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FURTHER STUDIES IN TERRESTRIAL RADIATION

BY

G. C. SIMPSON, C.B., F.R.S., F.R.MET.SOC.

[Throughout this paper the unit of radiation is the gram calorie per square centimetre per minute.]

Each year the earth receives from the sun a definite amount of energy in the form of solar radiation. A certain proportion of this energy is immediately returned to space by reflection from clouds, dust and the earth's surface without undergoing any change; the remainder is absorbed by the earth and its atmosphere, but is finally returned to space as terrestrial radiation. It is a fundamental problem of meteorology to determine how, when and where this return is made.

The temperatures found on the earth and in the atmosphere vary within relatively small limits, from approximately 200°a to 300°a, and we know that within this range of temperature the elementary gases of the atmosphere, oxygen, nitrogen, argon, etc., emit no radiation whatever.[1] We are therefore left with only two sources of terrestrial radiation, namely, the surface of the earth and the compound gases of the atmosphere: water vapour, carbon dioxide, ozone, etc.

The radiation from the surface of the earth can easily be calculated; for, to the accuracy with which we are concerned, the surface may be treated as a black body, and the energy emitted by a black body at different temperatures is well known and can be found in books of physical tables.

The radiation from the gases of the atmosphere is not so easily determined, for it depends on the distribution, temperature and absorption coefficients of the various gases. Of the compound gases in the atmosphere, ozone exists only in the extreme upper regions, and is generally neglected in discussions on terrestrial radiation; on the other hand, carbon dioxide is present in relatively large quantities throughout the atmosphere, but it absorbs only in a very narrow band of the long wave spectrum. Water vapour is the most important radiating gas in the atmosphere.

As the result of the large amount of work done during recent years on the investigation of the upper atmosphere, we know with a fair degree of accuracy the distribution of water vapour throughout the atmosphere and its temperature. The radiation could therefore be calculated if we knew the absorption coefficient of the vapour. Unfortunately the absorption coefficient of water vapour depends largely on the wave-length, and it is almost impossible to carry out any computations if this variation is taken into account. Nearly all writers on this subject have therefore neglected the variation of absorption with wave-length and assumed that, to a first approximation, the absorption of water vapour is the same in all wave-lengths emitted by bodies at atmospheric temperatures. Accepting this

[1] Burmeister, *Berlin, Verh. D. physik. Ges.*, **15**, 1913, p. 612.

assumption I recently attempted to calculate the outgoing radiation from the atmosphere, and the results were published by the Society in a paper entitled "Some Studies in Terrestrial Radiation."[2] The results of these studies were surprising and unsatisfactory. In the first place, they showed that no value of the uniform absorption coefficient would give the right amount of outgoing radiation to balance the effective incoming radiation unless Aldrich's value of the albedo were largely in error.[3] In the second place, the calculations showed that with a uniform absorption coefficient the outgoing radiation would be unchanged when the temperature of the earth's surface varied within wide limits, and that an increase in the incoming radiation would not produce a corresponding increase in the outgoing radiation. The concluding paragraph of the summary attached to my paper was as follows:

> As the outgoing radiation (assuming a uniform absorption coefficient for water vapour) is practically independent of the temperature of the surface, the problem arises as to how the temperature of the atmosphere readjusts itself to changes in solar radiation. This problem is considered in detail, but no solution is found.

An interesting problem of this nature naturally called for further investigation, and since my previous paper was written I have constantly had the problem in mind. It was clear that the variation of absorption coefficient with wave-length could not be neglected and that no progress could be made until this variation was taken into account. On the other hand, calculations which attempted to employ the complicated absorption spectrum of water vapour were entirely intractable. The further study given to the subject finally led to a method of determining the radiation which, without the use of difficult mathematics and without the use of long calculations, appears to give a simple and satisfactory solution of the whole problem. These further studies form the subject of this paper.

The Amount of Water Vapour and Carbon Dioxide in the Stratosphere.

It will be necessary in what follows to know the amount of water vapour and carbon dioxide in the stratosphere. Gold[4] has given good reasons for believing that at the base of the stratosphere the air must be completely saturated. Taking the temperature at the base of the stratosphere to be about 220°a, this gives a saturation pressure of about ·033 mb. Owing to the absence of convection within the stratosphere, the distribution of water vapour will obey Dalton's law, and therefore the total mass of water in the stratosphere will be that amount which can be supported by the vapour pressure at the base of the stratosphere.

Now a vapour pressure of ·033 mb. is 33 dynes per square centimetre. If m is the mass of water vapour in grams over each square centimetre of the base of the stratosphere we have $mg = 33$, *i.e.* $m = \cdot 034$ gram. Thus the water vapour in the stratosphere is equivalent to ·34 mm. of precipitable water. To be on the safe side we will assume that there is at least ·3 mm. of precipitable water in the stratosphere.

The amount of CO_2 in the stratosphere can be found in a similar way.

[2] G. C. Simpson, *Memoirs of the Royal Meteorological Society*, Vol. **2**, No. 16.

[3] On page 87 of *Some Studies* I stated that the original value of the albedo as determined by Abbot was ·337. This is a mistake for ·37, and the change from Abbot's value to Aldrich's value was 16 per cent and not 30 per cent as stated. The error is regretted.

[4] E. Gold, *Geophysical Memoirs*, No, 5, M.O. 210 e, p. 129, 1913.

The partial pressure of CO_2 throughout the troposphere is ·03 per cent of the total pressure. The total pressure at the base of the stratosphere is approximately 200 millibars; the partial pressure of the CO_2 is therefore 60 dynes per cm². If M is the mass of CO_2 over each square centimetre of the base of the stratosphere we have $Mg = 60$, *i.e.* $M = \cdot 06$ gram.

THE ABSORPTION COEFFICIENTS OF WATER VAPOUR.

The first requisite for the new study is an accurate knowledge of the way in which water vapour absorbs terrestrial radiation. In other words, as terrestrial radiation is all of wave-length longer than about 4 μ, we require to know the absorption coefficient for each wave-length greater than about 4 μ. This has been provided in the case of water vapour in a valuable paper by Hettner,[5] which gives a curve of the absorption coefficient of water vapour from 0·9 μ to 34 μ. From Hettner's curve Fig. 1 has been prepared. This gives, as a fraction of the incident radiation, the absorption which a beam of parallel radiation would experience in passing through a layer of water vapour equivalent to ·3 mm. of precipitable water.

From 3·5 μ to 4·4 μ water vapour is perfectly transparent; at 4·4 μ an absorption band commences which is so intense that at 5·5 μ the layer we are considering absorbs the parallel beam of radiation completely; complete absorption continues to 7 μ, where the absorption commences to decrease and rapidly falls to low values in the neighbourhood 8 μ. Low values are maintained to about 11 μ, where another band commences and the values rise again. From 12 μ the absorption rises, but the rise takes place in a number of oscillations due to secondary absorption bands. These are not important to us, and we may consider a general rise which leads to total absorption at 20 μ. Hettner's work shows that the absorption steadily increases as the wave-length increases beyond 20 μ, but this increase cannot be shown on our curve because the layer of water vapour which we are considering has already absorbed the whole of the radiation with a wave-length of 20 μ, and it cannot absorb more than the whole even of wave-lengths which are more absorbable.[6] The increase in absorption beyond 20 μ simply means that a layer containing less water vapour than ·3 mm. precipitable water would suffice to absorb it all. In fact, Hettner's curve shows that a layer containing as little as ·07 mm. of precipitable water in vapour form would absorb all the incident radiation having a wave-length of 34 μ. Hettner's observations did not extend beyond 34 μ. It is necessary to return to the portion of the curve between 8 μ and 12 μ. Hettner's curve shows some small absorption throughout this region even with so small a quantity of vapour as ·3 mm. On the other hand, observations of solar radiation show that in this region the atmosphere appears to be practically transparent. Observations made by Abbot and Aldrich could detect no certain absorption by the aqueous vapour in the atmosphere from wave-length 9 μ to 12 μ even when there was vapour in the path equivalent to 3 cm. of precipitable water.[7]

The layers of water vapour on which Hettner made his observations were columns of steam. The water vapour was therefore much more dense in his experiments than it ever occurs in the atmosphere; and we know that when gases are very dense their absorption increases. There is

[5] G. Hettner, *Ann. Physik. Leipzig*, 4th Folge, Band **55**, p. 476, 1918.

[6] Total absorption is of course never reached; but for practical purposes the curve in Fig. 1 has reached total absorption at 20 μ.

