

THESIS

CLOUDINESS, THE PLANETARY
RADIATION BUDGET, AND CLIMATE

Submitted by
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WE HEREBY RECOMMEND THAT THE THESIS PREPARED UNDER OUR
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ABSTRACT OF THESIS

CLOUDINESS, THE PLANETARY RADIATION BUDGET, AND CLIMATE

Satellite planetary radiation budget measurements from the Nimbus 3 satellite for four semi-monthly periods along with a 29 month composite of measurements from six satellites are applied in a quantitative study to evaluate the effect of cloudiness on the planetary radiation budget. Annual and seasonal results are expressed as zonal, hemispherical and global mean values.

The results show that for the planet as a whole the effect of "present day" clouds in reducing the absorbed shortwave flux is larger than their effect in reducing the longwave emitted flux. The difference between the two effects is significantly larger over oceans than over land. Similarly, the sensitivity to changes in cloud amount is greater in the shortwave absorbed flux than in the longwave flux emitted to space. It was also shown that the presence of clouds act to reduce the amplitude of the annual variation of the global planetary net radiation budget.

One may hypothesize from the results of this study and the works of others that a uniform increase in global cloud amount (in the absence of changes in cloud top height, cloud albedo, other atmospheric constituents, and

vertical temperature lapse rates) will decrease the global mean surface temperature until radiative equilibrium is restored.

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TABLE OF CONTENTS

	<u>Page</u>
ABSTRACT	iii
ACKNOWLEDGEMENTS	iv
LIST OF TABLES	vi
LIST OF FIGURES	viii
1. INTRODUCTION	1
2. DATA COLLECTION, PROCESSING AND ANALYSIS	5
2.1 Flux Estimates Before Satellite Measurements	6
2.2 Satellite Measurement of Planetary Radiation Flux	6
2.3 Cloud Free Flux	10
2.3.1 Cloud Free Albedo	12
2.3.2 Cloud Free Emitted Flux	25
3. CLOUD EFFECTS ON THE PLANETARY RADIATION BUDGET	36
3.1 Zonal Mean Cloud Effect	38
3.2 Global and Hemispherical Mean Cloud Effect	44
3.3 Seasonal Variations in the Cloud Effect	47
3.3.1 Zonal Variations	47
3.3.2 Global Variations	50
3.3.2.1 Indirect Analysis Method	51
3.3.2.2 Direct Analysis Method	66
4. SENSITIVITY TO A UNIFORM CHANGE IN CLOUD AMOUNT	74
4.1 Zonal Sensitivity	75
4.1.1 Cloud Amount	75
4.1.2 Absorbed Flux Sensitivity	78
4.1.3 Emitted Flux Sensitivity	80
4.1.4 Net Flux Sensitivity	83
4.1.5 Atmospheric Transport Sensitivity	85
4.2 Global and Hemispherical Sensitivity	96
5. CONCLUSIONS AND RECOMMENDATIONS	107
REFERENCES	110
APPENDIX A	116
APPENDIX B	119
APPENDIX C	127

LIST OF TABLES

	<u>Page</u>
2.1 Chronological list of earth-orbiting satellites from which the radiation measurements were taken.....	8
2.2 Uncertainty in mean monthly net radiation.....	9
2.3 Nimbus 3 MRIR semi-monthly data sets.....	11
2.4 Cloud-free albedo.....	24
3.1 Cloud effect in net flux (ΔN) for land-plus-ocean.....	46
3.2 Cloud effect in net flux (ΔN) for oceans.....	46
3.3 Cloud effect in net flux (ΔN) for semi-monthly periods..	67
3.4 Cloud effect in albedo ($\Delta \alpha$) for land-plus- ocean.....	67
3.5 Cloud effect in emitted flux (ΔI) for land-plus-ocean...	67
3.6 Cloud effect in emitted flux (ΔI) for oceans.....	70
4.1 Zonal values of $\frac{\partial N}{\partial Ac}$	89
4.2 Sensitivities associated with removal of stratus-type clouds from the Eastern Pacific Ocean in the 15N-35N latitude zone.....	92
4.3 Sensitivity of the atmospheric energy transport out of the 0-30N latitude zone to interannual variations of cloud amount in the 0-15N latitude zone.....	95
4.4 Annual global statistics (90N-90S).....	96
4.5 Summary of satellite derived statistics.....	97
4.6 Annual sensitivities (65N-65S).....	98
4.7 Annual global sensitivities.....	99
4.8 Annual global longwave sensitivity derived from models	104
A.1 Directional reflectance relative to the value at solar zenith angle of zero degrees for the Nimbus 3 ocean and cloud models.....	118

	<u>Page</u>
B.1 Regression coefficients (A) and standard error of estimate (e) for equation (B.1).....	121
B.2 Infrared spectral channels of the MRIR of the Nimbus 3 satellite.....	121
B.3 Spectral broadband empirical, broadband model and spectral model radiances calculated for a model atmosphere.....	123
C.1 Mean monthly global planetary radiation budget.....	128

LIST OF FIGURES

	<u>Page</u>
2.1 Albedo and cloud-free albedo over land-plus-ocean.....	16
2.2 Albedo and cloud-free albedo over oceans.	17
2.3 Photograph from aircraft showing partial cloud cover above Greenland's ice sheets on 22 June 1975 at 1600G.M.T.	20
2.4 Annual albedo and cloud-free albedo over land-plus- ocean, ocean and land.	22
2.5 Longwave flux and cloud-free longwave flux over land- plus-ocean.	28
2.6 Longwave flux and cloud-free longwave flux over oceans...	29
2.7 Annual longwave flux and cloud-free longwave flux over land-plus-ocean, ocean and land.	33
3.1 Cloud effect on net flux for land-plus-ocean.	39
3.2 Cloud effect on net flux for oceans.	40
3.3 Annual cloud effect on net flux over land-plus-ocean, ocean and land.	41
3.4 Time-latitude analysis of zonal mean profiles from the 29 month set: (a) albedo, (b) longwave flux, (c) net flux and (d) cloud amount.	48
3.5 Annual variation in global net flux.	52
3.6 Annual variation in global albedo.	56
3.7 Annual variation in global longwave flux.	59
3.8 Annual variation in global net flux with just an annual wave in longwave flux.	65

	<u>Page</u>
3.9 Annual variation in net flux and in the cloud effect on net flux for four semi-monthly periods.	73
4.1 Cloud amount from TIROS nephanalyses.	77
4.2 Absorbed flux sensitivity to changes in cloud amount.	79
4.3 Longwave flux sensitivity to changes in cloud amount.	81
4.4 Net flux sensitivity to changes in cloud amount.	84
4.5 Atmospheric mean meridional energy flux divergence for the 30N-to-90N polar cap.	87
4.6 Atmospheric mean meridional energy flux divergence for the equator-to-30N latitude zone.	87
4.7 Atmospheric mean meridional energy flux divergence for the equator-to-90N polar cap.	88
4.8 Longwave flux emitted by the earth-atmosphere system as a function of cloud amount for various heights of radiometrically black clouds.	101

1. INTRODUCTION

The objective of this study is to quantitatively determine the effect of clouds on the planetary radiation exchange with space. This effect is examined on spatial scales from zonal mean to global and temporally from months to an annual mean.

This study was undertaken to satisfy a need for quantitative information on the role of cloudiness in maintaining or changing climate regimes. Present climate study models cannot adequately incorporate clouds and cloud effects in an interactive mode because of the lack of observational and diagnostic studies of cloudiness at the larger planetary scale. Schneider (1972) clearly stated the problem... the global distribution of cloudiness is itself a consequence of the general circulation which, in turn, is driven by virtue of the latitudinal radiative imbalance between the incoming solar (minus the reflected flux) and the outgoing atmospheric IR (emitted longwave flux) radiation. Thus, in a study of the sensitivity of the global climate to changes in cloudiness, it is first necessary to determine the initial magnitude of such changes in cloudiness on the radiation balance. Simpson (1928) discussed the same problem but more from a point of view of required changes in cloud amount to offset changes in incoming solar radiation. Studies of this type have been more recently undertaken by Schneider (1972) and Cess (1976). Schneider found that a uniform change in global

cloud amount has a significant effect on the globally integrated planetary radiation budget while Cess (1976) found it to have an insignificant effect. Cess went on to conclude that clouds may not provide a significant climate feedback mechanism.

Cloudiness has traditionally been specified by cloud type, amount and base height above the ground. Even if these cloudiness specifications were exactly determined over the entire planet, they would not be sufficient for determining radiative properties of the clouds. Additional information needed are the cloud top height and temperature, the absorptance, emittance and reflectance or transmittance of the clouds. These are minimum requirements. Specification of microphysical processes and properties within the clouds would aid the specification of some of the cloud properties. The measurement of all aforementioned cloud properties is difficult to perform at the small scale (mesoscale 2 to 200 km). The technology has not been and may never be implemented or totally developed to make such measurements on the global scale. Yet such information is needed for the specification of the radiative properties of clouds.

The present study short-circuits the very detailed measurements of individual cloud properties by determining with satellite measurements the bulk effect of cloudiness on the planetary radiation budget. By bulk is meant a

taking together of all the properties without discriminating the effect of any one property. The importance of the present study is in the need for the information, the simplicity and accuracy of the method, and that the measurements are now available to quantitatively implement the method.

The joint Organizing Committee (JOC) of the World Meteorological Organization (WMO) and the International Council of Scientific Unions (ICSU) in a plan for advancing our knowledge of the climate and its variations (GARP No. 16, 1975) proposed observational efforts in three categories. Basically the three categories were directed at describing processes relevant for modeling, testing and validating models, and long-term monitoring programs of diagnostic variables for the detection of possible future climatic variations. A requirement for knowing the effect of the appearance and disappearance of various forms of extended cloudiness on the radiation budget is identified in their report (Chapters 5 and 6). Studies of the present type have applicability to all three categories but are most relevant to the latter two categories in testing and validating models, and monitoring diagnostic variables of a changing climate. Besides the immediate application to climate study models for initialization, constraint and validation of model output, the results of the study will be available to validate the radiative effect of clouds from detailed measurements as they become available.

The planetary radiation exchange with space is partitioned into incoming solar flux, reflected and scattered solar flux back to space, and emitted terrestrial flux. The incoming minus the other two component fluxes is the net flux. The components, other than solar flux, have been determined from measurements by earth orbiting satellites and are indicative of the fluxes at satellite height, or at some lower height above which the earth's atmosphere has no radiative properties significant to this study.

The effect of cloudiness on the radiation exchange with space is determined by measurements of the radiation budget components from space of a cloudy and cloud-free planet earth. The difference between the two states is the cloud effect for the present climate. Given the cloud effect for our present cloud cover, an extrapolation is made to other cloudiness states of both lesser and greater cloud cover.

The development of the satellite measured radiation budget components is discussed followed by development of the method and its application to zonal and global space scales. Finally, a discussion of the sensitivity of the planetary radiation budget to changes in cloudiness is given so that its significance as a climate feedback mechanism may be assessed.

2. DATA COLLECTION, PROCESSING AND ANALYSIS

Radiative energy flux between space and planet Earth, including its atmosphere, has been determined both through model calculations and direct measurement by earth orbiting satellites. Previous model calculations which normally relied on ground base measurements, some aircraft measurements and theory gave reasonable values for the fluxes. Since the middle 1960's earth orbiting satellites have been measuring the fluxes. A considerable number of satellite measurements collected since then have permitted the development of a planetary radiation budget climatology.

The planetary net radiation budget is composed of incoming solar flux, reflected solar flux to space and emitted terrestrial flux. It is mathematically defined at top of the earth-atmosphere boundary as:

$$N = (1-\alpha) S - I \quad (2.1)$$

where:

α = the ratio of reflected solar flux summed over all angles to the incoming solar flux

S = incoming solar flux based on a solar constant (SC) of 1360 w m^{-2} (after Drummond et.al. 1968)

$$S = SC \left(\frac{\bar{d}}{d}\right)^2 \cos \xi$$

I = emitted flux by earth-atmosphere system.

2.1 Flux Estimates Before Satellite Measurements

Prior to the availability of a planetary radiation budget climatology from satellite measurements, planetary heat budget studies required estimates of the radiation budget. Several extensive works published then computed the radiative energy fluxes by the application of theory to a combined set of ground based and aircraft measurements.

Two such studies include that of London (1957) for just Northern Hemisphere and that edited by Budyko (1963) for the entire planet. In both studies, the cloud effect on solar and terrestrial fluxes had to be incorporated into cloud-free planetary calculations. These cloud effects were estimated from ground based observations of cloud amount and type and aircraft and ground measurements of cloud radiative characteristics.

2.2 Satellite Measurement of Planetary Radiation Flux

Data from satellite measurements of reflected short-wave and emitted longwave radiances have been compiled into global radiation budget climatologies. They are in chronological order as follows: Bandeen et al. (1965), Raschke and Pasternak (1967), Winston and Taylor (1967), Vonder Haar and Suomi (1971), Raschke et al. (1973a), Vonder Haar and Ellis (1974), and Ellis and Vonder Haar (1976). The last two publications are the most extensive including as many as six separate satellite measuring systems. The last publication, Ellis and Vonder Haar

(1976), is a compilation of 29 months of measurements summarized into mean month, season and annual zonal mean-meridional profiles. This satellite data base is used for the present study.

The 29 month set is a selective compilation including many, but excluding some of the satellite measurements available at that time. Table 2.1 is a list of all measurements comprising the set. Analyzed measurements from the later two satellites, ITOS 1 and NOAA 1 are from Flanders and Smith (1974). All measurements are from polar orbiting satellites in sun synchronous orbits. Local daylight sampling time at the equator is also given in the table. There is considerable daytime diurnal variability in sampling with Experimental in the morning, Nimbus 2 and 3 near noon and ESSA7, ITOS1 and NOAA 1 in the afternoon.

The publication by Ellis and Vonder Haar (1976) gives a complete discussion of the data set and treats in detail uncertainties in the measurements. Uncertainty in the monthly means of net radiation summarized from that publication are shown in Table 2.2. The uncertainties are generally less than 10wm^{-2} with smallest values near polar regions of the winter hemisphere.

In the present study measurements from Nimbus 3 are used separately and as a part of the 29 month set. A complete description of the medium resolution infrared radiometer (MRIR) aboard the satellite is given in the NIMBUS III USER'S GUIDE (1969). It is a 5 channel

Table 2.1 Chronological list of earth-orbiting satellites from which the present radiation measurements were taken. The approximate local time at which each satellite crossed the equator during daylight hours is given in parentheses. EX = Experimental, N2 = Nimbus 2, N3 = Nimbus 3, E7 = ESSA 7, I1 = ITOS 1 and NO1 = NOAA 1.

MONTH	YEAR								SAMPLE SIZE
	1964	1965	1966	1967	1968	1969	1970	1971	
Jan		EX (10:30)				E7 (13:00)	N3 (11:30)		3
Feb		EX (10:35)				E7 (14:30)			2
Mar		EX (10:40)				E7 (14:30)			2
Apr						N3 (11:30)	I1 (15:00)		2
May			N2 (11:30)			N3 (11:30)	I1 (15:00)	NO1 (15:00)	4
Jun			N2 (11:30)			N3 (11:30)	I1 (15:00)		3
Jul	EX (08:30)		N2 (11:30)			N3 (11:30)			3
Aug	EX (08:55)					N3 (11:30)			2
Sep	EX (09:15)								1
Oct	EX (09:40)				E7 (14:30)	N3 (11:30)			3
Nov	EX (10:05)				E7 (14:30)				2
Dec	EX (10:30)				E7 (14:30)				2
ANNUAL	6	3	3	0	3	9	4	1	29

Table 2.2 Uncertainty in mean monthly net radiation (wm^{-2})

Latitude	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
North 85°	4	5	7	8	8	11	12	14	8	5	6	5
75	5	8	8	9	8	10	11	13	9	9	6	6
65	8	8	7	9	8	9	9	9	10	7	9	8
55	6	9	8	9	7	9	9	10	11	8	10	9
45	7	8	8	10	7	9	9	10	13	7	11	10
35	7	9	9	10	8	9	9	11	14	8	9	8
25	8	10	10	11	8	9	9	11	15	8	10	9
15	8	10	10	11	8	9	8	10	14	8	10	10
5	8	10	10	11	7	8	8	10	14	8	10	10
South 5	8	10	10	10	7	9	8	10	15	9	10	10
15	9	10	10	11	8	9	9	11	15	9	11	11
25	9	10	10	10	7	8	8	10	15	9	11	11
35	9	10	10	10	7	7	7	9	13	8	11	11
45	9	10	9	9	6	7	8	8	12	8	10	11
55	9	10	9	8	5	10	7	10	11	7	11	11
65	9	10	8	7	10	5	5	6	9	7	11	12
75	10	10	7	9	4	4	4	4	5	6	12	14
85	10	14	9	6	4	4	3	4	4	7	12	14

instrument with one shortwave channel (0.2 to 4.8 μm) and four longwave channels spanning the spectral range 6.0 to 23.9 μm). Total outgoing longwave radiance was computed by Raschke et al. (1973) through application of a multiple least-square regression formula to radiances in each of the four longwave channels. The field of view (FOV) of the instrument varied from 50 km of geocentric arc distance at satellite nadir to 110 km at an angle of 40° from nadir. The measurements were mapped to polar stereographic and mercator projections at 500 km to 200 km earth located grid spacing. This gridding allows grid scale to global scale studies.

Wide angle measurements by flat plate radiometers comprise 19 months of the 29 month set. Measurement resolution at the earth's surfaces varies with height of the satellite. For satellite heights in this set, the ground resolution of the wide angle sensors is 5900 km to 7770 km of geocentric arc distance in diameter. These data have been mapped on a 10 degree latitude-longitude grid. It is preferable to use these data for spatial scale studies of the 6000 km to global scale.

2.3 Cloud Free Flux

The satellite data set is composed of individual measurements of flux from cloud-free and cloudy atmospheres. To assess the effect of cloudiness on the components of the planetary radiation budget it is necessary to know the

flux components over a cloud free earth-atmosphere system. Cloud free reflected and emitted flux may be determined from satellite measurements. The small grid scale mapping, small FOV, and daily sampling during all seasons of the year by the Nimbus 3 MRIR permit a cloud-free determination. These gridded data are archived as 10 semi-monthly sets (table 2.3).

Table 2.3 - Nimbus 3 MRIR semi-monthly data sets*

April 16 to 30, 1969	July 1 to 15, 1969
May 1 to 15, 1969	July 16 to 31, 1969
May 16 to 31, 1969	August 1 to 15, 1969
June 1 to 15, 1969	October 3 to 17, 1969
June 16 to 30, 1969	January 21 to February 3, 1970

Four semi-monthly sets were selected from the 10 sets, one within each season of the year, to approximate the variation of the planetary radiation budget with season. The four periods having the best temporal and spatial sampling are: May 1-15, July 15-31, October 3-17, 1969 and January 21-February 3, 1970. The last period, representing winter in the Northern Hemisphere, is inadequate poleward of 45N due to large data void regions, particularly over the North Pacific Ocean (Raschke et al., (1973b) shows global maps).

* From Raschke, et al, (1973b)

2.3.1 Cloud Free Albedo

Cloud free albedoes may be determined at each horizontally spaced grid area by selecting the smallest measured albedo value from a time series of measurements at that grid area. The concept assumes that a cloudy grid area has a higher albedo value than does a grid area which is cloud-free. This minimum method of obtaining cloud-free measurements in the shortwave portion of the solar spectrum has previously been applied by Coburn (1971) and Miller and Feddes (1971) to satellite measured brightness values. Coburn found that a time series of 7 days was sufficient to observe most grid areas as cloud-free. In the present study a time series extending the duration of each semi-monthly period has been used.

One may conceive of partly cloudy cases with the albedo as seen from space being less than the cloud-free albedo. Two such cases could occur with scattered clouds over wet soil or a vegetated region versus cloud-free, dry soil or desertificated soil. Instances of this type are considered to be within the noise of the cloud effect study.

The minimum albedo is not a cloud-free albedo at all grid areas because of residual cloudiness not removed by the minimum method. Persistent stratus cloudiness over oceans east of subtropical high pressure centers are not always removed. Chen (1975) using data from a higher spatial resolution instrument, the NOAA 4 scanning radiometer,

also observed this. Some grid areas in regions of frequently occurring extratropical cyclones, such as the Aleutian Islands and Icelandic region, are not observed cloud-free during a semi-monthly period. Furthermore, in the intertropical convergence zone with frequent and persistent cloudiness, which are often smaller than the spatial gridding of Nimbus 3 MRIR measurements, clouds are not removed from all grid areas by minimum compositing.

To "clean-up" minimum albedo maps so as to obtain cloud-free zonal mean meridional profiles, further processing was required. The initial data reduction of the MRIR measurements required application of models to account for anisotropic reflections from various surfaces in the FOV (Raschke et al, (1973)). There were three models called cloud-land, snow, and ocean models. All three models accounted for the directional dependence of reflectance on solar zenith angle and bi-directional reflectance dependence on position of observer and solar zenith angle. A criterion for applying the ocean model, a model applicable to cloud-free ocean reflectances only, was that the measured reflectance be 0.10 or less. Presumably, one should be able to use the solar zenith angle dependence part of this model (the directional reflectance part) to compute an upper limit reflectance cut off value. All reflectance values above the cut off are cloudy values, and thus should be eliminated from the minimum albedo maps. (See Appendix A for ocean and cloud-land models.)

Conover (1965) determined cloud-free albedoes over the Gulf of Mexico to be 0.09 and over the Pacific Ocean to be 0.07. The values were taken as an average from the TIROS VII satellite measurements at solar zenith angles between 3 and 42 degrees. If 0.09 is taken as the instantaneous directional reflectance at a zero solar zenith angle ($\xi=0$), then application of the ocean directional reflectance model will overestimate the daily mean cloud-free albedo. The model calculations give a daily average albedo of 0.115 at the latitude of solar declination, which is significantly larger than Conover's 0.09. However, using the 0.09 value as an upper limit cloud-free reflectance value (at $\xi=0$) allows for varying ocean surface characteristics which may have larger planetary albedoes than an ocean with a smooth and homogeneous surface. It also allows for uncertainty in the ocean model itself. For the most part, albedoes in the tropics of the minimum albedo maps, before application of the upper limit cutoff value, were 0.08 to 0.10. The lowest values were observed in May over the North Indian Ocean as 0.06. This lowest value is at the limit of what molecular scattering theory predicts for a daily mean shortwave albedo over a black background with just molecular scattering (after Joseph, 1971).

Over land surfaces a simple reflectance model was not used to eliminate residual cloudiness because the spatial and temporal variability of cloud-free albedo is not directly dependent on the solar zenith angle. Results of

the combined ocean and land cloud-free albedoes have been reported on before without the reflectance cut off value over oceans by Vonder Haar and Ellis (1975) and with the cut off value by Ellis (1975).

Zonal mean profiles of albedo and cloud-free albedo for land-plus-ocean, and ocean are shown in Figures 2.1 and 2.2, respectively. For the most part, the large spike in albedo in the intertropical convergence zone (ITCZ) does not appear in the cloud-free albedo of the ocean profile; over the oceans the ITCZ cloudiness effect has been completely removed. Albedo and cloud-free albedo is smaller over the ocean than over the land-plus-ocean. However, differences between cloudy and cloud-free albedo, $\Delta\alpha$, are largest over the oceans, particularly in the middle to high latitudes of each hemisphere. The land surface being brighter than the oceanic surface gives a smaller difference between the cloudy and cloud-free land-plus-ocean profiles than is observed in the ocean profile.

