REFLECTION OF SUNLIGHT TO SPACE AND ABSORPTION BY THE EARTH AND ATMOSPHERE OVER THE UNITED STATES DURING SPRING 1962 [1]

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ABSTRACT

This study describes a method for determining the reflection of sunlight to space and absorption by the earth and atmosphere, using low-resolution radiometer data from earth satellites. The method has been used with TIROS IV data together with radiation measurements at the ground to determine the reflection and absorption of sunlight over the United States during the spring of 1962.

The results indicate that for this region and time, 40 percent of the incident sunlight at the top of the atmosphere was reflected to space, 13 percent was absorbed by the atmosphere and clouds, and the remaining 47 percent was absorbed at the earth's surface. Atmospheric absorption of sunlight varied from over 20 percent in the moist air in southeastern United States to less than 10 percent over much of the dry mountainous west and northern plains.

When atmospheric absorption values determined from this study are compared with earlier studies of absorption in a cloudless atmosphere, there is good agreement at *low* values of atmospheric water vapor; however, the present study gives significantly higher absorption at *high* values of water vapor.

Based on this study, an empirical relationship is determined for fractional absorption of sunlight in an atmosphere *with clouds* as a function of optical pathlength of water vapor: $q_a=0.096+0.045(u^*)^{1/2}\log_e u^*$. The fractional absorption of sunlight, q_a, is the fraction of the total amount incident at the top of the atmosphere. The optical pathlength, u^*, is given in cm.: $u^*=u\cdot\sec\zeta$. Here, u is total precipitable water in a vertical column, given in cm., and ζ is the solar zenith angle.

SYMBOLS

A_s/A_f	Ratio: albedo snow-covered ground to albedo snow-free ground
c_p	Specific heat of air at constant pressure
d	Direct distance from point p on ground to satellite at height h
g	Acceleration of gravity at earth's surface
h	Height of satellite
I_s	Solar constant
p	Point at earth's surface
q_a	Fraction of Q_0 absorbed in the atmosphere
q_e	Fraction of Q_0 absorbed at the ground
q_g	Fraction of Q_0 incident at the ground
q_r	Fraction of Q_0 reflected to space
Q_a	Radiation absorbed in the atmosphere
Q_g	Irradiance at the ground
Q_0	Irradiance at the top of the atmosphere
Q_r	Outward irradiance at the top of the atmosphere
Q_s	Irradiance reflected at the ground
r	Earth-sun distance
r_m	Mean earth-sun distance
u	Total precipitable water in the atmosphere
u^*	Optical pathlength of atmospheric water vapor
W_p	Weighting factor for radiance from point p to a spherical sensor at height h above the earth
w	Mixing ratio of water vapor
α	Surface albedo
ζ	Solar zenith angle at the subsatellite point at time of observation
ϕ	Zenith angle of satellite at height h, viewed from point p

1. INTRODUCTION

The albedo is defined as the ratio of reflected to incident radiation integrated over the solar spectrum from 0.2 to 4.0 microns. The albedo of the earth and atmosphere is an important component in the global heat budget, but, unfortunately, it is a difficult component to determine. Prior to 1957, there was no possibility of making direct measurements of this component on a global scale. In the ensuing nine years since satellite measurements have become available, there has been conflicting interpretation of them. One problem in determining the albedo of the earth and atmosphere is that cloudiness, which has a strong modulating influence on the albedo, varies widely over the planet both in time and space. This, of course, causes a sampling difficulty because measurements cannot be made everywhere, all the time. Another hindrance to determining the albedo is that the scattering properties of

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water droplets in clouds, and the variable solar-zenith angle over the spherical earth, further complicate the sampling.

Prior to the launching of earth satellites, the only "outside" approach to determining the amouut of sunlight reflected from earth was to measure the "earthlight" reflected from the shadowed side of the moon (Danjon [3], Dubois [4]). Computations of the planetary albedo have been made from those illumination measurements (Ångström [1]). Another approach to estimating the earth's albedo was through a knowledge of cloud distribution over the earth and estimates of the reflectivity of various cloud types (Simpson [15], Houghton [7]; London [12]).

Recently, Fritz et al. [5] described a method of combining satellite and surface radiation measurements to determine the reflection and absorption of sunlight. In that study, the data from medium-resolution radiometers of TIROS III were used. This gives the albedo of relatively small areas (50-km. radius) near surface radiation stations. The study reported in this paper is similar to that of Fritz et al. in that it combines simultaneous satellite and surface data to determine reflection and absorption terms. However, it differs in that low-resolution measurements are used. Thus, the resulting albedoes are representative of larger surface areas.

2. METHOD OF CALCULATION

From consideration of conservation of radiant energy, it is clear that sunlight reaching the top of the atmosphere is either reflected to space or is absorbed by the earth or atmosphere. To define these relationships for the present study, we have used a notation similar to that of Fritz.