[7] Fowle, *Smithsonian Misc. Collection*, Vol. **68**, p. 41, 1917.

reason, therefore, to believe that Hettner's curve would have reached the base line between 8 μ and 12 μ if he had been able to use water vapour having the density with which it occurs in the atmosphere. Hettner's curve has therefore been modified between 8 μ and 12 μ as shown by the dotted lines in Fig. 1, which it will be noticed simply continue the main

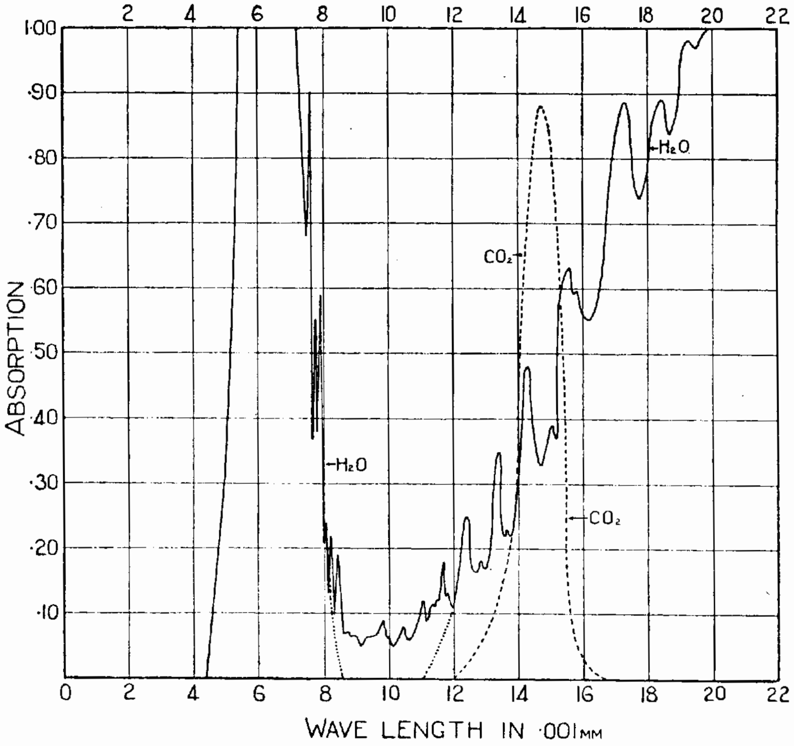

FIG. 1.—Absorption curves.

The full line is Hettner's curve for water vapour reduced to ·3 mm. precipitable water. The broken line is Rubens and Aschkinass' curve for CO_2 reduced to ·06 gram per cm^2. cross section.

branches on the two absorption bands to meet the abscissa at 8½ μ and 11 μ respectively. We shall therefore in the rest of this paper consider that the atmosphere is perfectly transparent to radiation between wave-lengths of 8½ μ and 11 μ.

THE ABSORPTION COEFFICIENTS OF CARBON-DIOXIDE.

Rubens and Aschkinass[8] have shown that carbon-dioxide has only one absorption band for wave-lengths greater than 4 μ. This band is narrow, but very intense, and has its maximum at 14·7 μ. Rubens and Aschkinass's curve of absorption for this band is shown in Fig. 1, reduced to an absorbing layer containing ·06 gram of CO_2 per square centimetre cross section.

[8] Rubens and Aschkinass, *Annalen der Physik und Chemie*, Vol. **64**, p. 584, 1898.

THE ABSORPTION OF RADIATION BY THE STRATOSPHERE.

In the sections dealing with the water vapour and CO_2 in the stratosphere we determined the quantity of the former to be the equivalent of ·3 mm. of precipitable water and of the latter to be ·06 gram per sq. cm. These values were used in preparing the curves in Fig. 1, and therefore Fig. 1 represents the absorption of radiation by water vapour and CO_2 in the stratosphere.

In Fig. 1, however, the two absorptions are shown separately; in order to combine the two it is only necessary to multiply together the proportion of each transmitted to give the proportion transmitted by the two together. Thus at the maximum of the CO_2 curve the absorption is ·88, and at the same wave-length the absorption of the water vapour is ·33; at this wave-length, therefore, the proportions transmitted separately are ·12 and ·67 and together $\cdot 12 \times \cdot 67 = \cdot 08$, which gives ·92 as the proportion absorbed. On combining the two curves in this way and removing the irregularities we

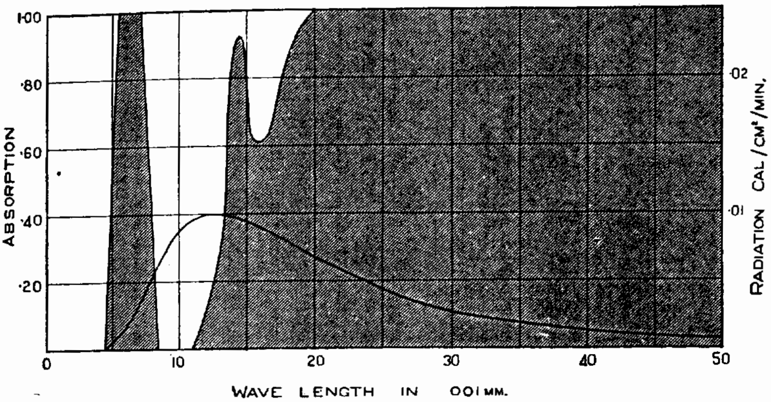

FIG. 2.—Absorption by the stratosphere and radiation curve for black body at 220°a.

have the absorption by the stratosphere shown in Fig. 2. On the same figure is shown the energy radiated in each wave-length by a black body at the temperature of the stratosphere 220°a. The absorption shown in Fig. 2 is, however, that of a beam of radiation traversing the stratosphere in a vertical direction. The actual radiation which enters the stratosphere comes from all directions, and as the absorption increases with the length of the path traversed it will be seen that most of the radiation which enters the stratosphere suffers a greater absorption than that represented in Fig. 2. It is impossible to calculate the excess of absorption due to the radiation being diffuse instead of parallel, but we can make no great mistake if we consider that all the radiation of greater wave-length than 14 μ, where the maximum of the CO_2 band occurs, is completely absorbed by the stratosphere. In addition to this region there is also the region between $5\frac{1}{2}$ μ and 7 μ in which complete absorption takes place.

THE RADIATION FROM A LAYER OF GAS.

The argument of the remainder of this paper depends on two very simple principles based on the theory of radiation, which may be set out as follows:

(*a*) If a layer of gas at a uniform temperature T throughout, completely absorbs radiation of wave-length λ, then it will emit radiation *of this wave-length* exactly as if it were a black body at temperature T.

(*b*) If a layer of gas rests on a black surface of infinite extent at temperature T_0, and the temperature within the gas decreases from the surface outwards so that at its outer surface the temperature of the gas is T_1, then the flux of radiation outwards of wave-length λ cannot be greater than that of a black body at temperature T_0 nor less than that of a black body at temperature T_1.

I do not propose to give a formal proof of these two principles, they will be perfectly obvious to those who have any acquaintance with the theory of radiation, and others may accept them with confidence.

The Outgoing Radiation from the Atmosphere in Latitude 50°.

The average mean annual temperature of the surface in latitude 50° is 280°a (7° C.) and the temperature of the stratosphere 218°a. In Fig. 3

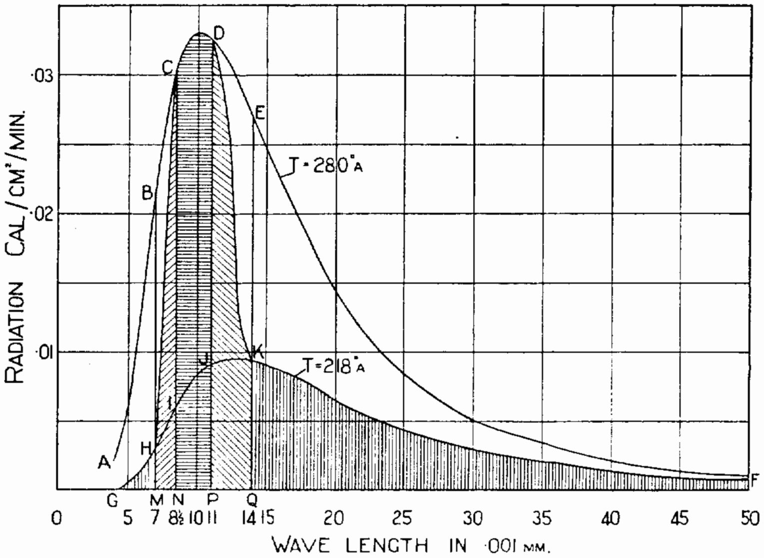

Fig. 3.—Outgoing radiation from latitude 50°.

two curves are drawn showing the black-body radiation for these two temperatures according to wave-length. The total area between the base of the diagram and the curve ABCDEF represents the total energy emitted by a black body at the temperature of the ground, 280°a, while the smaller area under the curve GHIJKF represents the energy which would be emitted by a black body at the temperature of the stratosphere, 218°a.

It has been pointed out above that the stratosphere absorbs all radiation between wave-lengths $5\frac{1}{2}$ μ and 7 μ, and from wave-length 14 μ to the end of the spectrum.

Therefore in these two regions the stratosphere radiates as a black body at its own temperature. The radiation emitted is therefore represented by the two areas GHM and QKF (F being supposed to be at infinity) which have been indicated in the diagram by vertical hatching.

The atmosphere is perfectly transparent to radiation having wave-lengths between $8\frac{1}{2}\ \mu$ and $11\ \mu$, therefore the radiation from the ground between these wave-lengths passes completely through the atmosphere. Thus the energy represented by the area NCDP passes into space.