Positions of the mean zonal albedo maxima in the tropics are in agreement with mean brightness composites of Kornfield and Hasler (1969) and relative cloud cover derived from satellite brightness of Miller and Feddes (1971). A narrow spike in albedo at 6N for May is from cloudiness of the ITCZ at 8N in the Pacific and 5N in the Atlantic. In July the peak is located at 9N with the base of the peak spreading north to 30N and south to the equator. This July albedo feature is characteristic of the cloudiness

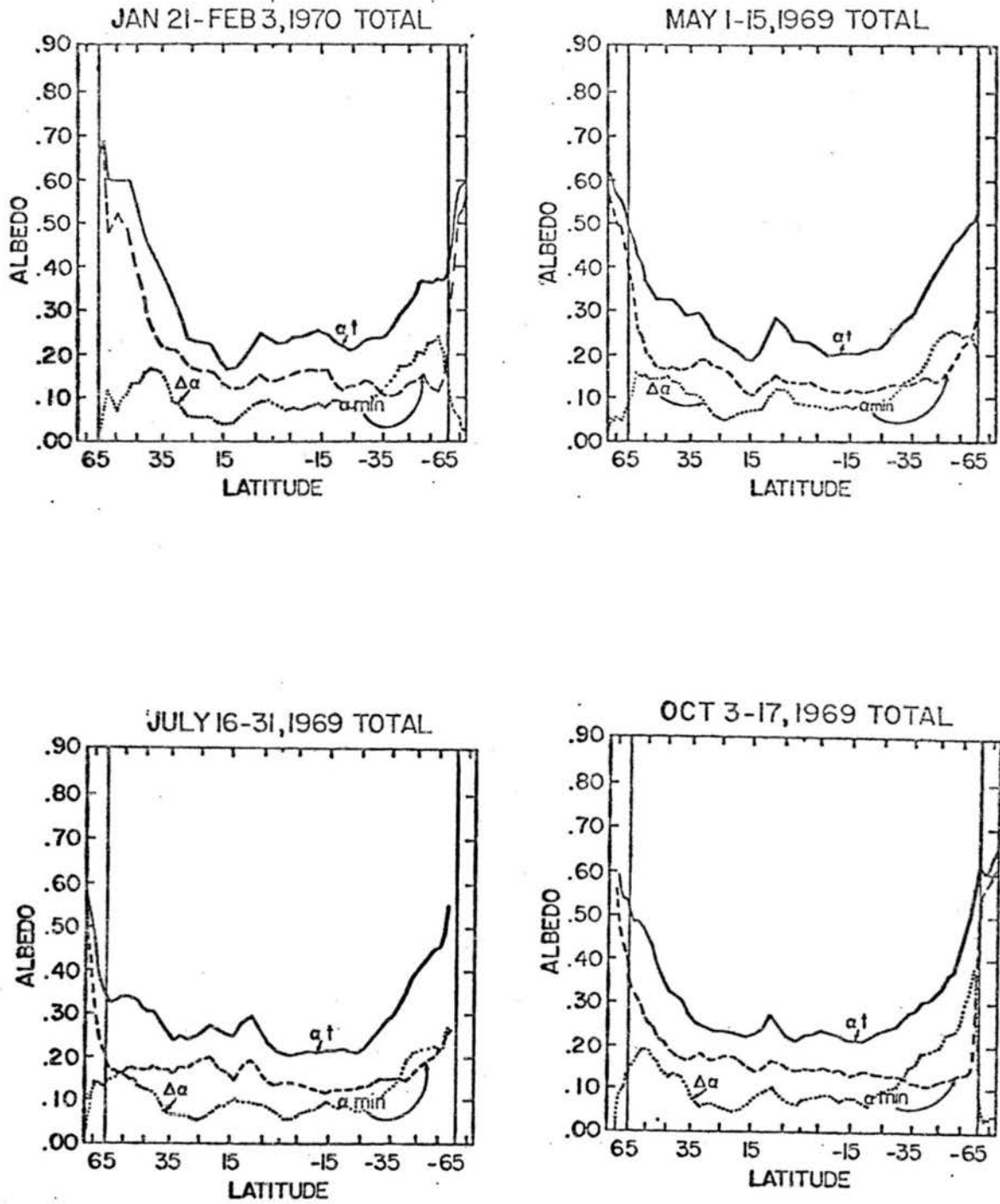


Figure 2.1.: Albedo (α_t) and cloud free albedo (α_{min}) over land-plus-ocean.

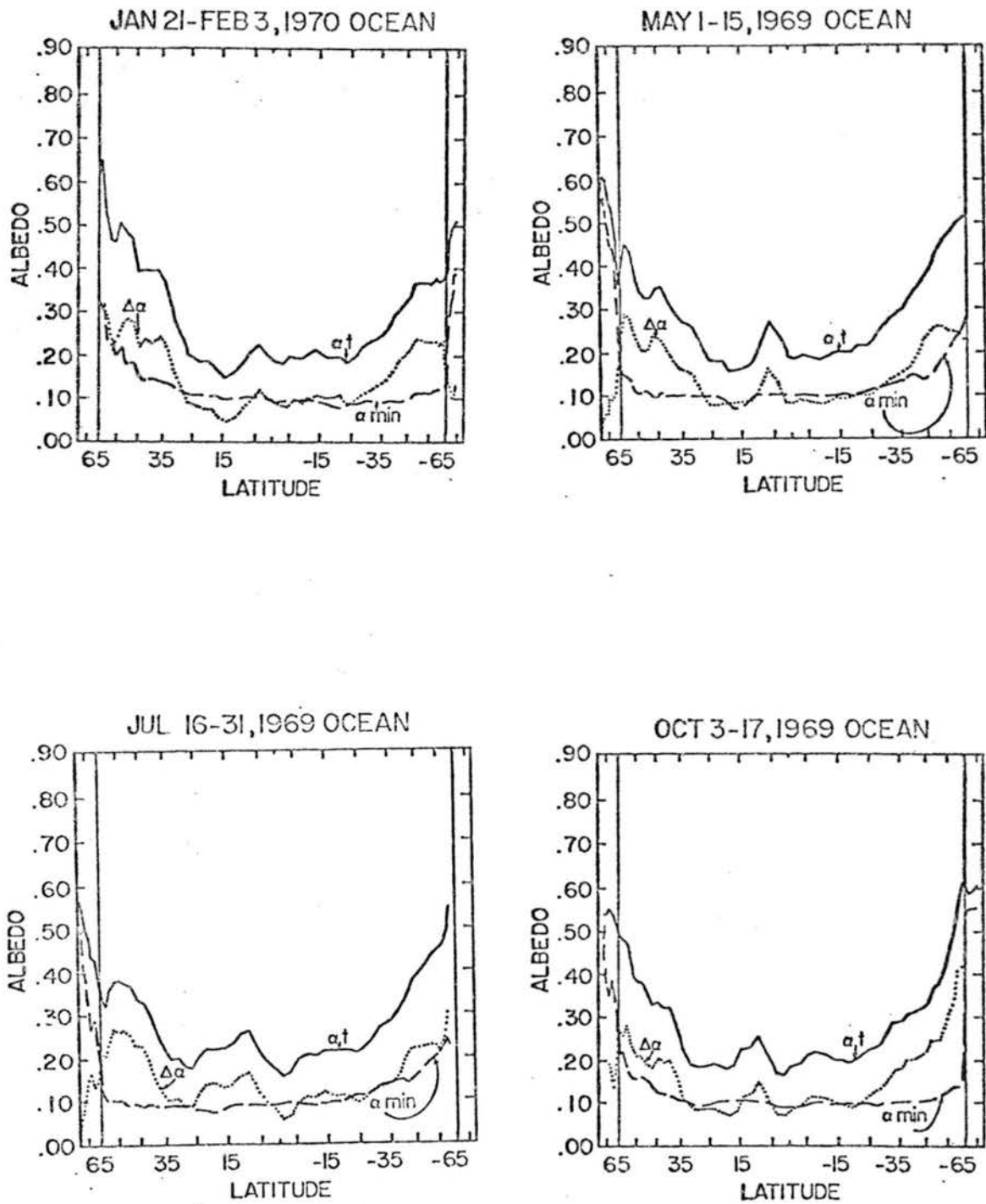


Figure 2.2.: Albedo (α_t) and cloud-free albedo (α_{min}) over ocean.

associated with the ITCZ at 9N in the Pacific, and at 10N in the Atlantic, the Southwestern Asian Monsoon and the stratus off of the southwestern coast of North America. The south to north albedo increase in the Southern Hemisphere tropics is also due to stratus off of the west coast of continents and a cloud band oriented northwest-to-southeast off of the northeast coast of Australia. In October, the predominant peak remains at 9N with a secondary ridge over the oceans at 9S. The secondary ridge is a combination effect of stratus off of the west coasts of South America and Africa, a cloud band off of the northeast coast of Australia as noted before, and the Indian Ocean ITCZ at approximately 5S. Finally, in the January-February period, the difference in the cloudy and cloud-free albedo is less in the total zonal profile over the Northern Hemisphere, than over the Southern Hemisphere, due to a brighter underlying surface. Again the bright peak of the tropics is in the Northern Hemisphere at 5N. However, a generally elevated albedo zone extends from 5N southwards in agreement with an ITCZ at 7N in the Pacific, 5N in the Atlantic, 5S-to-10S in the Indian Ocean and Malaysia, and the cloud band off of the Northeastern coast of Australia. The tropical cloudy albedo does not exceed 25 percent in this last period, where as in three other periods the highest values were 28 percent in May and July, and 26 percent in October.

The minimum albedo method as a tool for obtaining a cloud-free albedo is not valid over bright snow and ice surfaces at large solar zenith angles. As was shown by Hauth and Weinman (1968), using visible channel measurements from the Nimbus II MRIR in June 1966 over the Greenland ice cap, bidirectional reflectances from thick clouds were less than from the background surface for relative azimuth angles between sun and satellite greater than 90 degrees. The difference increased at larger satellite zenith angles.

The bidirectional reflectance from clouds has been observed to be less than the background ice cap even at relative azimuth angle less than 90 degrees. Figure 2.3 is a photograph taken from an aircraft over Southern Greenland within one hour of local noon on 22 June, 1975.¹ The slope of the ice cap in the area of the photograph is toward the sun and observer. The relative azimuth angle between sun and observer is approximately 30 degrees. The clouds cast a shadow which are the darkest areas; clouds are brighter; the background ice sheet is brightest.

In latitude zones poleward of 65 degrees in each hemisphere a snow model was applied during the initial data reduction of all Nimbus 3 measured reflectances greater than 0.50 (Raschke et al., 1973b). That model assumed diffuse reflection over bright snow and ice surfaces.

1 Photograph taken by Dr. Stephen Cox, Department of Atmospheric Science, Colorado State University.

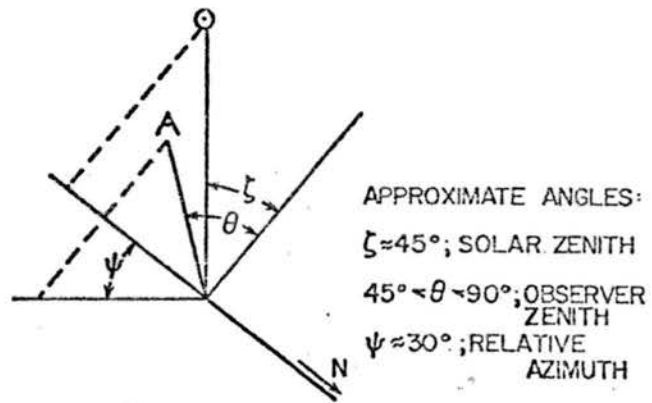
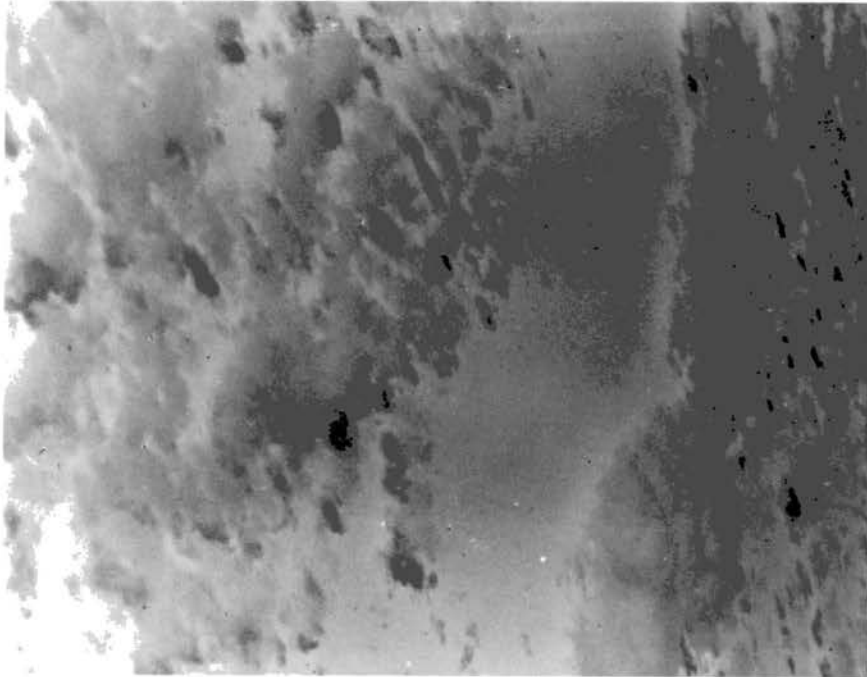


Figure 2.3.: Photograph from aircraft showing partial cloud cover above Greenland's ice sheets on 22 June 1975 at 1600 GMT.

Cloud reflectances were not distinguishable from background reflectances. If such a model could isolate clouds over the bright surface, it is quite likely that these cloud albedoes would, after angular corrections, be lower than the cloud-free albedoes. Thus all cloud-free albedoes poleward of 65 degrees in latitude in each hemisphere derived by the minimum method are not to be considered as cloud-free albedoes.

Annual profiles of albedo and cloud-free albedo were computed from the four semi-monthly profiles by assuming each period to be representative of a season and weighting it in the annual average by the mean season solar insolation. Annual profiles for total (land-plus-ocean), ocean and land from Nimbus 3 data are shown in Figure 2.4. The mean position of the cloudiness associated with the ITCZ at 7N-to-10N is apparent in both the total and ocean albedo profiles. An albedo of 23 percent in the ITCZ of the ocean albedo profile and the larger area of land surface in the Northern Hemisphere contribute to a brighter Northern Hemisphere. Oceanic albedo is essentially the same in corresponding latitude zones of each hemisphere except at the latitude of the ITCZ.

The annual land albedo profile shows a general increase in cloudy albedo from 35S to 65N latitudes. The cloud-free land albedo profile shows a lesser albedo increase from south-to-north, except for the 15N-to-30N zone, which shows a larger albedo increase. This zone includes the Sahara

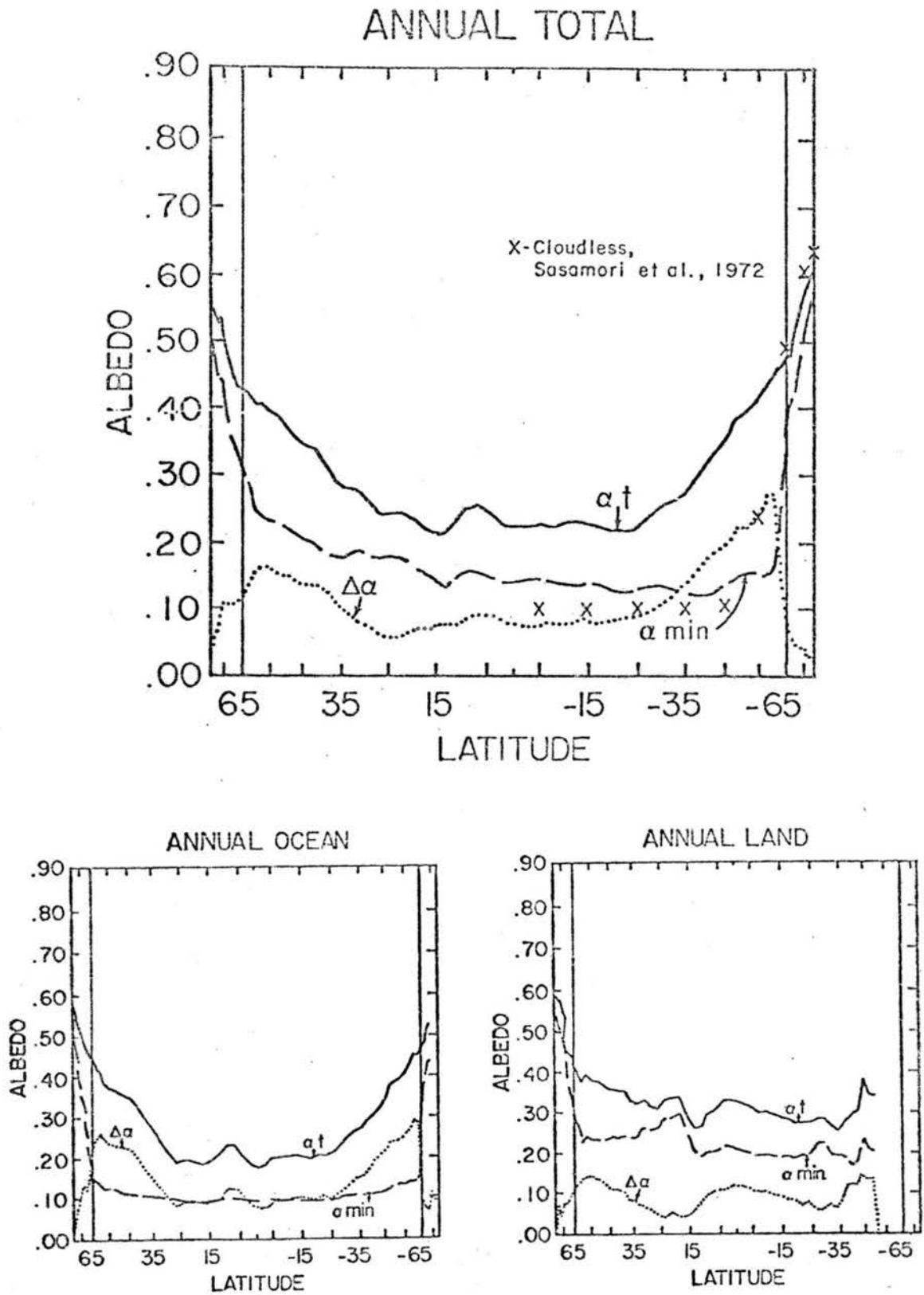


Figure 2.4.: Annual albedo (α_t) and cloud-free albedo (α_{min}) over land-plus-ocean, ocean and land.

Desert and Saudi Arabia, which are bright land surfaces. The latitude of largest cloud-free albedo occurs at 20N. The difference between cloudy and cloud-free albedoes over land ($\Delta\alpha$) is smallest in the 15N-to-25N zone, a zone of subtropical deserts which are nearly cloud-free, except during the period of the summer monsoon.

Cloudless albedo of the Southern Hemisphere from Sasamori et al (1972) is plotted on the land-plus-ocean albedo profile of Figure 2.4. The Nimbus 3 cloud-free albedo is greater by 2 to 3 percent from the equator-to-50S latitude zone. In the 50S-to-South Pole zone their values are considerably larger than the Nimbus 3 values. Undoubtedly this difference is due to differences between the two results in the equatorward extension of the sea ice around the Antarctica Continent and the albedo of the ice sheets.

Northern and Southern Hemisphere cloud-free albedoes from Nimbus 3 data are compared to cloudless albedo values of London (1957) and Sasamori et al (1972) for the Northern and Southern Hemisphere, respectively, in Table 2.4. The Nimbus 3 cloud-free values are larger in all seasons except for the Southern Hemisphere winter value which is identical to the value of Sasamori et al. Annual global cloud-free albedo from Nimbus 3 data is 16.9 percent as compared to a global value of 14.1 percent from their combined hemisphere values. London and Sasamori (1971) give an annual global value of 15 percent. This suggests that the 14.1

Table 2.4 - Cloud free albedo (percent)

	Winter	Spring	Summer	Fall	Annual
London, 1957 (90N-0)	17.3	15.0	13.5	13.4	14.7
Nimbus 3 (90N-0) (65N-0)	19.1 19.1	18.7 16.9	18.2 16.9	17.4 17.0	18.3 17.2
Sasamori et al (1972) 90S-0	13.8	14.0	14.8	11.1	13.6
Nimbus 3 (90S-0) (65S-0)	13.8 13.8	16.3 13.6	16.6 13.7	13.7 13.4	15.5 13.7

percent value should be considered as being too low. The difference between the Nimbus 3 data results and results of these other studies might be attributed to inadequate aerosol scattering to space and inaccurate land and ocean surfaces albedoes in the other studies, and residual cloudiness in the cloud-free albedoes of the present study. The true value of the annual global cloud-free albedo most likely falls between their value and the value of the present study.

2.3.2 Cloud Free Emitted Flux

The cloud-free emitted flux may be obtained in a fashion similar to cloud-free albedo by selecting the largest flux value at each grid area over each semi-monthly period. The concept assumes that without clouds a larger flux is measured from space with a standard atmosphere of decreasing temperature with increasing height above ground. Clouds are taken to be more opaque to terrestrial infrared and near infrared flux (3 to 30 μm) than other constituents are, and thus less terrestrial flux is emitted to space with clouds than without them.

It seemed prudent to test the empirical formula, used by Raschke et al (1973b) to extract total longwave radiance from the four Nimbus 3 IR channel radiances, for bias error with and without cloud cover. The four IR channels measure radiances in the 11 μm window, the 15 μm CO_2 absorption band and two water vapor absorption bands, one of which detects

emission largely from lower tropospheric water vapor. In the presence of middle and upper tropospheric opaque cloud cover the spectral composition of emitted flux to space will be different than cloud-free emission, principally because of the absence of emission from lower tropospheric water vapor. If the regression formula did not adequately consider this, then a bias error will be present in the difference between cloudy and cloud-free total longwave radiances.

A spectral IR model of Cox et al., (1976) was applied to a model atmosphere to compute channel radiances. The regression formula of Raschke et al., was applied to each model computed channel radiance to determine total broadband longwave radiance. The regression formula results were compared with broadband results of total longwave radiance. This comparison showed that the least squares regression formula handled the spectral radiances adequately both with and without middle tropospheric clouds, so that no significant bias error was introduced into the total longwave flux. (The model analysis appears in Appendix B).

An exception to using a "maximum" IR value at each grid point within a semi-monthly period is the case of clouds at the top of a lower tropospheric temperature inversion overlaying a cold surface. According to Vowinckel and Orvig (1970) such a condition is frequent and persistent over large areas of the Arctic particularly during winter. Large anticyclonic areas show a substantial

increase in radiation loss to space when clouds of this type are present because of higher temperature at cloud top than at the earth's surface. During the winter, in absence of appreciable solar flux, overcast low and middle clouds contribute to a larger radiation flux loss to space in the Arctic than during cloud-free conditions. As in the cloud-free albedo portion of this study, determining cloud-free IR by a "maximum" method will be limited to the latitude zone 65N-to-65S, inclusive.

Besides limiting the cloud-free zonal emitted flux profile to being the zonal average of "maximum" IR values at each grid area, all maximum IR values over ocean regions were excluded from the zonal mean whenever the ocean reflectance model, as applied to minimum albedoes, indicated the presence of clouds. This "throw out" criterion follows from a consideration that if, over some time interval, the minimum albedo is not a cloud-free albedo, then the maximum IR over the same time interval is not a cloud-free IR.

Zonal mean meridional profiles of cloudy and cloud-free IR are shown for each semi-monthly period in Figures 2.5 and 2.6 for land-plus-ocean and ocean, respectively. Notable in the May and July cloud-free profiles is the dip in flux values at the ITCZ. The dip in the July ocean profiles amounts to 25 wm^{-2} . This dip over the oceans at the mean position of the ITCZ may be a result of colder ocean surface temperature, or increased absorption and emission at a lower temperature due to abundant water

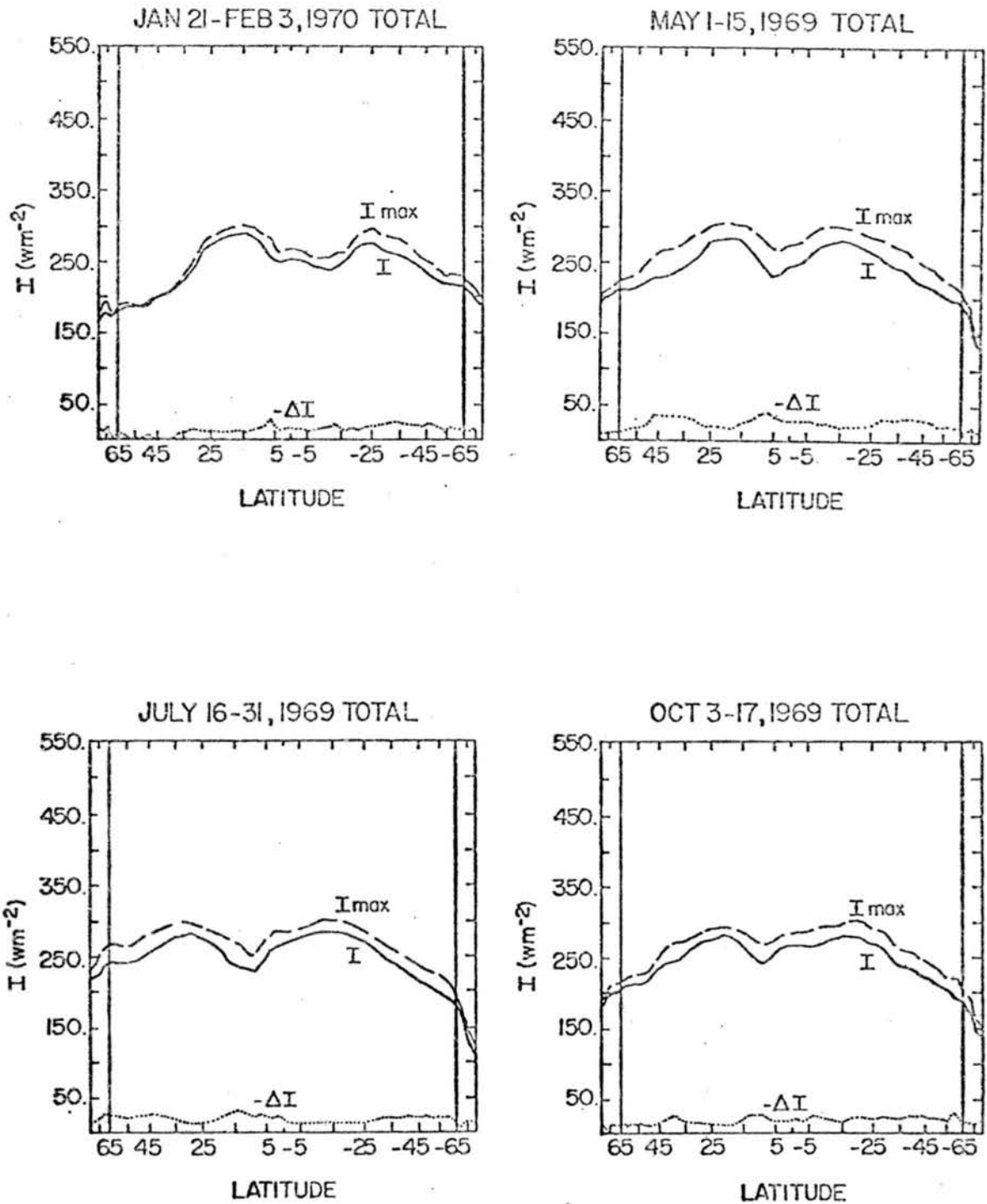


Figure 2.5.: Longwave flux (I) and cloud-free longwave flux (I_{max}) over land-plus-ocean.

