients of the overall cloud feedback problem, and in this sense ϵ should be regarded an indicator of net cloudiness feedback rather than cloud amount feedback.

For the purpose of obtaining estimates of ϵ from seasonal variability, we have employed zonal-average monthly means of F and α for the four latitude zones 20–10°N, 10°N–0°, 0°–10°S and 10–20°S. Recall that the present procedure is restricted to low latitudes. At mid- and high latitudes the seasonal variability in F due to changes in surface air temperatures, and in Q_a due to zenith angle variability, become too large to reasonably correct by present methods. The satellite data sets which we have employed are summarized below:

- Ellis and Vonder Haar (1976).
- Monthly means from the 45-month June 1974 to February 1978 NOAA-NESS data set as evaluated from a data tape obtained from NOAA through NCAR. These monthly means thus consist of averages of four months of data, except for March, April and May, for which only three months of data are available.
- Monthly means as evaluated from a data tape for the radiation budget data described by Cambell and Vonder Haar (1980).

We consider first linear regression analyses without correcting for seasonal variations in surface air temperature and zenith angle. Thus we employ the infrared data directly, whereas for absorbed solar radiation we use

$$Q_a' = \bar{Q}(1 - \alpha).$$

The seasonal variability in Q_a' is thus due solely to seasonal variations in α. Furthermore, to separate seasonal and latitudinal variations, we replace the

TABLE 1. Comparison of regression analyses and estimates of dF/dQ_a' as obtained by employing different earth radiation budget data sets.

Data set	$\sigma(F)$, (W m^{-2})	$\sigma(Q_a')$, (W m^{-2})	Correlation coefficient	dF/dQ_a' Min	Max	Wald
Ellis and Vonder Haar	9.9	4.6	0.75	1.60	2.84	2.08
Campbell and Vonder Haar	9.3	5.7	0.74	1.22	2.23	1.80
NOAA-NESS	6.9	9.4	0.87	0.63	0.83	0.74

F and Q_a' values by ΔF and $\Delta Q_a'$ which represent, for a given latitude zone, monthly-mean seasonal variations about the annual mean for that latitude zone.

The monthly means for the four latitude zones thus yield 48 pairs of ΔF and $\Delta Q_a'$ values, and linear least-square regressions of these data are summarized in Table 1. Listed here are the standard deviations of both ΔF and $\Delta Q_a'$, identified as $\sigma(F)$ and $\sigma(Q_a')$, the correlation coefficients, and the minimum and maximum regression estimates for dF/dQ_a'. Also, for comparative purposes, we have employed Wald's method (Johnston, 1963) of fitting a straight line when both variables are subject to error. These estimates, which we take as the "most likely" values for dF/dQ_a', are also listed in Table 1. A striking aspect of Table 1 concerns the differences between estimates using the NOAA-NESS data as compared to the other two data sets, and a discussion of this will be given later.

The same regression results are listed in Table 2 when corrections have been incorporated for seasonal variability in surface air temperature and zenith angle; a comparison of Tables 1 and 2 thus serves to show the magnitude of these corrections. Since there is little difference between $\sigma(F)$ and $\sigma(F^*)$, the correction for surface air temperature is not substantial, whereas the $\sigma(Q_a')$ and $\sigma(Q_a^*)$ comparisons indicate a significant effect due to the zenith-angle corrections.

We also have estimated dF^*/dQ_a^* by arbitrarily assuming that the ratio of error variances is unity, and then employing a linear least-square regression (Johnston, 1963). This yields $dF^*/dQ_a^* = 1.39$, 1.27 and 0.51 for the Ellis and Vonder Haar, Campbell and Vonder Haar, and NOAA-NESS data sets, respectively. These values do not differ appreciably from those given in Table 2 utilizing Wald's method.

Scatter diagrams of ΔF^* versus ΔQ_a^* are illustrated in Figs. 1–3 for the Ellis and Vonder Haar, Campbell and Vonder Haar, and NOAA-NESS data sets, respectively. In addition to treating all four latitude zones as a composite of 48 data points, we have additionally estimated dF^*/dQ_a^* using the 12 monthly means for each latitude zone. These results are summarized in Table 3. With the exception of the Ellis and Vonder Haar data set for 10°N–0°, these individual estimates for dF^*/dQ_a^* show modest latitudinal variability and are reasonably consistent with the composite values.

As previously discussed, available estimates of dF^*/dQ_a^* refer to minimum regression estimates. Ohring and Clapp (1980) and Hartmann and Short (1980) both employ the NOAA-NESS data and estimate that $dF^*/dQ_a^* \approx 0.4$ and $\leqslant 0.5$, respectively, while Cess (1976), employing the Ellis and Vonder Haar monthly means, suggests that $dF^*/dQ_a^* \approx 1.0$. Interestingly enough, the minimum dF^*/dQ_a^* values given in Table 2 are quite consistent with these estimates, when cognizance is taken of the satellite date set which has been employed. From Table 2 $(dF^*/dQ_a^*)_{min} \approx 0.5$ for the NOAA-NESS data, while $(dF^*/dQ_a^*)_{min} \approx 1.0$ for the Ellis and Vonder Haar data.