As illustrated in figure 1, Q_0 and Q_r are the values of solar irradiance incident and reflected from the top of the atmosphere. Likewise, Q_g and Q_s are values of solar irradiance incident and reflected from the earth's surface. These irradiance terms define the absorption in the atmosphere, Q_a, as

$$Q_a=(Q_0+Q_s)-(Q_r+Q_g). \tag{1}$$

Dividing equation (1) by Q_0 and rearranging terms gives,

$$1=q_a+q_r+q_g(1-\alpha) \tag{2}$$

where:

$q_a=Q_a/Q_o$, the fraction of incident sunlight absorbed in the atmosphere,

$q_r=Q_r/Q_0$, the fraction of incident sunlight reflected to space,

$q_g=Q_g/Q_0$, the transmission of the atmosphere,

$\alpha=Q_s/Q_g$, the surface albedo.

The last term on the right side in equation (2) is the fractional absorption at the earth's surface and will be denoted simply as q_e. Then,

$$1=q_a+q_r+q_e. \tag{3}$$

It is apparent from equation (3) that sunlight incident at the top of the atmosphere is divided between two

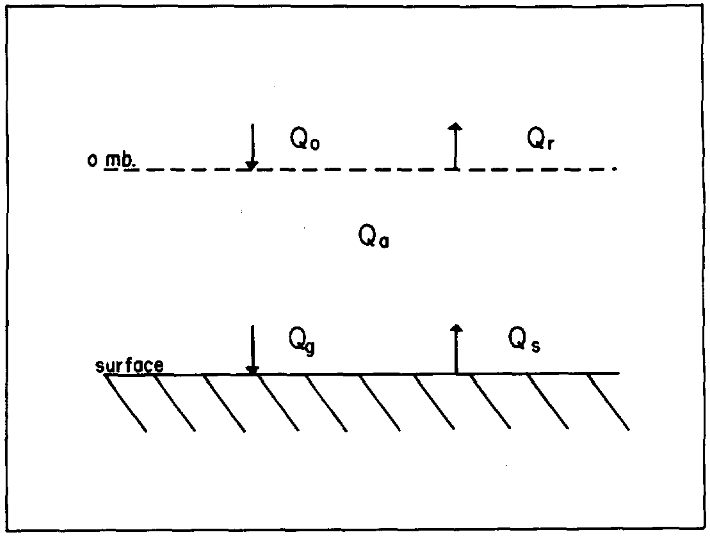

Figure 1.—Symbols for irradiance bounding the atmosphere and absorption of radiation in the atmosphere.

absorption terms, (1) the absorption in the atmosphere, q_a, and (2) the absorption at the earth's surface, q_e; and one reflection term, the reflection of sunlight to space, q_r, by the earth, atmosphere, and clouds.

Clearly, the fractional terms are completely specified if the four irradiance values are known and are compatible in time and space. The following section defines the irradiance terms to assure this compatibility.

IRRADIANCE AT THE UPPER BOUNDARY

The irradiance at the upper boundary of the atmosphere, Q_0, was calculated from

$$Q_0=I_s(r_m/r)^2 \cos \zeta. \tag{4}$$

The solar constant, I_s, was taken as 2.0 cal./cm.^{2} min.; r_m and r are the mean and actual earth-sun distances; and ζ is the solar zenith angle at the subsatellite point at the time of observation.

OUTWARD IRRADIANCE AT THE UPPER BOUNDARY

The outward irradiance at the upper boundary was obtained from TIROS IV low-resolution measurements of sunlight reflected to space by the earth, atmosphere, and clouds. These measurements were evaluated in prior studies at the University of Wisconsin and currently are available at the National Space Science Data Center [13]. The techniques used in reducing the measurements have been discussed previously by Suomi [16], [17], Bignell [2], and House [9], and more recently by Suomi et al. [18]. House [9] described the inflight calibration technique which was used in evaluating the TIROS IV low-resolution radiometer data used in the present investigation. He has also described an improved technique for obtaining zonal averages of albedo. However, the data obtained by the improved technique could not be used in the present investigation because they are zonally averaged values,

