# ON THE ANNUAL HEAT BALANCE OF THE NORTHERN HEMISPHERE

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ABSTRACT

The increased knowledge of atmospheric radiation during the last twenty years has made it desirable to re-compute the heat balance of the northern hemisphere. Insofar as possible the computations are based on observational data and, in particular, no assumption is made regarding the planetary albedo nor the albedo of clouds. The solar radiation reaching the surface is derived largely from the North American pyrheliometric network. These data are extended to the hemisphere on the basis of the differences between the cloud amounts along complete latitude circles and the cloud amounts along the North American segments. Absorption of solar radiation in the atmosphere is computed from the original Smithsonian data. Absorption by clouds is taken into account. The outgoing long-wave radiation is computed on the Elsasser radiation chart, with use of monthly mean data from a large number of individual stations. The planetary albedo is found to be 0.34, with a minimum of 0.28 in the subtropics and a maximum of 0.67 at the pole. About 19 per cent of the incident solar radiation is absorbed in the atmosphere, and 47 per cent is absorbed at the ground. The required poleward flux of heat is computed and is found to reach a maximum of $11.12 \times 10^{19}$ cal/day across the 40°N latitude circle. This is 20 to 25 per cent larger than earlier estimates. The hemispheric energy balance between the surface of the earth and the atmosphere indicates that the eddy flux of sensible heat is directed upwards, and that its magnitude is of the order of 10 per cent of the solar radiation incident on the outer atmosphere.

## 1. Introduction

Recent quantitative studies of the energy transport by the atmosphere, and improved knowledge of atmospheric radiation, have suggested a re-evaluation of the atmospheric heat balance. Our present knowledge of the heat balance is based largely on the classical treatments of Simpson (1928) and of Baur and Phillips (1934–1935). Of necessity their results were based on simplifying assumptions and inadequate information, particularly as regards the emission and absorption of long-wave radiation. Their computations of the absorbed solar radiation were based on the Smithsonian researches; Simpson adopted the Smithsonian planetary albedo of 0.43 at all latitudes, whereas Baur and Phillips made independent calculations that led to some variation of the albedo with latitude but to about the same mean albedo. These albedos were heavily dependent on an assumed cloud albedo of 0.78 that was derived by Aldrich (1919) from a single series of measurements on a stratus-cloud layer.

More recently, Elsasser (1942) has constructed a radiation chart based on more detailed information on the absorption spectrum of water vapor. It is known that the Elsasser chart does not properly account for the pressure-dependence of the absorption and that Elsasser's error-function expression is only an approximate representation of the transmission law for water vapor. On the other hand, his chart appears to be the best readily-available tool for the computation of the outgoing long-wave radiation from an atmosphere in which the distribution of water vapor with temperature is known. Such comparisons as have been made between direct observations of long-wave radiation and the chart values suggest that the chart is not grossly in error. Computations of the outgoing long-wave radiation made with the Elsasser chart yield values significantly larger than those given by Simpson (1928). This seems reasonable, since later information suggests that Simpson's emission layer should be extended down into the upper troposphere.

It also seems likely that the values of the absorbed solar radiation used by Simpson, and by Baur and Phillips, are too small. It is now almost certain that the mean cloud albedo is smaller than 0.78. In fact, Fritz (1949a) has suggested a mean cloud albedo of about 0.5 and a planetary albedo of 0.35.

It is the purpose of the present paper to present new computations of the outgoing long-wave radiation to space and of the absorbed solar radiation. Insofar as possible, these computations are based on observational data. In particular, no assumptions are made regarding the planetary albedo nor the cloud albedo; in fact, the planetary albedo and the variation of the albedo with latitude are among the results. Only the annual mean heat balance is considered, although many of the numerical quantities are derived from monthly or seasonal values.

## 2. Solar radiation reaching the surface

In accordance with the decision to use observational material whenever possible, the solar radiation reaching the surface was derived from pyrheliometric data. Only in North America is there a network of pyrheliometric stations which record the total of the direct and diffuse solar radiation on a horizontal surface. The annual means from all pyrheliometric stations with a period of record of one year or more were entered on a base map. A few stations were discarded as non-representative because of the effects of local atmospheric pollution. In construction of the radiation isolines, it was found helpful to consult a map of the mean cloud cover, since this is the largest single factor affecting the solar radiation received at the surface. No attempt was made to reduce the data to sea level, since it was desired to obtain the radiation at the actual surface of the earth.

Mean values of the solar radiation reaching the surface were determined from the analyzed map along each 5-deg circle of latitude from 30°N to 65°N. The results are presented in table 1. In view of the rather subjective nature of the analysis of the radiation map, this was repeated independently by a second analyst. There were no significant differences in the averages along the latitude circles. After the analysis reported here was completed, Fritz and MacDonald (1949) published solar radiation maps of the United States for each month. Annual means along the latitude circles derived from their maps agree with those of table 1 within 5 per cent. Since the data of table 1 cover a wider range of latitude than those of Fritz and MacDonald, it was decided to use the former in this study.