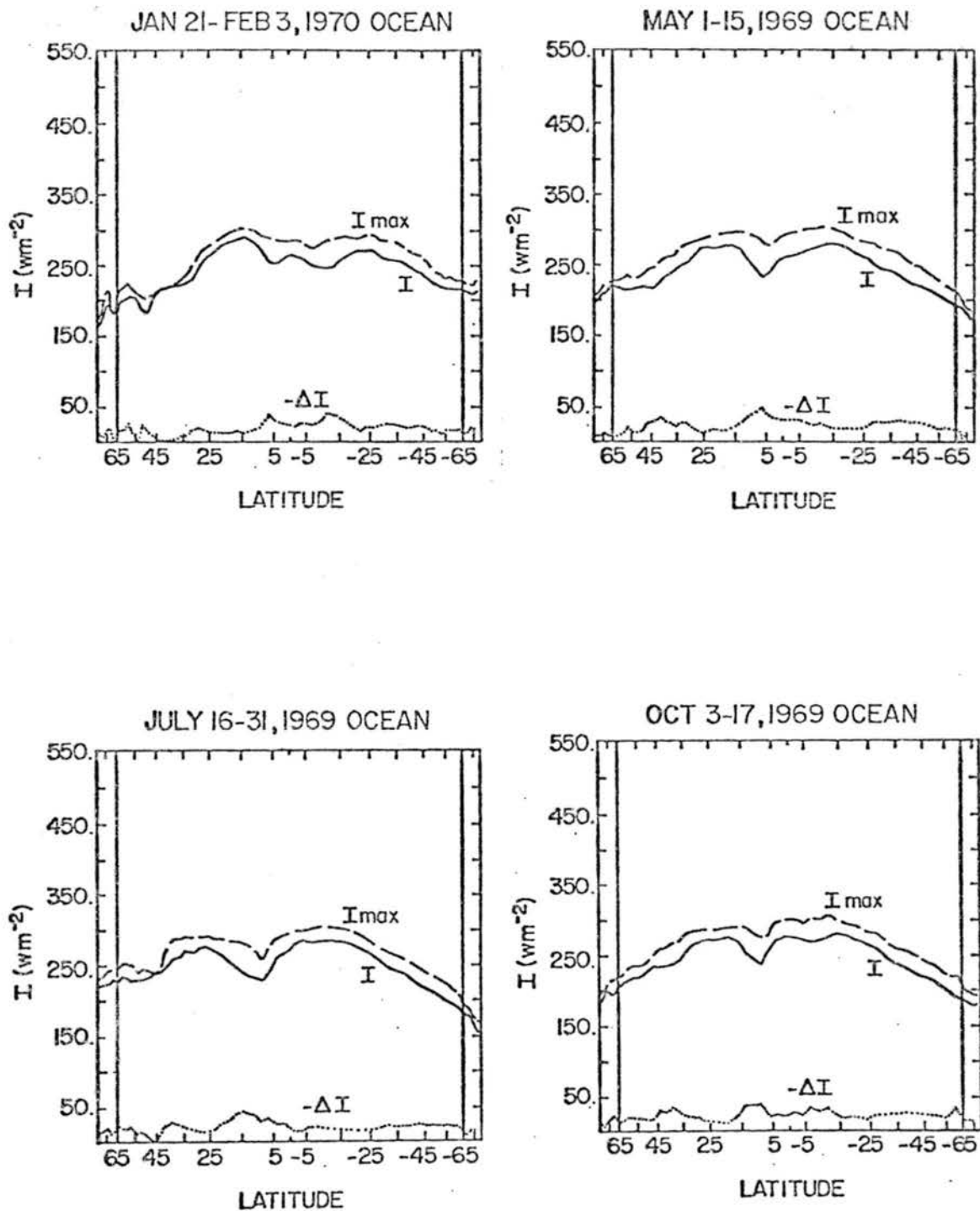


Figure 2.6.: Longwave flux (I) and cloud-free longwave flux (I_{max}) over oceans.

vapor, or cloudiness remaining in the profile. An analysis of this feature is made by considering differences between features at the ridge (30N) and trough (10N) of emitted flux.

Zonal average sea surface temperatures (SST) were computed from gridded data of Washington and Thiel (1970). The zonal mean SST at 10N is 1 to 2°C larger than at 30N in July. Thus emission from the sea surface will act to reduce the difference in 10N and 30N emitted flux. However, the SST effect on the difference may be minimal because of the nearly opaqueness of the atmospheric window at 8 to 12 μ due to large optical depths of water vapor as found in the tropics (Cox, 1975).

To demonstrate the magnitude of water vapor absorption along with the change in SST on the north-south flux gradient, a radiative transfer model was applied to July zonal mean vertical profiles of temperature and specific humidity of Oort and Rasmusson (1971) at 30N and 10N. The radiative transfer model is a total longwave flux model (broadband infrared) reported on by Cox et al. (1976). The computed flux was 10 wm^{-2} less at 10N than at 30N. To further assess the potential effect of the water vapor gradient, a vertical specific humidity profile of a composited Western Pacific cloud cluster (Gray et al., 1975 and Frank, 1976) was inserted into the radiative calculations. This profile is to be considered an extremely moist atmosphere with a good portion of the water vapor excess being

in the upper troposphere. This new computed value was 23 wm^{-2} less than the previously computed flux value at 30N. Thus the water vapor absorption-emission contribution to the north-to-south flux decrease appears to lie somewhere between 10 wm^{-2} , a space-time mean and zonal mean for land-plus-ocean, and 23 wm^{-2} , a time mean of very moist cloud clusters.

Finally, the existence of residual cloudiness, which was not removed by application of this maximum IR method, must be considered. Effects of optically thin cirrus cloudiness, which is not significantly detectable in the short-wave portion of the solar spectrum but is detectable in the IR, may not have been eliminated over the ITCZ. Theoretical work of Zdunkowski et al., (1965 and 1971) indicates that cirrus-cloud, invisible from the ground (or haze), can reduce the total longwave flux emitted to space by 10 percent (25 wm^{-2}). He further states that his results are consistent with those from radiometer measurements made in tropical regions (the radiometer measurements show a sudden decrease in the infrared total net outgoing flux in the upper troposphere). The extent and persistence of such cloudiness is not known but its presence could add to the magnitude of flux decrease in the cloud free profile. It is reasonable to assume that as much 10 wm^{-2} of the cloud-free flux profile dip at the ITCZ is caused by residual "haze" or thin cirrus cloudiness.

The cloud-free profile nearly coincides with the cloudy profile in the latitude zone 37N-to-55N for January-February. Even though the span is 13 days, at many grid areas there are less than 8 samples in time. Also the greater portion of the Pacific Ocean poleward of 45N was not sampled at all. Thus, the utility of the January-February period for this type of study certainly must be questioned. Over cold land surfaces the coincidence or near coincidence of the two profiles is entirely reasonable but not over an ocean surface.

Annual mean profiles for land-plus-ocean, ocean, and land were computed by averaging the four semi-monthly periods (Figure 2.7). As in the cloudy albedo, the cloudy IR shows the mean position of the ITCZ to be at 7N. The magnitude of the ITCZ dip is considerably reduced in the annual mean because of averaging the north-south migration of the zone with season. It is halved in the annual profiles from July values. Similarly, the albedo increase due to the ITCZ over oceans is halved in the annual mean from a magnitude of 5 to 2.5 percent.

The validity of the cloud-free emitted flux can only be suggested by radiative transfer calculations. Schneider (1972) applied radiative transfer calculations to a model atmosphere of the globe to obtain a cloud-free total infrared flux of 279 wm^{-2} . Using the same model atmosphere and applying a radiative transfer model as discussed in Cox et al., (1976), the cloud-free flux was

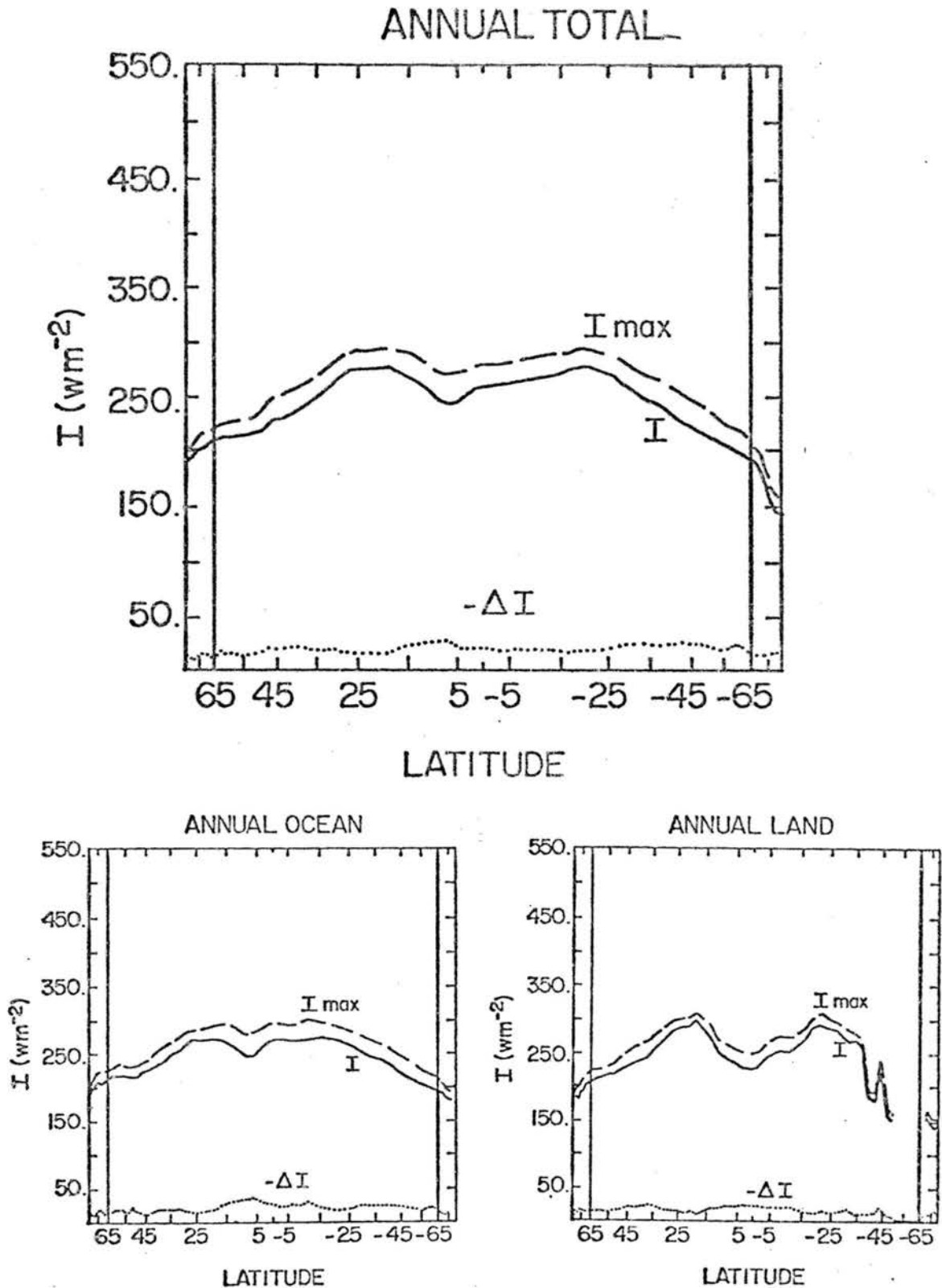


Figure 2.7.: Annual longwave flux, (I) and cloud-free longwave flux, (I_{max}) over land-plus-ocean, ocean and land.

computed to be a 257 wm^{-2} in the present study. With the inclusion of a mid-latitude ozone model of Krueger and Minzner (1976) in the model atmosphere, the cloud-free planetary emitted flux calculated with Cox's model was further reduced to 251 wm^{-2} .

The application of a middle latitude, Northern Hemisphere ozone model atmosphere is not a truly correct application but merely an approximation to show the gross effect of ozone emission on the terrestrial longwave flux to space. Total column ozone varies considerably with season in the higher latitudes. Not only does the total column ozone generally increase from near the equator to the higher latitudes, but the height of maximum ozone concentration decreases from the equator to the higher latitudes so as to further complicate the effect which ozone has on the total longwave flux to space, Dutsch (1971). In the higher latitudes over a very cold surface, such as might occur early in the spring season when total column ozone is largest, the ozone emission may increase the longwave flux to space. Not allowing for spatial and temporal variations is a shortcoming in using a one-dimensional model atmosphere. A similar discussion could follow on the effect of temporal and spatial variability in water vapor. However, it is quite reasonable to consider the ozone effect to be quite small (3 percent or less) in light of what will later be shown to be the cloud effect.

The global and annual mean value derived from the "maximum IR" method is 261 wm^{-2} . This value is within 4 wm^{-2} , or 1.5 percent, of the model atmosphere calculations using Cox's model without an ozone profile. The semi-monthly global satellite determined values range from 268 wm^{-2} in May to 251 wm^{-2} in the January-February period. Again a severe sampling limitation both in space and time of the latter period places that value in question. The next lowest value is 261 wm^{-2} in October. A recomputation of the mean, excluding the January-February period, yields a global mean value of 265 wm^{-2} . This new value is 14 wm^{-2} less than Schneider's, 8 wm^{-2} larger than results with Cox's model using the same model atmosphere, and 14 wm^{-2} larger than results computed from Cox's model with inclusion of a mid-latitude ozone profile. Quite apparent is that results of the "maximum IR" method are in the middle of the theoretical calculations. Thus it is concluded that the maximum IR method gives results which are reasonable.

3. CLOUD EFFECTS ON THE PLANETARY RADIATION BUDGET

In this section the effects of "today's" cloudiness in "today's" planetary radiation budget are examined. By "today" is meant the period for which satellite radiation budget measurements have been composited in this study, 1964 through 1971.

The heat budget of the planet is the long term control of the planet's climate. The planetary radiation budget is a part of the global heat budget. The role that clouds play in the radiation budget partially defines their role in the global heat budget. A discussion of the cloud role in the global heat budget should go concurrently with, or be preceded by a discussion of its role in the planetary radiation budget. Furthermore, the role of cloudiness in climate change, commonly referred to as cloud feedback processes, can be considered only after gaining thorough knowledge of its role in the global heat budget. This section will examine the single role of cloudiness in today's planetary radiation budget.

The cloud amount sensitivity may be defined as equation 3.1:

$$\frac{\partial N}{\partial A_c} = \frac{\partial A_b}{\partial A_c} - \frac{\partial I}{\partial A_c} \quad (3.1)$$

by operating on equation (2.1) with $\frac{\partial}{\partial A_c}$, a local derivative taken with respect to fractional cloud amount, A_c .

This derivative, or sensitivity factor, has been discussed in theoretical studies by Schneider (1972) and more recently by Cess (1976). This sensitivity factor, derived from the present diagnostic study, is compared to their theoretical results in the next major section. The cloud effect in the planetary radiation budget, ΔN , for "today's" radiation budget climatology is examined by differencing "today's" planetary radiation budget with a cloud-free, non-equilibrium radiation budget. It is mathematically defined as:

$$\Delta N = \Delta A_b - \Delta I \quad (3.2)$$

where: $\Delta N \equiv \Delta A_c \frac{\partial N}{\partial A_c}$
 or from equation (2.1)

$$\Delta N = - (S \Delta \alpha + \Delta I) \quad (3.3)$$

with a consideration that the solar constant does not vary with changes in cloudiness. The ΔN should be thought of as the radiative forcing in the heat budget which would exist with an instantaneous removal of all clouds. It will be examined on the zonal mean and global space scales, and on a seasonal time scale.

3.1 Zonal Mean Cloud Effect

The $\Delta\alpha$ and ΔI profiles, where Δ indicates cloudy minus cloud-free, were shown in Figures 2.1, 2.2, 2.4, 2.5, 2.6, and 2.7. Resulting profiles of ΔN , the cloudy planetary radiation budget minus the cloud-free radiation budget, are Figures 3.1 and 3.2 for total land-plus-ocean and ocean, respectively, for each semi-monthly period. Figure 3.3 shows the mean annual ΔN profile of total land-plus-ocean, ocean, and land.

All of the profiles show that ΔN is less than or equal to zero in all latitude zones except for occasional positive values in polar zones, zones comprising 10 percent or less of the global area. A negative ΔN indicates that clouds have a larger effect on ΔAb , the absorbed solar flux, than on ΔI , the emitted terrestrial flux. Largest negative values are in the hemisphere of solar declination. The values become more negative going from equator to approximately 60 degrees latitude in each hemisphere.

Contribution to ΔN from the two terms on the right side of equation (3.3) may be determined by examining the zonal profiles of $\Delta\alpha$ and ΔI in the previous section. Now $\Delta\alpha$ of Figures 2.1 and 2.2 increases away from the equator in each hemisphere in all seasons except for the Northern Hemisphere winter total land-plus-ocean profile. The profiles of ΔI , Figures 2.5 and 2.6, show no systematic change from equator poleward. Therefore, the negative increase in ΔN with latitude in the hemisphere of solar declination is

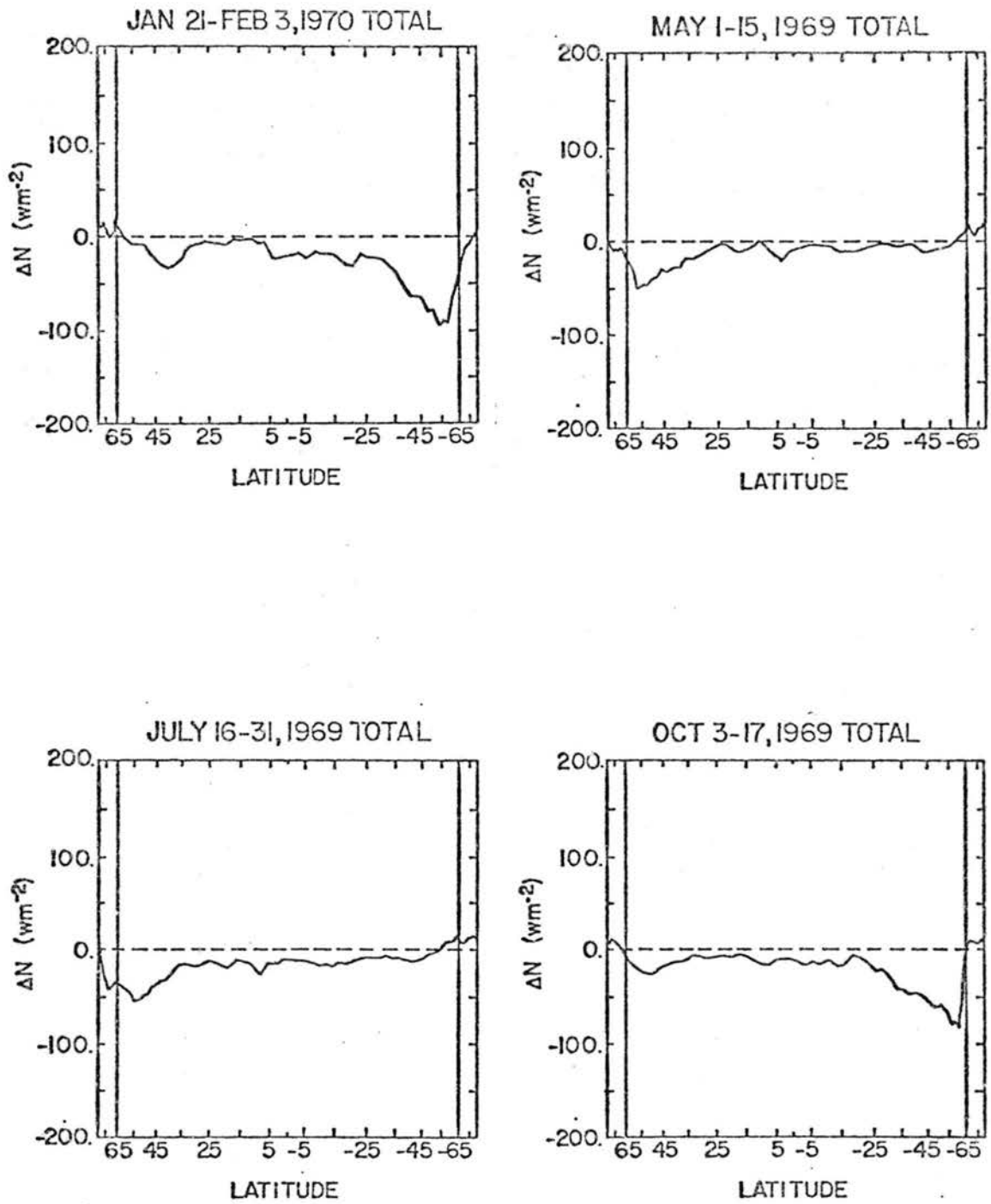


Figure 3.1.: Cloud effect on net flux for land-plus-ocean.

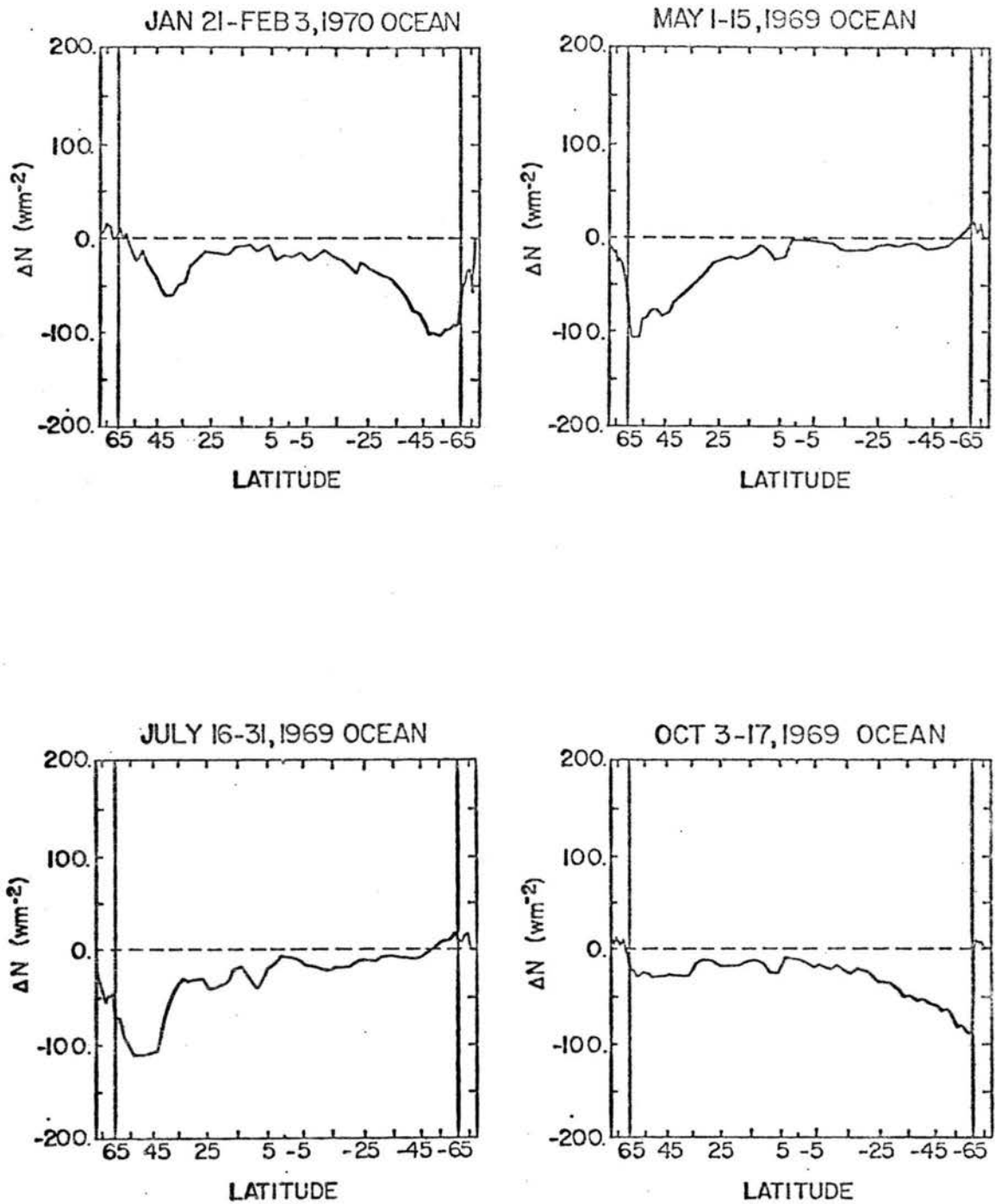


Figure 3.2.: Cloud effect on net flux for oceans.

