On the backscattering of global radiation by the sky

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(Manuscript received 13 February 1965)

ABSTRACT

The increase of the global radiation over a snow covered ground is an effect of the backscattering by the sky of the radiation reflected at the earth's surface. An attempt has been made to derive the coefficient of backscattering from observations both for clear and for overcast sky.

Introduction

It has been first shown by A. ÅNGSTRÖM (1931) that the global radiation (total radiation from sun and sky) increases under otherwise unchanged conditions if the ground is covered with snow. The radiation reflected by the earth's surface is partly reflected downward by the backscattering of the atmosphere and by reflection from clouds and, thus, increases the global radiation. The process is repeated *ad infinitum*. If r is the reflectance (albedo) of the surface, d the backscatterance of the sky the global radiation is

$$G_r = G_0(1 - rd)^{-1}, \qquad (1)$$

where G_0 is the global Radiation over an ideally black earth's surface. The quantity d is difficult to determine theoretically as well as by observations.

LOEWE (1961, 1963) has amended this equation by introducing the absorption of the radiation in the atmosphere. We will, however, use Ångström's original equation (1) here.

Theoretical determinations

The quantity d can be inferred from theoretical investigations of DEIRMENDJIAN & SEKERA (1954) in the case of a cloudless atmosphere of pure air without any contaminations and with scattering according to Rayleigh's law. They considered multiple scattering and determined the global radiation G and sky radiation H over a black surface and over surfaces with the reflectances 0.25 and 0.80. From their tables the intensification factors for G and H given in Table 1 may be derived. The global radiation increases by 4 to 6 % for a snow cover ($r = 0.80$) as compared to a black surface while the increase of the comparably weak sky radiation ranges from 10 to 140 %. Introducing these factors for the global radiation and the corresponding reflectances into equation (1) one obtains a value for the backscatterances of the Rayleigh sky between 6 and 7 %, decreasing somewhat towards low sun elevations. Equation (1) cannot be applied to the sky radiation H because both direct solar radiation and sky radiation are reflected at the earth's surface.

In a turbid atmosphere higher values of the backscattering may be expected. FEIGELSON *et al.* (1960) have presented comprehensive calculations of the radiation fluxes arising from multiple scattering in a turbid atmosphere. This is characterized by its optical thickness

TABLE 1. *Intensification factors of global radiation G and sky radiation H in a Rayleigh atmosphere over earth's surface of various reflectances after* DEIRMENDJIAN & SEKERA (*1954*).
In parentheses backscattering coefficients d %. z = Zenith distance of the sun.

z	0°	53°	84°	88.8°
$G_{0.8}/G_{0.25}$	1.042 (7.3)	1.040 (7.0)	1.029 (5.1)	1.033 (5.8)
$G_{0.25}/G_0$	1.016 (6.3)	1.015 (5.9)	1.016 (6.3)	1.011 (4.4)
$H_{0.80}/H_{0.25}$	1.72	1.50	1.13	1.075
$H_{0.25}/H_0$	1.38	1.24	1.075	1.075

TABLE 2. *Intensification factors of the global radiation $G_{0.6}/G_{0.1}$ and backscattering coefficients d (%, in parentheses) for various zenith distances z of the sun, optical thicknesses τ^* of the atmosphere, and shape of the indicatrices.*

After FEIGELSON *et al.* (1960).

z	30°	45°	60°	90°
$\tau^* = 0.2$, V	1.064 (12.0)	1.065 (12.2)	1.065 (12.2)	1.084 (15.5)
0.2, VI	1.064 (12.0)	1.064 (12.0)	1.065 (12.2)	1.083 (15.3)
0.4, VI	1.116 (20.8)	1.116 (20.8)	1.116 (20.8)	1.124 (22.1)
0.4, VII[a]	1.103 (18.7)	1.107 (19.4)	1.109 (19.7)	1.114 (20.5)
0.6, VII	1.160 (27.6)	1.158 (27.4)	1.158 (27.4)	1.165 (28.3)
0.6, VIII	1.117 (21.0)	1.117 (21.0)	1.116 (20.8)	1.120 (21.4)
0.8, VII	1.166 (28.4)	1.159 (27.4)	1.180 (30.5)	1.167 (28.6)
0.8, VIII	1.144 (25.2)	1.144 (25.2)	1.144 (25.2)	1.143 (25.0)

[a] Table IV of FEIGELSON *et al.* contains a misprint in the last column; the value of the integral was assumed 0.6043 instead of 0.2043.