In extending the data of table 1 to the entire hemisphere, it was assumed that, at any given latitude, the solar radiation received at the surface is dependent primarily on the cloud amount. The procedure was essentially to compare the mean cloud cover along an entire latitude circle with that along the portion of the circle in North America, and to correct the North American radiation accordingly. Since the two cloud amounts usually differ only slightly, the correction is not large and errors in the assumed relation between cloud amount and radiation have little effect on the final result. The data of table 1 cover only a portion of the region from equator to pole. It is not difficult to extrapolate from 65°N to the pole, but the region south of 30°N comprises one-half the area of the hemisphere, and blind extrapolation here is not justifiable.

To provide a sound basis for the extrapolations of the observed data and, simultaneously, to evaluate the solar radiation absorbed by the atmosphere, it was decided to compute the total solar radiation reaching the surface for clear skies. These computations were based on the results of the extensive investigations of the Astrophysical Observatory of the Smithsonian Institution. It was first planned to use the very convenient graph prepared by Kimball (1930), from the Smithsonian data, in these computations. Certain inconsistencies in this graph[1] led the writer to abandon it in favor of the original data. The pertinent data are scattered through the scores of publications of the Smithsonian Institution staff, principally by Abbott and Fowle, and no attempt will be made to list the references here. It is the practice of the Smithsonian workers to express the total transmission of the solar beam through a dust-free atmosphere as the product of three separate transmission ratios; these are the transmission due to dry-air scattering, the transmission due to water-vapor scattering, and the transmission due to water-vapor absorption. The three transmission ratios are plotted in fig. 1. The dry-air scattering is a function only of the air mass (unit air mass is a vertical path through the atmosphere from sea level), while water-vapor scattering and absorption are functions of the product of the air mass and the precipitable water. Computations of the total transmission from fig. 1 agree, as closely as can be expected, with the results of the Smithsonian Institution, *e.g.*, table 111 of the Smithsonian Meteorological Tables (1939). Any actual atmosphere contains dust, which introduces a fourth type of depletion. The dust content of the atmosphere is exceedingly variable. An examination of the available data on dust depletion led to the adoption of a "dust transmission" of $(0.95)^m$, where $m$ is the air mass.

TABLE 1. Annual means of total solar and sky radiation (ly/day) for North America from pyrheliometric data.

| Latitude | Radiation | Latitude | Radiation |
|---|---|---|---|
| 30°N | 412 | 50°N | 277 |
| 35 | 398 | 55 | 249 |
| 40 | 356 | 60 | 242 |
| 45 | 313 | 65 | 219 |

[1] For example, Kimball's curve showing the depletion by dry-air scattering is evidently based on the elevation of Mt. Wilson, where the pressure is 831 mb. This error is also included in others of Kimball's curves.

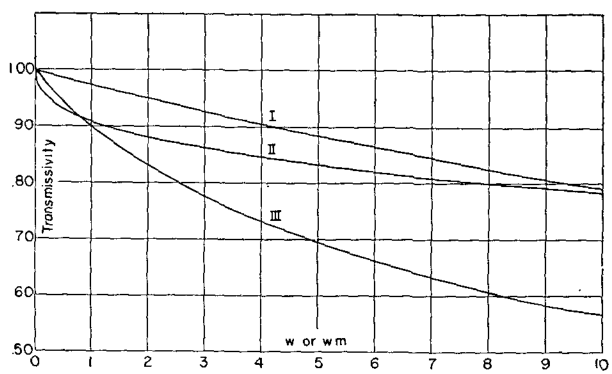


FIG. 1. Atmospheric transmission curves derived from data of Smithsonian Astrophysical Observatory; I: transmission due to water-vapor scattering; II: transmission due to water-vapor absorption; III: transmission due to scattering by dry, dust-free air. For curves I and II, abscissae are product of air mass and precipitable water; for curve III, abscissa is sea-level air mass. In figure, "w or wm" should read "m or wm".

Computations of the total solar radiation received at the surface of the earth were made for each 10 deg lat from the equator to the pole. Calculations were made for the two solstices and the two equinoxes, except in high latitudes where there is no sun in mid-winter. In the latter cases, sufficient computations were made to define the annual variation. The precipitable water at each latitude and time of the year was taken from a tabulation of Gabites (1950). The calculations were carried out for each 10 deg of hour angle through the day. The daily sums of the radiation were obtained by graphical integration, as were the annual means. The solar constant was taken as 1.94 ly/min. Absorption by ozone was not considered, because the Smithsonian procedures for determining the solar constant exclude that portion of the spectrum in which the strong ozone absorption occurs. For this reason, the heat balance considered in this paper is essentially the heat balance of the earth and the troposphere; the treatment of the outgoing long-wave radiation is consistent with this decision, as will appear below.