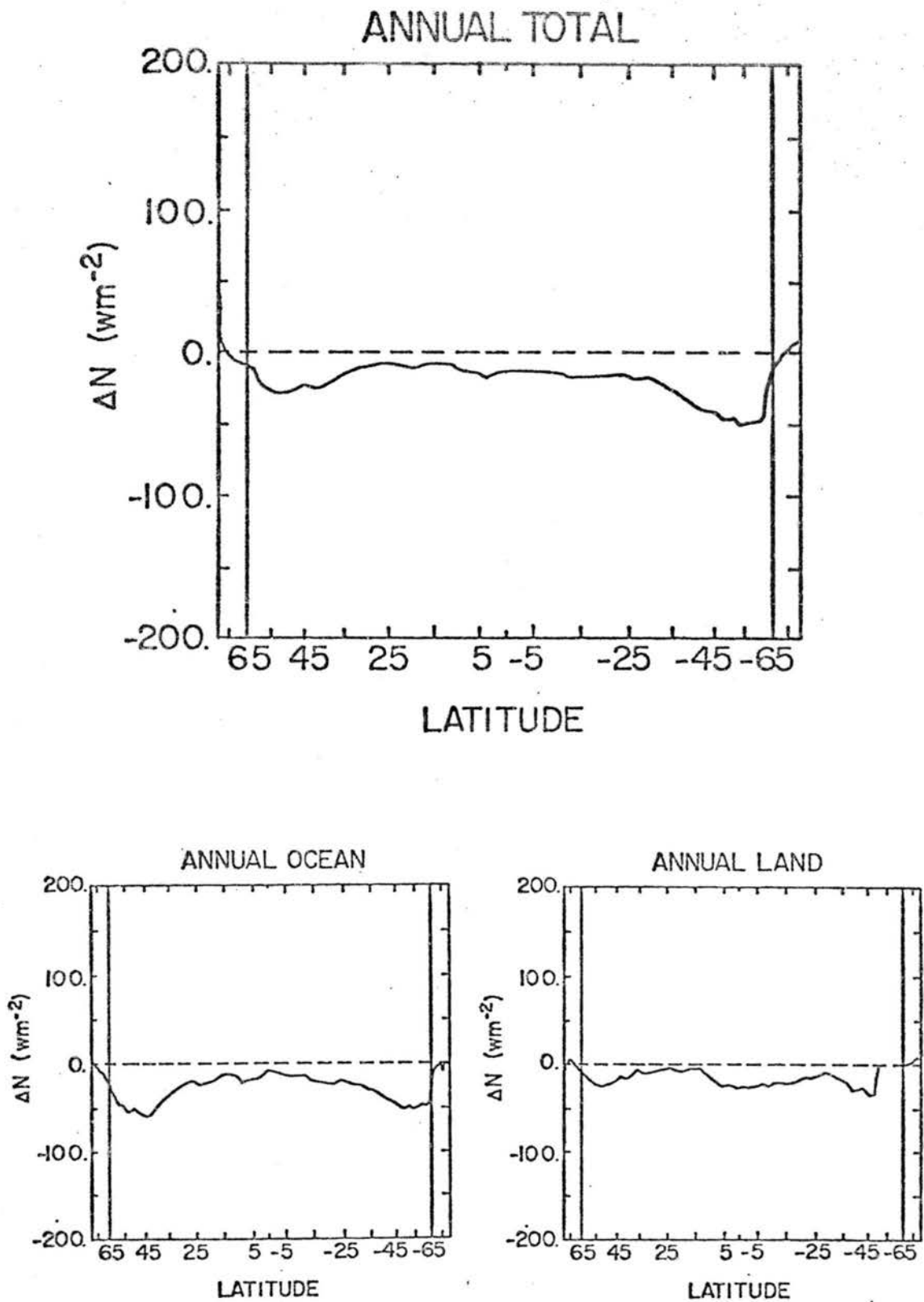


Figure 3.3.: Annual cloud effect on net flux over land-plus-ocean, ocean and land.

principally a result of the shortwave component, $s\Delta\alpha$, in equation (3.3). The ΔN over oceans, Figure 3.2, is greater than -100 wm^{-2} in the 50 to 60 degree latitude zone of the summer hemisphere. In the total (land-plus-ocean) profile the symmetry with time is absent because of the asymmetrical distribution of land surface between the two hemispheres. The largest negative value is -50 wm^{-2} in the Northern Hemisphere during the May and July periods, and -80 wm^{-2} to -100 wm^{-2} in the Southern Hemisphere during the October and January-February periods, respectively. A comparison of the ocean and land profile in Figure 3.3 shows a larger cloud effect in ΔN over ocean - a manifestation of lesser albedo over cloud-free-ocean than over land.

Also of considerable interest is the cloud effect in the tropics. A smaller ΔN in tropical latitudes than in middle latitudes illustrates a reciprocal nature of the absorbed and emitted flux above clouds in the tropics. This has also been noted by Winston (1967) for instantaneous satellite measurements. Increased reflected flux in the presence of a cloud is nearly offset in the net flux by a corresponding decreased emitted flux over tropical oceanic zones. Examination of oceanic ΔN profiles in Figure 3.3 show nearly complete compensation in May at 0-to-5S ($\Delta N = -3 \text{ wm}^{-2}$). In the same zone the ΔN for July is -8 wm^{-2}). This zone of smallest difference moves northward to 0-to-5N in October (-8 wm^{-2}); in January-February it is at 5N-to-15N. All of these minimum

differences occur in the hemisphere opposite the solar declination where incoming solar flux and $S\Delta\alpha$ decrease to barely balance ΔI .

Adjoining the minimum ΔN is a rather abrupt increase to a larger negative ΔN value. The position of this increase is precisely the mean position of the ITCZ as identified in the preceding section. The largest increase in the ITCZ anomaly, over what would exist with a smooth north-to-south profile across the zone, occurs in July. It is approximately 25 w m^{-2} . Now this value is of the same magnitude as the observed decrease in cloud-free IR over the ITCZ (Figure 2.6). As discussed in the previous section, this anomalous dip in cloud-free IR may be accounted for by both the presence of abundant water vapor and a haze type, or nearly invisible type, of cirrus cloudiness which affects only the cloud-free IR.

In polar zones during the polar night, ΔN is positive. The presence of clouds act to decrease the emitted flux loss to space. This holds true except for low and middle cloudiness formed just above a surface inversion in which case the opposite may be true. If the vertical lapse of temperature from the ground to some height above ground is greater than zero, $\frac{\partial T}{\partial Z} > 0$, then clouds at the inversion top will increase the emitted flux loss to space. Over ice free oceans, such an inversion may not be present because the free air temperature is most often lower than the

temperature of the ocean surface. A crossover to positive ΔN occurs, as a valid result, at latitudes 52S and 57N in the July and January-February ocean profiles, respectively. It cannot be ascertained that ΔN should remain positive all the way to the polar night pole because of the possible existence of strong surface inversions overlying ice and snow surfaces. If the effective cloud tops are warmer than the ground, then ΔN would be negative in the polar night zone.

The annual mean profiles in Figure 3.3 show that for all latitudes within the 65N-to-65S zone, ΔN is negative. The minimum cloud effect in the total land-plus-ocean profile is -7 wm^{-2} at 12N with a maximum value of -49 wm^{-2} at 55S. Asymmetry in ΔN between the two hemisphere's is largely the result of asymmetry in land mass distribution. The annual mean ocean profile is quite symmetrical about the equator. Thus, the cloud effect on the planetary radiation budget over oceans appears to be the same in both hemispheres. The annual mean ΔN profile over land surface lacks symmetry between hemispheres largely because of differences in land surface albedo. In the zone 30N-to-15N, ΔN is less than 10 wm^{-2} because of bright and relatively cloud-free land surfaces.

3.2 Global and Hemispherical Mean Cloud Effect

The global and hemispherical mean cloud effects are summarized in Table 3.1. Seasonal ΔN values were computed

by applying a seasonal solar insolation to equation (3.3) with the semi-monthly albedo and emitted flux values both taken as being representative of seasonal values. The annual value is merely the average of the four seasonal values. The values in Table 3.1 are not quite global and hemispherical, but are limited to 65N-65S and 0-65 degrees latitude, respectively, because of uncertainty in the cloud effect polewards of 65 degrees latitude.

The annual mean global value is -20 wm^{-2} indicating that global cloudiness has a larger effect on absorbed solar flux than on emitted terrestrial flux. Instantaneous removal of clouds over the 65N-65S latitude zone would have an immediate effect of a net gain of 20 wm^{-2} in the global planetary radiation budget. The cloud effect is 9.5 wm^{-2} larger in the Southern Hemisphere than the Northern Hemisphere or, in other words, the Northern Hemisphere's cloud effect is only 62 percent of the Southern Hemisphere's.

The cause of the asymmetry between the values of the two hemisphere's is elucidated by Table 3.2. Each latitude zone is considered to be composed entirely of ocean so that over a 65N-65S oceanic zone an annual mean ΔN of -30.6 wm^{-2} would be observed. This value is 1.5 times larger than the land-plus-ocean value of Table 3.1. Thus, whether clouds are over land or water is of significant concern in studying the cloud effect on the planetary radiation budget. Furthermore, the difference in annual means of the two hemispheres in Table 3.2 is less than 4 wm^{-2} and of

Table 3.1 - Cloud effect in net flux (ΔN) for land-plus-ocean (wm^{-2})

	MAM	JJA	SON	DJF	ANNUAL
65N-65S	-14.0	-16.0	-23.5	-26.7	-20.0
65N-0	-14.2	-23.5	-12.0	-11.3	-15.3
65S-0	-13.8	- 8.5	-34.9	-42.0	-24.8

Table 3.2 - Cloud effect in net flux (ΔN) for oceans (wm^{-2})

	MAM	JJA	SON	DJF	ANNUAL
65N-65S	-30.1	-29.2	-30.0	-33.2	-30.6
65N-0	-44.8	-48.2	-18.0	-19.1	-32.5
65S-0	-15.3	-10.2	-42.0	-47.2	-28.7

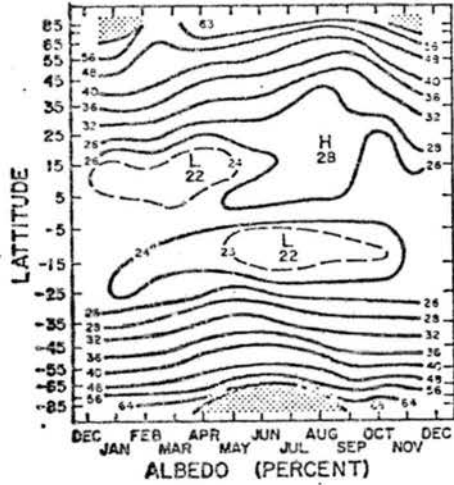
opposite sign to the difference in annual mean hemispherical totals of Table 3.1. The difference between ocean values is most likely not significant or real in light of limited ocean sampling in higher latitudes of the Northern Hemisphere, but the difference in annual hemispherical values of Table 3.1 is undoubtedly due to differences in land-ocean distribution between the hemispheres. Also of interest is the variation in global values with season. For the land-plus-ocean values this variation amounts to 12.7 wm^{-2} whereas over oceans it is 4 wm^{-2} indicating that seasonal variations in ΔN are principally a result of asymmetry in land and ocean distribution. The season with largest global cloud effect is DJF. Largest negative values appear in the summer season with the southern summer value being nearly twice as large as the northern summer value.

3.3 Seasonal Variations In the Cloud Effect

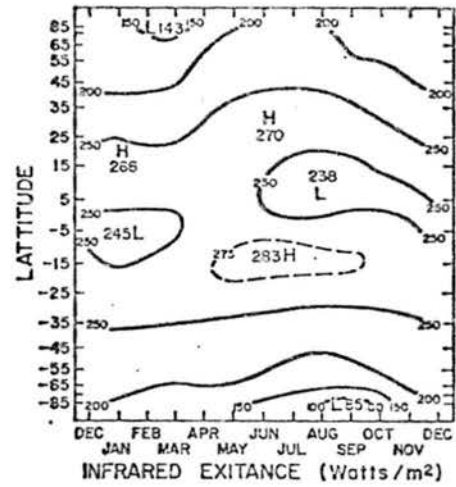
Seasonal variations, are examined both qualitatively and quantitatively in zonal profiles and global mean values.

3.3.1 Zonal Variations

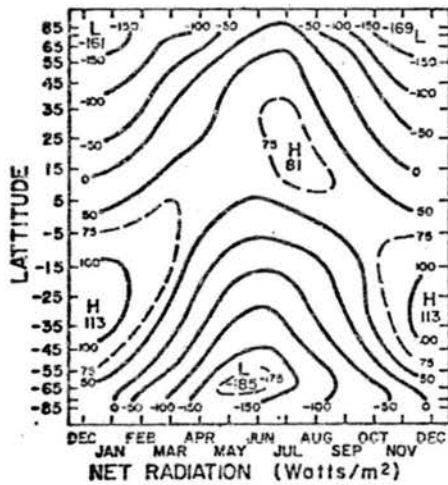
Time-latitude displays of zonal mean profiles of albedo, longwave emitted flux and net flux are shown in Figure 3.4. These are from the composited 29 months set of satellite measurements discussed in Ellis and Vonder Haar (1976). Also shown in Figure 3.4d is a time-latitude



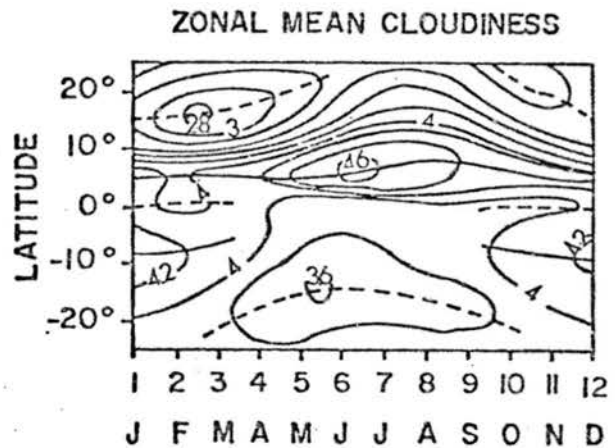
(a)



(b)



(c)



(d)

Figure 3.4.: Time-latitude analysis of zonal mean profiles from the 29 month set: (a) albedo, (b) long-wave flux, (c) net flux and (d) cloud amount.

display of tropical cloud amount from Murakami (1975). Large seasonal variations in albedo and longwave flux occur in the 30N-30S latitude zone. Of particular interest is the albedo increase in the Equator-to-20N zone from a low value of 24 percent in May to 28 percent in August, followed by an abrupt decrease in October to 25 percent (Figure 3.4a). Undoubtedly the variation is associated with the cloudiness change during the Southwest Asian Monsoon. The same seasonal variation appears in cloud amount. Similarly, south of the equator in the tropics, a variation of lesser magnitude and of opposite sign takes place.

In latitude zones both north and south of the tropics, large changes in albedo take place throughout the year. By August and September the extratropical zones of the Northern Hemisphere attain their minimum albedo. Undoubtedly, this is an effect of a minimum in ice-snow and cloud coverage associated with the northward retreat of the polar front and a decrease in cyclogenesis. At 45N latitude the seasonal variation amounts to 17 percent while at 45S latitude the variation is but 7 percent.

It must be recalled that these profiles are greatly smoothed in the north-south direction because the individual measurements were resolved at a geocentric arc distance at the earth's surface as large as 7770 km for some satellites, while as small as 50 km for others (Ellis and Vonder Haar, 1976). Some of the smaller wiggles in the isolines are artifacts of the satellite sampling. The emitted longwave

flux of Figure 3.4b shows seasonal variation but of opposite sign to the albedo at corresponding temporal and spatial positions.

Net flux time-latitude depiction, derived from the albedo, emitted longwave flux and solar insolation, does not show the tropical zone anomalies as previously discussed (Figure 3.4c). To a large extent, the albedo and longwave anomalies largely cancel each other in the net. The overwhelming seasonal varying pattern in the net follows closely the pattern of extraterrestrial solar flux as shown in Sellers (1965). Seasonal variations due to cloud effects in net radiation as observed in zonal mean profiles are apparently quite small compared to the effect of the north-south march of extraterrestrial solar flux with time.

3.3.2 Global Variations

The effect of clouds in the annual (intra-annual) variation of the global net radiation budget is assessed by two methods. The first is an indirect method which evaluates contributions to the annual variation from other than the cloud contribution. The residual variation not explained by the other than cloud contribution is taken to be the cloud contribution. The second method, a direct method, examines the annual variation of the globally integrated values from the four semi-monthly zonal profiles of cloud effects. Each of the methods are discussed in order, first the indirect and then the direct.

3.3.2.1 Indirect Analysis Method

The radiation budget components of incoming solar flux and outgoing reflected shortwave and longwave flux were globally integrated and composited into monthly mean values from the 29 month satellite data set compiled by Ellis and Vonder Haar (1976). A discussion of the integration method along with a monthly mean tabulated listing of the global incoming solar insolation, albedo, reflected shortwave flux, outgoing longwave flux and net flux are given in Appendix C.

Global values of the planetary net radiation flux, the composited mean values (N) and the individual monthly values, are plotted in Figure 3.5a. There is a considerable amount of month-to-month variability in both the composited mean curve and in the envelope of all values. Even though the uncertainty in the global values is as large as 10 wm^{-2} (Ellis and Vonder Haar, 1976), which is more than one half the amplitude of the annual variation, the fact that the shape of the envelope and the composite curve are so very similar gives credibility to the phase and amplitude of the annual variation. Another way of looking at the credibility of the annual variation is to observe that the variability of the individual monthly values about the composite curve does not exceed $\pm 12 \text{ wm}^{-2}$ for any monthly period, while the variability of the individual monthly values about the annual mean value is greater than $\pm 25 \text{ wm}^{-2}$. The difference, 13 wm^{-2} , can be attributed to annual variation. The credibility of annual variation in global net radiation

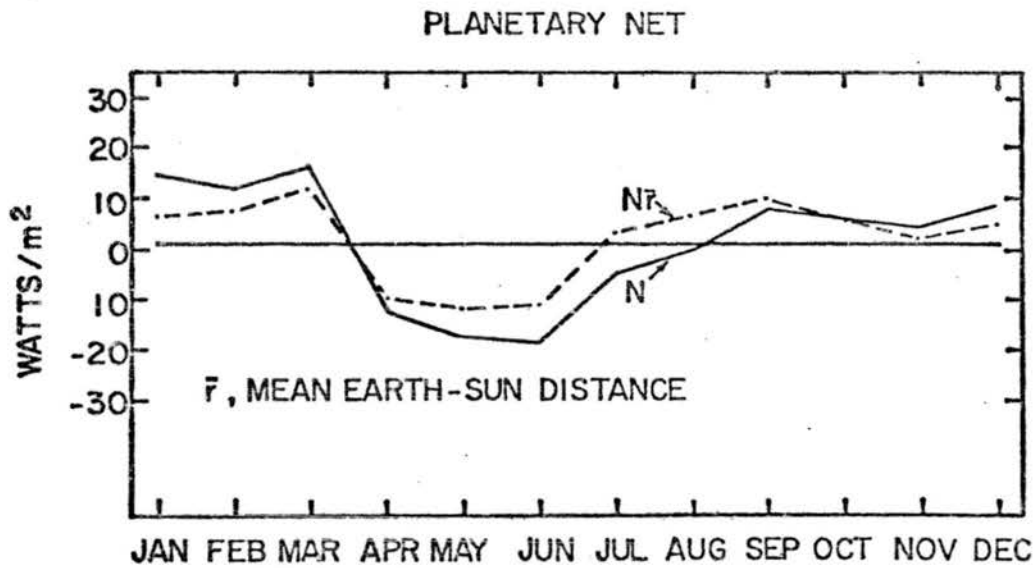
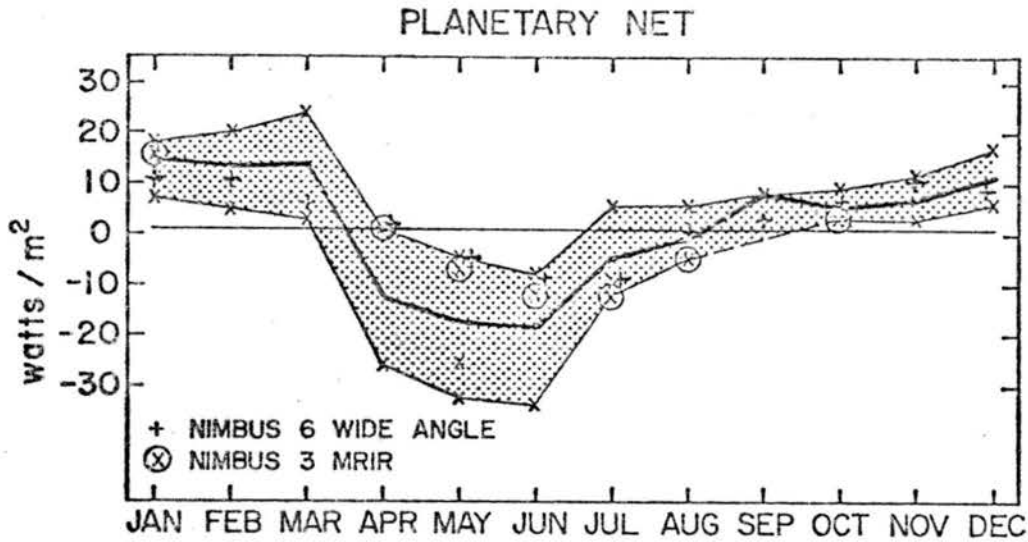


Figure 3.5.: Annual variation in global net flux.

is further validated by planetary heat budget studies. Annual variation in the rate of change in heat content (storage) of the planet Earth was shown by Ellis et al, (1978) to be of the same amplitude and phase as the seasonal variation in the planetary net radiation budget. The global heat storage had been computed from independent, in situ measurements of temperature.

That there should be an annual variation in the global planetary net radiation budget was discussed from a theoretical viewpoint by Simpson (1929) prior to the advent of artificial earth satellites for its measurement. Raschke (1973) concluded from the data of the medium resolution infrared radiometer (MRIR) onboard the Nimbus 3 satellite, that during the ten semi-monthly periods examined, there was appreciable annual variation. The Nimbus 3 values, which are included in the composite profile, are uniquely identified in Figure 3.5a. Also identified are preliminary values from the wide angle earth radiation budget sensor aboard the Nimbus 6 satellite (Smith et al, 1977). The Nimbus 6 data have not been included as part of the composite data set, since they are preliminary values. The agreement between the Nimbus 3 and Nimbus 6 value is surprisingly good.

Annual variation in the planetary radiation budget is hypothesized to receive principal contributions from the following:

1. Annual variation in the earth-sun distance.

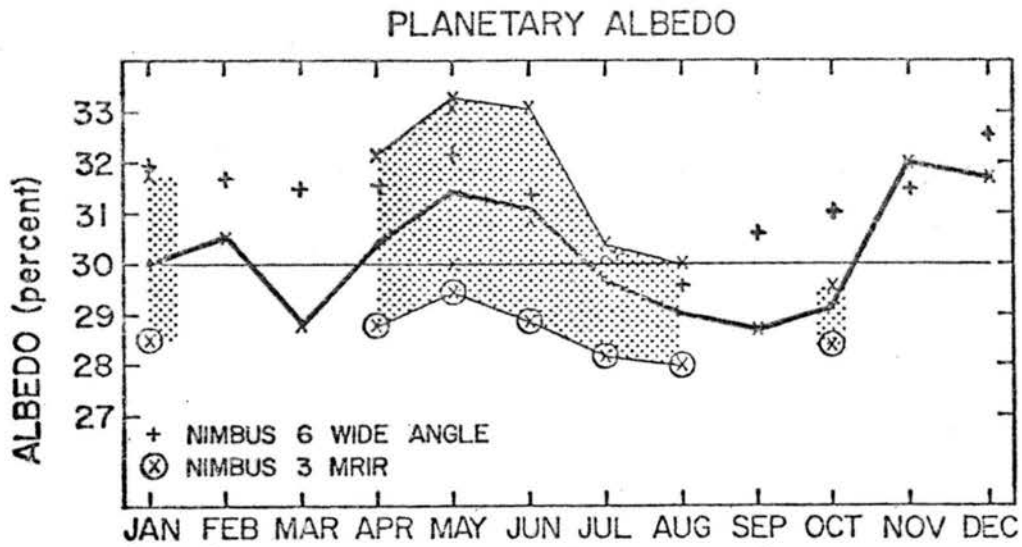
2. A semi-annual cycle in the global albedo because of enhancement from the highly reflective polar regions near the time of solstice.
3. Annual variation in the global albedo because of the seasonal advance and retreat of sea ice and continental snow cover.
4. Annual variation in the longwave emission to space because of unequal temperature response between the land and water surfaces and the asymmetrical distribution of these surfaces between the Northern and Southern Hemispheres.
5. Seasonal variation in the distribution and amount of global cloud cover.