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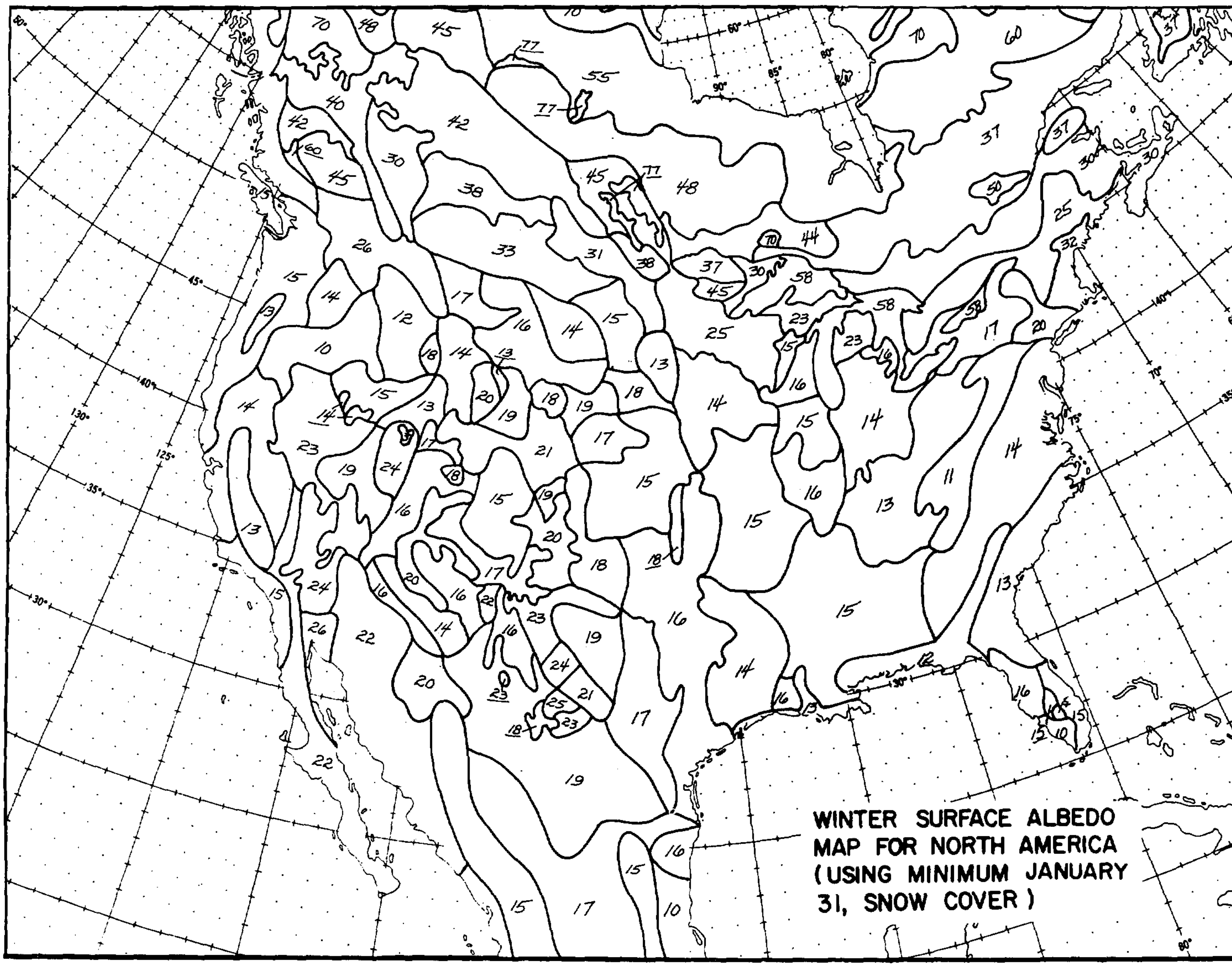

FIGURE 2.—Winter minimum albedoes based on minimum snow cover (from Kung et al. [11]). (Values shown are percentages.)

whereas we required individual measurements over the United States. Thus, data for the present investigation were evaluated by use of the standard reduction techniques given in the reference above.

Although these TIROS sensors respond to sunlight reflected from a relatively large area below the spacecraft (radius of 2890 km.),[3] the radiation reaching the sensor is strongly weighted in favor of a relatively small area around the subsatellite point. For example, about 50 percent of the irradiance comes from an area with radius of 734 km., centered at the subsatellite point.

IRRADIANCE AT THE GROUND

The irradiance at the ground, Q_0, was calculated from radiation measurements at many surface stations in the

[3] Distance from the subsatellite point to the earth's horizon, as viewed from the satellite.

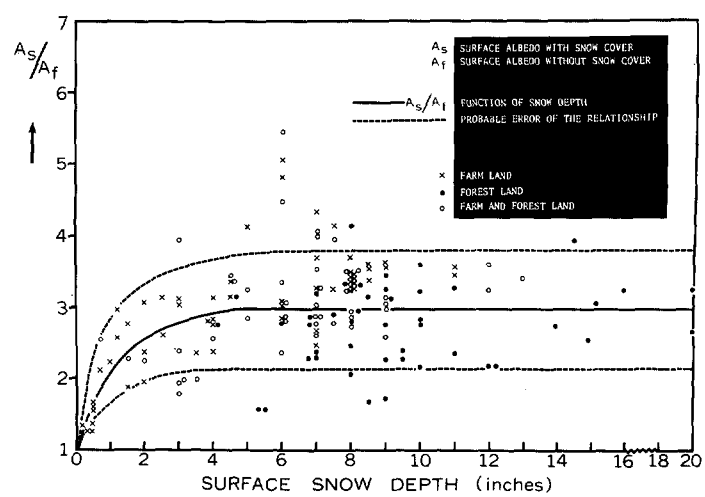

FIGURE 3.—A factor A_s/A_f for modifying the albedo of snow-free ground as a function of snow depth (from Kung et al. [11]).

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region viewed by the satellite. Data were weighted by a factor that was symmetrical about the subsatellite point and varied with distance from the subsatellite point in the same manner as described below for satellite measurements of outward irradiance. Identical geographic weighting is an important consideration of this study, because it is essential that irradiance determined for the upper and lower boundaries of the atmosphere represent identical areas. This is necessary because atmospheric absorption is only a small difference between large irradiance values. The appropriate weighting factor W_p for radiation from a point p on the ground reaching a spherical sensor at height h above the ground has been derived by House [8], as

$$W_p = (h/d)^2 \cos \phi. \tag{5}$$

Here, d is the direct distance from p to the satellite; ϕ is the zenith angle of the satellite, viewed from point p; and h is the height of the satellite.