Turning now to the radiation of wave-length between $11\ \mu$ and $14\ \mu$. This radiation is partially absorbed in the atmosphere, partially transmitted, and also the water vapour itself distributed throughout the atmosphere emits radiation in this region. The absorption in this region varies so rapidly from wave-length to wave-length (see Fig. 1) that a calculation of the outgoing radiation is quite impracticable. Our second principle (*b*), however, gives us definite limits between which the radiation must lie: it must be greater than the radiation emitted by a black body at the temperature of the stratosphere, and less than the radiation emitted by a black body at the temperature of the ground. This means that the radiation must be greater than that represented by the area PJKQ, and less than that represented by the area PDEQ. Some portion of the area JDEK has therefore to be added to the area PJEQ to represent the outgoing radiation between these wave-lengths. We also know that except for the irregularities on the absorption curve, which may be neglected, the absorption varies regularly from the clear band at $11\ \mu$ to total absorption at $14\ \mu$. Therefore the point D has to be connected to the point K by a regular curve. Although we do not know exactly how this curve should be drawn, there is really very little latitude; for any reasonable curve will divide the area JDEK into two nearly equal parts. In other words, a very near approximation to the total outgoing radiation will be obtained by adding half the area JDEK to the area PJKQ. This is equivalent to $\frac{1}{2}$(PDEQ + PJKQ); which is simply the mean of the radiation emitted between $11\ \mu$ and $14\ \mu$ by two black bodies, one at the temperature of the stratosphere and the other at the temperature of the surface. In the same way we obtain the radiation emitted between the wave-lengths $7\ \mu$ and $8\frac{1}{2}\ \mu$. We have now accounted for the radiation of all wave-lengths emitted by the earth and the atmosphere, and we are in a position to determine the total outgoing radiation. From Table I. (p. 8) we extract the values to carry out the following computations:

Wave-lengths $5\frac{1}{2}\ \mu$ *to* $7\ \mu$

Outgoing radiation from black body at 218°a. $= \cdot 003$ cal./cm.²/min.

Wave-lengths $7\ \mu$ *to* $8\frac{1}{2}\ \mu$

Mean of outgoing radiation from black bodies at 280°a and 218°a. $= \dfrac{\cdot 041 + \cdot 007}{2} = \cdot 024.$

Wave-lengths $8\frac{1}{2}\ \mu$ *to* $11\ \mu$

Outgoing radiation from black body at 280°a. $= \cdot 079.$

Wave-lengths $11\ \mu$ *to* $14\ \mu$

Mean of outgoing radiation from black bodies at 280°a and 218°a. $= \dfrac{\cdot 091 + \cdot 028}{2} = \cdot 059.$

Wave-length greater than $14\ \mu$

Outgoing radiation from black body at 218°a. $= \cdot 128.$

Total outgoing radiation . . $\cdot 293.$

TABLE I.—INTENSITIES OF THE BLACK-BODY SPECTRUM.

Based on Table 276 *Smithsonian Physical Tables*, Seventh Edition, 1920.

If the figures given in the table are plotted in cm. as ordinates to a scale of abscissæ of 1 cm. to 1 μ, then the area in cm.^{2} between the smooth curve through the resulting points, and the axis of abscissæ, is equivalent to the radiation in calories per minute from 1 cm.^{2} of a black body at the corresponding temperature radiating to absolute zero.

λ.	200°a.	207°a.	212°a.	218°a.	220°a.	261°a.	280°a.	292°a.	298°a.	299°a.
μ										
4	·0000	·0000	·0000	·0000	·0000	·0006	·0016	·0027	·0032	·0034
5	·0000	·0003	·0004	·0006	·0006	·0030	·0061	·0094	·0118	·0122
6	·0003	·0008	·0010	·0014	·0015	·0071	·0137	·0193	·0234	·0238
7	·0011	·0017	·0022	·0028	·0031	·0129	·0218	·0292	·0339	·0348
8	·0022	·0032	·0039	·0050	·0054	·0172	·0284	·0368	·0409	·0417
9	·0032	·0044	·0055	·0068	·0073	·0207	·0319	·0400	·0440	·0452
10	·0043	·0057	·0068	·0083	·0088	·0226	·0331	·0408	·0453	·0460
11	·0050	·0063	·0076	·0090	·0096	·0230	·0325	·0400	·0440	·0441
12	·0056	·0069	·0080	·0094	·0100	·0226	·0314	·0376	·0409	·0415
13	·0060	·0071	·0081	·0095	·0100	·0215	·0295	·0347	·0377	·0381
14	·0061	·0073	·0082	·0094	·0098	·0205	·0266	·0314	·0339	·0344
15	·0061	·0072	·0079	·0090	·0095	·0188	·0244	·0280	·0300	·0300
20	·0049	·0053	·0058	·0064	·0067	·0116	·0145	·0162	·0171	·0172
25	·0034	·0037	·0040	·0043	·0044	·0069	·0083	·0091	·0095	·0095
30	·0023	·0026	·0028	·0030	·0031	·0043	·0050	·0055	·0058	·0058
40	·0011	·0011	·0012	·0013	·0013	·0019	·0021	·0022	·0023	·0023
50	·0005	·0005	·0006	·0006	·0006	·0009	·0010	·0011	·0011	·0011
<5½	·000	·000	·000	·000	·000	·004	·008	·012	·015	·016
5½ to 7	·001	·002	·002	·003	·003	·012	·024	·033	·040	·040
7 to 8½	·003	·004	·005	·007	·008	·025	·041	·053	·057	·060
8½ to 11	·010	·013	·016	·020	·021	·055	·079	·100	·114	·115
11 to 14	·017	·021	·025	·028	·029	·066	·091	·107	·117	·118
14 to ∞	·101	·111	·119	·128	·132	·222	·266	·295	·309	·312
Total σT^4	·132	·151	·167	·186	·193	·384	·509	·602	·652	·661

Computations have been made in this way for the outgoing radiation in latitudes 0°, 15°, 30°, 50° and 70°, with the following results:

TABLE II.—OUTGOING RADIATION FROM CLEAR SKIES.

R in cal./cm.^{2}/min.

Latitude.	0°	15°	30°	50°	70°
Stratosphere temperature	200°a.	207°a.	212°a.	218°a.	220°a.
Surface temperature	299°a.	298°a.	292°a.	280°a.	261°a.
Wave-lengths	R	R	R	R	R
5½ μ to 7 μ	·001	·002	·002	·003	·003
7 μ to 8½ μ	·032	·031	·029	·024	·016
8½ μ to 11 μ	·115	·114	·100	·079	·055
11 μ to 14 μ	·067	·069	·067	·059	·047
>14 μ	·101	·111	·119	·128	·132
Total	·316	·327	·317	·293	·253

These values of the outgoing radiation have been plotted against the sine of the latitude as Curve I, Fig. 4, and it will be seen that they indicate a maximum near latitude 15°, the values falling off slightly towards the equator, and more appreciably towards the poles.

So far we have only considered clear skies through which the radiation from the surface in the band $8\frac{1}{2}\ \mu$ to $11\ \mu$ can pass without obstruction. It is, however, not difficult to find the effect of the clouds. A layer of clouds of any appreciable thickness cuts off all radiation from the atmosphere and ground below it; but radiates itself almost like a black

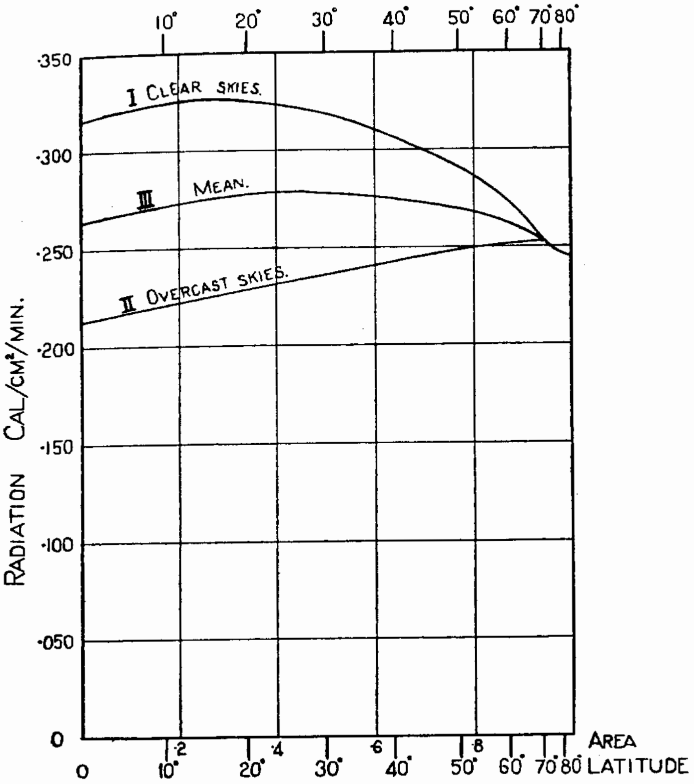

FIG. 4.—Outgoing radiation according to latitude.

body at the temperature of its upper surface. In my previous paper I discussed this question, and came to the conclusion that, on the average, the effective temperature of the clouds may be taken as about 260°a. There will be little variation in the effective temperature of the clouds in different latitudes, for the thick clouds which we are considering always form in layers of the atmosphere of approximately the same temperature in all latitudes. To calculate the outgoing radiation from overcast skies we therefore carry through the same calculation as that given above for cloudless skies, except that we use the temperature of the cloud layer instead of that of the ground. It will be noticed that the temperature chosen for the clouds happens to be practically that of the ground in latitude 70°, and to save work I have taken the temperature of the clouds to be 261°a instead of 260°a, as values for the latter temperature are already contained in Table I. The following table contains data for overcast skies similar to that contained in Table II. for clear skies:

TABLE III.—OUTGOING RADIATION FROM OVERCAST SKIES.