The orbit of the planet Earth about the Sun is eccentric. The solar flux received by the planet as it moves in its orbit varies as a sine wave with a period of one year because of the eccentric orbit. The amplitude in annual variation of globally integrated solar flux received by the earth is approximately 11 wm^{-2} (assuming a solar constant equivalent to 1360 wm^{-2}). The annual cycle was removed from the incoming flux and the global net radiation was recomputed. These artificial global net radiation values are plotted as a composited mean curve, N_r , in Figure 3.5b. The difference between the N and N_r curves illustrates the contribution to the annual variation in net radiation flux from the annual variation in incoming solar flux. The contribution is slightly less than one half the

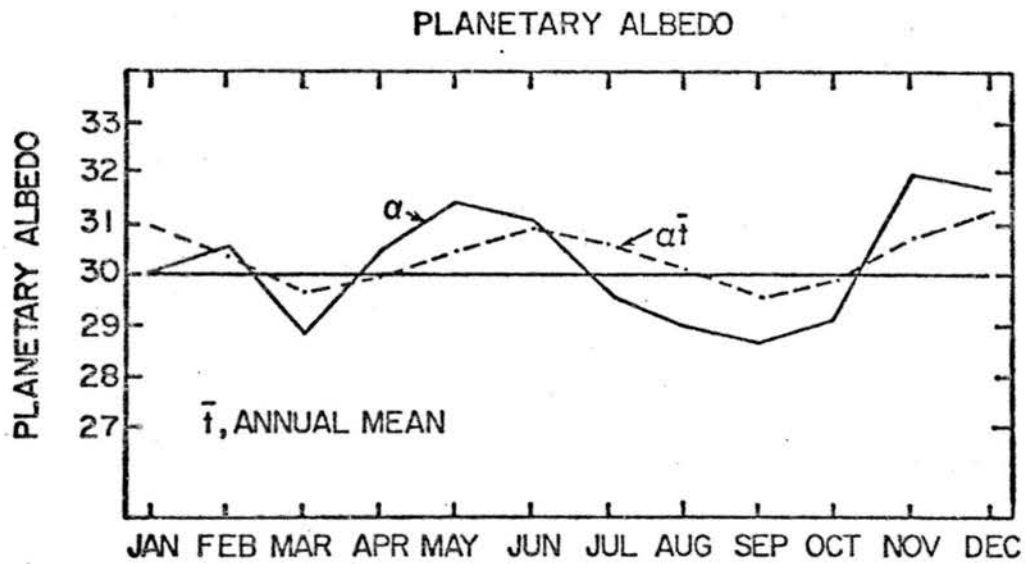
amplitude of annual variation of the global net flux. Therefore the annual variation in global net radiation is only partially explained by the contribution from solar flux.

The other two components of the global net radiation balance, the global albedo and global outgoing longwave flux, will be examined for annual variation. The global albedo composited mean profile and the envelope of individual monthly values are plotted by month in Figure 3.6a. There is but one value plotted for five of the monthly periods, because the albedo values from the ESSA 7 satellite data were excluded from this analysis (discussed in Appendix C). The global albedo values from the preliminary Nimbus 6 data set are plotted, but are not included in the composite curve. The Nimbus 6 albedo values are systematically larger than the composite curve, but generally show phase agreement with the composite. Largest discrepancies in month-to-month variation between the two curves appear in the months of March and November.

The contribution to the annual variation in the global albedo due to north-to-south differences in the earth-atmosphere planetary albedo can be assessed, ie. the snow and ice covered polar latitude zone in each hemisphere (60 to 90 degrees) has a larger zonal mean albedo than the tropical latitude zone (20N to 20S degrees latitude) (Vonder Haar, 1968; Vonder Haar and Soumi, 1971; Ellis and Vonder Haar, 1976).



(a)



(b)

Figure 3.6.: Annual variation in global albedo.

The annual albedo in each latitude zone ($\alpha(\phi)$) can be computed with equation (3.4)

$$\alpha(\phi) = \frac{\int_{\text{12 months}} \bar{\alpha}(\phi, t) s(\phi, t) dt}{\int s(\phi, t) dt} \quad (3.4)$$

where:

$\bar{\alpha}$ = the zonal mean planetary albedo

s = incoming solar flux

t = time in months

ϕ = latitude

The global albedo may now be computed for each month, assuming that the annual albedo in each latitude zone prevails for that month,

$$\alpha_{\bar{t}} = \frac{\int_{-\pi/2}^{\pi/2} \alpha(\phi) s(\phi, t) d(\sin \phi)}{\int s(\phi, t) d(\sin \phi)} \quad (3.5)$$

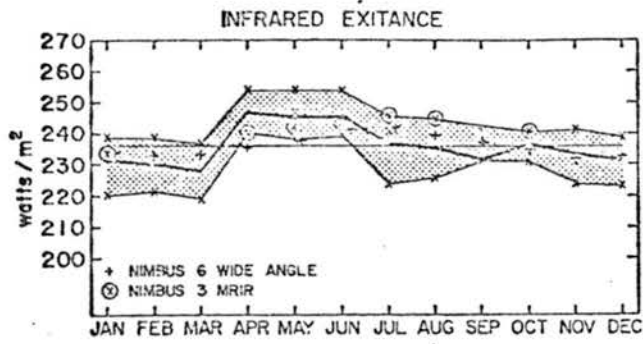
The $\alpha_{\bar{t}}$ values are plotted by month along with the actual composite global albedo values (α) in Figure 3.6b.

The $\alpha_{\bar{t}}$ curve shows semi-annual periodicity with maxima at the summer and winter solstice and minima at the vernal and autumnal equinox. The largest maxima occurs at winter solstice. The difference between maximum and minimum values amount to 1.5 percentage values. The $\alpha_{\bar{t}}$ annual variation contributes approximately 4 wm^{-2} to the annual variation in the absorbed shortwave flux.

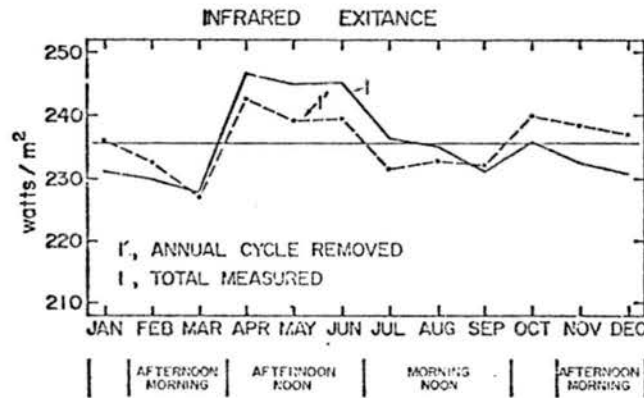
The $\alpha_{\bar{t}}$ contribution to the variation in absorbed shortwave flux is negative near winter solstice. The earth is nearest to the sun several weeks later (perihelion) when the solar flux contribution is greatest. Six months later, when the earth is farthest from the sun, the $\alpha_{\bar{t}}$ contribution is again negative. Thus, the positive solar contribution to the absorbed shortwave flux ($+8 \text{ wm}^{-2}$) is diminished by 2 wm^{-2} near winter solstice, while the negative solar contribution to absorbed shortwave flux (-8 wm^{-2}) is amplified by an additional 2 wm^{-2} .

The annual variation in $\alpha_{\bar{t}}$ accounts for one half the variation in α . The residual albedo variation, 1.5 percentage values (4 wm^{-2}), can be attributed to varying albedo within the latitude zones. Sampling bias, directional reflectance properties of an earth-atmosphere column, changes in cloud, snow, ice and vegetation cover contribute to the temporal albedo variation within latitude zones, and thus to the residual variation in global albedo. The cloud effect appears to be less than 4 wm^{-2} in the annual variation of the global shortwave component of net radiation provided that the other temporally varying contributors are not masking the cloud effect.

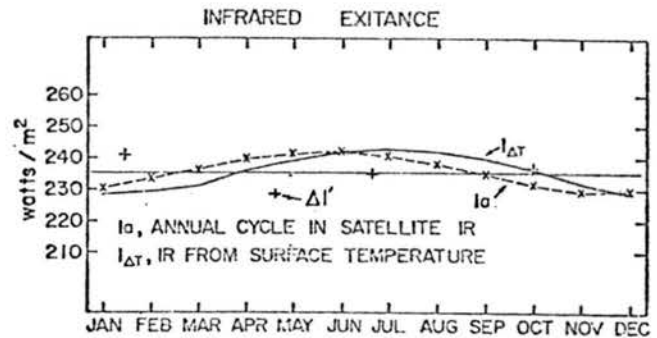
The global integrated emitted longwave flux composite curve (1) is plotted in Figure 3.7a. Also shown is the range of values within each month and the individual Nimbus 3 values. The Nimbus 3 values fall along a curve which can be fitted with an annual wave. Raschke (1973)



(a)



(b)



(c)

Figure 3.7.: Annual variation in global longwave flux.

has also shown the annual wave in Nimbus 3 measurements. Preliminary results from Nimbus 6 ERB wide angle measurements, which are not included in the 29 month averages, agree quite well with the Nimbus values, but with smaller amplitude. This agreement suggests that the global mean values from the 29 month composite may have a sampling bias.

A fourier analysis applied to the 29 month values decomposed the time plot into an annual cycle (Ia) and higher frequency cycles. Figure 3.7b shows the residual emitted flux (I') after removal of an annual cycle from I. Along the abscissa are bracketed daylight sampling times during which the measurements were taken. Now if it is assumed that global planetary emitted flux values increase from the morning to afternoon, then the sign of the I' perturbations agree with the sampling times. In other words, there is a possibility that a diurnal sampling bias is present in the composite 29 month longwave flux data.

The annual cycle in the satellite composited mean values of outgoing longwave flux is shown as I in Figure 3.7c. Shown in the same figure is $I(\Delta T)$, a representation of the annual cycle in global surface and surface air temperature as it would be measured with a radiometer at the top of the earth's atmosphere in the absence of compensations by the earth's atmosphere.

Global integration of the combined ocean surface temperature (Washington and Thiel, 1970) data and land

surface air temperature data (Crutcher et al., 1970 and Taljaard et al., 1969) showed an annual cycle in the global surface temperature with amplitude of $\pm 2^{\circ}\text{C}$. The effect of the annual cycle in global surface temperature on the outgoing longwave flux to space was assessed with the aid of the radiative transfer equation. The equation in a broadband flux form for an atmosphere in thermodynamic radiative equilibrium containing no scattering is,

$$F(\text{Tr}) = \epsilon\sigma T_s^4 \tau_s + \int_0^{z_t} \left[\alpha T(z)^4 \frac{\partial \tau(z)}{\partial z} dz \right] \quad (3.6)$$

where ϵ , the surface emissivity will be assumed to be equal to unity, σ is the Stefan-Boltzmann constant, T is temperature, τ_s is the transmittance of the atmosphere to upwelling surface flux, z is height and z_t is the height of the top of the radiating atmosphere. Equation (3.6) may be represented by,

$$F(\text{Tr}) = F(\text{Ts})\tau_s + F(\text{Ta}) \quad (3.7)$$

where F is the upward directed longwave flux at the top of the atmosphere at the effective blackbody radiating temperature of the earth-atmosphere system (Tr), the surface temperature (Ts) and the atmospheric temperature (Ta). Divide equation (3.7) by $F(\text{Ts})$ to get,

$$\frac{F(\text{Tr})}{F(\text{Ts})} = \tau_s + \frac{F(\text{Ta})}{F(\text{Ts})} \quad (3.8)$$

It will be assumed, as a first order approximation to the radiative processes, that τ_s and the ratio $\frac{F(\text{Ta})}{F(\text{Ts})}$ do not change with small changes in surface temperature. Call the right hand side of equation (3.8) K , and differentiate with respect to surface temperature,

$$\frac{dF(\text{Tr})}{dT_s} = K4\sigma T_s^3 \quad (3.9)$$

where:

$$K = \frac{I_m}{\sigma T_s^4} \quad (3.10)$$

with I_m defined as $F(\text{Tr})$. Combine equations (3.9) and (3.10) so that,

$$\frac{dI_m}{dT_s} = \frac{4I_m}{T_s} \quad (3.11)$$

The annual global longwave flux to space, I_m , from the 29 month satellite data set is 235.7 wm^{-2} , while the annual global surface temperature is 288°K . Thus,

$$\frac{dI_m}{dT_s} = 3.3 \text{ wm}^{-2} \quad (3.12)$$

and K is equal to 0.60.

It is known that T^4 is approximately linear in T for the normal range of atmospheric and surface temperatures, so that no serious error is introduced by applying equation (3.12) over the 4°K range of global surface temperatures. If ϵ in equation (3.6) were taken to be 0.9 rather than 1.0, then $\frac{dI}{dT_s}$ would be 3.7 wm^{-2} rather than 3.3 wm^{-2} . Either value is sufficient for the purpose of this discussion.

The effect of the annual wave in global surface and surface air temperature was inserted into the annual global longwave flux at monthly time intervals (t) by,

$$I(t) = I_m + \frac{dI_m}{dT_s} \Delta T_s(t) \quad (3.13)$$

where ΔT_s is the algebraic difference between monthly mean T_s and the annual mean T_s . $I(t)$ is plotted as $I(\Delta T)$ in Figure 3.7c. Both $I(\Delta T)$ and I_a are of the same amplitude but differ in phase by one month. It should be noted that $I(\Delta T)$ has the same phase as the annual profiles of I for the Nimbus 3 and 6 satellites (Figure 3.7a). The equivalence in phase between the $I(\Delta T)$ and the Nimbus satellite profiles and the phase difference between $I(\Delta T)$ and I_a further supports the argument that the composited mean profile in Figure 3.7a may contain temporal sampling bias. If one concludes that all of the I' (Figure 3.7b) and the phase shift in wave number one of I are due to sampling bias, then one must also conclude that there is no apparent

effect on the annual variation of I which is due to an annual variation in cloudiness, ie. month-to-month variations in cloud amount, and/or effective radiating temperatures. However, cloudiness modulates the amplitude of I even though cloudiness may not show a month-to-month effect on I . It is evident from equation (3.6) that τ_s , the transmissivity of the atmosphere to the net upward long-wave flux from the earth's surface, varies directly with large changes in cloud amount and changes in the transparency of the clouds to the upwelling flux. If the effect of cloudiness on T_s and on the second term of equation (3.6) is neglected, then the elimination of all cloudiness will increase τ_s and thus increase the magnitude of response in $F(\text{Tr})$, or I , to an annual wave in global surface temperature. The converse argument with an increase in cloud amount (all other cloud properties held constant) shows that the annual amplitude in I will be less. The magnitude of the cloud effect on the amplitude of the annual wave quantitatively will be shown as part of the discussion on the direct method for resolving the effects of clouds on the global planetary radiation balance (Section 3.3.2.2).

The planetary net radiation curve has been reconstructed in Figure 3.8 using just the annual cycle of the longwave flux (I_a) from Figure 3.7c as the longwave contribution to the net radiation flux. (Note: the phase of the net radiation curve in Figure 3.8 would be shifted one half month later into the year if the I (ΔT) curve were used rather than

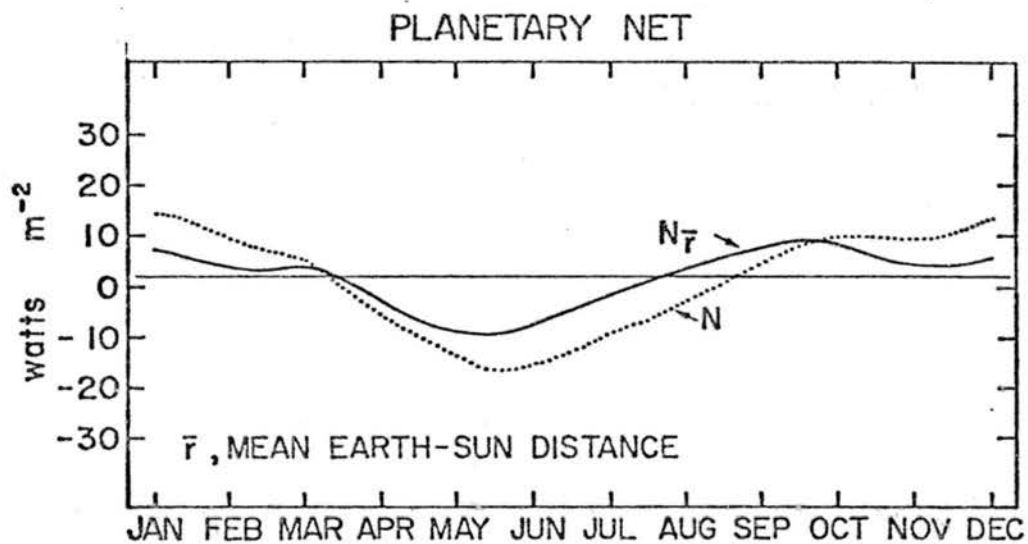


Figure 3.8.: Annual variation in global net flux with just an annual wave in longwave flux.

the Ia curve). A comparison of the curves of Figure 3.5b and 3.8 reveals that the principal differences occur in the months of March and April. Both curves have the same general shape over the other monthly periods.

The indirect method of examining the month-to-month variations in the cloud effect on the global planetary radiation budget has shown the cloud effect to be small. No annual variation in the cloud effect was discerned in the composited global longwave flux. A residual of 4 wm^{-2} in global absorbed shortwave flux may account for the effects of cloud, ice-snow and directional reflectance on the annual variation. However, there could be contributions to the annual variation of the net radiation flux from sources other than clouds which are of the same magnitude but opposite in sign to the cloud contribution. The cloud effect would be masked in such cases.

3.3.2.2 Direct Analysis Method

A second method, a direct analysis of the globally and hemispherically integrated cloud effects, will be used to assess the cloud effect on the annual variation of the net radiation budget. Global and hemispherical values of the cloud effect on the albedo ($\Delta\alpha$) longwave flux (ΔI) and net radiation flux (ΔN) were derived by integrating the zonal profiles of the cloud effects from Figures 2.1, 2.5 and 2.8, respectively. The integrated values are shown in Tables 3.3, 3.4 and 3.5. The cloud effect on the net

Table 3.3 - Cloud effect in net flux_z (ΔN) semi-monthly periods for land-plus-ocean (wm^{-2})

	MAY	JULY	OCT.	JAN-FEB
65N-65S	-12.0	-15.8	-23.1	-26.7
65N-0	-18.6	-23.8	-13.1	-11.7
65S-0	- 5.3	- 7.8	-33.2	-41.6

Table 3.4 - Cloud effect in albedo (ΔN) for land-plus-ocean

	MAY	JULY	OCT.	JAN.-FEB.	ANNUAL
65N-65S	+0.112	+0.105	+0.118	+0.112	+0.112
65N-0	+0.108	+0.105	+0.093	+0.084	+0.100
65S-0	+0.119	+0.105	+0.137	+0.127	+0.125

Table 3.5 - Cloud effect in emitted flux (ΔI) for land-plus-ocean (wm^{-2})

	MAY	JULY	OCT.	JAN.-FEB.	ANNUAL
65N-65S	-27.1	-19.9	-19.5	-14.1	-20.1
65N-0	-28.5	-22.6	-17.5	-10.6	-19.8
65S-0	-25.7	-17.3	-21.5	-17.6	-20.5

radiation is related to the cloud effect on the radiation budget components (the albedo and longwave flux) by equation (3.3).

The cloud effect on the albedo increases in both hemispheres going from the winter season to the spring season (Table 3.4). This increase is probably related to the increase in illumination of the clouds and the ice-snow areas of the higher latitudes. A decrease, similar in magnitude to the increase, occurs in both hemispheres going from the summer season to the fall season. The decline in the cloud effect on the albedo continues into the winter season of each hemisphere. The sign of the change in the cloud effect on the albedo is the same for the globe and the Southern Hemisphere. Thus, the cloud effect on the albedo of the Southern Hemisphere is predominant in the global integral over the same effect of the Northern Hemisphere. The predominance of the Southern Hemisphere cloud effect is most apparent in the annual mean values.

The annual variation in $\Delta\alpha$ may be expressed as a departure from the annual mean ($\overline{\Delta\alpha}$),

$$\Delta\alpha' = \Delta\alpha - \overline{\Delta\alpha} \quad (3.14)$$

so that $\Delta\alpha'$ varies as ± 0.007 , or $\pm 2.4 \text{ wm}^{-2}$ in reflected flux. This variation is small, and does not significantly differ from the residual value of $\pm 4 \text{ wm}^{-2}$ derived by the indirect method. Thus the annual variation in the global albedo due

to cloud effects appears to contribute less than $\pm 3 \text{ wm}^{-2}$ to the annual variation in the net radiation flux.

The annual variation in emitted longwave flux due to clouds can be expressed in the same fashion as,

$$\Delta I' = \Delta I - \overline{\Delta I} \quad (3.15)$$

where $\overline{\Delta I}$ is the annual mean cloud effect in the longwave flux. The $\Delta I'$ variation is slightly less than $\pm 7 \text{ wm}^{-2}$. The $\Delta I'$ values have been added to the I_a value and are plotted on Figure 3.7c as plus signs for a comparison with the annual variation of the longwave flux. Notice that the $\Delta I'$ values are of the same magnitude as the annual variation (I_a) but are generally of opposite sign. Of course, the four points along the curve are hardly sufficient to evaluate the annual variation in $\Delta I'$. Since $\Delta I'$ is of opposite sign to I_a , the cloud effects act to reduce the amplitude of I_a .

Table 3.6 shows ΔI values for a globe hypothetically covered entirely by ocean. The table values were generated by globally integrating the zonal ΔI ocean profiles of Figure 2.6. Note that a global average $\Delta I'$ from Table 3.6 has an annual amplitude of less than $\pm 2.5 \text{ wm}^{-2}$ and is not in phase with the $\Delta I'$ of $\pm 7 \text{ wm}^{-2}$ derived from Table 3.5. The larger $\Delta I'$ of Table 3.5 is largely a result of the asymmetry of the land-ocean distribution between the Northern and Southern Hemispheres. Indeed, it is reasonable

Table 3.6 - Cloud effect in emitted flux (ΔI) for oceans (wm^{-2})

	MAY	JULY	OCT.	JAN-FEB
65N-65S	-20.9	-21.7	-23.1	-18.8
65N-0	-15.6	-23.6	-22.4	-14.2
65S-0	-26.2	-19.8	-23.9	-23.5

to conclude that the removal of clouds over the continental regions would increase the amplitude of I_a due to an increase in the atmospheric transmittance (τ_s in equation 3.6) to thermal radiation upwelling from the earth's surface, ie. an increase in K of equation (3.9). It is not surprising that clouds act to dampen the annual variation in emitted longwave flux. The $\Delta I'$ variation derived from Table 3.5, may be actually less than 7 wm^{-2} because of a data sampling deficiency in the January-February data. Elimination of that semi-monthly period from the study will reduce the variability of $\Delta I'$ to less than $\pm 4 \text{ wm}^{-2}$. This 4 wm^{-2} added to the 6 wm^{-2} amplitude of I_a is equivalent to 10 wm^{-2} amplitude. This is the total annual variation which would be obtained apparently with an instantaneous removal of all clouds. This is the same annual variation which would be computed by equation (3.9) with K equal to 1.0. Thus, the annual variation of the longwave flux (in the absence of all clouds) is nearly equivalent to the upwelling longwave flux at the earth's surface. The results of this direct analysis indicate that as much as $\pm 4 \text{ wm}^{-2}$ of the I' annual variation shown in Figure 3.7b can be accounted for by annual variations in cloud effect on I_a .

Annual variations in the planetary net flux due to annual variations in cloudiness can be elucidated by plotting N' , the departure of semi-monthly values from the annual mean net flux (\bar{N}) with $\Delta N'$ for the four semi-monthly periods.

The symbols are defined as:

$$\Delta N' = \Delta N - \overline{\Delta N} \quad (3.16)$$

and
$$N' = N - \overline{N} \quad (3.17)$$

Figure 3.9 shows that the two plots are nearly 180 degrees out of phase. The difference between N' and $\Delta N'$ in the January-February period is 18.3 wm^{-2} , while it is -18.1 wm^{-2} in the July period. The near equivalency in absolute magnitude illustrates an element of symmetry during the year. It also illustrates that the amplitude in N' , $\pm 13 \text{ wm}^{-2}$, could be increased to $\pm 18 \text{ wm}^{-2}$ with an instantaneous removal of all clouds (without a simultaneous adjustment to a condition of radiative equilibrium).

In summary (for the 29 month period studied) the annual variation in the global planetary net radiation budget due to the effect of clouds is small. The cloud effect on the annual variation of shortwave absorbed flux is less than $\pm 3 \text{ wm}^{-2}$, on the emitted longwave flux it is less than $\pm 6 \text{ wm}^{-2}$ (and most probably less than $\pm 4 \text{ wm}^{-2}$ in the longwave component). Clouds generally act to reduce the amplitude of the annual variation in the net radiation budget.

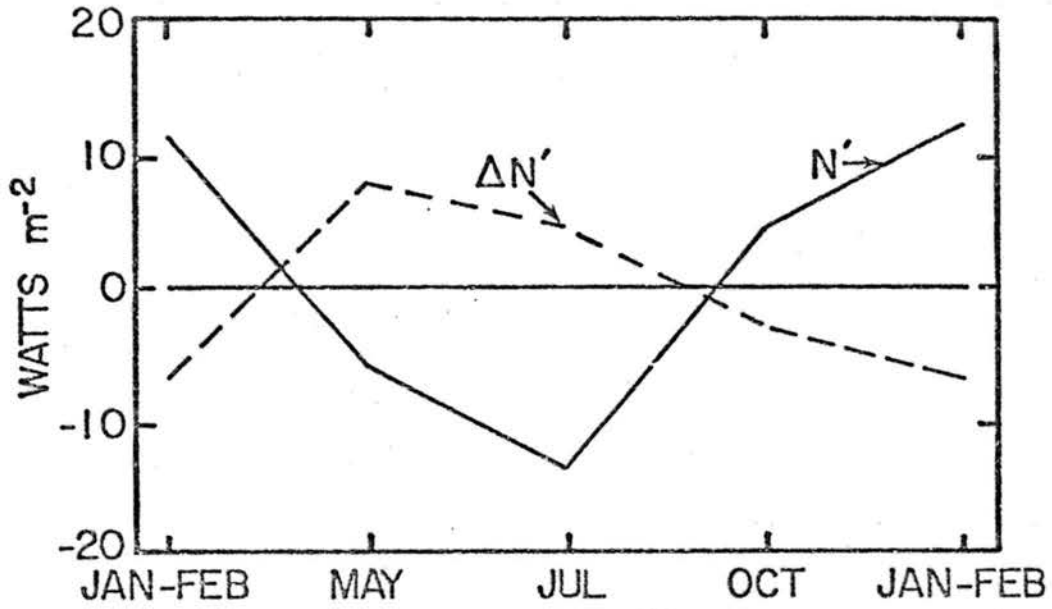


Figure 3.9.: Annual variation in net flux (N') and in the cloud effect in net flux ($\Delta N'$) for four semi-monthly periods.