IRRADIANCE REFLECTED AT THE GROUND

The irradiance reflected at the ground, Q_s, was obtained as the product of irradiance at the ground and the surface albedo. Albedo values are representative for an area around the station with a radius of 280 km. This particular radius value was selected in order to include all of the area between surface stations. The values used were the "winter-minimum" albedoes (fig. 2) for the United States as determined by Kung et al. [11]. Since these are representative of snow-free conditions, it is necessary to correct them if snow exists at a particular time. From observed daily snow depth, a best estimate of the surface albedo at each station each day was calculated. The nomogram (fig. 3) developed by Kung was used for translating snow depth to albedo.

BOUNDARY CONDITION

A limitation of this study is the fact that outside the surface-station network it is no longer possible to determine an irradiance at the ground which is representative of the area seen by the satellite. As a result, measurements along the periphery of the network may be in error—particularly where a significant change in cloudiness or surface albedo occurs across the boundary. Inside the boundary, and over most of the continental United States, this is not a problem.

3. MEASUREMENTS

IRRADIANCE VALUES

A total of 775 observations of outward irradiance, Q_r, were obtained by TIROS IV low-resolution radiometers over the United States from February 8 to June 3, 1962. For calculating the partitioning of sunlight, it was required that the other three irradiance terms correspond in time and space with the satellite observations. To obtain this, the irradiance at the ground was calculated from hourly radiation measurements at surface stations, weighted geographically as previously defined. Hourly surface observations were used for convenience, since a large number of satellite observations have no systematic time bias within 1-hour intervals. The two other irradiance terms were calculated as previously described.

Because of the precession of TIROS IV, daytime observations over the United States are not on a continuous basis, but are for intermittent periods. These periods are shown in figure 4. In addition to time bias, there is geographical bias in the sample. This is due to: (1) the orbit inclination of 48.3° which has the effect of increasing the relative frequency of observations in the northern latitudes, and (2) the location of the read-out station in the eastern United States which has the effect of decreasing the number of observations in that region. Clearly, the data used in this study are biased in time and space. To surmount this problem we have selected the data to have closest agreement in time and space to the satellite observations. Thus, the *same* bias is introduced in all variables being studied. In this way, it is possible to obtain meaningful relationships among them.

OPTICAL PATHLENGTH OF ATMOSPHERIC WATER VAPOR

Atmospheric water vapor is an effective absorber of sunlight and a variable constituent of the atmosphere. As a result, absorption by water vapor is significantly large, but not geographically uniform. In this paper, the absorption of sunlight as a function of precipitable water is examined in order to make an estimate of its nonuniformity.

A calculation of total precipitable water, u, was made to correspond with each satellite observation. The radiosonde observation closest to the subsatellite point, in time and space, was used. The total precipitable water was

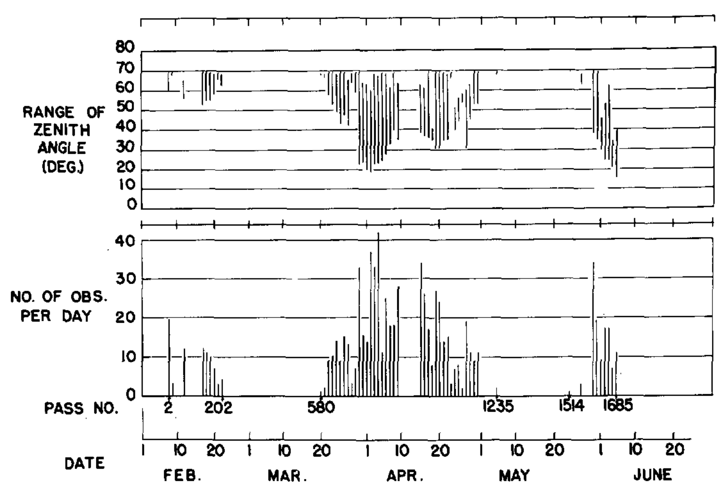

FIGURE 4.—The number of daytime observations from TIROS IV low-resolution radiometers and associated solar zenith angles at times when the satellite was over the continental United States, February–June, 1962.

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computed as,

$$u = g^{-1} \int w dp \tag{6}$$

where g is the gravitational acceleration at the earth's surface, w is the mixing ratio of water vapor, and p is pressure. The integration was done from the surface to 300 mb.