The total fractional depletion of the direct solar beam is one minus the total transmission. For the present purposes, this total loss must be apportioned between absorption and scattering. Atmospheric scattering occurs primarily in the short-wave portion of the solar spectrum, whereas the absorption occurs almost entirely in the long-wave portion. Furthermore, the scattered radiation is not lost but continues to traverse the atmosphere as diffuse radiation, either down or up. Radiation scattered above the center of gravity of the atmosphere has a total path somewhat less than the entire atmosphere, while radiation scattered below the center of gravity has a path somewhat greater than the entire atmosphere. From these considerations it was decided to assume, in effect, that the absorption occurs first, followed by the scattering. This means that the absorptivity was applied to the undepleted solar beam to obtain the absorbed radiation. It is believed that dust absorbs as well as scatters. In the absence of quantitative information, one-half of the dust depletion was counted as absorption. Finally, it was assumed that one-half of the scattered radiation reaches the surface as diffuse radiation or sky light. This is strictly true for molecular scattering, but larger particles scatter more radiation into the forward hemisphere than to the rear. The assumption made will lead to a small underestimate of the sky light and a corresponding overestimate of the planetary albedo.

The results of the computations are presented in table 2. They are given in more detail than is strictly necessary for the present purposes, in the hope that they will be useful to other investigators. Fritz (1949b) has published charts of the clear-sky radiation reaching the surface over the United States, based largely on pyrheliometric data. According to Fritz, the precipitable water on clear days is about 85 per cent of the mean for all days. When corrected for this and the reduction in precipitable water due to the elevation of the continent, the data of table 2 agree with those of Fritz within 2 per cent. A number of comparisons were also made with detailed observations of the direct radiation and of the diffuse radiation as a function of the zenith angle of the sun. As anticipated above, the computed diffuse radiation is slightly smaller than that observed. The computed direct radiation is slightly larger than the observed. All differences were small, and it is concluded that the

TABLE 2. Computed values (ly/day) of disposition of solar radiation incident on a clear atmosphere.

| | | Latitude (deg. N) | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 0 | 10 | 20 | 30 | 40 | 50 | 60 | 70 | 80 | 90 |
| Spring equinox | Outside atmosphere | 892 | 879 | 839 | 771 | 682 | 573 | 446 | 305 | | |
| | Direct beam at surface | 509 | 508 | 505 | 485 | 425 | 354 | 259 | 159 | | |
| | Diffuse at surface | 93 | 90 | 83 | 72 | 66 | 61 | 53 | 43 | | |
| | Absorbed by atmosphere | 198 | 190 | 169 | 142 | 123 | 99 | 82 | 61 | | |
| Summer solstice | Outside atmosphere | 791 | 873 | 934 | 973 | 992 | 989 | 978 | 1010 | 1058 | 1076 |
| | Direct beam at surface | 434 | 495 | 545 | 588 | 608 | 609 | 600 | 593 | 609 | 609 |
| | Diffuse at surface | 89 | 93 | 95 | 95 | 98 | 97 | 98 | 110 | 124 | 128 |
| | Absorbed by atmosphere | 182 | 195 | 201 | 195 | 192 | 189 | 184 | 198 | 203 | 211 |
| Fall equinox | Outside atmosphere | 883 | 871 | 830 | 764 | 675 | 568 | 441 | 302 | 174 | 60 |
| | Direct beam at surface | 504 | 497 | 477 | 445 | 387 | 318 | 232 | 137 | 60 | 4 |
| | Diffuse at surface | 92 | 92 | 87 | 78 | 72 | 65 | 55 | 44 | 30 | 10 |
| | Absorbed by atmosphere | 196 | 192 | 182 | 162 | 143 | 122 | 100 | 78 | 60 | 44 |
| Winter solstice | Outside atmosphere | 842 | 731 | 603 | 461 | 318 | 175 | 49 | | | |
| | Direct beam at surface | 462 | 398 | 331 | 256 | 164 | 78 | 12 | | | |
| | Diffuse at surface | 95 | 83 | 69 | 52 | 42 | 27 | 9 | | | |
| | Absorbed by atmosphere | 194 | 170 | 135 | 100 | 71 | 45 | 19 | | | |
| Annual mean | Outside atmosphere | 850 | 838 | 801 | 743 | 668 | 576 | 469 | 406 | 369 | 353 |
| | Direct beam at surface | 477 | 475 | 465 | 444 | 396 | 340 | 270 | 225 | 197 | 176 |
| | Diffuse at surface | 93 | 89 | 83 | 74 | 70 | 62 | 55 | 51 | 48 | 48 |
| | Absorbed by atmosphere | 193 | 187 | 172 | 150 | 132 | 114 | 95 | 82 | 81 | 81 |

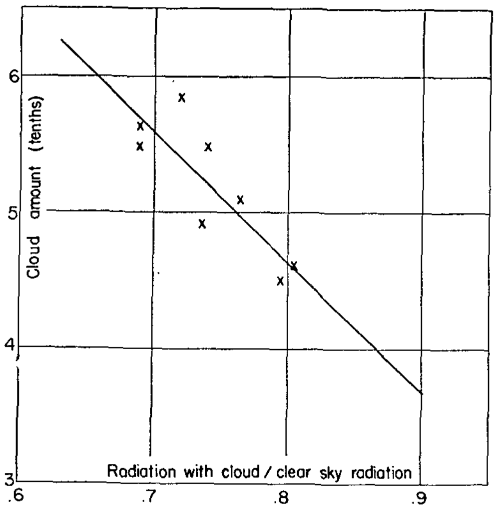


FIG. 2. Ratio of solar radiation reaching surface with normal cloud cover to computed radiation with clear skies *vs.* cloud amount. Data are from North America. Straight line is fitted to plotted points by method of least squares.

data of table 2 are entirely adequate for this study.