Of course, such agreement could be in part coincidental. As previously discussed, ϵ is a measure not only of the albedo versus infrared modifications associated with changes in cloud amount, but it also includes effects due to changes in cloud structure, height and optical properties. Since the seasonal low-latitude increase in cloud amount coincides with an increase in deep convective clouds, it is by no means

TABLE 2. Comparison of regression analyses and estimates of dF^*/dQ_a^* as obtained by employing different earth radiation budget data sets.

Data set	$\sigma(F^*)$, (W m^{-2})	$\sigma(Q_a^*)$, (W m^{-2})	Correlation coefficient	dF^*/dQ_a^* Min	Max	Wald
Ellis and Vonder Haar	10.6	8.2	0.79	1.03	1.65	1.28
Campbell and Vonder Haar	10.4	8.5	0.84	1.03	1.46	1.32
NOAA-NESS	7.9	13.9	0.80	0.46	0.72	0.59

clear that the present approach should be compatible with the other three empirical approaches which employ interannual, day-to-day and latitudinal variability. Nor, for that matter, is it clear that any of these methods are actually indicative of cloudiness change as a climate feedback mechanism. About all that one can hope for is to attempt to find a consensus from a number of imperfect "climate experiments."

The estimates shown in Table 2 which employ the Campbell and Vonder Haar data do not differ appreciably from those using the Ellis and Vonder Haar data. Both data sets employ essentially the same set of satellite experiments from the 1960's and early 1970's, while the Campbell and Vonder Haar data set additionally incorporates two years of Nimbus 6 data.

What is worrisome are the substantially different results for dF^*/dQ_a^* which are obtained employing the NOAA-NESS data as opposed to the other two data sets. This roughly factor of 2 difference can be traced to the following causes:

- About 70% of the difference is due to a considerably larger variation in Q_a^* as evaluated from the NOAA-NESS data, which may be noted by comparing the $\sigma(Q_a^*)$ values in Table 2.
- The remaining 30% of the difference is due to a smaller seasonal variation in F^* as determined from the NOAA-NESS data.

This infrared difference is consistent with an alternate interpretation. Applying Cess' (1976) procedure to the NOAA-NESS data yields $\partial F/\partial A_c = -68$ W m^{-2}, which would reduce dF^*/dQ_a^* by 25% relative to the use of $\partial F/\partial A_c = -91$ W m^{-2} as determined from the Ellis and Vonder Haar data.

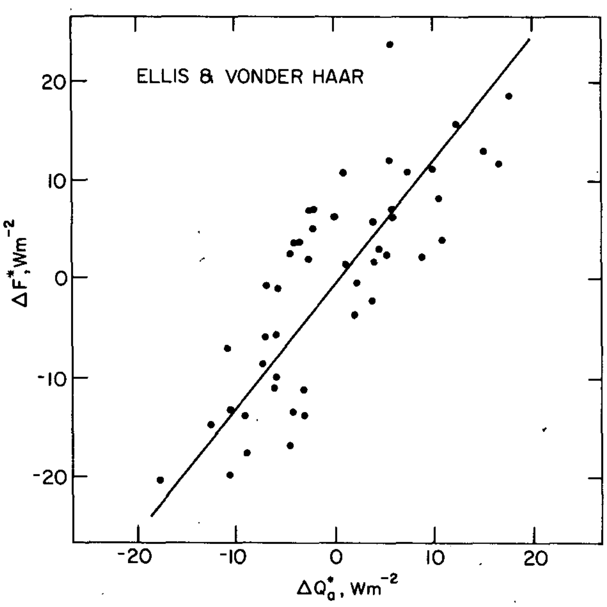

FIG. 1. Scatter diagram of ΔF^* versus ΔQ_a^* for the satellite data set of Ellis and Vonder Haar. Also shown is the straight-line fit using Wald's method.

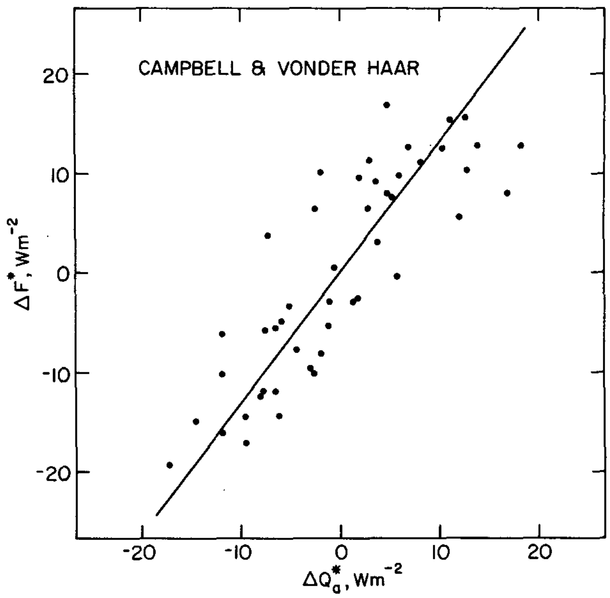

FIG. 2. As in Fig. 1 except for the Campbell and Vonder Haar data set.