Over most of the atmosphere, sunlight enters on a slant path rather than vertically. Then, if ζ is the zenith angle of the sun, the optical pathlength, u^*, for sunlight penetrating a clear atmosphere is given as

$$u^* = u \cdot \sec \zeta. \tag{7}$$

4. GEOGRAPHICAL DISTRIBUTION

REFLECTION AND ABSORPTION OF SUNLIGHT OVER THE UNITED STATES

The reflection and absorption data were calculated for the hours indicated on figure 4. For the period March through May, averages of reflection, absorption, and total precipitable water data are shown in figures 5 and 6. Spatial smoothing was done by averaging observations within 330 km. of fixed grid points over the United States. A 2° by 2° latitude-longitude grid was adopted for this purpose. In the southeastern United States where there are sampling limitations, a minimum number of 15 samples was required at a grid-point before the average was calculated. Over the remainder of the United States, the data have been shown up to the border, even though the boundary condition may influence the representativeness of the analysis.

Several relationships between the partitioning terms seem significant in figures 5 and 6. For example, the patterns for atmospheric absorption and absorption at the ground are similar—particularly over the eastern two-thirds of the United States. It is interesting that neither of these patterns resembles that of reflection to space, except over the western United States where absorption at the ground is inversely correlated with reflection to space.

A physical interpretation of these results suggests that in areas where atmospheric water vapor tends to be uniform (e.g., in western United States), the variation in atmospheric absorption is correspondingly small. Under these conditions, the reflection term has a dominant effect on the amount of sunlight reaching the ground. On the other hand, in regions where atmospheric water vapor is nonuniform (e.g., in eastern United States), the variation in atmospheric absorption is increased considerably. In fact, it appears to be increased to the extent that the dominant effect on the amount of sunlight reaching the ground is created by water vapor rather than by the modulating effect of clouds which would otherwise control it.

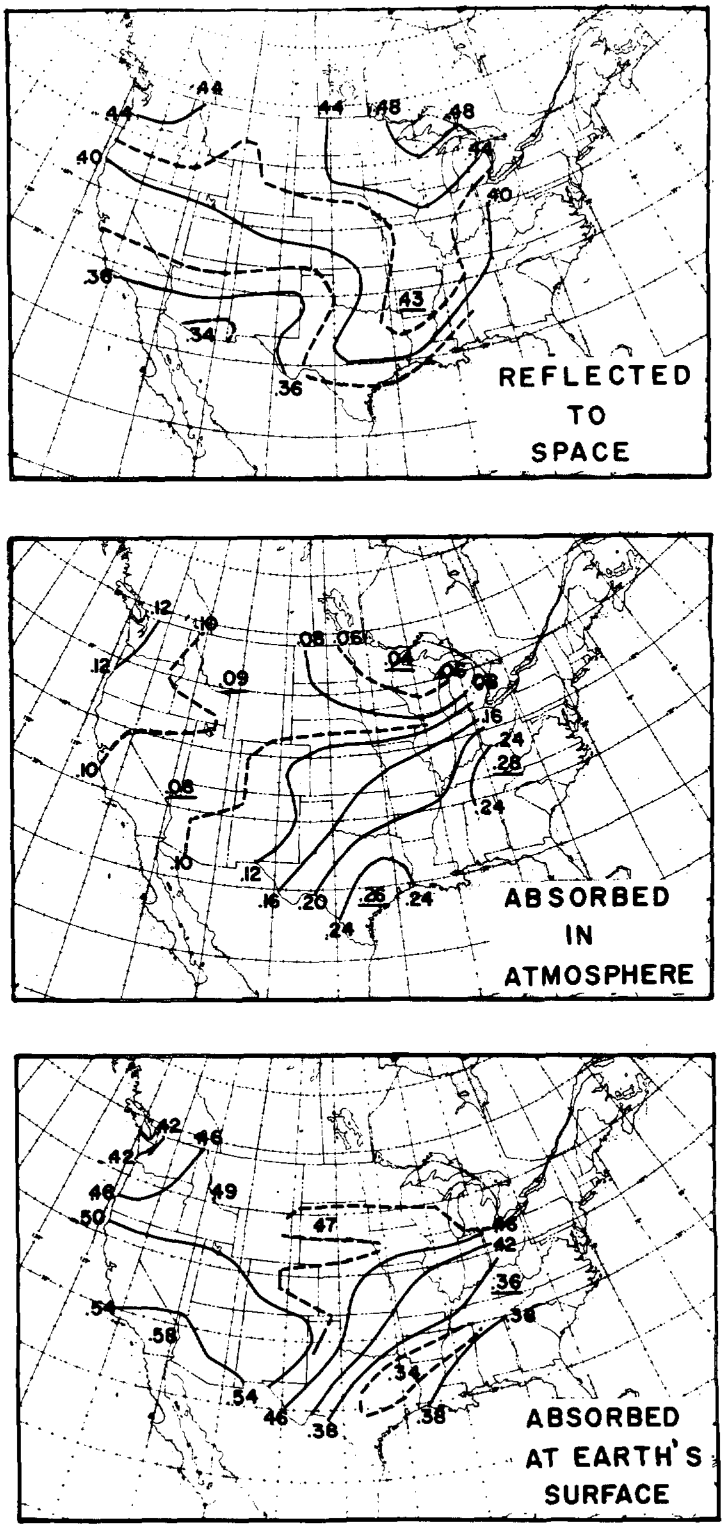

FIGURE 5.—The reflection and absorption of sunlight during the period March through May, 1962, based on TIROS IV low-resolution radiometer and surface radiation data. (Values are fraction of incident sunlight at the top of the atmosphere.)