For each latitude circle, the ratio of the observed radiation from table 1 was taken to the annual mean computed clear-sky radiation from table 2. These ratios of radiation with normal cloud cover to radiation with clear skies were plotted against the radiation-weighted mean cloud cover, and the best straight line was fitted by the method of least squares, as shown in fig. 2. The cloud-cover data for North America were taken from Brooks *et al* (1936), who give the daytime cloudiness for January and for July. The annual mean was obtained by weighting the January data in accordance with the 21 December clear-sky radiation, and the July data in accordance with the 21 June radiation. Since an extension of the straight line of fig. 2 to zero cloud cover would give a ratio in excess of unity, it appears that the relationship is non-linear. This is unimportant, since the relation is to be used only within the range of mean cloudiness along latitude circles. The linear relation is a means of adjusting the observed radiation over North America to the radiation over the hemisphere in proportion to the corresponding differences in mean cloudiness. At most latitudes the adjustment is small and the shape of the curve is of minor importance. A more serious uncertainty lies in the inherent assumption that a given cloud amount has the same effect on the radiation over the entire northern hemisphere as it does in the North American segment.

The weighted mean cloud cover for the northern hemisphere was derived from the data of Brooks (1927), supplemented at high latitudes with data given by Haurwitz and Austin (1944) and from *Preliminary climatic atlas of the world* (Anonymous, 1943). As described above for North America, the cloud cover was weighted in proportion to the clear-sky radiation, to obtain a mean cloud cover appropriate for use with the annual mean of the solar radiation. These mean cloud amounts for the northern hemisphere are presented in the third column of table 3.

Fig. 2 was utilized to determine the factors by which the clear-sky surface radiation must be multiplied to obtain the values of the surface radiation for normal cloud cover, entered in column 4 of table 3.

## 3. The surface albedo

To determine the solar radiation absorbed at the surface of the earth, it is necessary to know the albedo of the surface. No previous study of the latitudinal variation of the surface albedo was found. From Goode (1947), the land areas of the northern hemisphere were classified into six types: grasslands (including tundra), forest, scrub lands, savanna, sand (desert) and water. The fractions of each of the latitude circles occupied by each of these six types of land surface were determined. When present, snow was assumed to cover grasslands and sand but to have no effect on the albedo of forests. Scrub was taken as half bushes and half sand; snow cover was assumed to affect only the sand. Savanna lands were treated as one-fifth trees and four-fifths grassland.

Appropriate values of the albedo of each type of surface were obtained from the literature. Based on

TABLE 3. Disposition of solar radiation (ly/day) with normal cloud cover.

| Lat. (deg N) | Solar rad. at surface with clear sky | Weighted cloud cover | Solar rad. at surface with normal cloud | Surface albedo | Solar rad. absorbed at surface | Absorbed in normally cloudy atmos. | Total absorption | Planetary albedo |
|---|---|---|---|---|---|---|---|---|
| 0 | 570 | 0.54 | 410 | .071 | 381 | 192 | 573 | .326 |
| 10 | 564 | 0.50 | 429 | .080 | 395 | 183 | 578 | .310 |
| 20 | 548 | 0.44 | 452 | .098 | 408 | 166 | 574 | .283 |
| 30 | 518 | 0.42 | 438 | .110 | 390 | 142 | 532 | .284 |
| 40 | 466 | 0.49 | 359 | .102 | 322 | 122 | 444 | .335 |
| 50 | 402 | 0.57 | 277 | .092 | 252 | 100 | 352 | .389 |
| 60 | 325 | 0.63 | 204 | .091 | 185 | 76 | 261 | .443 |
| 70 | 276 | 0.67 | 161 | .168 | 134 | 58 | 192 | .527 |
| 80 | 245 | 0.65 | 148 | .36 | 95 | 52 | 147 | .602 |
| 90 | 224 | 0.59 | 150 | .56 | 66 | 51 | 117 | .669 |
| Mean 0–90 | 475 | 0.52 | 361 | .100 | 325 | 136 | 461 | .34 |

Büttner's (1929) data, the albedo of forests was taken as 0.07. The albedo of grass was taken as 0.15, on the basis of data given by Büttner (1929) and Richardson (1919). An albedo of 0.25 was adopted for sand, on the authority of Büttner (1929) and Bartels (1930). A value of 0.10 for bushes is based on Büttner (1929). The snow cover was assumed to be fresh in the winter, and its albedo was taken as 0.70 in accordance with the measurements of Ångstrom (1929). From the same source, an albedo of 0.55 was adopted for old snow. For sea ice the same values were used as for snow. Winter snow and ice were assumed to be "new," and snow and ice areas in the summer were considered to be "old." For lakes and other water areas, an albedo of 0.07 was selected on the basis of the study of ocean albedos discussed below.