4. SENSITIVITY TO A UNIFORM CHANGE IN CLOUD AMOUNT

A local uniform increase in cloud amount at all altitudes without a change in other cloud characteristics, including cloud top height, will give rise to an increase in the shortwave flux reflected to space and a decrease in the longwave flux emitted to space. The magnitude of change in the individual reflected and emitted flux components determines the magnitude of change in the planetary net radiation budget. It was established in the previous section that ΔN , or $\Delta A_c \frac{\Delta N}{\Delta A_c}$, is negative and equivalent to -33 wm^{-2} for the planet as a whole with the present cloudiness climatology. The sensitivity to uniform changes in cloud amount, $\frac{\partial N}{\partial A_c}$, will now be established so that a measure of the magnitude of the cloud effect may be determined for various other cloud amounts. However, it must be remembered that $\frac{\partial N}{\partial A_c}$ may not be valid for other climate states which have a mix of clouds with radiative characteristics different than "today's". These characteristics include: height, the vertical and horizontal distribution of finite clouds, the land-ocean spatial distribution, the cloud reflectance, transmittance and emittance. Given the same radiative characteristics as "today's" clouds, then $\frac{\partial N}{\partial A_c}$ is a constant over the full range of A_c values. The $\frac{\partial N}{\partial A_c}$ will be examined at both zonal and global mean scales.

4.1 Zonal Sensitivity

The net flux sensitivity to cloud amount, $\frac{\partial N}{\partial A_c}$, has been written as δ by both Schneider (1972) and Cess (1976) so that equation (3.1) is simply,

$$\delta = \frac{\partial A_b}{\partial A_c} - \frac{\partial I}{\partial A_c} \quad (4.1)$$

The denominator in equation (4.1) is equivalent to A_c , if the derivatives are taken as the difference between "today's" cloud amount and cloud-free conditions. Since ΔN of equation (3.3) was derived as the difference between the cloud-free and "today's" cloudy net radiation budget, equation (4.1) is equivalent to equation (4.2) for uniform changes in cloud amount,

$$\delta = \frac{\Delta N}{A_c} \quad (4.2)$$

4.1.1 Cloud Amount

Zonal mean profiles of A_c are available from a number of studies; profiles have been deduced from satellite measurements and ground based observations. In keeping with a study based on satellite measurements, A_c is taken from the former source. The global coverage of satellites permit measurements in remote regions, where surface observations are not routinely available. Values of A_c determined satellite data, may not be accurate in an absolute sense, however, they are potentially more accurate in a relative sense (such as a north-south zonal mean profile).

Values of A_c have been taken from studies of Sadler (1969) and Clapp (1964). Both authors collected their statistics from TIROS satellite nephanalyses. Clapp's statistics are drawn from just one year of data (March 1962 through February 1963) while Sadler's are from two years of data (February 1965 through January 1967). Sadler's values are restricted to the 30N-to-30S latitude zone at a 2.5 degree latitude-longitude gridding. Clapp's values are in the 60N-to-60S latitude zone (except for the winter hemisphere in which the poleward limit is 55 degrees latitude) and are at 5 degrees latitude-longitude gridding. Values of A_c were selected from Sadler in the 30N-to-30S zone, from Clapp poleward of the tropical zone and from the ground based observations of Landsberg (as shown by Clapp) in the 65-to-60 latitude zone (65-to-55 latitude zone in the winter hemisphere). Clapp (1964) suggested that the cloudiness derived from the TIROS nephanalyses "tends" to underestimate the amount of scattered cloudiness but to overestimate cloud amount in the broken to overcast range of cloud cover. His values are seasonal averages while values taken from Sadler are averages for the February, May, July and October months. The blended profiles are shown for the four periods and for the annual mean in Figure 4.1. Note that the cloudiness associated with the ITCZ is distinctly shown as a spike in all periods. This spike can be identified with the spike in the total albedo of Figures 2.1 and 2.2. Plotted along with the annual

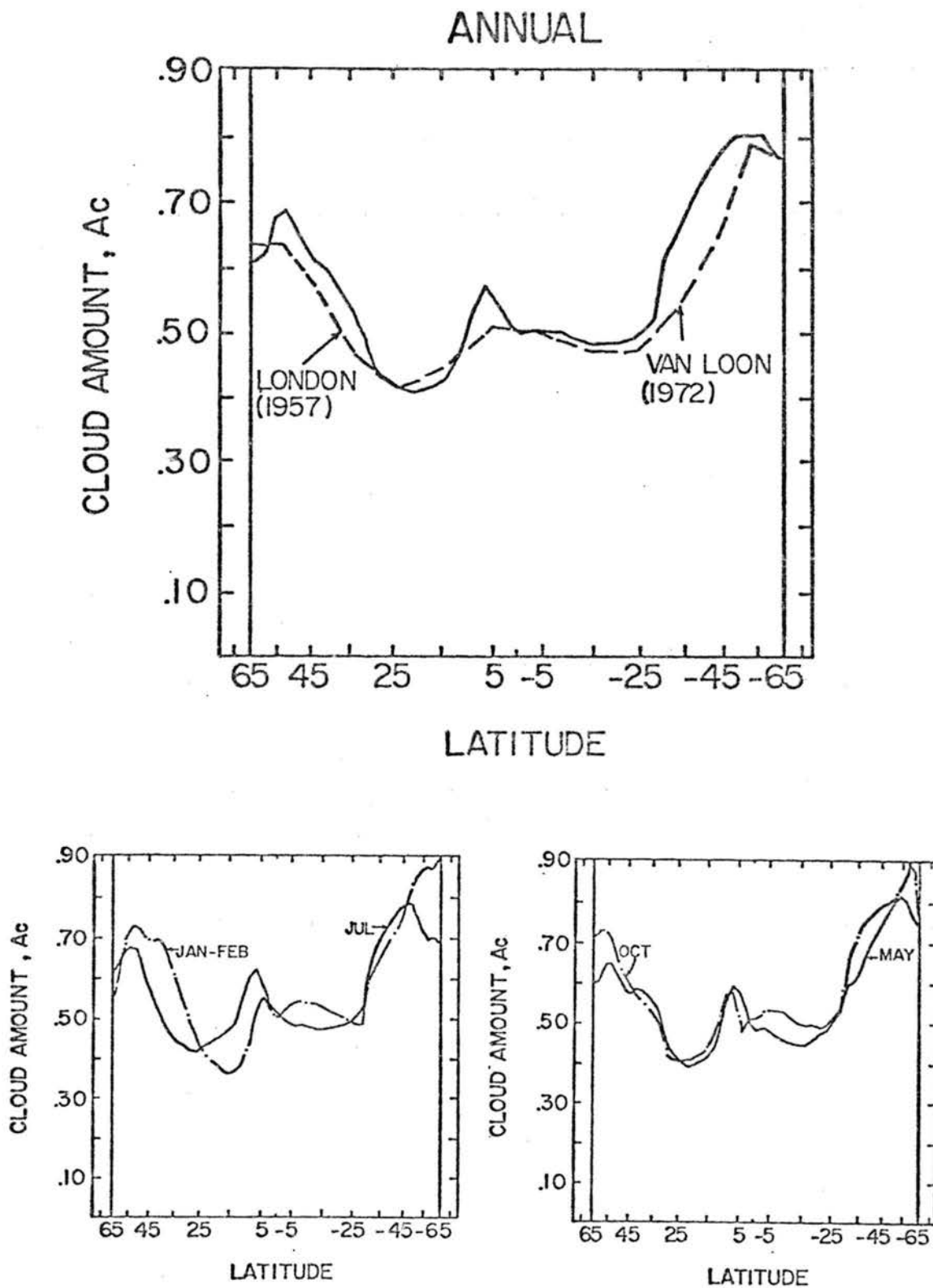


Figure 4.1.: Cloud amount from TIROS nephanalyses.

mean Ac curve derived from the satellite data, is the annual cloud amount of London (1957) for the Northern Hemisphere and an average of the July and January cloud amount of van Loon (1972) for the Southern Hemisphere. Noticeable is the ITCZ spike in the 2.5 degree gridded Ac values derived from satellite brightness data versus the smoothness of the 5 degree gridded ground based Ac values. Cloud amount differences between satellite and surface based data in the latitudes 30 to 65 degree of each hemisphere, are quite possibly due to sampling deficiencies in both sets of statistics.

4.1.2 Absorbed Flux Sensitivity

The zonal mean profiles of the first term on the right side of equation (4.1), $\frac{\partial Ab}{\partial AC}$, is displayed in Figure 4.2. Values range from zero at the polar night to -128 wm^{-2} at 58S latitude in January-February and to -110 wm^{-2} at 58N latitude in July. In the annual mean, $\frac{\partial Ab}{\partial AC}$ drops to -65 wm^{-2} at 50N latitude and becomes larger negative at 58S latitude, a maximum negative value of -86 wm^{-2} . The larger negative annual values in the Southern Hemisphere's extratropics indicate a greater sensitivity of absorbed solar flux to cloudiness over oceans than over land. Plotted along with the May and October profiles, is a zonal profile taken from the radiation model results of Adem (1967). His values are more than twice the satellite results in the tropical zones.

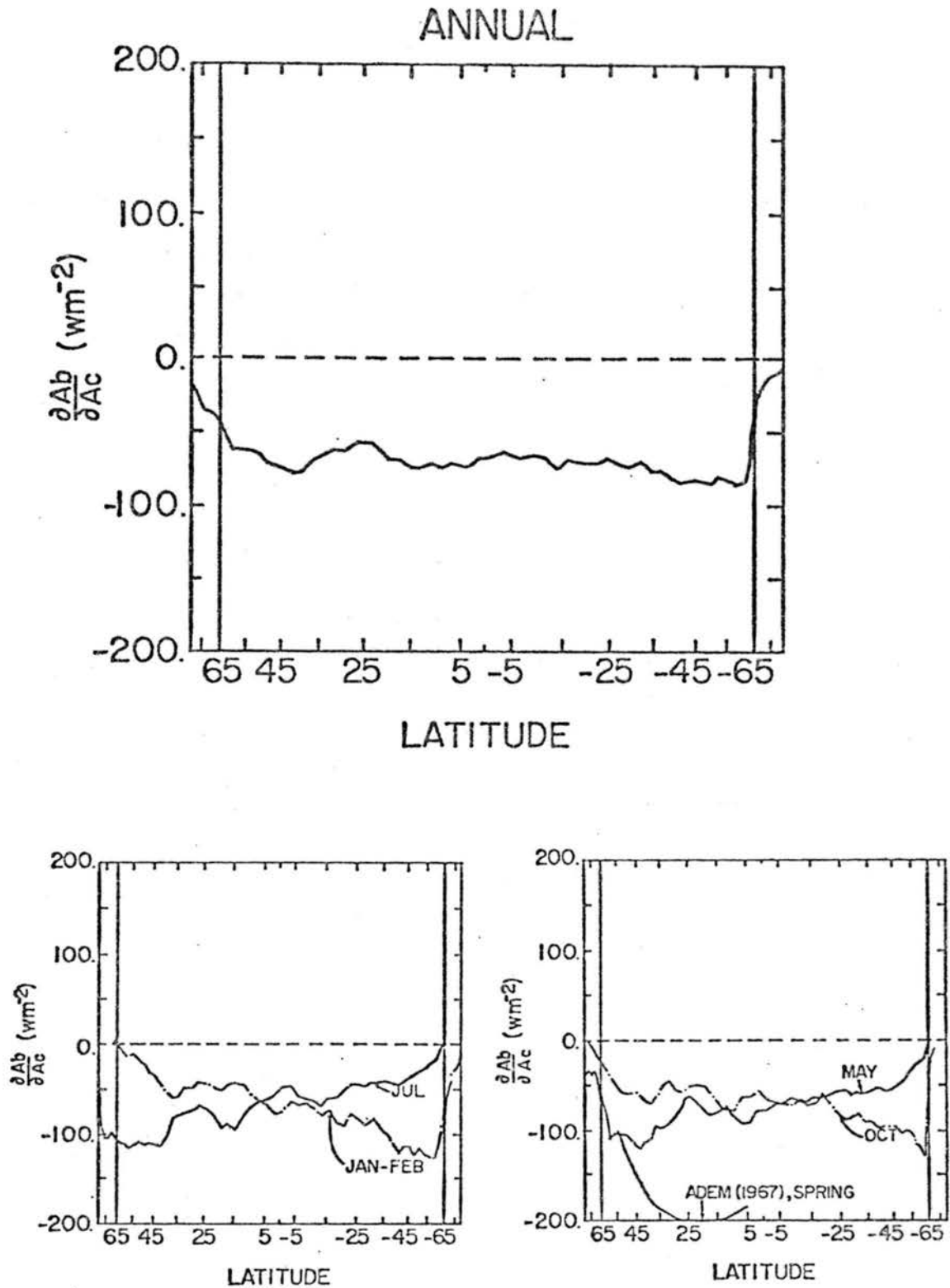


Figure 4.2.: Absorbed flux sensitivity to changes in cloud amount.

4.1.3 Emitted Flux Sensitivity

Emitted Flux Sensitivity ($\frac{\partial I}{\partial A_c}$) is shown in Figure 4.3. Examination of the annual profile reveals that $\frac{\partial I}{\partial A_c}$ values are 30 to 60 percent of the $\frac{\partial A_b}{\partial A_c}$ values in the 65N-to-65S latitude zone. Thus δ , the net radiation budget sensitivity, will be negative in sign for a uniform change in cloud amount (Equation 4.1).

Longwave sensitivity is largest in the latitude zone 17N-to-5N with the peak sensitivity occurring at 14N. A similar peak is not apparent in the tropics of the annual $\frac{\partial A_b}{\partial A_c}$ profile of Figure 4.2. The lack of similarity between $\frac{\partial I}{\partial A_c}$ and $\frac{\partial A_b}{\partial A_c}$ indicates a potentially real increase in long-wave sensitivity, which is not due to error in A_c . Now if the apparent residual cloudiness effect remaining in the cloud-free longwave profile in vicinity of 8N latitude (discussed in Section 2.3.2 Figure 2.6) were removed, then the peak in the $\frac{\partial I}{\partial A_c}$ annual profile would occur at 8N, the mean position of the ITCZ rather than at 14N. The value of $\frac{\partial I}{\partial A_c}$, being less at 8N than at 14N, may be due partially to an inadequacy in using cloud amount to describe thin cirrus cloud cover without including some type of cloud emittance description.

The cause of the systematic difference between the $\frac{\partial I}{\partial A_c}$ profiles in May and October is not clear. The cloud amount profiles in Figure 4.1 do not show this systematic difference. One may conclude:

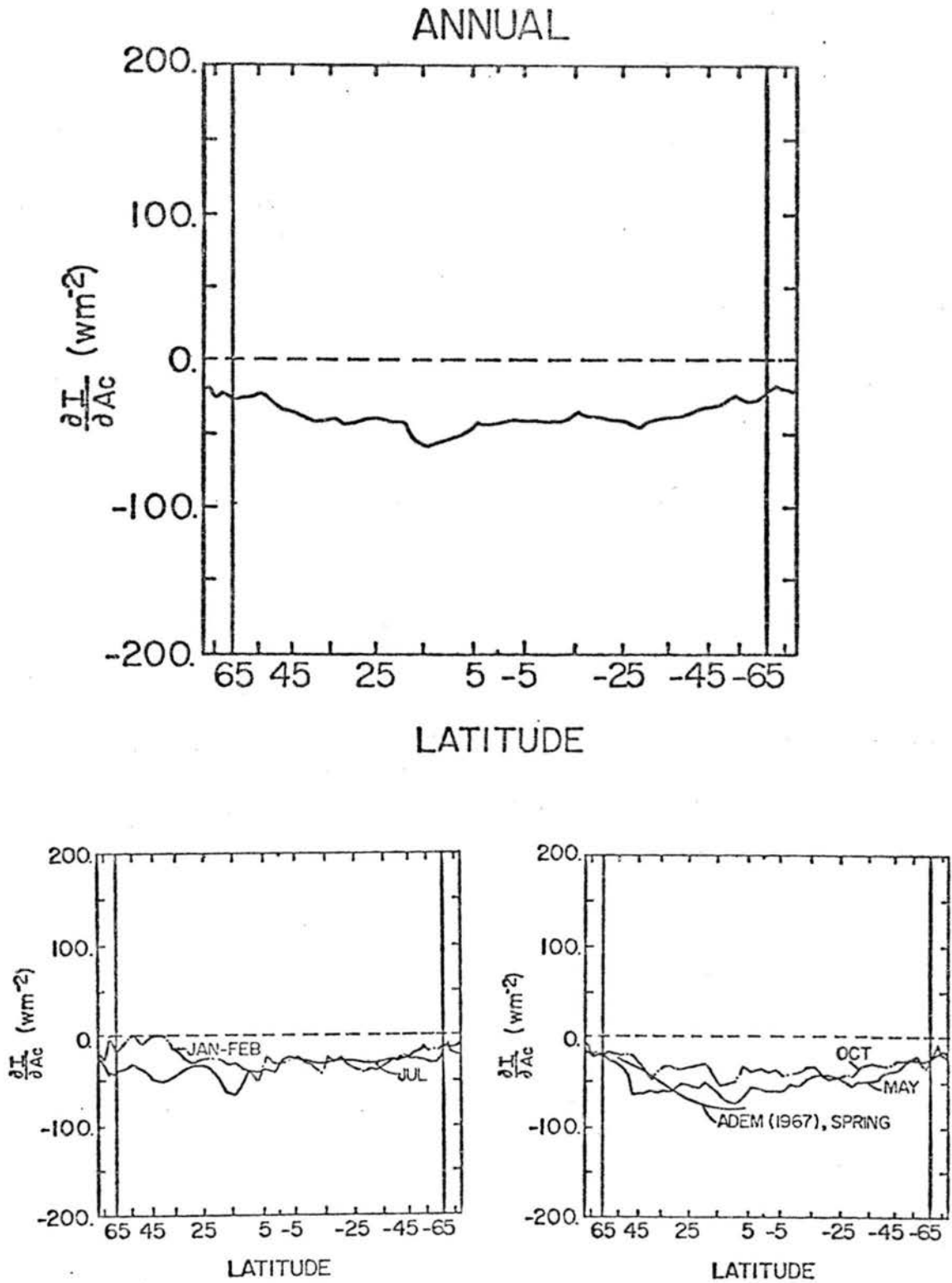


Figure 4.3.: Longwave flux sensitivity to changes in cloud amount.

1) Clouds, in general, are higher during the October period and/or have different emittance values between the two periods, or 2) A_c and/or ΔI are in error. The global $\frac{\partial A_b}{\partial A_c}$ does not show a significant difference between May and October (-72 versus -71 wm^{-2}), whereas $\frac{\partial I}{\partial A_c}$ does show a significant difference (-51 versus -35 wm^{-2}) (See Table 4.5 in Section 4.2). This global result gives support to confirming that the zonal difference between the May and October profiles of $\frac{\partial I}{\partial A_c}$ is due to ΔI and not to A_c . Thus, the cloud characteristics, other than A_c , appear to differ between the two periods.

The radiation model results of Adem (1967) gave a $\frac{\partial I}{\partial A_c}$ profile for the spring season in the Northern Hemisphere which is in close agreement with the satellite results (Figure 4.3). Particularly good agreement is found in the tropical regions.

Differences between the July and January-February $\frac{\partial I}{\partial A_c}$ profiles are understandable. Spatial sampling north of 45N in the latter period was extremely poor. However, a lesser sensitivity to clouds during the January-February period is acceptable, because the outgoing longwave flux would not be as sensitive to clouds over cold continental regions during the winter as it would be over warm continents during the summer.

4.1.4 Net Flux Sensitivity

The annual mean $\frac{\partial Ab}{\partial Ac}$ values were larger negative than the $\frac{\partial I}{\partial Ac}$ values between 65N and 65S latitude, so that application of equation 4.1, yields negative values for δ over the same latitude zones (Figure 4.4). The magnitude of the annual δ values decrease from a maximum in the Northern Hemisphere of -42 wm^{-2} at 55N to a minimum of -16 wm^{-2} at 11N. From 11N δ again increases to a maximum value in the Southern Hemisphere of -60 wm^{-2} at 55S. The minimum value appearing at 11N would appear at 8N (the mean ITCZ position in the Nimbus 3 data) if $\frac{\partial I}{\partial Ac}$ were to be adjusted for apparent residual cloud effects as previously discussed. The δ profile does not show perfect symmetry between hemispheres. A comparison of corresponding latitude zones between hemispheres show greater cloud sensitivity in all zones of the Southern Hemisphere.

An examination of the semi-monthly profiles reveals a negative δ at all latitudes, except at latitudes in and near the polar night. The region of the polar night comprises less than 10 percent of the global area, so that small positive values of δ have little effect on the globally integrated δ value.

Plotted along with the July and January-February profiles are Northern Hemisphere values which were computed by Schneider (1972). As he stated, his values ... are not intended to be taken as quantitatively realistic ... However, they are qualitatively realistic as for the sign

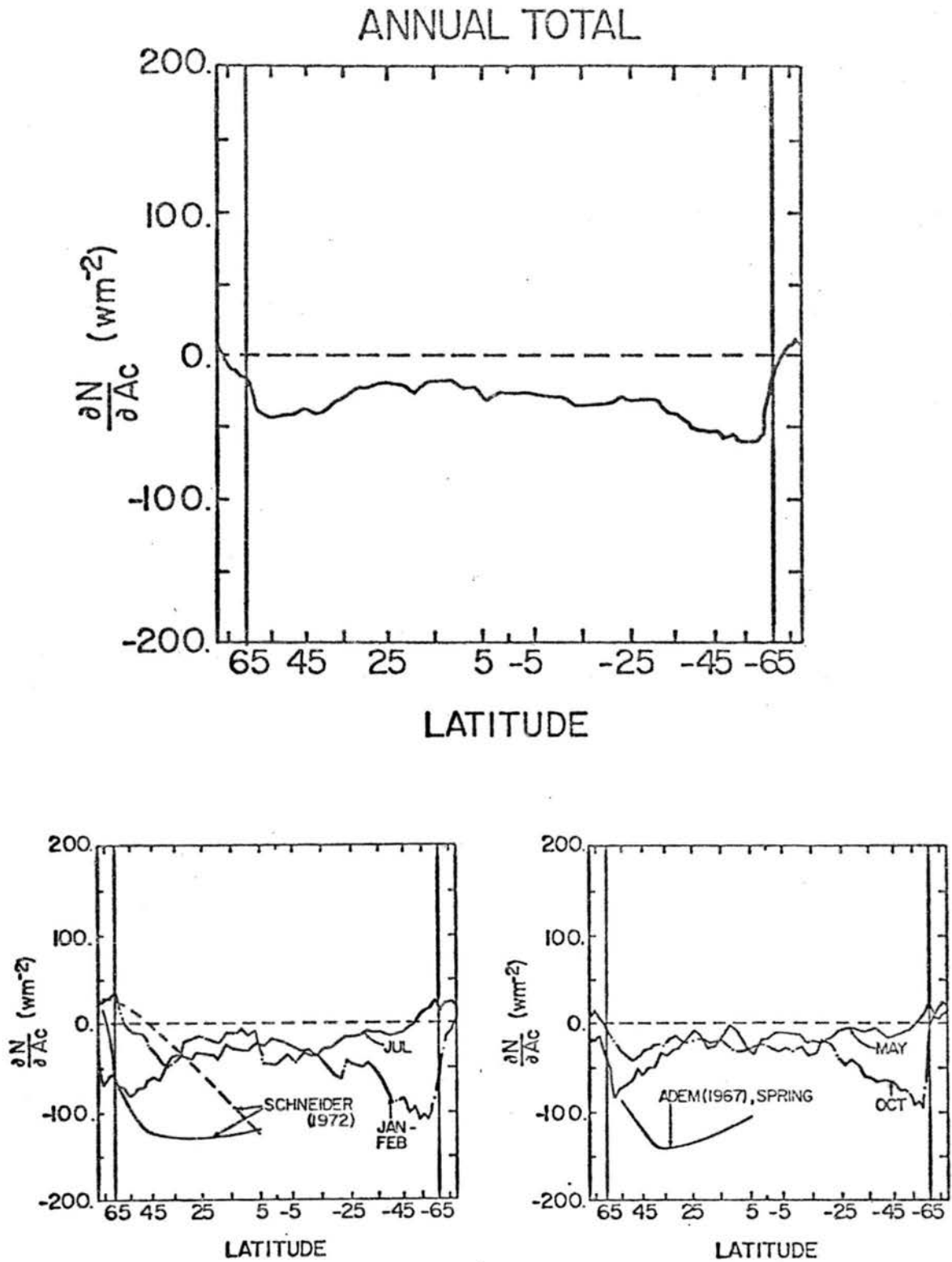


Figure 4.4.: Net flux sensitivity to changes in cloud amount.

on δ . The difference at 5N between Schneider's value and the "quantitatively realistic" value of this study is approximately 100 wm^{-2} . The difference is large because of the near cancellation between the satellite absorbed and longwave flux sensitivities in the tropical latitude zones.