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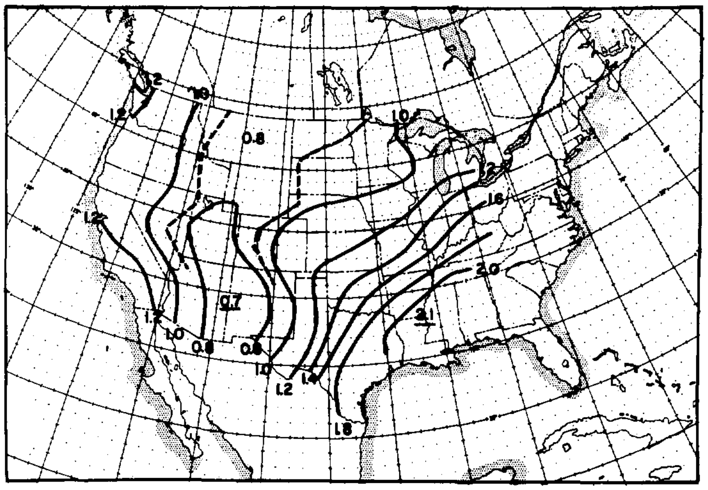

FIGURE 6.—Total precipitable water (cm.) in the atmosphere during the period March through May, 1962. Sampling times correspond to the data illustrated in figure 5.

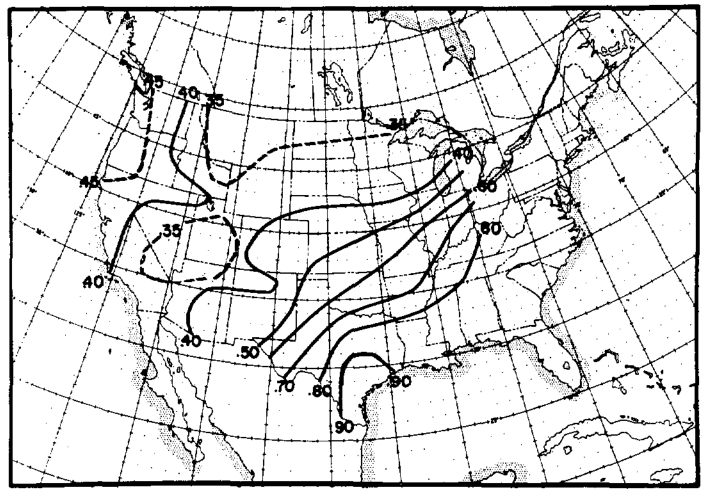

FIGURE 7.—Atmospheric heating (°C./day) due to absorption of solar radiation in the atmosphere during the period March through May, 1962, based on q_a values of figure 5.

A striking feature of figures 5 and 6 is the similarity between the patterns of total precipitable water and atmospheric absorption—one that would be expected in view of the results of previous work (Gates [6], Yamamoto [19], Houghton [7], and Kimball [10]), among others. The two patterns are remarkably similar over nearly all of the United States, except for the upper Great Lakes region where absorption values are low. One possible explanation is that surface albedo estimates may not be representative of the surface area seen by the satellite, particularly in the snow-covered country to the north. Consideration of a boundary condition of this type would tend to reduce the calculated absorption values for the atmosphere.

It is interesting to note in figure 5 that absorption in the atmosphere has a considerable variation across the United States, in spite of the smoothing technique used. For example, at the same geographical latitude, absorption varies by a factor of three from the minimum values in the Great Basin to the maximum values in southeastern United States. With latitude, it varies by a factor of seven between the western Great Lakes region and the southeastern United States. Apparently atmospheric absorption has as large a spatial variation as absorption at the ground, and has an even larger variation than does the reflection of sunlight to space.

Finally, from the data illustrated in figures 5 and 6, overall spatial averages of these terms for the continental United States were derived and summarized in table 1.

Reitan [14] has found that for the period March–May the precipitable water over the *entire* United States averaged 1.44 cm., based on 11 yr. of data. When Reitan's data are limited to the area of the present study, the average is 1.39 cm. This indicates the average precipitable water in the present study is low by 0.2 cm., and the fractional absorption of sunlight in the atmosphere is low by about 1 percent in relation to the 11-yr. period of Reitan.

SOLAR HEATING RATE OVER THE UNITED STATES

The atmospheric absorption values which were determined in this study can be expressed as the amount of heating of an atmospheric column. For the values given in figure 5, the resultant heating rate was calculated from

$$\text{Average Heating Rate (°C./day)} = (g\overline{Q}_0 q_a)/(10^3 \cdot c_p \cdot \Delta p). \tag{8}$$

$\overline{Q}_0$ is the average for the period March through May for the United States and Δp was taken as 1000 mb. The results shown in figure 7 indicate the maximum warming exceeded 0.8°C./day in the moist belt across the southeastern United States. The minimum warming rate of about 0.35°C./day is found over the northern Great Plains and throughout the western mountain region.