τ^* and by the shape of the indicatrix numbered by roman numbers. These numbers indicate that the radiation scattered by a unit volume in the forward direction (angular range $0 < \varphi < \pi/2$) has a ratio γ to the backward scattered portion. The numbers correspond to $\gamma = 1.85$ (V), 2.48 (VI), 4.69 (VII), 8.46 (VIII).

From the tables of FEIGELSON *et al.* intensification factors $G_{0.6}/G_{0.1}$ have been derived and are given in Table 2. The backscattering coefficients d which can be inferred from these are given again in parentheses. The values lie between 12 and 30 % and evidently increase if the optical thickness τ^* or the extinction coefficient increases with constant indicatrix. A decreasing coefficient, d, is found with increasing forward scattering, if the optical thickness is kept constant. The variation with the zenith distance z of the sun is very small which demonstrates that the consideration which leads to equation (1) is a very good approximation of the complex mechanism of multiple scattering. It is, however, only suited to the particular problem of determining the global radiation and not for calculations of the brightness of singular points of the sky. In all the computations for the pure and the turbid atmospheres (Tables 1 and 2) the absorption of the solar and sky radiation by atmospheric gases such as water vapor, carbon dioxide, and ozone has been neglected.

If in the turbid but cloudless atmosphere backscatter coefficients up to 0.30 can be found then even higher values may be expected for cloudy or overcast skies. Among the several theoretical calculations of this problem, we refer to the investigation by KORB (1961) who calculated the global radiation through an overcast by a method which he calls the $(2 + 2\pi)$—stream method. He takes into account the upward, downward and horizontal radiation fluxes and their mutual influence. Table 3 has been compiled from his results. Each one of the given numerical values holds with but very little variations for 4 different zenith distances of the sun ($z = 0°$, 30°, 60°, 80°) and for 3 different reflection coefficients ($r = 30$, 60, 90 %) if the global radiation is compared with the one over black ground. The absorption by water vapor has been taken into account. One can take from this table that a backscatterance of about 50 % is valid for clouds of moderate transparency, 80 % for clouds with large optical thickness, and 95 % for very dense and thick clouds such as nimbostratus or cumulonimbus.

Empirical determinations

It is difficult to investigate the backscattering by evaluation of observations. For it appears hardly possible to compare the global radiation directly with low and with high albedo of the ground under otherwise unchanged conditions. Only at a place where a snow-free area borders a snow-covered one, could one obtain comparable values. Such an opportunity is given at the snow-covered coast of an open ocean, where two albedometers may be mounted one over the land, the other one over the sea. One could also obtain these values from an aircraft flying across the coast. I am indebted to Dr. H.

TABLE 3. *Backscatterances d of overcast skies.* Calculated after KORB (1961).

Cloud	Optical thickness	d
Low cloud, 1 km thick	5	0.48
	30	0.80
High cloud, 1 km thick	5	0.46
	30	0.78
Tall cloud, 5 km thick	25	0.78
	150	0.95

WEICKMANN (1964) for providing me with a series of measurements of downcoming global radiation under a dense overcast sky and of the albedo of the underlying earth's surface taken on such a flight across the snow-covered coastline of New Jersey. Fig. 1 shows the sharp decrease of the albedo from about 50 to 7 % and the smoother variation of the global radiation. Taking the corresponding values of the global radiation 2.5 and 1.75 cal cm^{-2} min^{-1} we obtain from eq. (1) a backscattering coefficient $d = 67\,\%$ for the altostratus cloud at the time of observation.