R in cal./cm.^{2}/min.

Latitude.	0°	15°	30°	50°	70°
Stratosphere temperature	200°a.	207°a.	212°a.	218°a.	220°a.
Cloud layer temperature	261°a.	261°a.	261°a.	261°a.	261°a.
Wave-length	R	R	R	R	R
$5\frac{1}{2}\ \mu$ to $7\ \mu$	·001	·002	·002	·003	·003
$7\ \mu$ to $8\frac{1}{2}\ \mu$	·014	·014	·015	·016	·016
$8\frac{1}{2}\mu$ to $11\ \mu$	·055	·055	·055	·055	·055
$11\ \mu$ to $14\ \mu$	·042	·043	·045	·047	·047
$>14\ \mu$	·101	·111	·119	·128	·132
Total	·213	·225	·236	·249	·253

These values have been plotted as Curve II on Fig. 4. It will be seen that the radiation from overcast skies is least at the equator and most at the poles; this is because the radiation from the clouds is the same in all latitudes, while the radiation from the stratosphere increases as the temperature of the stratosphere increases from the equator to the poles.

The actual radiation in any latitude will be between the curve for clear skies and overcast skies according to the average amount of cloud in that latitude. In this preliminary study it is not desirable to complicate the problem by allowing for the variation of cloud amount according to latitude; but a mean cloud amount of 5, which is near to the mean for the earth as a whole, will be assumed for each latitude.

The mean radiation for each latitude can therefore be obtained by taking the mean value for the clear and overcast skies. This has been done and the result plotted as Curve III of Fig. 4. By reading off the values of the three curves I, II and III at each 10° of latitude, the following table has been compiled:

TABLE IV.—RADIATION FROM THE EARTH AND ATMOSPHERE TO SPACE.

cal./cm.^{2}/min.

Latitude.	Clear Skies.	Overcast Skies.	Mean.
0°	·316	·213	·264
10°	·324	·220	·274
20°	·325	·228	·277
30°	·319	·236	·277
40°	·307	·243	·275
50°	·291	·249	·270
60°	·274	·252	·265
70°	·253	·253	·253
80°	...	...	(·246)
90°	...	...	(·245)

The values of the radiation were plotted against the sine of the latitude in Fig. 4 in order to facilitate the computation of the total radiation from the earth as a whole, for the total energy from a zone of latitude is proportional to the average energy and the sine of the latitude. The total energy from a hemisphere is therefore proportional to the area under each curve in Fig. 4, and the average outgoing radiation per sq. cm. is equal to the average ordinate of each curve. In this way we find that the average outgoing terrestrial radiation from the earth as a whole is ·307 cal./cm.^{2}/min. for clear skies, ·235 cal./cm.^{2}/min. for overcast skies, and ·271 cal./cm.^{2}/min. for an average cloud amount of 5.

So far we have not taken the incoming radiation into account in any way, we have obtained our values for the outgoing radiation from consideration only of the observed temperatures, and Hettner's values for the absorption coefficients of water vapour; we have not even had to take into account the distribution of water vapour except to assume that the air is saturated at the base of the stratosphere. A test of the work is therefore now available, for if it is along the right lines the average value found for the outgoing radiation, ·271 cal./cm.^{2}/min., should be, approximately, equal to the average incoming solar radiation.

The intensity of solar radiation at the distance of the earth from the sun is 1·953 cal./cm.^{2}/min. This is the energy received on a square centimetre exposed at right angles to the solar ray. The total energy intercepted by the earth is, therefore, $\pi R^2 \times 1{\cdot}953$. This amount, however, has to be spread over the whole surface of the earth, therefore,

$$\left.\begin{array}{r}\text{average solar radiation}\\ \text{received by the earth}\end{array}\right\} = \frac{\pi R^2 \times 1{\cdot}953}{4\pi R^2} = {\cdot}488 \text{ cal./cm.}^2\text{/min.}$$

Of this incoming radiation a large proportion is reflected back by the clouds and dust in the atmosphere without taking part in the heat exchanges of the atmosphere. Aldrich has determined this proportion (the albedo) to be 0·43; therefore the average effective incoming radiation is $\cdot 488 \times \cdot 57 = \cdot 278$ cal./cm.^{2}/min. Thus the value for the average outgoing radiation which we have found, ·271 cal./cm.^{2}/min. is within three per cent of the average effective incoming radiation.

This very close agreement is really not so significant as it would seem, for we have left several factors out of consideration. We have neglected any scattering of the radiation from the ground as it traverses the whole thickness of the atmosphere; we have neglected the uneven distribution of the clouds, and we have assumed an arbitrary value for the temperature of the radiating clouds. But any allowances made for these factors would be small and not all in the same direction, and no change in any of these factors could materially alter the result. In any case, the agreement between the value for the outgoing radiation which we have found, and the accepted value for the incoming solar radiation is gratifying, and encourages us to proceed further with the investigation.

Nocturnal Radiation.

In the absence of sunlight all radiation from bodies at atmospheric temperatures is of the long wave type which we have been studying. A black horizontal plate exposed to the sky radiates towards the sky the full energy of black-body radiation. On the other hand it receives radiation from the atmosphere. The amount radiated and the amount received are seldom equal; generally the black plate radiates more than it receives, but in certain conditions the reverse may be the case.[9]

The difference between the two is called the nocturnal radiation because it is generally observed after the sun has set, but it is quite possible to make observations of the long wave radiation from the sky during the day time. The easiest way to do this is to measure the total radiation from the sky; then insert a screen of plane glass which cuts out the long wave radiation without appreciably diminishing the short wave solar radiation, and make another measurement. The difference between the two observations gives the amount of the long wave radiation received from the sky.

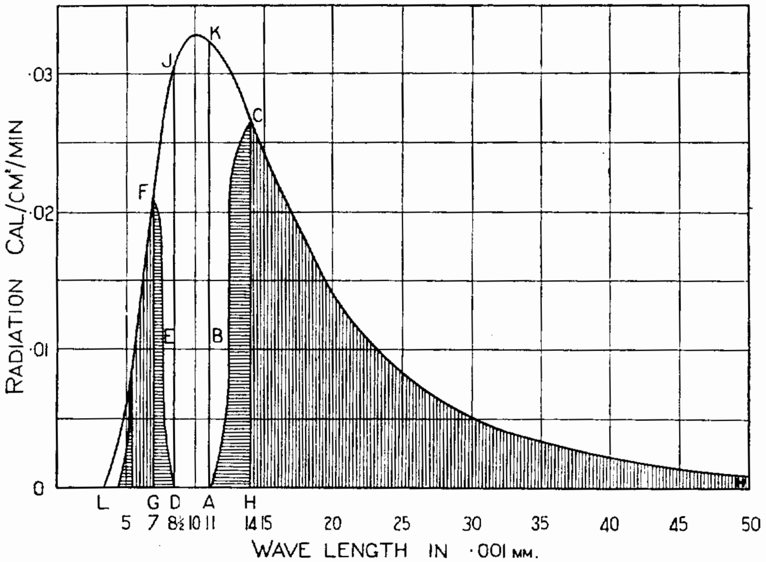

FIG. 5.—Nocturnal radiation.

The nocturnal radiation then is the difference between the radiation sent out per square centimetre by a horizontal plate at the temperature of the surrounding air and the radiation received per square centimetre from the sky. Now we know the amount of radiation sent out by a black body at any temperature, and, therefore, to find the nocturnal radiation we have only to determine the amount of radiation received from the sky.

Let us consider an evening with the air temperature T; then the energy emitted by the plate is represented by the Curve LFJKCM shown in Fig. 5.