5. ABSORPTION OF SUNLIGHT IN THE ATMOSPHERE

There has been considerable work on absorption of solar radiation in the atmosphere, based on laboratory measurements (e.g., Yamamoto [19], Houghton [7]). These studies were concerned with the absorption in a *cloudless* atmos-

TABLE 1.—*Average values of partitioning of sunlight over the continental United States during spring 1962 (fraction of total incident sunlight)*

Reflection to space, q_r	0.40
Absorbed in atmosphere, q_a	0.13
Absorbed at ground, q_s	0.47
	1.00
Total precipitable water	1.19 cm.

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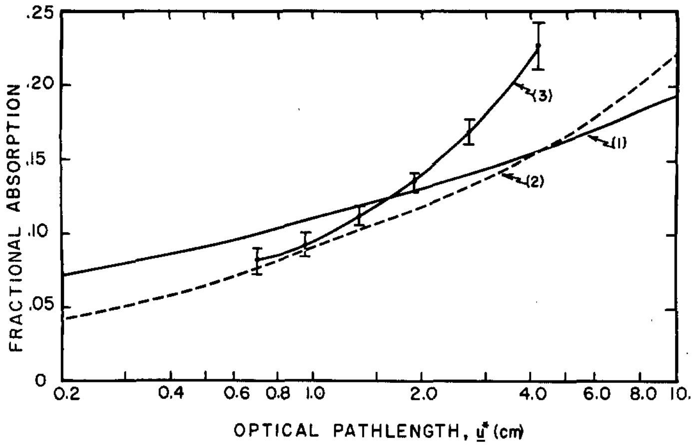

FIGURE 8.—Absorption of solar radiation in the atmosphere as a function of optical pathlength, u^*, in cm. Curve (1) is the absorption in a clear atmosphere by H_2O, CO_2, and O_2, from Yamamoto [19]. Curve (2) is the absorption in a clear atmosphere by H_2O only, from Houghton [7]. Curve (3) is the absorption in the atmosphere over the United States with mixed cloudiness, based on the present study. Brackets indicate the confidence limit of the function values.

phere. Yamamoto, and also Houghton, relate absorption to total precipitable water, for an optical air mass of one (i.e., when the solar zenith angle is zero), as shown in figure 8, curves 1 and 2.

The absorption values of the present study have been averaged by class intervals of the natural logarithm of u^*, in order to provide results comparable to those mentioned above. The results are shown by curve 3 of figure 8.

The work of Yamamoto and Houghton shows the absorption of sunlight in "artificial" atmospheres which depart from the real case because they (1) are cloudless, (2) are dustless, (3) have fixed vertical distribution of absorbing constituents, and (4) have a nonreflecting surface at the lower boundary. Clearly, the curves in figure 8 are not directly comparable. Nevertheless there is significant similarity in the shape of the curves.

The important feature of figure 8 is that the absorption of sunlight in an atmosphere *with* clouds (curve 3) departs from the clear-sky absorption curves (1 and 2) in the region of *higher* water vapor content. At lower values of water vapor, there is much closer agreement among the