Another possibility is given when, standing at the coast, one scans the sky by a radiometer with narrow aperture in a circle perpendicular to the coastline and, if possible, perpendicular to the sun's vertical. Viewing at points 15° over the horizon, WEICKMANN (1964) found a ratio of the sky brightnesses between spots over the snow surface and over the water of 2.0 on 9 Feb. 1961, 12.45–13.45 EST close to Asbury Park, N.J., when a large snow ice float was still present about 1 mile off-shore thus disturbing pure conditions. A schematical calculation using eq. (1) would give an even higher value for d but the equation does not hold for single points at the sky. BOLLE *et al.* (1964) measured from the Jungfraujoch Observatory (3570 m) the sky radiance in the infrared of points in opposite directions where one viewing beam was situated over a glacier and the other one over the snow-free lower country. The ratio of the radiances exactly corresponded to the calculations of COULSON *et al.* (1960) for a Rayleigh sky. The agreement in the very clean high-Alpine atmosphere is very satisfactory. This kind of measurement also confirms the well-known observations of navigators in polar seas that the clear sky

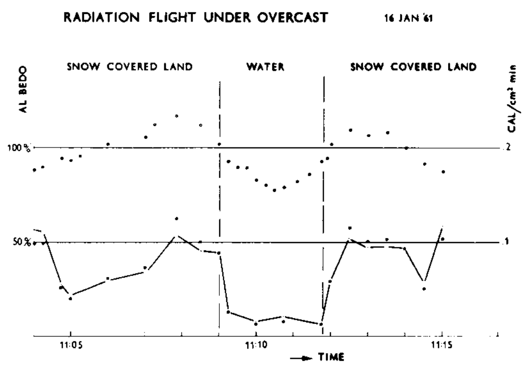

FIG. 2. Histogram of sky radiation H cal cm^{-2} h^{-1} at Moosonee for snow covered ground and solar heights between 20 and 30°; non-overcast sky. Curve A comprises cases of cloudless sky, B those of partly cloudy sky.

as well as the overcast sky is brighter over an icefield than over open water even if the ice field is still below the horizon and invisible. The phenomenon is called "ice blink" (MÖLLER, 1953).

An attempt was made to obtain numerical values of comparison on a statistical basis by comparing measurements of the sky or global radiation which were obtained in winter over a snow-covered surface with others obtained at the same place in summer with snow-free ground. The Canadian station Moosonee ($\varphi = 51°16'$ N, $\lambda = 80°39'$ W) and to a smaller extent Toronto–Scarborough ($\varphi = 43°43'$ N, $\lambda = 79°14'$ W) have suitable pyranometer records with sufficiently long snow cover periods. We only selected the hourly values with sun elevations of 10°–20° and 20°–30° over the horizon. Because there is no information of the snow cover, the ratio of reflected to global radiation was used as a criterion of the presence of snow. Snow-free ground was assumed when the ratio r of the daily sums was smaller than 0.25, snow cover when it was larger than 0.40. Information of the cloudiness during the single hours was not available in the published tables. It is, however, easy to find the hours of overcast sky because then and only then the diffuse sky radiation H is equal to the global radiation G. A tolerance of 1 cal cm^{-2} h^{-1} was permitted so that $H \geqslant G - 1$ was counted overcast, $H < G - 1$ cloudy or cloudless.

For Moosonee during the time from 1.1.1960 to 25.6.1961 (without February 1961) a total

TABLE 4. *Average global radiation G and sky radiation H cal cm^{-2} h^{-1} for snow-free ($r < 25\%$) and snow-covered ($r > 40\%$) ground at Moosonee and Toronto–Scarborough, intensification factors and backscatterances d of the sky derived from them.*