To find the energy received from the sky we proceed as follows:

We have already seen that a layer of air containing ·3 mm. of precipitable water in vapour form absorbs all radiation of wave-length between 5·5 μ and 7 μ, and all radiation of longer wave-length than 14 μ. For these wave-lengths the layer acts as a black body, and, if it is at a uniform temperature, it emits the same radiation as a black body at that temperature. Now the humidity of the lower atmosphere is such that a very thin

[9] Ångström, *Stockholm Medd. Statens Met.-Hydro. Anst.*, Band 3, No. 12, p. 10, 1927.

layer of air contains the required amount of water vapour. It is easy to show that if the vapour pressure is p millibars, then the length of a column of air containing ·3 mm. of precipitable water is given by the following expression:

$$L = 21/p \text{ metres.}$$

From this we see that with a vapour pressure as low as 1 millibar (17 per cent relative humidity at 0°C.), the thickness of the layer which absorbs all the radiation in the specified wave-lengths is only 21 metres. Now a layer as thin as this may be considered to be at a uniform temperature, and, therefore, the atmosphere radiates like a black body at the air temperature for the specified wave-lengths. Thus, in Fig. 4, the plate receives from the atmosphere the energy represented by the area which has been indicated by vertical hatching.

For wave-lengths between $8\frac{1}{2}$ μ and 11 μ the air is completely transparent, therefore it cannot radiate in these wave-lengths, and the plate receives no radiation from the atmosphere, assumed free from clouds, within this region.

Turning now to the wave-lengths between 11 μ and 14 μ where the absorption coefficient of water vapour varies rapidly with the wave-length. The radiation from the sky within this region will vary not only with the amount of water vapour near the ground, but also throughout the atmosphere. On the whole, however, the water vapour near the ground will be the predominating factor for three reasons: first, because vapour pressure decreases rapidly with height in normal conditions, therefore changes in the vapour pressure of the upper atmosphere are less important than at the ground; secondly, the temperature decreases with height, and, therefore, a given amount of water vapour emits more radiation near the ground than higher in the atmosphere; and thirdly, the water vapour near the ground cuts out a great proportion of the radiation coming from above.

It is impossible to take all these factors into account and calculate the radiation of these wave-lengths received from the atmosphere; we can, however, set limits to the possible amount of the radiation.

The radiation within the wave-length we are considering must be less than that of a full radiator at the air temperature, but it may be anything between this amount and zero according to the amount of water vapour in the atmosphere. In Fig. 5 the Curve ABC has been drawn to indicate a possible distribution of the energy received by the plate in the region 11 μ to 14 μ, and a similar Curve DEF for the region 7 μ to $8\frac{1}{2}$ μ (the small area for shorter wave-lengths than 5·5 μ is neglected). The area between these lines and that for which the radiation is full has been indicated by horizontal hatching. With the humidity conditions represented in Fig. 5 the black plate will receive from the atmosphere an amount of energy represented by the shaded area, if the humidity increases the shaded area will increase, and if the humidity diminishes the shaded area will decrease. On the other hand, the black plate radiates the energy represented by the whole area under the thick curve, and, therefore, the difference between what the plate radiates and what it receives—the nocturnal radiation—is represented by the unshaded area in the figure. This area must obviously be greater than the area DJKA, and less than the area GFJKCH, and as these areas simply represent the radiation of a black body within the specified wave-lengths, we may compute a table giving the limits within which the nocturnal radiation must lie for given temperatures at the observing station. This result, however, only strictly applies when the vapour pressure at the observing station is not appreciably less than about 1 millibar.

TABLE V.—LIMITS OF NOCTURNAL RADIATION.
cal./cm.²/min.

Temperature.	Nocturnal radiation.		
	Maximum.	Minimum.	Mean.
°a.			
261	·146	·055	·100
271	·176	·066	·121
280	·211	·079	·145
292	·262	·100	·181
299	·293	·115	·204

These values have been plotted as the thick curves on Fig. 6. The interpretation of this figure is that for each of the temperatures on the abscissa the nocturnal radiation will fall somewhere between the two outer curves, and the higher the humidity the nearer the radiation will fall to the lower curve.

As in the case of the outgoing radiation, the investigation has been built up solely from considerations of the absorption coefficients of water vapour, and the test of the investigation is how the observed values of the nocturnal radiation fit into the limits given by the theory. To make this test we have the excellent set of observations of nocturnal radiation made by W. H. Dines at Benson, near Oxford, and published by the Society.[10] Mr. Dines measured the radiation from the clear sky during five years, 1922 to 1926, and the results are given for each month in Table I. of the paper referred to. The Table contains also the total radiation from a black body at the temperature of the air at the time of observation of the radiation from the sky. The values converted to the units used in this paper are reproduced in Table VI.

TABLE VI.—MEAN LONG WAVE RADIATION FROM CLOUDLESS SKIES AT BENSON IN CAL./CM.²/MIN.

	Jan.	Feb.	Mar.	April.	May.	June.
Black body radiation	·479	·501	·505	·541	·565	·588
Sky radiation	·339	·350	·350	·382	·420	·442
Difference	·140	·151	·155	·159	·145	·146
Temperature (°a.)	276	279	280	285	288	291

	July.	Aug.	Sept.	Oct.	Nov.	Dec.
Black body radiation	·601	·576	·567	·538	·490	·483
Sky radiation	·466	·438	·422	·390	·349	·343
Difference	·135	·138	·145	·148	·141	·140
Temperature (°a.)	292	289	288	283	278	277

[10] W. H. Dines and L. H. G. Dines, *Memoirs of the R. Meteor. Soc.*, Vol. 2, No. 11, 1927.

In Table VI. the line marked "Difference" gives the average nocturnal radiation for clear skies at Benson, and the line marked "Temperature" gives the mean air temperature at the time of observation. These values have been entered on Fig. 6. It will be seen at once that they all fall within the two extremes. At first sight it would appear that there is little relationship between the nocturnal radiation and temperature, the nocturnal radiation for July, the warmest month, being less than for January, the coldest month. This is because humidity is also a factor, the absolute humidity in July being considerably greater than in January. It will be noticed that the points for each month from March to July

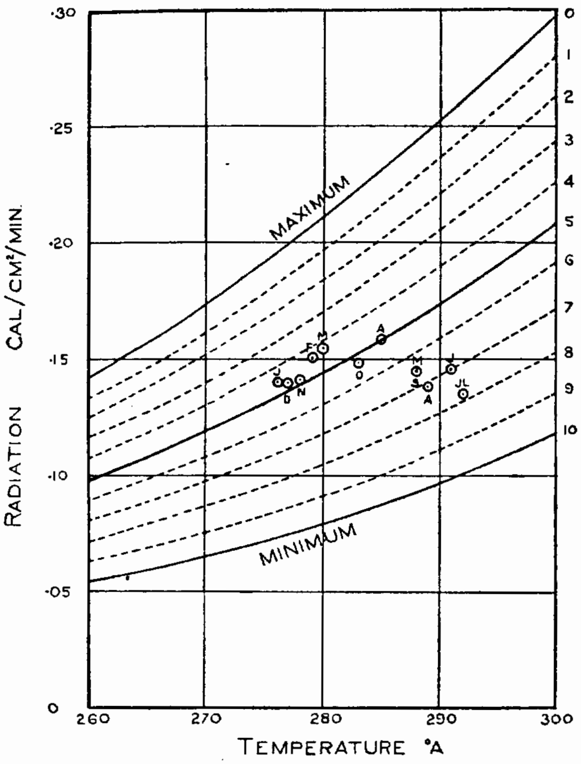

FIG. 6.—Limits of nocturnal radiation and monthly values for Benson.

move progressively towards the lower limit indicated by the lower thick line, while from July to December the movement is in the opposite direction. This is exactly the effect we should expect humidity to produce; for the absolute humidity in England increases from March to July, and decreases from July to December. Unfortunately, Mr. Dines did not give the humidity at the times of his observations, otherwise we might be able to test this point from the observations. Humidity is a very variable factor, and varies rapidly from day to day, and even from hour to hour, so there is no certainty that the mean humidity at the time of the observations would be the same as the mean humidity of the month. Still, it appeared worth while to investigate this point further.

Observations of humidity not being available for Benson, the mean humidity for Kew for each month during the five years over which the radiation observations extended was taken for comparison. The intervals between the maximum and minimum values of the radiation for each temperature on Fig. 6 were then divided into ten equal parts, and the

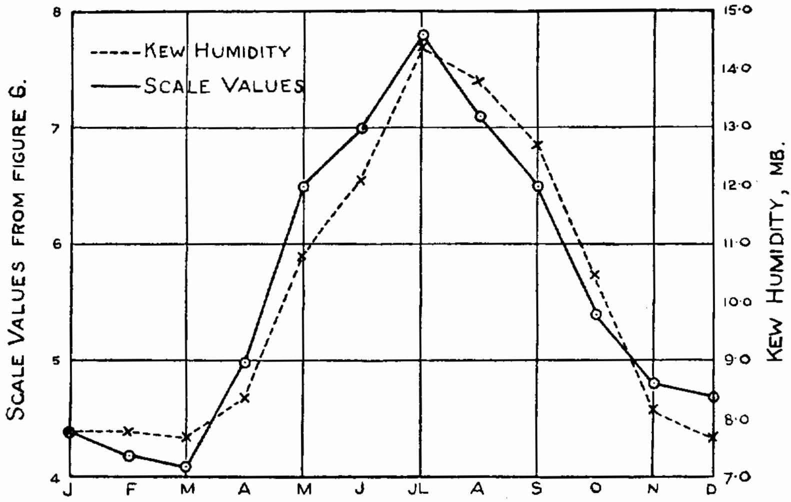

FIG. 7.—Nocturnal radiation and humidity.

points joined by dotted lines. The position of each monthly point on Fig. 6 was then read off against this arbitrary scale, numbered 0 to 10, and compared with the corresponding value of the absolute humidity at Kew; the result being given in Table VII., and plotted in Fig. 7.