The albedo of the oceans was computed at three latitudes from data on the direct and diffuse solar radiation at Apia (Samoa), Washington (D. C.) and Uppsala (Sweden). The reflection coefficients of the sea surface for the direct solar beam as a function of zenith angle were taken from Ångstrom (1925) and Powell and Clarke (1936). A constant reflection coefficient for diffuse radiation of 0.065 was adopted from Powell and Clarke (1936). It is realized that larger albedos for diffuse radiation have been reported, but the measurements of Powell and Clarke were accepted for the present study. The mean daily ocean albedo for clear skies was obtained by graphical integration of the curves of the diurnal variation of the total incident and total reflected radiation. For overcast skies the radiation was assumed to be diffuse, and the albedo was taken as 0.065 independent of the solar zenith angle. The surface albedo for the mean cloud cover of the latitude circle was obtained by linear interpolation, in proportion to cloud amount, between the clear sky and overcast sky albedos. In this way, the ocean albedo at latitude 14 deg was found to be 0.064; at latitude 39 deg, 0.068; and at latitude 60 deg, 0.083.

The albedos of the several types of ground surface and of the ocean were combined in proportion to the segments of each along each 10-deg latitude circle. For all latitudes higher than the southern extent of winter snow cover, both a winter and a summer albedo were determined. The differences between summer and winter are due to the change in the area covered by snow, the change from "old" to "new" snow and the freezing of certain water areas. The winter and summer albedos were weighted in proportion to the December and June radiation to yield the weighted annual mean surface albedo to be found in the fifth column of table 3. One minus the albedo, multiplied by the solar radiation incident on the surface, gives the solar radiation absorbed at the surface, entered in the last column of table 3.

## 4. Solar radiation absorbed in atmosphere

The computed values of the solar radiation absorbed in a clear atmosphere, as given in table 2, cannot be used for an atmosphere with normal cloud cover without considering the effect of clouds on the absorption. A cloud layer reflects a substantial fraction of the incident radiation, and thus prevents it from encountering the water vapor below the cloud. The radiation reflected by the cloud, and that leaving the base of the cloud, is diffuse and is absorbed at a different rate than parallel-beam radiation. Substantial absorption is to be expected within the cloud, both because of the liquid water (or ice) and because multiple scattering may greatly increase the effective water-vapor path length. The magnitude of these several effects is dependent on the cloud height, depth and density, on the vertical distribution of the water vapor, and on the zenith angle of the sun. It seemed impossible to judge the net effect of clouds on the absorption. Even though the requisite data are exceedingly sparse and often inconsistent, it was decided to attempt a direct computation of the effect of clouds on the absorption.

Low, middle, high and cumuliform clouds were considered separately. For each class, an albedo and an absorptivity were chosen. The available data on cloud albedos are conveniently summarized by Fritz (1951). The only available data on cloud absorptivity are those given by Fritz and MacDonald (1951) for three very deep cloud systems, and those of Neiburger (1949) for California stratus. Hewson (1943) computed the absorption of clouds of specified particle size and liquid-water content, but his results give only the absorption in the liquid water and not the water-vapor absorption. Haurwitz (1948) gives values of the transmission of clouds of various types at Blue Hill. All of these data were used in a somewhat subjective fashion, to arrive at reasonable values for the albedo and absorptivity for the four cloud categories. The values adopted are given in table 4. Cloud heights and depths were taken from Hann-Süring (1939). The sky cover for each cloud type as a function of latitude was taken from an unpublished study based on data from the last Polar Year. Computations were made for the equinoxes at latitudes 0, 20, 40 and 70°N, with use of the same data employed in computing the absorption in a clear atmosphere. As a result of an analysis of the absorption of diffuse solar radiation by water vapor, an effective path length of

TABLE 4. Albedo and absorptivity of clouds.

| | Cloud type | | | |
|---|---|---|---|---|
| | Low | Middle | High | Cumuliform |
| Albedo | .69 | .48 | .21 | .70 |
| Absorptivity | .06 | .04 | .01 | .10 |

1.62 times the vertical path length was used for diffuse radiation. [Note that this is almost the same as the similar figure for diffuse long-wave radiation given by Elsasser (1942).] The absorption was computed as the sum of the absorption of the radiation reflected from the cloud, the absorption by the cloud, and the absorption of the radiation transmitted by the cloud. The absorption was integrated over the zenith angle of the sun, to obtain the daily absorption for 10/10 coverage of each cloud type. The absorption for the normal cloud cover was computed by weighting the absorption for each cloud type (and the clear-sky absorption) in proportion to the fractional coverage of each type. The ratio of the absorption with normal cloud cover to the clear-sky absorption was then determined. These ratios were found to be 0.995 at 0°N, 0.964 at 20°N, 0.921 at 40°N, and 0.705 at 70°N. It therefore appears that the normal cloud cover results in a smaller absorption than with clear skies, particularly at high latitudes. The reduction in absorption is greatest with high and middle clouds; in low latitudes, a cumuliform cloud cover results in greater absorption than with clear skies. The reduction in the ratio of cloudy-sky to clear-sky absorption with latitude is due to the change in frequency of low and cumuliform clouds and to the increase of the average direct-beam path length with latitude; the diffuse radiation has a constant path length of 1.62, which is smaller than the mean path length of the direct beam in middle and high latitudes.