Plotted on the May and October profiles are the results from Adem's (1967) radiation model calculations for the spring season. His values derived from a quantitative parameterized radiation model, compare no better with the results of the present study than did Schneider's qualitative results.

4.1.5 Atmospheric Transport Sensitivity

The sensitivity of the horizontal flux divergence in various latitude zones to changes in cloud amount may be investigated. The energy balance equation integrated throughout the vertical extent of the earth-atmosphere system is written for each latitude zone as:

$$N = \frac{\partial E}{\partial t} + \text{divFa} + \text{divFo} \quad (4.3)$$

where: N = the net radiation flux at the top of the atmosphere.

$\frac{\partial E}{\partial t}$ = the time rate of change of energy content within the zone, or rate of heat storage by atmosphere, land, ocean and cryosphere.

divFa = meridional atmospheric energy flux transported out of the zone, or flux divergence.

divFo = meridional oceanic energy flux transported out of the zone, or flux divergence

Now just the atmospheric flux divergence term, divF_a , will be permitted to respond to changes in cloud amount. All rate of heat storage terms, $\frac{\partial E}{\partial t}$ and the oceanic flux divergence, divF_o , are assumed not to vary with cloud amount changes. This assumption has some validity only if the immediate or near simultaneous response of the system to a change in cloud amount is considered. Thus, the ocean transport and heat storage terms are permitted to respond only at some later time. This type of analysis permits one to observe the atmosphere response as an upper limit type of response.

Under the assumed conditions, the sensitivity to a cloud amount change, ΔA_c , may be expressed as:

$$\frac{\partial N}{\partial A_c} \Delta A_c = \frac{\partial \text{divF}_a}{\partial A_c} \Delta A_c \quad (4.4)$$

or in simpler notation as:

$$\Delta N = \Delta \text{divF}_a \quad (4.5)$$

Figures 4.5, 4.6 and 4.7 show the response in ΔdivF_a for three different latitude zones with a $\Delta A_c = \pm 1$. The mean curves for divF_a are taken from Oort and Vonder Haar (1976). The vertical bars represent a $\pm \Delta \text{divF}_a$ response to a $\pm \Delta A_c$. They were computed from the tabulated $\frac{\partial N}{\partial A_c}$ in Table 4.1. They show that a change in cloud amount,

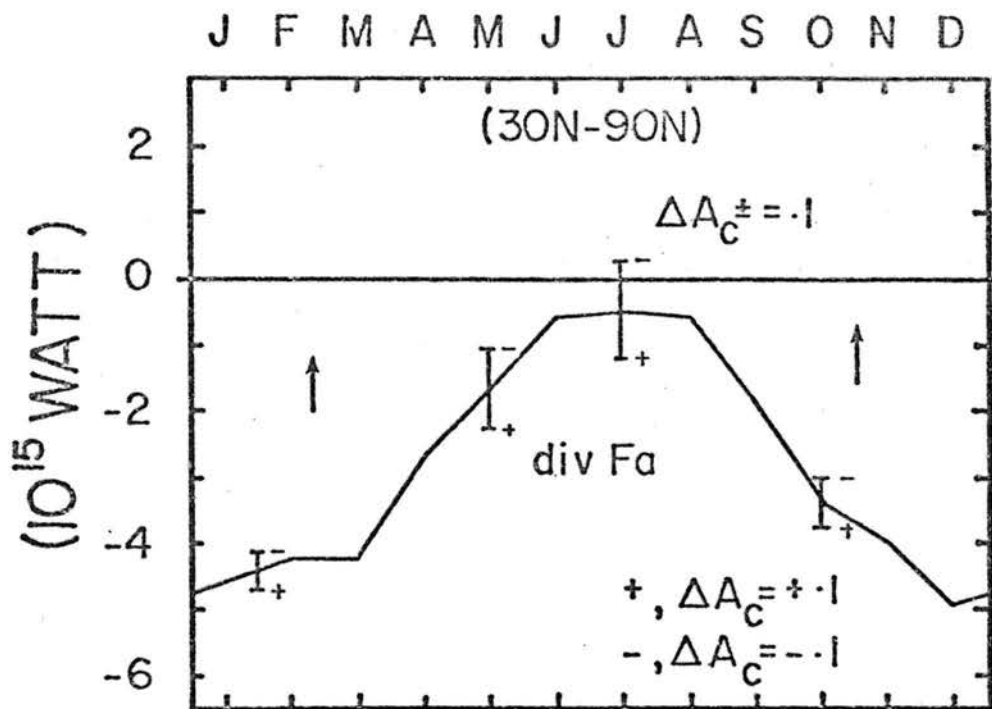


Figure 4.5. Atmospheric mean meridional energy flux divergence for the 30N-to-90N polar cap.

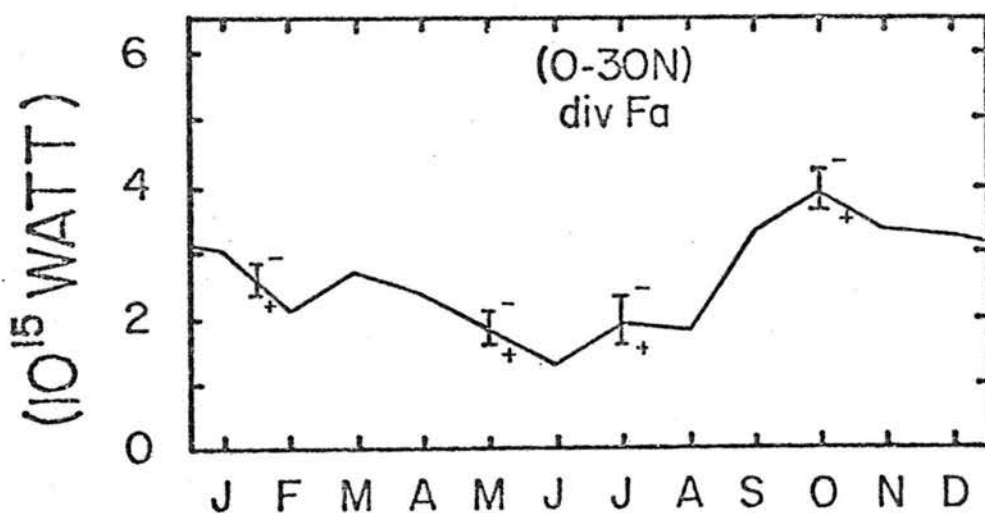


Figure 4.6.: Atmospheric mean meridional energy flux divergence for the equator-to-30N latitude zone.

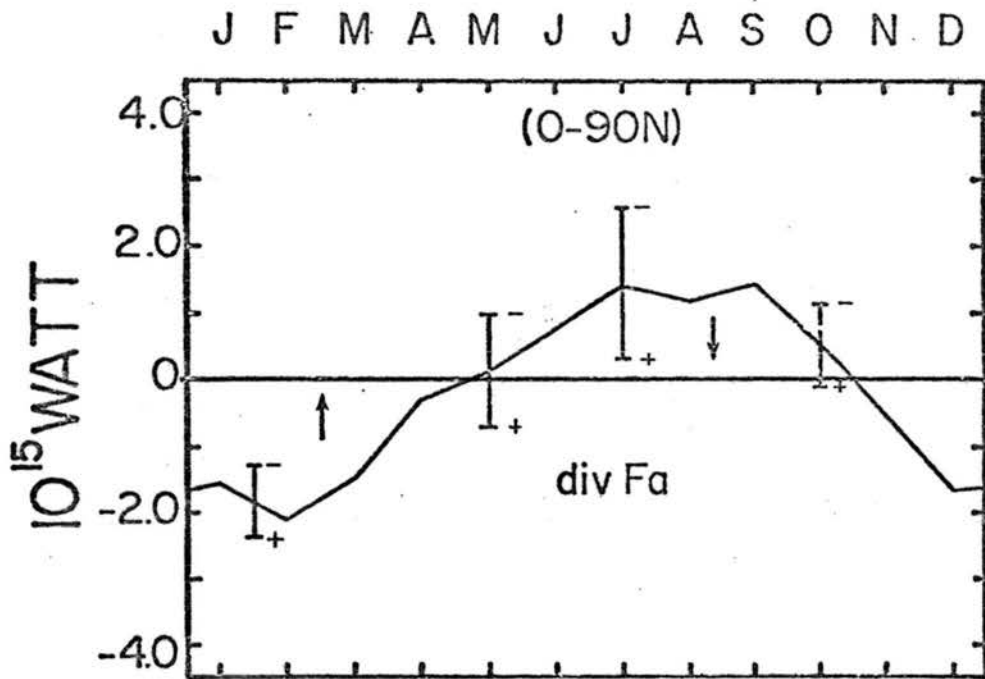


Figure 4.7.: Atmospheric mean meridional energy flux divergence for the equator-to-90N polar cap.

Table 4.1: Zonal Values of $\frac{\partial N}{\partial AC}$ (wm^{-2})

<u>North Latitude</u>	<u>May 1-15</u>	<u>July 16-31</u>	<u>Oct. 3-17</u>	<u>Jan. 21-Feb. 3</u>
60-70	-34.3	-60.5	-5.6	+21.6
50-60	-72.8	-81.5	-32.6	-9.1
40-50	-54.1	-64.0	-31.5	-30.1
30-40	-33.4	-35.4	-20.3	-41.9
20-30	-11.1	-33.8	-19.7	-14.8
10-20	-18.8	-26.8	-19.4	-10.9
0-10	-29.1	-24.8	-27.4	-27.9

which is quite large in magnitude, will not change the basic shape of the annual divFa profile.

The northward energy transport (T_n) by the atmosphere is opposite in sign to divFa for the polar cap 30N-90N and 0-90N in Figures 4.5 and 4.7, respectively; a positive divergence is an outflow from the cap and a negative divergence is an inflow to the cap, or a convergence. For the polar cap,

$$T_n = -\text{divFa} \quad (4.6)$$

A $\Delta\text{Ac} = +.1$ in the 30N-90N cap of Figure 4.5 will require of the atmosphere, as an upper limit, a T_n increase of $.3 \times 10^{15}$ watt in January-February and $.75 \times 10^{15}$ watt in July. Likewise, for a $\Delta\text{Ac} = -.1$ a response in T_n of the same magnitude and opposite in sign would be required. Note that in July the sign of T_n changes with $\Delta\text{Ac} = -.1$ so that a southward energy transport is required of the atmosphere.

A similar analysis for the entire north polar cap, 0-90N, for a $\Delta\text{Ac} = \pm .1$ uniformly distributed over 0-90N is shown in Figure 4.7. The sign of T_n would change during the equinoxes with a ΔAc change; however, the overall shape of the divFa or T_n profile with time throughout the year would not change even with such a large change in cloud amount.

The effect on atmospheric transport of removing stratus and stratocumulus clouds east of the subtropical high pressure center in the Eastern Pacific Ocean along the coast of North America is examined. However, in this case the contribution of $\frac{\partial I}{\partial A_c}$ to $\frac{\partial N}{\partial A_c}$ is ignored, so that

$$\frac{\partial N}{\partial A_c} = \frac{\partial A_b}{\partial A_c} \quad (4.7)$$

Ignoring $\frac{\partial I}{\partial A_c}$ is a reasonable assumption in light of this cloud type which prevails near the top of a temperature inversion in the subtropical high pressure regions over a relatively cold ocean surface. Thus, $\frac{\partial I}{\partial A_c}$ would be quite small compared to $\frac{\partial A_b}{\partial A_c}$.

Shown in Table 4.2 is the cloud amount fraction of the type associated with the eastern Pacific high in the latitude zone, 15N-35N (after Miller and Feddes, 1971). The ΔA_c for each semi-monthly period is negative which indicates a removal of the cloud. The persistence and area of coverage of the stratus type cloud is largest in July and smallest in January of the four periods. Similarly, $\frac{\partial A_b}{\partial A_c}$ is largest in July, which may be associated with the high, bright clouds of the southwestern Indian Monsoon, as much as with the lower stratus type clouds. The zonal mean sensitivities of the present work do not permit distinguishing sensitivity by cloud type and by longitude. For the most part, the change in $\frac{\partial A_b}{\partial A_c}$ with season is due to seasonal solar radiation weighting on the cloud albedo. The January solar

Table 4.2: Sensitivities associated with removal of stratus-type clouds from the Eastern Pacific Ocean in the 15N-35N latitude zone.

	<u>Jan.-Feb.</u>	<u>May</u>	<u>July</u>	<u>Oct.</u>
ΔA_c	-.017	-.022	-.045	-.028
$\frac{\partial A_b}{\partial A_c} (10^{15}_w)$	-3.84	-6.42	-6.46	-4.02
$\Delta \text{divFa} (10^{15}_w)$.07	.14	.29	.11
$T_n (40N) (10^{15}_w)$	3.7	2.4	1.6	3.5
% $T_n (40N)$	1.8	5.9	18.2	3.2

weighting in the 15N-35N zone is approximately 65 percent of the July solar weighting; it accounts for 85 percent of the difference in $\frac{\partial Ab}{\partial Ac}$ between the two periods.

Application of equations (4.7), (4.4) and (4.5) yields ΔdivFa as shown in Table 4.2. The ΔdivFa is four times larger in July than in the January-February period. The present day Northward transport of energy by the atmosphere across the 40N latitude circle (shown in Table 4.2 as T_n) was taken from Oort and Vonder Haar (1976). Inspection of their Table 9 reveals that all of the divFa in the 15N-35N zone goes into positive northward transport across the 40N latitude circle. If it is assumed that the direction of energy transport by the atmosphere remains identical to that of the present climate regime with a removal of the subtropical high pressure stratus type clouds in the eastern Pacific Ocean, then the percent increase in required energy transport may be calculated. (It must be assumed under a global radiative equilibrium constraint that a ΔN of equal magnitude but opposite in sign takes place north of 40N, so that this energy transport may occur). These percentages range from 1.8 in January-February to 18.2 in July (Table 4.2).

A more realistic study than has been discussed in the previous case studies is to determine the sensitivity of divFa in the 0-30N latitude zone to year-to-year changes in cloud amount which are associated with the intertropical convergence zone (ITCZ). Cloud amount interannual

variability for various latitude zones within the 30N-30S zone has been determined by Murakami (1975). He derived it from 7 years of gridded TIROS nephanalyses. The variability of ΔAc in the 0-15N latitude zone for each of the four semi-monthly periods was taken from his Figure 6. This zone was selected as being representative of the variability in cloud amount associated with the ITCZ. These ΔAc values are shown in Table 4.3. Cloud amount variability in the July period is twice that in the January-February and October periods.

The $\Delta divFa$ has been computed for the entire 0-30N latitude zone for a ΔAc in the 0-15N latitude zone with equations (4.4) and (4.5). As expected, ΔAc and $\Delta divFa$ are opposite in sign and $divFa$ is two or more times larger in July than in any other period. Using the same reasoning as used in the previous case on stratus clouds in partitioning $\Delta divFa$ into northward energy transport in the same proportion as the $divFa$ of today's climate regime, the percent change in northward energy transport at both 30N and the equator were computed as shown in Table 4.3. A decrease in cloud amount by 0.05 in the 0-15N zone during July may be balanced by both a northward energy transport across 40N and a southward energy transport across the equator, which are nearly equal in magnitude. A similar argument holds for the other periods in which the sensitivities are considerably less.

Table 4.3: Sensitivity of the atmospheric energy transport out of the 0-30N latitude zone to interannual variations of cloud amount in the 0-15N latitude zone.

	<u>Jan.-Feb.</u>	<u>May</u>	<u>July</u>	<u>Oct.</u>
ΔAc	$\pm .025$	$\pm .040$	$\pm .050$	$\pm .025$
$\frac{\partial N}{\partial Ac} (10^{15}_w)$	-1.10	-1.15	-1.81	-1.67
$\Delta \text{divFa} (10^{15}_w)$	$\bar{\mp} .028$	$\bar{\mp} .046$	$\bar{\mp} .090$	$\bar{\mp} .042$
fn	1.00	.94	.26	.87
$T_n(30N)(10^{15}_w)$	4.4	1.7	0.5	3.4
% $T_n(30N)$	$\bar{\mp} 0.6$	$\bar{\mp} 2.5$	$\bar{\mp} 4.6$	$\bar{\mp} 1.1$
$T_n(\text{Eq.}) (10^{15}_w)$	1.9	-.1	-1.4	-.5
% $T_n(\text{Eq.})$	0	± 3.0	± 4.8	± 1.0

The foregoing hypothetical cases were presented to demonstrate what may be considered as an upper limit to the immediate response of the atmosphere energy transports to changes in cloud amount. The significance of the atmospheric response should be determined through climate model sensitivity studies.

4.2 Global and Hemispherical Sensitivity

Global and hemispherical intergrated values of various parameters appear in Table 4.5. The poleward limits of integration used were 65N and 65S latitude. The 29 month set of measurements because of the large sample size, allows for better global integrated mean statistics than the Nimbus 3 data set. If global (90N-90S) annual mean cloudy statistics are compared between the Nimbus 3 and the 29 month data sets, one sees that the 29 month set shows a planet which is both brighter and colder than the Nimbus 3 set. This comparison is shown in Table 4.4.

Table 4.4 - Annual Global Statistics (90N-90S)

	ALBEDO	LONGWAVE (wm^{-2})
29 month	.301	235.7
Nimbus 3	.280	242.0

Table 4.5 - Summary of satellite derived statistics.

	MAY	JULY	OCT.	JAN-FEB	NIMBUS 3 ANNUAL	29 MONTHS ANNUAL	
65N-65S	$\frac{\partial Ab}{\partial Ac}$	-72.0	-65.8	-70.9	-67.5	-69.8	-83.8
	$\frac{\partial I}{\partial Ac}$	-50.5	-36.4	-34.8	-26.0	-36.6	-46.7
	$\frac{\partial N}{\partial Ac}$	-21.5	-29.4	-36.1	-41.5	-33.2	-37.1
	αc	0.359	.347	0.355	0.350	0.354	0.389
	Ac	0.553	0.557	0.579	0.578	0.568	0.568
	αT	0.268	0.263	0.269	0.268	0.268	0.288
	αm	0.156	0.158	0.151	0.156	0.155	0.155
65N-0	$\frac{\partial Ab}{\partial Ac}$	-88.3	-87.1	-58.1	-43.3	-66.2	
	$\frac{\partial I}{\partial Ac}$	-55.3	-43.4	-34.1	-22.4	-38.5	
	$\frac{\partial N}{\partial Ac}$	-33.0	-43.7	-24.0	-20.9	-27.7	
	αc	0.376	0.369	0.345	0.348	0.360	
	Ac	0.522	0.525	0.530	0.534	0.528	
	αT	0.277	0.274	0.263	0.275	0.271	
	αm	0.169	0.169	0.170	0.191	0.172	
65S-0	$\frac{\partial Ab}{\partial Ac}$	-55.6	-44.5	-83.7	-91.6	-73.4	
	$\frac{\partial I}{\partial Ac}$	-45.7	-29.3	-35.5	-29.5	-34.8	
	$\frac{\partial N}{\partial Ac}$	-9.9	-15.2	-48.2	-62.1	-38.6	
	αc	0.337	0.316	0.354	0.341	0.346	
	Ac	0.585	0.590	0.629	0.622	0.608	
	αT	0.253	0.243	0.273	0.264	0.264	
	αm	0.134	0.138	0.136	0.137	0.137	

The annual mean global albedo and longwave flux of the 29 month data set is 7.5 percent higher and 2.6 percent lower, respectively, than that of the Nimbus 3 data set. These differences are assumed to result from the effects of cloudiness differences between the two data sets. Therefore, the sensitivity of the radiation budget to cloud amount within the 65N to 65S latitude zone may be estimated for the 29 month data from the Nimbus 3 statistics within that latitude band. The Nimbus 3 cloudy albedo and cloudy longwave flux values are increased by 7.5 percent and decreased by 2.6 percent, respectively. Nimbus 3 cloud-free statistics are taken together with the cloudy statistics to derive radiation budget sensitivity values for the 29 month set. Table 4.6 compares sensitivity factors derived from both sets of data.

Table 4.6 - Annual Sensitivities (65N-65S) (wm^{-2})

	$\frac{\partial \text{Ab}}{\partial \text{Ac}}$	$\frac{\partial \text{I}}{\partial \text{Ac}}$	$\frac{\partial \text{N}}{\partial \text{Ac}}$
29 month	-83.8	-46.7	-37.1
Nimbus 3	-69.8	-36.6	-33.2

Sensitivities from the 29 month data set are larger than those from the Nimbus 3 data set in all radiation budget components, ie. larger by 14 wm^{-2} , 10 wm^{-2} , and 4 wm^{-2} in absorbed shortwave, emitted longwave, and net flux, respectively.

Global sensitivity factors have been derived in theoretical studies by others. Annual global cloud amount used in this study is 0.568 (Table 4.5, $A_c = 0.568$ for 65N-to-65S). The global mean value of A_c is quite uncertain. Schneider (1972) used $A_c = 0.500$, Hoyt (1976) in his model used $A_c = 0.532$, and Cess used $A_c = 0.540$. So that a comparison might be made on an equal base cloud amount, all sensitivity factors have been adjusted to an $A_c = 0.500$, and to a solar constants of 1360 wm^{-2} . These normalized sensitivity values are shown in Table 4.7.

Table 4.7 - Annual Global Sensitivities (wm^{-2})

	$\frac{\partial A_b}{\partial A_c}$	$\frac{\partial I}{\partial A_c}$	$\frac{\partial N}{\partial A_c}$
29 Month** (present study)	-95.2	-53.1	-42.4
Nimbus 3** (present study)	-79.3	-41.6	-37.7
Adem (1967)*	-178.1	-58.4	-119.7
Schneider (1972)	-129.2	-74.6	-54.6
Cess (1976) NH	-88.4	-91.0	+2.6
SH	-81.6	-81.0	-0.6
Hoyt (1976)		-36.7	
Wang & Dimoto (1974)		-66.2	
* Equator to 60N			

** 65N-to-65S

The $\frac{\partial AB}{\partial AC}$ computed by Cess agrees well with results of the present study; they should be in agreement since both studies used cloud-free planetary albedo data, which were derived from the same four Nimbus 3 semi-monthly periods. The $\frac{\partial I}{\partial AC}$ values in Table 4.7 increase from Hoyt's -36.7 wm^{-2} to the largest value, -91.0 wm^{-2} of Cess. The larger $\frac{\partial I}{\partial AC}$ of Cess is quite interesting since it nearly cancels $\frac{\partial Ab}{\partial AC}$ thus giving virtually no global net flux sensitivity to cloud amount. The net flux sensitivities, $\frac{\partial N}{\partial AC}$, of all other studies are large negative values.