TABLE VII.—NOCTURNAL RADIATION AND HUMIDITY.

	Jan.	Feb.	Mar.	April.	May.	June.
Position of point on radiation scale of Fig. 6	4·4	4·2	4·1	5·0	6·5	7·0
Mean vapour pressure at Kew (mb.)	7·8	7·8	7·7	8·4	10·8	12·1

	July.	Aug.	Sept.	Oct.	Nov.	Dec.
Position of point on radiation scale of Fig. 6	7·8	7·1	6·5	5·4	4·8	4·7
Mean vapour pressure at Kew (mb.)	14·4	13·8	12·7	10·5	8·2	7·7

The agreement between the two curves in Fig. 7 is surprising, and is excellent evidence that the method here employed does correctly interpret the physics of nocturnal radiation. It will be noticed that in Fig. 7 the curve for the radiation is higher than the curve for the humidity in the first half of the year and lower during the second half. This is no doubt due to the fact that the humidity is measured only near the ground while the radiation is affected by the humidity of the upper atmosphere which probably has a somewhat different annual variation. It is interesting to note that the radiation curve is more symmetrical about its maximum than is the humidity curve, but it would probably be unwise to lay much stress on such points in this preliminary investigation.

We have, so far, confined our study of nocturnal radiation to occasions of cloudless skies. The effect of the presence of clouds is, however, easily ascertained. A cloud acts as a black body at the temperature of its lower surface. Therefore the black plate will receive radiation in the band 7 μ to 14 μ which is absent during clear skies. The amount of this radiation will depend on the temperature of the cloud. Thus the effect of the clouds will be to reduce the nocturnal radiation; and if the clouds are low and heavy their effective temperature will be only a degree or two lower than that of the air at ground level, and so the nocturnal radiation will practically disappear. Occasionally the temperature of the clouds is higher than the temperature of the air near the ground; in this case the black plate will receive from the sky more radiation than it emits, and the nocturnal radiation will be negative.[11]

The success of the method developed in this paper, as judged by the values obtained for the average outgoing radiation from the earth, and for the nocturnal radiation and its variations with humidity, justifies us in using it to throw light on several old problems of meteorology which have remained unsolved because of our want of knowledge of how to deal with the radiation of the atmosphere.

The Transport of Heat in the Atmosphere.

A problem of considerable meteorological importance is the transfer of heat from equatorial to polar regions. In line 1 of Table VIII. are entered the values for each 10° of latitude of the average solar radiation received per square centimetre of the earth's surface, taking the solar constant to be 1·953 cal./cm.^{2}/min. and Aldrich's value of the albedo, 0·43. This gives ·278 as the average value for the effective incoming solar radiation.

In Table IV. (p. 10) values of the outgoing terrestrial radiation, according to latitude, are given in the column headed "Mean." These values were obtained as the result of our investigation, and, as stated above, they give a mean outgoing radiation of ·271. This is slightly less than the mean incoming effective solar radiation, while for balance it should be the same. In order to effect this balance the values for the terrestrial radiation in Table IV. have been increased by the factor $\frac{278}{271}$. The values so adjusted are entered in line 2 of Table VIII.

[11] Since writing this paper I have received from Dr. Anders Ångström an advance proof of a paper which is shortly to be published. In this paper Dr. Ångström states that, as the result of his observations of nocturnal radiation, he concludes that terrestrial radiation must be divided into three groups according to wave-length: (*a*) a group which suffers no appreciable absorption by the atmosphere, (*b*) a group which is absorbed according to the amount of water vapour, and (*c*) a group for which the atmosphere is remarkably opaque. He gives the relative proportions of these three groups as 25 per cent, 30 per cent and 50 per cent, respectively, at temperatures near 273°a. The corresponding proportions of the similar groups discussed in this paper are 16 per cent, 25 per cent and 59 per cent.

TABLE VIII.

Latitude.	Equator.	10°	20°	30°	40°	50°	60°	70°	80°
Effective incoming solar radiation cal./cm.²/min.	·339	·334	·320	·279	·267	·232	·193	·160	·144
Outgoing terrestrial radiation cal./cm.²/min.	·271	·282	·284	·284	·282	·277	·272	·260	·252
Difference between incoming and outgoing radiation cal./cm.²/min.	+·068	+·052	+·036	+·013	−·015	−·045	−·079	−·100	−·108
Total horizontal flow of heat across circles of latitude per min. cal./$2\pi R^2$	·0000	·0106	·0183	·0221	·0219	·0183	·0122	·0066	·0020
Horizontal flow of heat per cm. of circle of latitude per min. cal./10^7	·00	0·69	1·24	1·63	1·83	1·82	1·56	1·23	0·73

The values of the solar and terrestrial radiation are plotted as Curves I and II, respectively, in Fig. 8. It will be seen that while there is a large decrease of the effective solar radiation with latitude the terrestrial

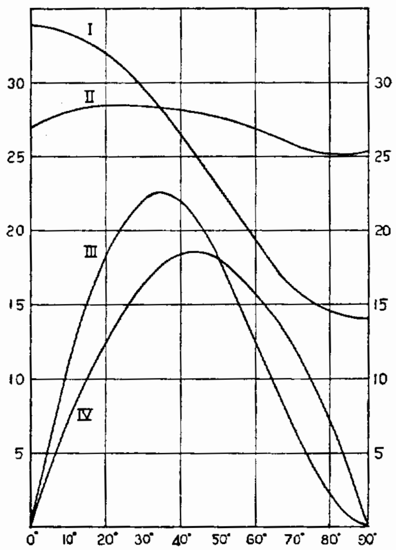

FIG. 8.—Curve I. Effective solar radiation.
Curve II. Outgoing terrestrial radiation.
Curve III. Total horizontal heat flow across circles of latitude.
Curve IV. Horizontal heat flow per cm. of circle of latitude.
(For unit of ordinates see Table VIII.)

radiation is nearly uniform, the outgoing radiation from polar regions being only slightly less than at the equator, while the maximum is in about latitude 25°. The curves show that the solar and terrestrial radiation are equal in latitude 35°; at lower latitudes the incoming radiation exceeds the outgoing, while in higher latitudes the reverse is the case. The consequence is that there must be a flow of heat from lower to higher latitudes, and this flow is brought about by the general circulation of the atmosphere and oceans. By plotting the values of the incoming and outgoing radiation against the sine of the latitude, it is possible to calculate the excess or defect of heat received in each zone, and as the excess must be transferred to regions of defect, the flow of heat across each circle of latitude necessary to effect this can easily be evaluated. The result of such a computation is shown in line 4 of Table VIII., and plotted as Curve III of Fig. 8. It will be seen that starting from zero at the equator the stream of heat across each circle of latitude increases to latitude 35°, after which the stream decreases until at the pole it reaches zero again. These totals are useful for certain purposes, but a more useful quantity is the density of the flow, that is, the flow across each centimetre of the length of the circle of latitude. Values are given in line 5 of Table VIII., and these are plotted on Curve IV of Fig. 8.

The decrease in the length of the circles of latitude as one passes from the equator to the pole results in the density of flow reaching its maximum in about latitude 45°, *i.e.* 10° higher than the latitude of maximum flow.

Effect of Changes in Solar Radiation.

One of the most interesting problems of meteorology, especially for those who have studied the apparent changes of climate during past ages, is to determine what would be the effect on the climate of the world of a change in solar radiation. The method of this paper makes it possible to throw new light on this problem.

An increase in solar radiation would necessitate an increase of the same amount in the outgoing radiation from the earth. This could be effected in three ways: (*a*) a rise in the temperature of the surface would result in more radiation leaving the atmosphere in the band of wave-lengths 7 μ to 14 μ, mainly through the band $8\frac{1}{2}$ μ to 11 μ; (*b*) a rise in the temperature of the stratosphere would result in more radiation being emitted from the stratosphere chiefly of longer wave-length than 14 μ; and (*c*) an increase in cloud amount would return the excess radiation by reflection. We will study each of these possibilities in turn; finding the effect of changing one, while the other two remain unchanged.

Taking first the effect of changing the surface temperature, it is easily found by the method described in the first part of this paper that a change of 1°C. in the temperature of the surface increases the average outgoing radiation by ·0015 cal./cm.^{2}/min. This is a small amount compared with the ·007 cal./cm.^{2}/min. which a change of 1°C. produces in the radiation of a black body at the same average temperature.

A change of 1 per cent in the solar radiation would involve an increase of 1 per cent in the outgoing terrestrial radiation assuming no change in the albedo. That is, a change of 1 per cent in the solar radiation would produce an average change in the outgoing radiation of ·0028 cal./cm.^{2}/min. To supply this the temperature of the surface would have to change by ·0028/·0015 = 2°C., approximately.