Moosonee	$r < 25\%$ ($\bar{r} = 16.8\%$)			$r > 0.40\%$ ($\bar{r} = 71.2\%$)					
	n	G	H	n	G	H	G_{71}/G_{16}	H_{71}/H_{16}	$d\%$
$10° < h_{\odot} < 20°$									
Overcast	242	5.40	—	389	8.62	—	1.60	—	69
Cloudy + cloudless	241	16.60	6.68	362	19.86	8.19	1.20	1.23	31
Partly cloudy	100	16.45	9.36	166	17.84	11.30	1.08	1.21	14
Cloudless	141	16.71	4.77	196	20.56	5.51	1.23	1.16	34
$20° < h_{\odot} < 30°$									
Overcast	212	9.82	—	185	14.31	—	1.46	—	58
Cloudy + cloudless	235	27.51	10.81	266	33.88	13.23	1.23	1.22	34
Partly cloudy	145	24.73	13.71	140	30.74	18.31	1.24	1.34	36
Cloudless	90	32.02	6.14	126	37.27	7.58	1.16	1.23	25

Toronto–Scarborough	$r < 25\%$ ($\bar{r} = 21.4\%$)			$r > 40\%$ ($\bar{r} = 65.4\%$)					
	n	G	H	n	G	H	G_{65}/G_{21}	H_{65}/H_{21}	$d\%$
$10° < h_{\odot} < 20°$									
Overcast	356	5.66	—	101	6.75	—	1.19	—	36
Cloudy + cloudless	487	18.00	7.24	73	21.41	8.64	1.19	1.19	36
$20° < h_{\odot} < 30°$									
Overcast	300	9.16	—	137	10.44	—	1.14	—	28
Cloudy + cloudless	569	30.10	11.57	151	31.53	12.80	1.05	1.11	11

of 2132 hourly values could be used, for Toronto from 17.8.1961 to 31.12.1962 a total of 2174. Table 4 shows the averaged values of the global radiation G and, in the cases of the non-overcast sky, of the sky radiation H. In Moosonee the groups with equal cloud state and sun height are fairly equally filled for large and small albedo, while in Toronto a shortage of snowy cases occurs because of its lower latitude and higher winter temperatures. The comparisons of Moosonee are therefore based on more data and are considered to be more reliable. The averaged values of the albedo for the selected days are at Moosonee $r = 16.8\%$ and 71.2% for bare and snow covered ground resp., at Toronto 21.4% and 65.4%.

It is well known that global radiation with scattered clouds is, on the average, 2 to 3 times larger than with overcast sky while the sky radiation H is of the same order of magnitude, as it is also shown in Table 4. The intensification factors for G of snowy to bare ground have in Moosonee for overcast sky values of 1.60 and 1.46 for lower and higher sun resp., while in Toronto they are much lower namely 1.19 and 1.14 resp. The values for non-overcast sky are generally lower. Table 4 gives for Moosonee 1.20 and 1.23, for Toronto 1.19 and 1.05. The latter value appears statistically unreliable.

Using the averaged reflectance values r of the ground as given in Table 4 it is possible to determine from eq. (1) the mean backscatterance d of the ocercast or non-overcast sky. It is found to be about 70 or 60 % for overcast, 31 or 34 % for cloudy + cloudless sky, at the mean sun heights 15° and 25° resp. These numbers may be compared with the values based on Korbs' theoretical calculations (Table 3) in the case of an overcast sky. Then, they are comparable with Korb's values for a cloud of 1 km thickness and lie between those for clouds of mean and high optical thickness. This confirms the correctness of the theory as well as the reliability of the statistics. The numerical values

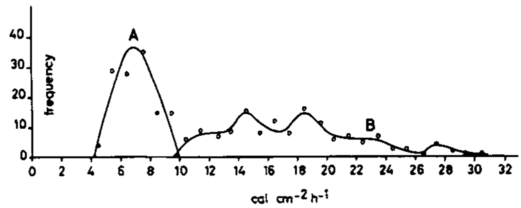

FIG. 2. Measurements of global radiation G (upper circles) and upward reflected radiation R (lower circles) taken from a flight over snow-covered land and open ocean. Full line: albedo $r = R/G$. (By courtesy of Dr. H. Weickmann.)

of d for the cloudy + cloudless skies have about half this size.