We have no quantitative explanation for these different dF^*/dQ_a^* estimates, but a possible cause pertains to the fact that the NOAA-NESS data are based on 0.5–0.7 μm solar and 10.5–12.5 μm infrared measurements, the latter being converted through an algorithm to total infrared values. It is possible that these constraints underestimate dF^*/dQ_a^* for the following reasons, of which some have previously been discussed by Ramanathan and Briegleb (1980):

- Cirrus clouds are relatively transparent in the 10.5–12.5 μm wavelength interval, with the actual opacity being strongly dependent upon cloud thickness, whereas they become quite black at longer wavelengths (Haurwitz and Kuhn, 1974). Thus, it is possible that the NOAA-NESS data tend to underestimate variations in F due to cirrus cloud variability, whereas Cess and Ramanathan (1978) have suggested that variability in outgoing infrared radiation, due to changes in cloud amount, is due principally to variability in cirrus clouds.
- Cloud albedos decrease at wavelengths > 0.7 μm, such that a 0.5–0.7 μm measurement would, relative to a solar average, increase the contrast between cloudy-sky and clear-sky albedos. This effect would be even more pronounced over surfaces covered by vegetation, since the albedo of most types of vegetation increases dramatically for wavelengths > 0.7 μm. To cite some examples, from 0.7 to 0.8 μm the albedo increases from 0.06 to 0.50 for Sudan

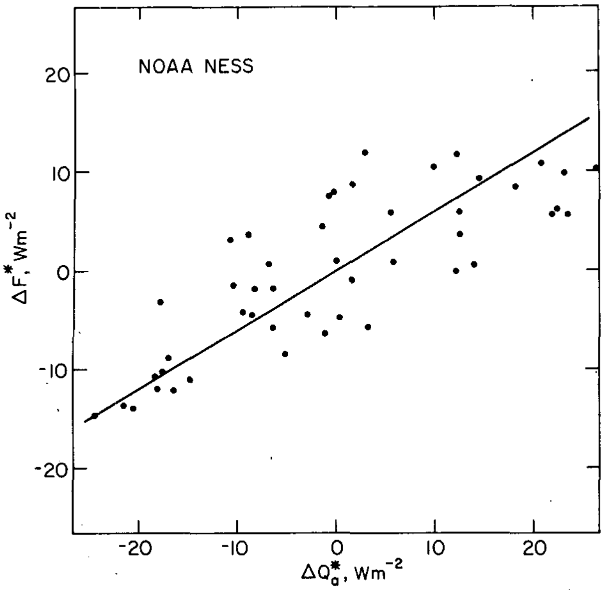

FIG. 3. As in Fig. 1 except for the NOAA-NESS data set.

grass, from 0.12 to 0.40 for alfalfa, and from 0.08 to 0.30 for clover (Mironova, 1973).

Note that the above items act in the same direction, that is, they would reduce dF^*/dQ_a^*, as is consistent with our results using the NOAA-NESS data as compared to the two other data sets.

But the above list is by no means complete. For example, additional items include the effects of both Rayleigh scattering and solar absorption by atmospheric water vapor on the contrast between cloudy-sky and clear-sky albedos, as given by the NOAA-NESS data versus a solar average. It is difficult to predict whether the net effect would be an increase or a decrease in contrast.

A further problem, which pertains to all three data sets, is the fact that the albedo measurements do not correspond to diurnal averages. Ellis and Vonder Haar employ data from a number of satellites with albedo measurements ranging in time from 0845 to 1440, but no attempt was made to weight these for a diurnal average. The two years of Nimbus 6 data, which were additionally incorporated by Campbell and Vonder Haar (1980), refer to local noon albedos, although they made an effort to empirically correct their albedos to a diurnal average. The NOAA-NESS data represent 0900 albedos.

4. Concluding remarks

The purpose of the present study has been to 1) illustrate intercomparisons, as well as potential problems, associated with empirical deductions of the relative roles of the albedo and infrared components

TABLE 3. Estimates of dF^*/dQ_a^* for individual latitude zones using Wald's method.

Latitude	Data set: Ellis and Vonder Haar	Campbell and Vonder Haar	NOAA-NESS
20–10°N	1.2	1.4	0.7
10°N–0°	0.9	1.6	0.5
0°–10°S	1.6	1.5	0.6
10–20°S	1.5	1.1	0.5
20°N–20°S	1.3	1.3	0.6

of cloudiness feedback, and 2) to emphasize the uncertainties in these estimates associated with using available earth radiation budget data sets. Our "most likely" estimates of dF^*/dQ_a^* indicate that, for cloudiness variability at low latitudes, the albedo modification is somewhat less than that of the infrared when we employ either the Ellis and Vonder Haar or Campbell and Vonder Haar data sets. But when use is made of the NOAA-NESS data set, the albedo modification is found to dominate over that of the infrared by nearly a factor of 2. We suggest that these differences might be attributable to the NOAA-NESS data being derived from narrow spectral measurements. We further emphasize that our results are intended solely for comparative purposes; it would be premature to interpret them in terms of a conclusion concerning the relative roles of the albedo and infrared components of cloudiness feedback.

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