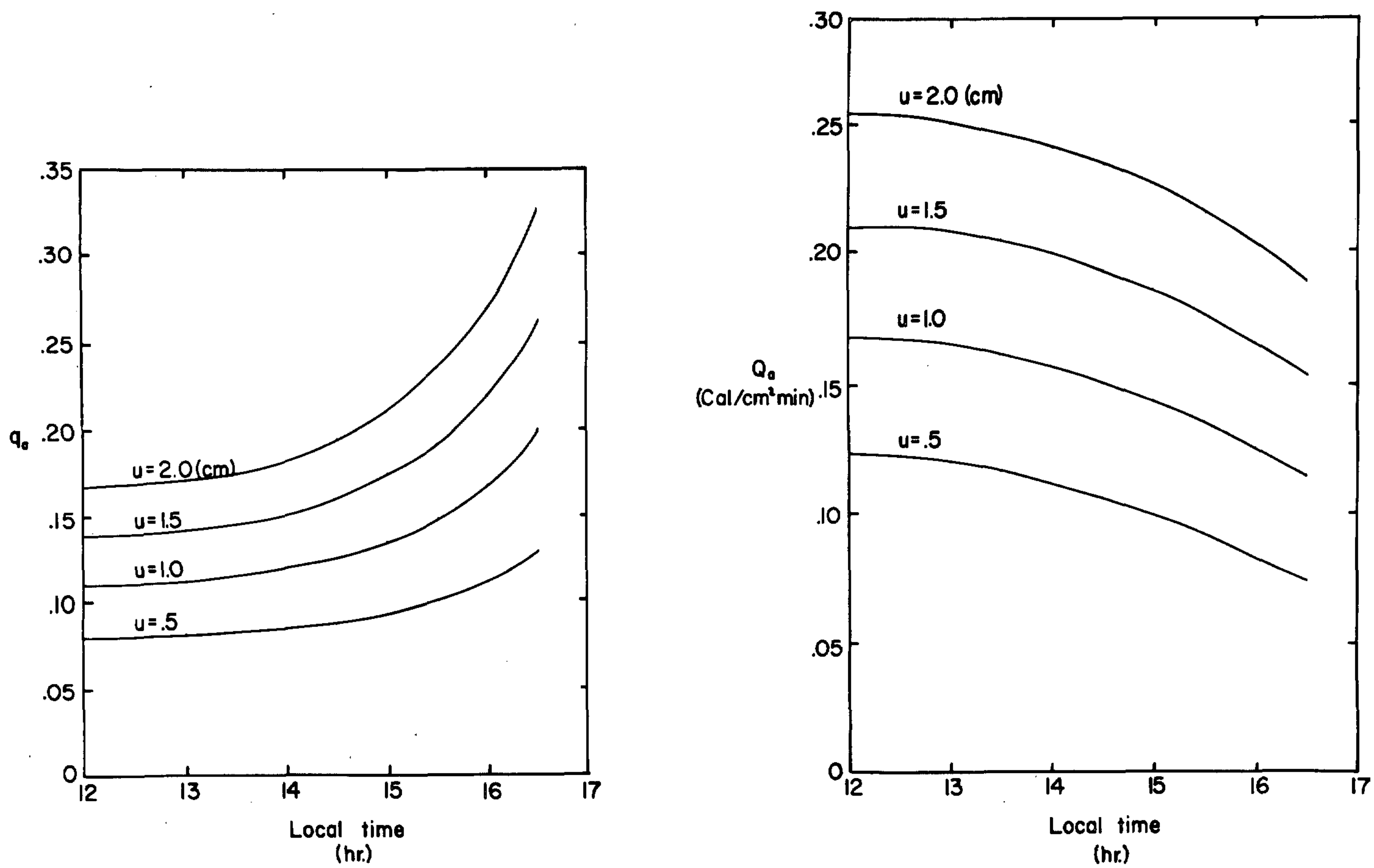

FIGURE 9.—Fractional absorption, q_a, and total absorption, Q_a, at 40° N. latitude on vernal equinox based on equation (9).

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curves. Thus, it appears that clear-sky absorption estimates are least applicable in areas with more moisture (and presumably more cloudiness), and the tendency in these cases is to *underestimate* the actual absorption that occurs with clouds.

This is particularly important because infrared cooling of the atmosphere is relatively less in moist, cloudy areas; thus, *both* the infrared and solar radiation have a positive effect on the generation of available potential energy in mid-latitude, synoptic-scale disturbances. In addition, the solar absorption term is greater than previously thought.

For further studies it is useful to know the function which represents curve 3 (fig. 8). An approximation was found to be

$$q_a = 0.096 + 0.045(u^*)^{1/2} \log_e u^*. \tag{9}$$

The variance of the dependent data from this function is 1.0×10^{-4}.

An example illustrating the use of equation (9) was calculated for 40° N. latitude at the time of the vernal equinox. Since both time and position on earth are specified, the fractional absorption, q_a, and total absorption Q_a are simply functions of the optical depth of water vapor. When optical depth is held constant, the fractional and total absorptions are specified, as shown in figure 9.

Limitations of this example are: (1) the empirical data obtained over the United States must be used with caution in other areas, e.g., oceanic areas, and (2) because of scattering of sunlight by cloud droplets, the optical pathlength is not simply a function of sec ζ, although this may be a good approximation.

6. CONCLUSIONS

This study has combined TIROS IV low-resolution radiometer data with surface pyranometer data in order to determine the distribution and amount of sunlight absorbed in the atmosphere over the United States during the spring of 1962. The results show that, for this region and time, 40 percent of the incident sunlight at the top of the atmosphere was reflected to space, 13 percent was absorbed by the atmosphere and clouds, and the remaining 47 percent was absorbed at the earth's surface. Atmospheric absorption of sunlight varied from over 20 percent in the moist air in southeastern United States to less than 10 percent over much of the dry mountain west and northern plains.

Values of atmospheric absorption determined from this study have been compared with earlier studies of absorption in a *cloudless* atmosphere. Although there is good agreement between the results of these studies at *low* values of atmospheric water vapor, the present study (which includes clouds) shows significantly higher absorption at *high* values of water vapor. Apparently, clear-sky absorption curves tend to underestimate absorption in a real atmosphere—particularly in moist, tropical air masses.

An empirical relation has been determined in this study for parameterizing atmospheric absorption as a function of water vapor. It is given in equation (9). Also, an example is given to illustrate the general use of equation (9) for other regions and times. In this example, figure 9, which is representative for 40° N. at the vernal equinox, the fractional absorption, q_a, and total absorption, Q_a, of sunlight in the atmosphere are given simply as functions of optical depth of water vapor and time of day. Until better estimates of atmospheric absorption are available for a wider range of conditions, absorption values for use in numerical models could be obtained by the relationship derived here, providing it applies reasonably well to other areas. This is a reasonable assumption, however, because both polar and tropical air masses were represented in the data sample from which the relationship was derived.

Of course, the reflected solar radiation measurements used in this study were obtained from one of the earliest thermal radiation experiments flown on a satellite. As such, they are representative of albedo observations available at the present time. Future measurements by improved radiometers aboard ESSA and Nimbus satellites should provide additional information on the conclusions presented in this paper.

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