Turning now to the effect of changing the temperature of the stratosphere, a similar calculation shows that, on the average, a change of 1 per cent in the solar radiation would involve an average change of 1·4°C. in the temperature of the stratosphere, assuming no change in surface temperature or in the amount of cloud. Thus, changing the temperature of the stratosphere produces a slightly larger effect on the outgoing radiation than a similar change of the surface temperature.

If a change of solar radiation did take place, and to fix our attention we will consider an increase of 1 per cent, the first effect would obviously be a change in the temperature of the surface. But whether the temperature of the stratosphere would rise in consequence it is impossible to say with our present knowledge. At first sight one would expect the mean temperature of the atmosphere in all its parts to rise; but there are reasons which indicate that this consequence would not necessarily follow. The temperature of the stratosphere is lowest where the temperature of the surface is highest, and vice versa. This indicates that the tendency might be for the temperature of the stratosphere to decrease when the temperature of the surface increased. If this happened, the average temperature of the surface would have to rise further than the 2°C. which would suffice if the temperature of the stratosphere remained unchanged. We can, at least, say that the initial effect of a change of 1 per cent in solar radiation would be to raise the temperature of the surface of the earth; and the tendency would be for this change to reach, at least, 2°C. on the average, and possibly somewhat more.

Now the increase of radiation would not be distributed uniformly over the earth's surface, but the equatorial regions would receive per unit area a larger increase than polar regions. This means that, other things remaining the same, equilibrium would not be reached until the surface temperature in equatorial regions had risen considerably more than 2°C. Now there is very good reason to believe that variations in solar radiation of the order of magnitude of 1 per cent have occurred in recent years without anything like a change of 2°C. in the mean temperature of equatorial regions. It is obvious, therefore, that there is another factor to be taken into account, and it is equally clear that this is the change in cloud amount which would accompany any rise in the surface temperature. That the cloud amount would change follows at once from the fact that the increase in solar radiation being more effective in equatorial than in polar regions would result in an increase in the temperature gradient between the equator and the poles. This would lead to greater activity of the general circulation of the atmosphere, with a consequent increase of evaporation and condensation which would involve an increase in the amount of cloud.

The effect of an increase in cloud amount on the heat balance of the atmosphere is easily shown by the method employed in this paper.

We derived above (page 11) values for the outgoing radiation from the earth's atmosphere under existing temperature conditions with clear and with overcast skies, and found average values of ·307 and ·235 cal./cm.^{2}/min., respectively. These values, however, are to be adjusted to give an average outgoing radiation of ·278 cal./cm.^{2}/min. instead of ·271. When the adjustment has been made we find outgoing radiation from clear skies to be ·314 cal./cm.^{2}/min., and outgoing radiation from overcast skies ·242 cal./cm.^{2}/min.

In Fig. 9 the abscissæ represent the proportion of the whole sky covered by cloud, so that a half-clouded sky is represented by ·5. The ordinates represent radiation in cal./cm.^{2}/min. On the ordinate for 0 cloud the point A has been marked at ·314, which is the average outgoing radiation for clear skies, similarly on the ordinate for 1·0 cloud the point B has been marked at ·242, the corresponding value for overcast skies. As a first approximation we may assume that, between these limits, the outgoing radiation will be a linear function of the cloud amount, so that the line AB gives the average outgoing radiation for the various cloud amounts from 0 to 1. On page 11 above, we found that the average intensity of the incoming solar radiation when distributed over the whole surface of the earth is ·488 cal./cm.²/min. This is the average amount of radiation which the earth would receive if the sky were clear of clouds, and this amount would have to be returned as terrestrial radiation (this neglects the reflection from the surface and scattering in the atmosphere which, however, are relatively only small amounts). Aldrich[12] has measured the reflecting power of clouds, and concludes that a cloud returns 78 per cent of the incident radiation, and that the amount of

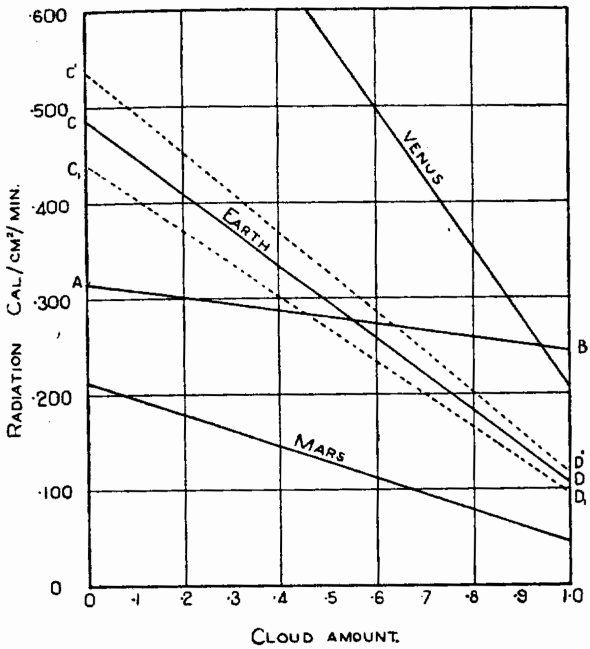

FIG. 9.—Radiation and amount of cloud.

reflection is practically independent of the angle of incidence. Thus, with a totally overcast sky, 78 per cent of the solar radiation would be reflected, and only 22 per cent would enter the atmosphere to be returned as terrestrial radiation. In Fig. 9 the point C represents the incoming solar radiation with a clear sky, ·488, and the point D the incoming solar radiation with a completely overcast sky, ·488 × ·22 = ·108. The line CD represents the amount of effective solar radiation for different cloud amounts. If, therefore, the temperature of the earth's surface and the stratosphere remain unchanged, and the outgoing radiation is regulated to the incoming solar radiation by change in cloud amount, the balance would be effected at the cloud amount where the two lines AB and CD cross. In the figure this is seen to be ·55, which is in remarkable agreement with the generally accepted value for the mean cloud amount of the

[12] Aldrich, *Smithsonian Misc. Collections*, Vol. **69**, No. 10, 1919.

earth 5·4 tenths (Aldrich and Brooks).[13] It is interesting to note that, if we had known nothing about the mean amount of cloud this investigation would have determined it; for the lines AB and CD were obtained from a knowledge of the solar constant, the reflectivity of clouds, the temperature of the atmosphere, and the absorption of water vapour.

We are now in a position to see what change in cloud amount would be necessary to balance a change in solar energy if the temperature of the surface and stratosphere remained unchanged. In Fig. 9 C represents the present incoming solar radiation in the absence of clouds. C^1 and C_1 are corresponding points when the solar radiation is increased and decreased by 10 per cent. Similarly, D^1 and D_1 represent the effective solar radiation with overcast skies. Where the lines C^1D^1 and C_1D_1 cut AB, we have the cloud amounts which would balance an increase and decrease of solar radiation by 10 per cent, respectively. These are ·46 and ·64, respectively, giving a mean change of 2 tenths, approximately, for a 20 per cent change of solar radiation. Thus a change in solar radiation of 1 per cent would be balanced by an extension of the cloud over one-hundredth part of the sky.

Collecting our results we find that if only one of the three factors, surface temperature, stratosphere temperature and cloud amount, varied while the other two factors remained constant, the following changes would be necessary to balance a change of one per cent in the solar radiation:

Surface temperature, 2°C.
Stratosphere temperature, 1·5°C.
Cloud amount, ·01 of the area of the sky.

Which of these factors is the most likely to predominate in regulating the heat exchanges is seen by considering a fairly large change in solar radiation—say 10 per cent. This would necessitate a change of surface temperature of 20°C., or of stratosphere temperature of 15°C., if either acted alone. Let us assume that each provides half the extra outgoing radiation, then the surface temperature would rise by 10°C., and the stratosphere temperature by 7½°. Such a large change in the temperature of the atmosphere accompanied by increased circulation could not occur without increasing largely the amount of cloud. Actual figures cannot be given, but it seems reasonable to conclude that long before the temperature could change by the amounts specified, the cloud amount would increase by the tenth necessary to balance the incoming radiation alone. Everything points to changes in cloud amount being by far the most important factor in establishing thermal equilibrium between incoming and outgoing radiation.

Some support is given to this conclusion by a comparison of the existing conditions on the three planets, Mars, the Earth and Venus. Mars and Venus are both known to have water vapour in their atmospheres. The distance of Mars and Venus from the Sun are 1·52 and ·723 times the Earth's distance, respectively. Thus the intensity of solar radiation at Mars is, approximately, a half that at the Earth; while at Venus it is twice as much as at the Earth.