As discussed in Section 2, radiative transfer calculations with models of Schneider (1972) and Cox et al., (1976) did not give a global mean value of cloud free longwave flux equivalent to that derived from the satellite measurements. The global atmospheric temperature and moisture vertical profiles used in the model calculations were undoubtedly not identical to the global atmospheric conditions existing at the time of the Nimbus 3 measurements. However, large discrepancies did exist between the two model results; the satellite measurements fell between them. Given identical model atmospheres and an opaque cloud top at 5.5 km, the model results discussed by Schneider gave $\frac{\partial I}{\partial AC} = -74.6 \text{ wm}^{-2}$, while the model discussed by Cox et al. (1976) gave $\frac{\partial I}{\partial AC} = -52.3 \text{ wm}^{-2}$. These values differ significantly. The principle difference between the model results may be seen in Figure 4.8. Longwave flux out the top of the atmosphere for a cloud-free

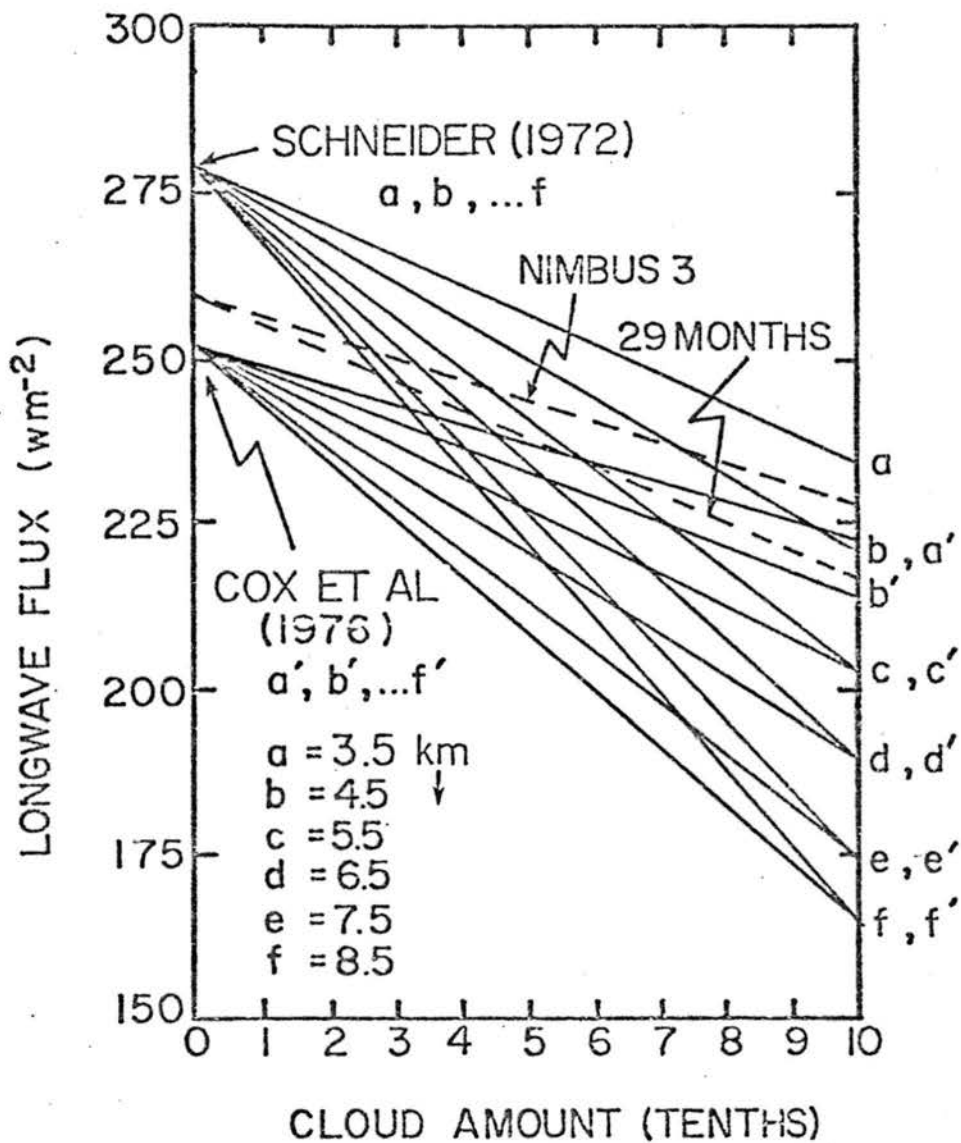


Figure 4.8.: Longwave flux emitted by the earth-atmosphere system as a function of cloud amount for various heights of radiometrically black clouds.

atmosphere differs between the models by 28 wm^{-2} . Differences decrease as both cloud amount and cloud top height of opaque cloudiness increase. In fact, the model results converge between cloud top heights of 5.5 km and 6.5 km for a total cloud cover case (10/10).

Curves of $\frac{\partial I}{\partial A_c}$ for both the Nimbus 3 and the 29 month set are plotted in Figure 4.8. The curves were determined by applying the $\frac{\partial I}{\partial A_c}$ of each set (Table 4.6) to the global integrated (90-90S) longwave flux value of each set (Table 4.5) at a 0.568 cloud amount. A comparison of model derived longwave flux values, for various cloud top heights at 10/10 cloud amount, with the satellite values, extrapolated to 10/10 cloud amount, yields an estimate of the effective cloud top height in the satellite data. The effective cloud top height for the Nimbus 3 data set is between 3.5 and 4.5 km using the results from the model of Cox et al, and below 3.5 km from Schneider's model results. The cloud tops are between 4.5 km and 5.5 km in the 29 month data set from both results.

Schneider (1972) found that he could obtain a zero net radiation balance at the top of the atmosphere by inserting into his longwave transfer calculations opaque cloud tops at a 5.5 km height. The satellite results indicate that effective opaque cloud top heights are at some height below 5.5 km. However, if global cloud amount is taken as 0.500 rather than 0.568, the slope of the

satellite curves in Figure 4.8 will increase. Under this condition, the 29 month curve will indicate effective opaque cloud top heights to be at 5.5 km.

The difference in $\frac{\partial I}{\partial A_c}$ between Schneider's and the 29 month satellite values (Table 4.7) might be largely attributed to uncertainty in the global model atmosphere used in the radiative transfer calculations to compute cloud-free longwave flux. Cess (1976) obtained $\frac{\partial I}{\partial A_c}$ by differentiating a parametric function, which described the emitted longwave flux as a function of surface temperature and cloud amount. He derived similar empirical relationships for both the Northern and Southern Hemispheres by applying a least squares fit to annual-mean-zonal dependant variables. Differentiating both expressions with respect to cloud amount yielded -91 wm^{-2} and -81 wm^{-2} for $\frac{\partial I}{\partial A_c}$ of the Northern and Southern Hemispheres, respectively. The latitude dependence of his $\frac{\partial I}{\partial A_c}$ within each hemisphere was less than $\pm 4 \text{ wm}^{-2}$, where as in the present study, it is greater than $\pm 10 \text{ wm}^{-2}$ in the annual mean case. It is not clear as to why his $\frac{\partial I}{\partial A_c}$ values are so much larger in absolute value than those of the present study.

The $\frac{\partial I}{\partial A_c}$ derived from both radiative transfer model results are shown in Table 4.8 for various effective cloud top heights. The $\frac{\partial I}{\partial A_c}$ value of the satellite method (-53 wm^{-2} with $A_c = 0.50$) is equivalent to a sensitivity with effective cloud tops at 5.5 km in Cox's model and between 3.5 km and 4.5 km in Schneider's model.

Table 4.8.: Annual and Global Mean Longwave Sensitivity derived from Models

Cloud Top Height (km)		3.5	4.5	5.5	6.5	7.5	8.5
Cox et al (1976)	$\frac{\partial I}{\partial \Delta C}$ (wm^{-2})	-31.0	-40.0	-52.0	-65.0	-80.0	-91.0
Schneider (1972)	$\frac{\partial I}{\partial \Delta C}$ (wm^{-2})	-45.0	-59.0	-75.0	-89.0	-104.0	-113.0

The $\frac{\partial I}{\partial A_c}$ of Cess allows for effective cloud tops in the Northern Hemisphere at 8.5 km and 6.5 km and in the Southern Hemisphere at 7.5 and 5.5 km, using the Cox and Schneider model results, respectively. Not only is there a large disparity between $\partial I/\partial A_c$ values of various studies but there is a large disparity between the effective cloud top heights of the various studies.

The differences in $\frac{\partial A_b}{\partial A_c}$ in Table 4.7 may be examined in terms of cloud albedo (α_c) by differentiating equation (4.8) with respect to cloud amount,

$$\alpha_t = \alpha_c A_c + \alpha_{min} (1 - A_c) \quad (4.8)$$

to obtain

$$\frac{\partial \alpha_t}{\partial A_c} = \alpha_c + A_c \frac{\partial \alpha_c}{\partial A_c} - \alpha_{min} + (1 - A_c) \frac{\partial \alpha_{min}}{\partial A_c} \quad (4.9)$$

It is assumed that the albedo of clouds and of the cloud-free atmospheres are not functions of the cloud amount, so the equation (4.9) reduces to:

$$\frac{\partial \alpha_t}{\partial A_c} = \alpha_c - \alpha_{min} \quad (4.10)$$

Thus, it is seen that $\frac{\partial \alpha_t}{\partial A_c}$ is a function of just the albedo above clouds and the albedo of the cloud-free earth and atmosphere system. Global values of α_c and α_{min} which were used by Schneider are 0.50 and 0.12, respectively. Values of α_c and α_{min} derived from the 29 month and Nimbus 3 data sets for the latitude zone 65N-to-65S are 0.389 and 0.155, respectively. The satellite derived α_c is

based on a cloud amount of 0.568. Now if an A_c of 0.50 is used in equation (4.8) along with the quasi-global albedo data of the 29 month satellite data set, one obtains an α_c of 0.421, and a $\frac{\partial \alpha_c}{\partial A_c}$ of 0.266 from equation (4.10). The 0.266 value is 30 percent less than Schneider's value of 0.380. Therefore, accuracy in the sensitivity of absorbed shortwave flux to cloud amount is critically dependent upon an accurate measure of the planetary albedo of clouds and of the cloud-free planet.

Schneider (1972) demonstrated that a variation in cloud top height by 0.6 km, while conserving the vertical temperature lapse rate and cloud amount, could cause a 2°K change of the same sign in surface temperature. He also demonstrated that an increase (decrease) in cloud amount, while conserving the vertical temperature lapse rate and the cloud top height, should decrease (increase) the surface temperature. The magnitude of his temperature change result is model dependent. Since both his study and the present one show $\frac{\partial N}{\partial A_c}$ to be negative, then the sign of the surface temperature change under identical conditions should be the same in both studies. Thus, the results of the present study indicate that an increase (decrease) in cloud amount, while conserving the vertical temperature lapse rate and the cloud top height, should decrease (increase) the surface temperature.

5. CONCLUSIONS AND RECOMMENDATIONS

A number of conclusions are drawn from this study. The principle conclusion is that a uniform increase (decrease) in global cloud amount, with all other factors held constant, will cause a negative (positive) planetary radiation budget to prevail. If the present atmospheric vertical temperature lapse rate is conserved, then the global mean surface temperature will decrease (increase) until radiative equilibrium is restored. Without compensating changes in cloud albedo, emittance, transmittance and height, a uniform change in cloud amount does appear to be a significant climate feedback mechanism.

The cloud effect for the globe as a whole is larger in the absorbed shortwave flux than in the longwave emitted flux. Similarly, the sensitivity of the absorbed shortwave flux to changes in cloud amount is larger than the sensitivity in longwave emitted flux to changes in cloud amount. The cloud effect in the absorbed shortwave flux is significantly larger over ocean surfaces than over land surfaces.

The presence of clouds act to reduce the amplitude of the annual variation of the planetary net radiation budget. The amplitude reduction is mostly in the planetary longwave flux term, because the presence of clouds decrease the transmittance of the atmosphere to longwave flux upwelling from the earth's surface. Thus, the full amplitude of the annual variation in surface temperature is not seen in the

planetary longwave flux. The annual variation of the planetary net radiation budget is approximately $\pm 15 \text{ wm}^{-2}$. This variation is apparently real and is caused by :

- 1) an annual cycle in the incoming solar flux, due to the eccentricity of the earth's orbit around the sun,
- 2) an annual cycle in the longwave flux, due to large amplitude continental surface temperature changes of the Northern Hemisphere,
- 3) a semi-annual variation in albedo due to apparent north-south seasonal migration of the sun's path with latitude over dark tropical zones relative to brighter extra tropical zones, and
- 4) a small residual variation, due to measurement and sampling biases and to time varying atmospheric and surface albedoes.

The cloud-free planetary albedo determined from satellite shortwave radiance measurements was shown to be higher than the value commonly associated with planetary heat budget studies.

It is recommended that a similar study be carried out with the scanning measurements from Nimbus 6 and Nimbus G Earth Radiation Budget (ERB) experiments. This would permit verification of the results derived from the small satellite sample used in this study. An assessment of inter-annual variability, if any exists, in the cloud effect on the planetary radiation budget could also be done.

It is further recommended that the results of this study be implemented in planetary heat budget and "climate" models to test model sensitivity to them.

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APPENDIX A

The Ocean Reflectance Model

"Ocean" and "Cloud-land" directional reflectance and bi-directional reflectance models which were applied to Nimbus 3 MRIR satellite measurements are described in Raschke et al. (1973). The directional reflectance part of the "cloud-land" and ocean models gives the dependence of directional reflectance on solar zenith angle. It is a model derived from a limited number of measurements by a number of authors.

The effects of application of the directional reflectance model to a directional reflectance value, $r(\xi)$, may be simulated. This is accomplished by assuming a value for the directional reflectance at zero solar zenith angles, $r(\xi=0)$, and then applying the model. It is described mathematically by:

$$r(\bar{\xi}) = R(\bar{\xi}) r(\xi=0) \quad (\text{A.1})$$

with ξ = solar zenith angle

$r(\bar{\xi})$ = the daily mean directional reflectance

$r(\xi=0)$ = directional reflectance at zero solar
zenith angle

and

$$R(\bar{\xi}) = \frac{\int_{T_d}^{T_n} \frac{r_{\xi}(t)}{r(\xi=0)} \cos \xi(t) dt}{\int_{T_d}^{T_n} \cos \xi(t) dt} \quad (\text{A.2})$$

with T_d, T_n = Time of sunrise and sunset, respectively.

The expression, $\frac{r(\xi)}{r(\xi=0)}$, is the directional reflectance model as given in Table A.1. The table values were taken from Sikula and Vonder Haar (1972), which were tabulated from graphs in Raschke et al. (1973). In Table A.1., there are two reflectance models: a cloud-land model and a cloud-free ocean model. For the application of the cloud-land model in estimating albedo over clouds as a function of solar zenith angle, a value of 0.25 was selected for $r(\xi=0)$. This value was taken from earlier work of Ruff et al. (1967) in which they investigated the angular dependence of solar radiation from clouds with TIROS IV measurements.

The ocean model was applied to obtain an upper limit for albedo over a cloud-free ocean. This upper limit is applied as an albedo cutoff; values larger than it are considered to contain cloud contamination while those lower are considered cloud-free. The value for $r(\xi=0)$ was taken from Conover (1965) as 0.09. Of course, this value is high, but it is used only as an upper limit.

Table A.1.: Directional reflectance relative to the value at solar zenith angle of zero degrees for the Nimbus III ocean and cloud models

ζ	$\cos \zeta$	Ocean $\left(\frac{r(\zeta)}{r(\zeta=0)} \right)$	Cloud $\left(\frac{r(\zeta)}{r(\zeta=0)} \right)$
0°	1.00	1.00	1.00
18°	0.95	1.00	1.00
26°	0.90	1.00	1.00
32°	0.85	1.00	1.00
37°	0.80	1.00	1.05
41°	0.75	1.03	1.10
45°	0.70	1.10	1.14
49°	0.65	1.20	1.18
53°	0.60	1.30	1.22
57°	0.55	1.40	1.28
60°	0.50	1.60	1.32
63°	0.45	1.80	1.38
66°	0.40	2.00	1.42
69°	0.35	2.20	1.48
72°	0.30	2.50	1.52
75°	0.25	2.80	1.55
78°	0.20	3.10	1.58
81°	0.15	3.40	1.60
84°	0.10	3.70	1.60
87°	0.05	4.00	1.60

APPENDIX B

Sensitivity of the Broadband Longwave Radiance Model to Clouds

The method by which radiance values from the four longwave spectral channels of the MRIR (Medium Resolution Infrared Radiometer) aboard the Nimbus 3 satellite were converted to a longwave broadband radiance value (3 to 30 μm) has been discussed by Raschke et al. (1973). A least squares relationship between spectral radiances and broadband radiances was derived by using radiances which were calculated from model atmospheres. Radiances were calculated at various zenith angles and at eight different cloud levels from a set of 160 model atmospheres. The purpose in this section is to assess the sensitivity of that derived empirical relationship to changes in the spectral radiances as clouds, at various levels, are introduced into a model atmosphere.

The empirical relationship between spectral longwave radiances in the four channels and the broadband longwave radiance derived by Raschke et al. is:

$$N_t = A_0 + A_1 N_2 + A_2 N_2^2 + A_3 N_2^3 + A_4 N_4 + A_5 N_1 + A_6 N_3$$

(B.1)

where N_t is the broadband radiance, $N_1 \dots 4$ are the spectral radiances and $A_1 \dots 6$ are the regression coefficients. The coefficients and the standard error of estimate are given in Table B.1.

A global model atmosphere was selected for this experiment. (The model atmosphere is discussed in Section 2.3.2). Broadband and spectral longwave radiances were calculated for this atmosphere for various cloud top heights by applying broadband and spectral radiative transfer models, as discussed in Cox et al. (1976). The spectral resolution of the spectral model was 10 cm^{-1} . The spectral intervals used in the radiance calculation could not be precisely matched to the channel spectral intervals. The channel spectral intervals, defined by the half power spectral interval of the channel response function, and the spectral intervals used in the model calculation are shown in Table B.2. The radiance measured by each channel is represented by:

$$N_i = \int_{\nu_1}^{\nu_2} f_i(\nu) N(\nu) d\nu \quad (\text{B.2})$$

where f_i is the normalized spectral response function for channel "i", N is the spectral radiance and ν_1 to ν_2 denote the spectral interval of the channel response. Equation (B.2) was applied to the model calculated spectral radiances in an approximate form, as

$$N_i = \int_{\nu_1(\frac{1}{2})}^{\nu_2(\frac{1}{2})} N(\nu) d\nu \quad (\text{B.3})$$

Table B.1. Regression coefficients (A) and standard error of estimate (e) for equation (B1) (Raschke et al., 1973)

<u>A0</u> (calcm ⁻² min ⁻¹)	<u>A1</u>	<u>A2</u>	<u>A3</u>	<u>A4</u>	<u>A5</u>	<u>A6</u>	<u>e</u> (calcm ⁻² min ⁻¹)
0.0160	0.00385	0.000317	-0.951x10 ⁻⁵	0.0139	0.00215	0.003144	0.00070

Table B.2. Infrared spectral channels of the MRIR of the Nimbus 3 satellite.

<u>Channel</u>	<u>Half Power Interval</u>		<u>Model</u>	<u>Weighting function peak</u>
	<u>um</u>	<u>cm⁻¹</u>	<u>cm⁻¹</u>	
1	6.35-6.72	1488-1575	1485-1585	Upper tropospheric water vapor
2	10.1-11.2	893-990	895-995	Window channel
3	14.5-15.8	633-690	635-695	Lower stratospheric CO ₂
4	20.8-23.2	431-481	435-485	Lower tropospheric water vapor

where a normalized rectangular response function, $f_i = 1$, was applied across the spectral interval of the half power channel response, defined between wave number ν_1 ($\frac{1}{2}$) and ν_2 ($\frac{1}{2}$). The approximate equation (B.3) was used because the limited spectral resolution of the radiative transfer model did not justify using the exact equation (B.2).

The outgoing longwave radiance values, computed by equation (B.1) and by the broadband radiance transfer model for a no-cloud case and for three cloud top altitude cases, are shown in Table B.3. The differences between equation (B.1) computed values (empirical) and the broadband radiative transfer model values (model) also are shown in the table. The difference between the model results and the empirical results are quite large. Some of the difference may be attributed to differences between the prescribed radiative characteristics of the absorbers and the numerical methods applied within the three different radiative transfer models used to compute broadband radiances, spectral radiances and the coefficients of equation (B.1). The differences may also be attributed to not computing radiance values, using Cox's spectral model, over spectral intervals identical to those intervals for which the coefficients in equation (B.1) were derived.

Even though the difference between the empirical and model computed broadband radiances is quite large for each cloud case, the sensitivity of the empirical results to

Table B.3. Spectral broadband empirical, broadband model and spectral radiances calculated for a model atmosphere.

Cloud Top Height (km)	Broadband Radiance ($\text{wm}^{-2}\text{sr}^{-1}$)			Spectral Radiances ($\text{wm}^{-2}\text{sr}^{-1}$)			
	Model	Empirical	Difference	Channels 1	2	3	4
no clouds	86.103	99.863	-13.760	0.302	8.381	3.059	4.893
3.5	72.820	85.755	-12.935	0.301	5.796	3.058	4.720
5.5	64.200	76.364	-12.164	0.294	4.499	3.058	4.345
8.5	51.249	62.245	-10.996	0.244	2.935	3.051	3.551

cloud top height can be determined by examining the difference of the differences between the cloud cases. This difference of a difference will be called a D2 difference. The broadband differences for the no-cloud and the 8.5 cloud top cases differ (a D2 difference) by $2.8 \text{ wm}^{-2}\text{sr}^{-1}$ (Table B.3). This difference exceeds the $0.05 \text{ wm}^{-2}\text{sr}^{-1}$ ($0.00070 \text{ calcm}^{-2}\text{sr}^{-1}$) standard error of regression of equation (B.1) (Table B.1) by nearly two orders of magnitude. This D2 difference value might be attributed to differences in the radiative transfer models used, as was previously discussed. How significant is this D2 difference with respect to the sensitivity of equation (B.1) to changes in height of opaque cloud tops?

The sensitivity of the D2 difference to changes in the spectral radiances input to equation (B.1) should be the largest for no cloud and the 8.5 km cloud top cases. Thus, the no cloud and the 8.5 km cloud top cases have been selected for the sensitivity study. Each channel radiance will be examined for its effect on equation (B.1) and on the D2 differences between the two cases. The channel 3 radiance value does not contribute to the D2 difference, since it does not change significantly between the two cases. The effect of the channel 1 radiance value on the empirical results can be assessed by holding it constant to the no cloud value, and recomputing the empirical value from equation (B.1) for the 8.5 km cloud top case. The

empirical value for the 8.5 km cloud top case increases by a mere $0.1 \text{ } \mu\text{m}^{-2} \text{ sr}^{-1}$ when the channel 1 no-cloud radiance value is used in the calculation. Equation (B.1) is quite insensitive to cloud effects in the channel 1 radiances. The bulk of the sensitivity of equation (B.1) to clouds is caused by changes in the radiances of channel 2, the long-wave window channel, and channel 4, the lower tropospheric water vapor channel. The sensitivity of equation (B.1) to channel 4 radiances is tested in the same way as was done with channel 1. Holding the channel 4 radiance constant at the no-cloud case value, $4.893 \text{ } \mu\text{m}^{-2} \text{ sr}^{-1}$, and recomputing the 8.5 km cloud top case with this channel 4 radiance value and equation (1), causes the 8.5 km cloud top broadband empirical value to increase by $13.1 \text{ } \mu\text{m}^{-2} \text{ sr}^{-1}$. Channel 4 radiances account for 35 percent of the broadband difference between the no-cloud and the 8.5 cloud top cases. This increase is 4.8 times greater than the $2.8 \text{ } \mu\text{m}^{-2} \text{ sr}^{-1}$ D2 difference. The channel 2 radiance accounts for the remaining 65 percent of the difference. Equation (B.1) is indeed significantly sensitive to the change in spectral composition of the outgoing spectral radiances due to cloud changes. The sensitivity is significantly larger than the possible errors of this simple test.

If the D2 difference is taken as a lack of sensitivity of equation (B.1) to cloud changes rather than as an

experimental error, then it should be examined for its effect on ΔI , the cloud and no-cloud broadband irradiance difference. The 5.5 km cloud top height may be considered as the approximate global mean opaque cloud top height. The D2 value at 5.5 km computed from Table B.3 values is $1.6 \text{ wm}^{-2} \text{sr}^{-1}$. If 50 percent global cloud cover is assumed, then the D2 difference is reduced to $0.8 \text{ wm}^{-2} \text{sr}^{-1}$. The earth-atmosphere system is assumed to be a homogeneous and isotropic emitter, so that multiplication of the radiance difference by π yields an irradiance difference of 2.5 wm^{-2} . This irradiance difference is less than 10 percent of the 26 wm^{-2} global ΔI value (see Section 2.3.2).

APPENDIX C

Global Integration

Global integrated values of planetary net flux, albedo, and longwave flux emitted to space were obtained from the 29 months of satellite measurements. These measurements were averaged into zonal profiles of mean months (Ellis and Vonder Haar, 1976).

Global integrated values of longwave flux were computed from the zonal profiles of the 29 month set by:

$$I(t) = \int_{-\pi/2}^{\pi/2} I(\phi, t) d(\sin \phi) \quad (C.1)$$

where:

t = month

ϕ = latitude

They appear in Table C.1.

Global integrated values of albedo were also computed from zonal profiles by:

$$\alpha(t) = \frac{\int_{-\pi/2}^{\pi/2} \alpha(\phi, t) S(\phi, t) d(\sin \phi)}{\int_{-\pi/2}^{\pi/2} S(\phi, t) d(\sin \phi)} \quad (C.2)$$

where symbols are the same as in equation (C.1) and

$$S = SC \left(\frac{d}{d_0} \right)^2 \cos(\xi),$$

Table C.1. Mean monthly global planetary radiation budget

	S* INSOLATION (wm^{-2})	α ALBEDO (Percent)	R REFLECTED (wm^{-2})	I EMITTED (wm^{-2})	N NET (wm^{-2})
January	350.7	30.2	105.9	231.1	13.7
February	347.6	30.5	106.0	230.0	11.6
March	342.5	28.8	98.6	227.8	16.1
April	336.8	30.4	102.4	246.8	-12.4
May	332.0	31.4	104.2	245.0	-17.2
June	329.1	31.1	102.4	245.4	-18.7
July	328.8	29.6	97.3	236.5	- 5.0
August	331.1	29.0	96.0	235.2	- 0.1
September	335.7	28.7	96.3	231.1	8.3
October	341.6	29.1	99.4	235.8	6.4
November	347.1	32.0	111.1	232.6	3.4
December	350.5	31.7	111.1	230.7	8.7
Annual	340.0	30.2	102.7	235.7	1.6

* Based on a Solar Constant = 1360 wm^{-2}

the solar insolation. The denominator of equation (C.2) is the global integrated solar insolation, $S(t)$.

The global integrated albedoes were not computed from the 29 month set as presented in Ellis and Vonder Haar (1976), but from a 23 month set (Table C.1). The 23 month set is the 29 month set less 6 months of ESSA 7 measurements. These were excluded because the albedoes from the ESSA 7 were derived by Mac Donald, 1970 with an assumption that radiative equilibrium existed in global net flux. The Table C.1 values show that planetary radiative equilibrium does not generally exist on time scales less than a year.

From the component albedo and longwave flux values, global average net flux, as shown in Table 1, was computed by:

$$N(t) = (I - \alpha(t)) S(t) - I(t) \quad (C.3)$$