These results are dependent on uncertain assumptions and inadequate data. In particular, the assumed cloud absorptivities are little more than an educated guess. On the other hand, the mean cloud albedo derived from the data used here is 0.55, in close agreement with the estimate made by Fritz (1949a) in an entirely different way. Also, the radiation transmitted by the clouds was made to fit the data of Haurwitz (1948) and is in fair agreement with the data of fig. 2. Since the sum of the cloud albedo, the cloud transmissivity, and the cloud absorptivity is unity, the values chosen for the cloud absorptivity cannot be greatly in error.

## 5. The planetary albedo

The solar radiation absorbed in the atmosphere with normal cloud cover was obtained by multiplying the computed clear-sky absorption from table 2 by the factors given above. The results are entered in the seventh column of table 3. The planetary albedo is then easily derived as a function of latitude from the total absorption and the radiation incident on the outer atmosphere. The minimum albedo in the subtropical high-pressure belt is clearly evident. The increase in the albedo at higher latitudes is due to the greater cloudiness and, at very high latitudes, to the snow and ice cover. The mean albedo of the hemisphere is about 0.34, in excellent agreement with the independent estimate of 0.35 made by Fritz (1949a). Since the mean cloud cover over the southern hemisphere is greater than that over the northern hemisphere, it is probable that the albedo of the whole earth is somewhat larger than 0.34.

## 6. Long-wave radiation to space

As indicated in section 1 above all computations of long-wave radiation to space were performed on the radiation chart devised by Elsasser (1942). The writer recognizes the validity of the objections that have been raised to this chart on theoretical grounds, but feels that is is the best available tool for such computations. The data used were the monthly mean soundings of some 30 aerological stations, including weather ships, ranging from the equator to 71°N. The precipitable water above the highest level at which humidities were given was obtained by a smooth extrapolation (of the precipitable water) to zero at the tropopause. This is consistent with the treatment of the solar radiation in which ozone absorption was neglected, and means that the heat balance presented here refers to the earth and the troposphere. This will probably introduce no great error, since the stratosphere is at least approximately in radiative equilibrium. Several other means of extrapolating the water vapor to the tropopause were tried, and all reasonable methods gave essentially the same results.

The square-root pressure correction was used, as by Elsasser. It is now well established that the half-width varies directly with the pressure, but the pressure dependence of the absorption of a band depends on the form of the transmission function of the band. In view of the several fundamental deficiencies of the Elsasser chart, it seems best to consider the chart and the pressure correction as empirical. It was found that, on the average, a linear pressure correction increased the outgoing long-wave radiation by 3 per cent. This is almost certainly within the precision of the results and even of the sounding data.

For many of the stations separate computations were made by Degani (1942) for each of the twelve months, to determine the annual mean. From his results it was found that the mean of four months (January, April, July and October) was adequate, and this procedure was followed for the balance of the stations. The aerological stations used in this study were primarily from North America and the North Atlantic. When the outgoing long-wave radiation from each of the stations was plotted against latitude, it was found possible to draw a smooth curve through the points such that the extreme deviations were ±5 per cent, and the mean deviation without regard to sign was 0.4 per cent. There were no apparent dif-

TABLE 5. Annual mean outgoing long-wave radiation (ly/day) with normal cloud cover.

| Latitude | Outgoing radiation | Latitude | Outgoing radiation |
|---|---|---|---|
| 0°N | 488 | 50°N | 442 |
| 10 | 502 | 60 | 419 |
| 20 | 503 | 70 | 400 |
| 30 | 492 | 80 | 385 |
| 40 | 469 | 90 | 380 |

ferences among east-coast, west-coast, continental and oceanic stations, a fact which lends support to the decision to assume that the curve could be used for the entire hemisphere.

It was necessary to estimate the effect of the mean cloud cover on the outgoing long-wave radiation. To this end, the distribution of clouds by type was obtained from an unpublished study for January, April, July and October at latitudes 17, $45\frac{1}{2}$ and $71\frac{1}{3}$°N. The mean heights of the tops of the several cloud types were taken from Hann-Süring (1939). For each cloud type the outgoing long-wave radiation was computed on the Elsasser chart, with the assumptions of 10/10 cloud cover and that the cloud acts as a black radiator. The outgoing long-wave radiation with normal cloud cover was then found by weighting the results in proportion to the fraction of sky coverage by each cloud type (including clear sky as a "cloud type"). In this way, it was found that the mean cloud cover reduces the clear-sky outgoing long-wave radiation by 7.2 percent at 17°N, 7.6 per cent at $45\frac{1}{2}$°N, and 5.9 per cent at $71\frac{1}{3}$°N. The appropriate reductions for other latitudes were obtained by interpolation and applied to the outgoing radiation computed for clear skies. These results are presented in table 5 and, in modified form, in fig. 3. The hemispheric mean of the outgoing long-wave radiation is 472 ly/day, which is to be compared to the hemispheric mean of the absorbed solar radiation of 461 ly/day. The agreement, within 3 per cent, of these two figures,