In Fig. 9 we drew the lines C^1D^1 and C_1D_1 to represent the effect of increasing and decreasing the solar radiation by ten per cent, respectively. If it were legitimate, which is doubtful, to draw similar lines for the intensity of the solar radiation at Mars and Venus, we should get the lines

[13] C. E. P. Brooks, *Memoirs, R. Meteor. Soc.*, Vol. 1, No. 10, 1927, p. 135.

marked with the names of the planets on Fig. 9. The line for Mars does not cut AB at all, therefore there should be no clouds on Mars; while the line for Venus cuts AB between ·9 and 1·0, therefore Venus should have nearly completely overcast skies. Now observations show that there are no clouds on Mars, except possibly very occasional light mists, while Venus is completely covered by clouds.

We thus arrive at the result that the cloud amounts on the three planets Mars, the Earth and Venus are in the same relative order as the intensity of the solar radiation at their positions, which supports the suggestion that cloud amount plays a predominating part in adjusting the balance of incoming and outgoing radiation in an atmosphere containing water vapour.

It is interesting to consider in further detail what would happen to the Earth's atmosphere if solar radiation increased at all appreciably. As already stated, the first effect of an increase of solar radiation would be an increase in the temperature difference between the equator and the poles; but it would also be accompanied by an increase in the temperature difference between the sea and the land. In consequence there would be an increase in the activity of the general circulation of the atmosphere; in particular there would be stronger monsoons, and the cyclonic depression of middle latitudes would be intensified. The increased circulation and the increased temperature would lead to increased cloud amount and increased precipitation.

The increase in cloud amount would naturally occur chiefly in the three main cloud belts, one over equatorial regions and two in high latitudes. In between these belts the skies would probably remain relatively clear. The temperature condition would probably be something as follows: In the relatively clear belts the temperature would rise due to the increased solar radiation. Under the cloud belts the day and summer temperatures would be decreased; but the night and winter temperatures might be higher than at present. This would be particularly the case in high latitudes. Under the cloud belts precipitation would be increased appreciably.

The effect of increasing the solar radiation would therefore be a definite decrease of summer temperatures in high latitudes with an increase of precipitation. Now these are the conditions which are necessary for an ice age. It has constantly been pointed out that a decrease of temperature alone would not produce an ice age; for it would reduce the precipitation to such an extent that large accumulations of ice could not be built up.

Antevs in his recent book, *The Last Glaciation*, sums up his discussion of climates during glaciation with the following paragraph (p. 20):

> Thus, since in some regions with abundant precipitation, drop in summer temperature seems to have been the sole factor of glaciation, and since even in the glaciated areas whose centres have deficient precipitation, primary fall of temperatures is a necessary condition for the large expansion of the ice sheet, low summer temperature may be regarded as the most prominent factor in glaciation, with heavy snowfall playing a very important rôle. The growth of the ice sheets in low latitudes evidently demanded considerable nourishment.

Thus, provided the temperature of the polar regions remained below the freezing point for a considerable part of the year, the conditions which we have seen to be the natural consequence of an increase in solar radiation are just those which are required to produce an ice age in high latitudes.

There is also abundant evidence that the ice ages were accompanied by abnormal precipitation in lower latitudes. In fact, the period which is referred to as the ice age in high latitudes is called the pluvial period in other parts of the world.

Whether the solar energy can undergo the relatively large changes required to produce the increased cloud for an ice age—probably less than a twenty per cent increase would be sufficient—it is impossible to say. All that I wish to stress here is that such an increase, if it did occur, would probably produce the conditions we associate with an ice age.

There is no need to go into a detailed discussion of what would happen if the present amount of solar energy decreased; it is clear that the chief effect would be a decrease in cloud amount and less precipitation, with probably changes of temperature insufficient to be readily appreciated without really large changes in the solar radiation.

Review.

The results obtained in this paper are so different from those obtained in the previous paper, *Some Studies in Terrestrial Radiation*, that it is necessary to review the whole position, to find the causes of the differences and to estimate their significance. The chief difference in the procedure adopted in the two papers is that in the former we assumed that water vapour absorbs like a grey body, that is, the absorption is the same for all wave-lengths greater than $2\ \mu$; while in this paper each wave-length has been assigned its appropriate absorption.

The former method may almost be dignified by the adjective classical. It was used in all the papers which have laid the foundation of our ideas on terrestrial radiation, amongst other by Humphreys,[14] Gold (in part),[15] Emden,[16] Hergesell,[17] and Mügge.[18] While most of the previous writers approached the problem from considerations of radiative equilibrium, Hergesell and I approached it from the point of view of the existing distribution of temperature and water vapour, and calculated the radiation which would result from these conditions without any consideration of how the temperature conditions are maintained.

Two outstanding results followed from my previous paper which are now shown to be wrong, because of the assumption that water vapour absorbs like a grey body. These are (*a*) the outgoing radiation originates in the layers of the atmosphere which have temperatures between 220°a and 286°a, and therefore are well within the troposphere; and (*b*) the stratosphere provides an insignificant amount of terrestrial radiation. The new investigation shows that with clear skies in middle latitudes the radiation is provided by the surface, the atmosphere and the stratosphere in the following proportions: from the stratosphere alone, 38 per cent; from the surface alone, 32 per cent; and from the surface atmosphere and stratosphere in different proportions, 30 per cent. The proportions are slightly different in different latitudes, while with overcast skies, the contribution from the ground is replaced by a smaller contribution from the clouds.

14 Humphreys, *Astrophysical Journal*, Vol. **29**, p. 14, 1909.
15 Gold, *London, Proc. R. Soc.*, Vol. **82** (A), p. 43, 1909.
16 Emden, *München Sitzber. Bayr. Akad. Wiss.*, p. 55, 1913.
17 Hergesell, *Lindenberg Arbeit. Preuss. Aero. Obs.*, Vol. **13**, Wiss. Abh.
18 Mügge, *Zs. Geophysik, Braunschweig*, Vol. **2**, p. 63, 1926.

The new results affect previous work materially. Emden found that the stratosphere sends no radiation downwards, and of course the same result came out of my previous work. The new investigation shows that the stratosphere sends on the average a downward flux of long-wave radiation of more than ·120 cal./cm.^{2}/min., which is more than 43 per cent of the effective solar radiation. This agrees with the observations made by Ångström on mountain peaks and in balloons, which revealed a downward radiation of between ·13 and ·16 cal./cm.^{2}/min. at heights between 4000 and 5000 metres, where, according to Emden,[19] there should have been less than ·05 cal./cm.^{2}/min.

Any radiation which the stratosphere sends downwards has to be sent out again, therefore this amount has to be added to the flux of terrestrial radiation which would be necessary to return the effective solar radiation alone. In other words, the effective solar radiation being ·278, the outgoing radiation crossing the base of the stratosphere must be $> \cdot 278 + \cdot 120$, *i.e.* $> \cdot 398$ cal./cm^2./min. Emden and Humphreys have derived expressions which purport to give the temperature of the stratosphere from considerations of the radiation crossing the base of the stratosphere. Both have neglected the long-wave radiation from the stratosphere which crosses its lower layers just as the terrestrial radiation crosses them, but in the opposite direction. Both have assumed grey or nearly grey radiation, so neglecting the fact that a large proportion of the terrestrial radiation is not absorbed at all by the stratosphere. In so far as both Emden and Humphreys simply equate the amount of energy absorbed to the amount emitted their methods are correct in principle; but as the numerical values they employ far from represent the facts, their results are of little value. It is little more than a coincidence that the expressions used by these investigators give even an approximate value for the temperature of the stratosphere.

The lesson to be learnt from this work is that totally misleading results follow from the assumption that water vapour absorbs like a grey body, and that even qualitative results cannot be obtained on that assumption.

Many problems of atmospheric radiation have apparently been solved by the use of this assumption, and in all these cases the problems must be re-examined using the known absorption of water vapour in the various wave-lengths. At present we have no satisfactory answers to any of the following questions:

(*a*) Why does not the temperature in the stratosphere decrease with height?

(*b*) Why does the temperature of the stratosphere increase as we pass from low to high latitudes?

(*c*) Why is the base of the stratosphere higher over equatorial than over polar regions?

The answer to the first problem will probably involve consideration of the high temperatures, at 40 to 50 kilometres above sea-level, which we now associate with the ozone layer; this was not mentioned by Emden, but was referred to by Humphreys in his first paper on this subject.

As to the two latter questions we have as yet no solution in sight; but the controlling factor will probably be found to be in the dynamics of the troposphere rather than in the thermo-dynamics of the stratosphere.

19 Emden, *Loc. cit.*, p. 129.

SUMMARY.

By using the observed temperatures of the earth's surface and of the stratosphere, and observed values for the absorption coefficients for water vapour and carbon dioxide, approximate values have been obtained for the outgoing radiation from the earth and its atmosphere, and the laws governing nocturnal radiation have been indicated. Values have been found for the horizontal transfer of heat across the circles of latitude which is necessary to obtain radiative equilibrium of the atmosphere as a whole. The consequences of changes in the intensity of solar radiation have been investigated, and the conclusions drawn that change in cloud amount would be the chief agency by which radiative equilibrium would be restored. An increase in solar radiation would result in increased cloud and precipitation, while a decrease in solar energy would lead to less cloud and less precipitation. The possibility of increased solar activity leading to an "ice age" is discussed.