In the case of a cloudless and non-turbid atmosphere we had seen that the intensification factors of sky radiation H are much larger than those of the global radiation G (Table 1). That means that the brightening of the sky by the sun radiation reflected from the earth's surface is rather effective. For the cloudy sky, however, the factors for sky and global radiation are equal, except Toronto shows for the higher sun elevation a somewhat larger factor for H than the unreliable appearing factor for G. An explanation for this fact may be that the cloudy sky is already very bright—brighter than the overcast and the cloudless sky—because of the sun light which is transmitted through thin cloud edges or reflected at the bright side walls of cumuliform clouds. This part of the sky radiation is not increased by the larger albedo of the ground and therefore the intensification factor of H is not greater than that of G. A separation into cloudy and cloudless sky would make these relations clearer.

Unfortunately, an exact separation of the cases of non-overcast skies into partly cloudy and cloudless cases is impossible because the needed information is missing. An attempt was made to use the records of a Campbell-Stokes sunshine recorder being in operation on both places. The verification whether or not an hour with full sunshine is really cloudless can be obtained by considering the value of H or of the ratio H/G for the same hour. These quantities are very sensitive to small amounts of scattered clouds. It turned out that in quite a number of cases when the recorder gave 1.0 hours of sunshine H was remarkably great and H/G was even larger then 0.5. Therefore a selection of the cloudless hours appeared impossible in that way.

A histogram of H, however, seems to give good statistical evidence. Fig. 2 shows an example for Moosonee and snowy surface. It appears easy to separate the range of small values of H corresponding to cloudless skies from the irregular distribution of H for the different degrees of cloudiness. The separation between cloudless and partly cloudy skies has been made in that way for all cases in spite of the uncertainity which adheres to it particularly for G. Table 4 gives also these results for Moosonee. The number of data for Toronto is too small for the separation. Even for Moosonee the results are not consistent. The backscatterance is smaller for cloudless sky than for the cloudy one if the sun elevation is in the higher range 20° to 30°, while in the range of 10° to 20° the opposite (and certainly incorrect) result was obtained.

There are only very few other determinations of d from observations. ALBRECHT (1951) derived values between 7% and 19% for cloudless sky which are lower than ours and also in the lower range of those obtained from Feigelson's calculations (Table 2). ÅNGSTRÖM & TRYSELIUS (1934) as well as ETIENNE (1940) used the assumptions first given by Ångström of $d = 0.25$ for clear and $d = 0.75$ for overcast sky. An extensive study by BENER (1963) for different cloud types and cloudinesses unfortunately cannot be used for our purposes because it uses the sky radiation H only and not the global radiation G. KALITIN (1931) applied equation (1) on measurements of the global illumination. From the values he published rather high reflection coefficients of sky and clouds can be derived. For 10/10 cloudiness it follows as average value for sun elevations of 7°, 15° and 30°; ci 37%, as 76%, ac 32%, sc 64%, st 74%, ns 85%, cb 82%, cloudless 3%. The high values for the different clouds may be influenced by differing wheather conditions as thick dark clouds in summer and thinner, better transparent clouds in winter.

Another interesting fact relying on the same principle is the increase of the upward going radiation flux on top of a cloud layer at places where the earth's albedo is increased. HOINKES (1960) has taken very interesting photographs from an aircraft over a stratus deck covering the borderline between the antarctic snow covered continent and the open ocean and showing the strong contrast of the brightness within the cloud surface. From Korb's calculations similar figures may be derived giving intensification factors of 1.25 to 1.70 strongly increasing with the sun's height. But intensifications of this size only occur if the clouds are thin, i.e. in the first and third examples of Table 3. The thicker clouds show increase in the brightness of a few per cent only, the thickest cloud (last example in Table 3) does not show any perceptible increase.

Acknowledgements

The author wishes to thank Dr. H. Weickmann, Fort Monmouth, N.J., for the permission to use his unpublished measurements in this paper, and Dr. W. L. Godson, Met. Office Toronto, for making available the unpublished data of sunshine recorders at the arctic Canadian stations. Particular thanks are expressed to Prof. J. London for giving the opportunity to finish this little study during a stay at the Department of Astrogeophysics of the University of Colorado, Boulder/Colorado.

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