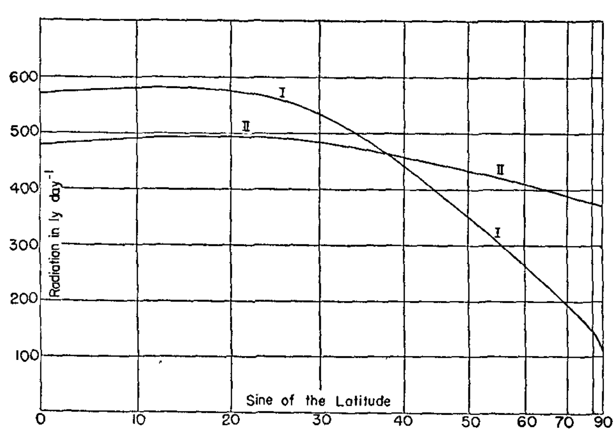


FIG. 3. Annual mean of solar radiation absorbed by earth and atmosphere (curve I), and outgoing long-wave radiation leaving atmosphere (curve II), as functions of latitude.

arrived at in entirely independent fashions, is gratifying. It is worthy of note, however, that earlier investigators (such as Simpson, and Baur and Phillips) obtained equally good agreements, although at significantly lower radiation levels. Perhaps there is a special law that insures the approximate balance of such estimates.

## 7. Nocturnal radiation

Although it does not appear in the overall radiation balance, the net upward flux of long-wave radiation at the surface of the earth is an important item in the heat balance between the earth and the atmosphere. This quantity, which is often called the nocturnal radiation, may be evaluated on the Elsasser radiation chart. Cloud cover has a much more important effect on the nocturnal radiation than on the outgoing radiation at the tropopause. Low clouds have the greatest effect, and it was not considered satisfactory to use only the cloud data employed in the study of the effect of clouds on the outgoing radiation. This information was supplemented by data showing the frequency of ceiling heights at a number of selected stations. In this way, it was found that the hemispheric mean of the annual nocturnal radiation is 96 ly/day. The nocturnal radiation with clear skies was also evaluated and found to be 145 ly/day. The latter figure is considered to be more reliable than the one for normal cloud cover, and was obtained primarily to establish an upper limit. Since the average cloud cover is about 0.5, and since a cloud cover of average height will not reduce the nocturnal radiation to zero, it is to be expected that the value for the normal cloud cover should lie between one-half and one times the clear-sky value. The linear mean is 109 ly/day, which lends further support to the computed result of 96 ly/day.

## 8. The radiation and heat balance

The total absorbed solar radiation from table 3, and the outgoing long-wave radiation at the tropopause, are plotted in fig. 3. For this comparison the long-wave outgoing radiation at each latitude has been multiplied by 461/472, so that the areas under the two curves are exactly equal. It will be noted that the two curves intersect at latitude 38°N, in reasonably good agreement with Simpson's similar intersection at 35°N, and Baur and Phillips' at 37°N. With the exception of the absorbed solar radiation in high latitudes, the ordinates of the curves of fig. 3 are significantly larger than those of Simpson's similar curves.

From fig. 3, the poleward flux of heat required to maintain the overall balance has been computed. Table 6 shows this flux in terms of the total heat

TABLE 6. Required poleward flux of heat ($10^{19}$ cal/day) across latitude circles.

| Latitude | Flux | Latitude | Flux |
|---|---|---|---|
| 0°N | 0 | 50°N | 9.61 |
| 10 | 4.05 | 60 | 6.68 |
| 20 | 7.68 | 70 | 3.41 |
| 30 | 10.46 | 80 | 0.94 |
| 40 | 11.12 | 90 | 0 |

transport across a complete circle of latitude, for each 10 deg of latitude. The heat flux is found to be larger at all latitudes than those of Simpson and of Baur and Phillips. The greatest differences occur at higher latitudes.

The heat balance between the surface of the earth and the atmosphere is of considerable interest. This balance involves a flux of latent heat and of sensible heat, in addition to the radiational items. The mean upward flux of latent heat can be computed from the mean rainfall. The flux of sensible heat is a micro-meteorological problem, and no attempt has been made by the writer to make an independent estimate of its magnitude. The heat balance, derived from the results of this paper, is presented in schematic form in fig. 4, in which the unit is 1 per cent of the solar radiation incident on the outer atmosphere. The flux of sensible heat has here been selected to produce an overall balance at the surface. It thus includes the integrated errors of the other quantities entering into the heat balance, and no great confidence can be placed in its numerical value. On the other hand, there seems little reason to doubt its sign. It is now generally recognized that, in the mean, the eddy flux of heat is directed upwards, but its magnitude is still in debate.

## 9. Discussion and conclusions

It has already been pointed out that the planetary albedo of 0.34 obtained in this study is in good agreement with the recent independent estimate of Fritz (1949), derived from the lunar observations of Danjon.

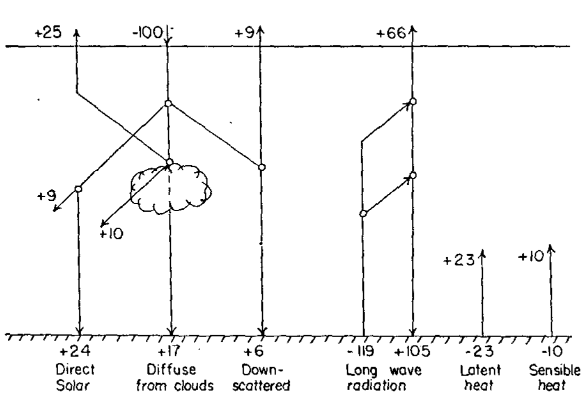


FIG. 4. Schematic representation of heat balance of earth and atmosphere. Solar radiation reflected from earth's surface is not shown separately, but is included in solar radiation returned to space.

In consequence, the absorbed solar radiation and the outgoing long-wave radiation are some 16 per cent larger, on the average, than those obtained by the earlier investigators who used albedos of 0.40 to 0.43. An even larger increase in the required flux of heat across the latitude circles has resulted from the derived latitudinal variation of albedo that is due to variations in cloud cover and to the increase in surface albedo at high latitudes.

In the writer's opinion, the most uncertain item entering into the present radiation balance is the absorption of solar radiation in the atmosphere. The use of the Smithsonian data without a pressure correction is generally thought to overestimate the water-vapor absorption. There is an additional uncertainty about the assumption that one-half of the dust depletion is absorption. On the other hand, a recent study by Fritz (1953) suggests that the cloud absorption has been underestimated here. If the total absorption of solar radiation has been overestimated, it is probable that the planetary albedo has been slightly underestimated, since the absorption at the surface is so closely tied to pyrheliometric observations.

It is of interest to compare the results of this study with those given by London (1951) for the month of March, and tentatively extended by him to an annual basis. London finds a planetary albedo of 0.355 and an absorption in the atmosphere of 0.145 of the incident radiation. In estimating the atmospheric absorption, London used somewhat smaller water-vapor absorptivities than those adopted here. He included absorption due to clouds and ozone, but not that due to dust. In part, his lower absorption is due to the fact that the mean annual precipitable water is greater than that for March. London's value for the mean outgoing long-wave radiation agrees within 2 per cent with that of the present study. In general, the agreement of the hemispheric mean values of all the items in the radiation budget as arrived at in these independent studies is as close as present basic uncertainties permit. On the other hand, there appear to be important discrepancies between the latitudinal distributions of the absorbed solar radiation and of the outgoing long-wave radiation. As compared to the results presented in this paper, London finds a smaller latitudinal variation of the absorbed solar radiation and a greater latitudinal variation of the outgoing long-wave radiation. These differences both act to decrease the radiational excesses and deficits, and hence to reduce the required heat transport across the latitude circles. Thus, London computes a maximum heat flux of $5.67 \times 10^{19}$ cal/day at 40°N, which is to be compared to the $11.12 \times 10^{19}$ cal/day found in the present study. It is of interest that Simpson gives $8.04 \times 10^{19}$ cal/day, and Baur and Phillips $9.11 \times 10^{19}$ cal/day, at 40°N. London's results for the absorbed solar radiation are slightly smaller than those presented here in low and middle latitudes, and increasingly larger at latitudes north of 60°N. The difference at low and middle latitudes is due to London's smaller atmospheric absorption; at high latitudes, the difference results from the higher surface albedos used here for the snow- and ice-covered areas. It is more difficult to explain the larger latitudinal gradient of the outgoing long-wave radiation found by London. His values are both larger at low latitudes and smaller at high latitudes than those of this study. Inspection of the data for March obtained here suggests that London's adjustment of his March results to represent the annual mean values was not entirely adequate. The major part of the difference must be attributed to London's use of a mean atmospheric cross-section for March, whereas the writer used data from individual sounding stations. Both techniques are based on the same data, and it is difficult to say which is the better approach. It must be concluded that, although the hemispheric mean values entering into the heat budget are reasonably well-established, considerable uncertainty remains regarding the latitudinal distribution and the resultant meridional transport of heat.

The results presented in this paper should be considered simply as new estimates and not as definitive. When further substantial strides have been made in our quantitative knowledge of atmospheric radiation, the radiation balance should be studied anew. It should be emphasized that annual means as presented in this paper have only limited application to studies of the general circulation. The next step is to consider monthly or seasonal values. This involves the additional problem of heat storage. Gabites (1950) has made a preliminary investigation of the heat balance month by month. Although this is beset by even greater uncertainties than those encountered here, his results clearly indicate the profound differences between the annual mean picture and the monthly patterns. Studies of this type are essential to a quantitative understanding of atmospheric circulations, and it is hoped that certain portions of this paper will be useful in efforts along these lines.

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