Atmospheric turbidity, global illumination and planetary albedo of the earth

By ANDERS ÅNGSTRÖM, *The Eppley Foundation for Research, Newport, R.I.*

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ABSTRACT

On the basis of measurements by A. J. Drummond of the natural illumination from sun and sky in South Africa, a formula has been derived, expressing the global illumination at sea level as a function of air mass and turbidity. From this formula the present author has derived the relative decrease in illumination caused by the intervention of the atmosphere, or what is here termed the "*photometric loss factor*". The dependence of this factor on turbidity and air mass is studied. From its value at various latitudes the *planetary photometric albedo* of the system earth + cloudfree atmosphere is derived. If the planetary photometric albedo (atmosphere partly cloudy) of Danjon is accepted, we derive a value for the planetary energy albedo of the earth of 0.333. It is suggested that this value is slightly too low. With application of here-derived albedo values for the clear atmosphere and accepting Houghton's value 0.55 for the cloud albedo we obtain a planetary energy albedo of the earth of 0.380. This value implies that the photometric albedo ought to be found slightly higher than the value derived by Danjon from measurements of the brightness of the moon, namely 0.43 instead of 0.39.

Introduction

The problem as regards the albedo of the earth as a whole has in recent time been the object of great interest. Among the authors which have devoted special attention to this question may here be mentioned SIMPSON (1928, 1929), DANJON (1936), HOUGHTON (1954), FRITZ (1951, 1954), DRUMMOND & DEVENTER (1954), BLACKWELL, ELDRIDGE & ROBINSON (1954), and ROBINSON (1956, 1958). The main cause behind this attention is an evident one: the albedo of the earth is one of the fundamental factors in determining the temperature conditions of the atmosphere and of the surface layers of the earth. We may make a concentrated summary of the conditions on the basis of the simple formula:[1]

$$\pi r^2 \cdot I(1-A) = 4\pi r^2 \cdot E(T)$$

or

$$I(1-A) = 4 \cdot E(T), \tag{1}$$

where I is the solar constant integrated over all wave-lengths, r is the radius of the earth. $E(T)$, the energy, which the system earth and atmosphere is losing per unit surface, is a function of what may be termed the effective temperature (T) of the named system. Sometimes E is defined through the arbitrary equation:

$$E(T) = \sigma T^4, \tag{2}$$

where σ is the constant of Stefan-Boltzmann.

Under these conditions T is the mean temperature which a spherical body with an albedo A at the earth's mean distance from the sun would assume under the influence of solar radiation. The variation of T with the albedo A is demonstrated in Table 1. At the albedo value 0.40 for the earth as a whole, a change of 1 percent in the albedo (from 0.40 to 0.41) corresponds to a change in T of close to 1°C. If all the atmospheric layers would be affected in the same way, this would be the change to be expected, also at the surface of the earth. Several conditions imply, however, that this can only be taken as a rather rough estimate. Especially at the surface of the earth, an increase of the incoming radiation is not compensated by an increase also in the outgoing terrestrial radiation, this because of the increase in the water vapor content of the atmosphere, which must follow a rise in temperature. If no increase in convection and evaporation would take place, the effect of a change in

[1] For symbols see the end of the paper.

TABLE 1.

A	T°K	t°C
0.30	254.5	−18.5
0.325	252.2	−20.8
0.35	249.8	−23.2
0.375	247.3	−25.7
0.400	244.9	−28.1
0.425	242.4	−30.6

the albedo on the temperature of the lowest air layers would thus be far above the value mentioned above.

The albedo of the earth as a whole, may be regarded as composed by mainly three different albedos, namely:

I. the albedo of the surface of the earth,
II. the albedo of the cloudless atmosphere, and
III. the albedo of the clouds and of the cloudy atmosphere.

As regards the various albedo values there are some basic conditions to be considered. If the albedo refers to the solar radiation, it is by definition the fraction of the total solar radiation which is reflected, viz. the fraction which is not absorbed and turned into heat at the surface in question. Generally the measurements are carried out in this way: they are taken with a pyranometer, mounted successively in two different positions, in both cases with the receiving element in a plane parallel with the surface, the albedo of which shall be measured. In the first case we measure the vertical component of the radiation falling upon the surface. In the second case the instrument is turned upside down, and also in this case the vertical energy component is measured, now of the specular and diffuse reflection. The albedo is defined as the ratio between these two radiation values.

For a given surface the albedo is in general a function, as well of the angle of incidence of the light falling upon the surface, as also of its wave-length. This means, that if we denote the angle of incidence with θ, the wavelength with λ, the total albedo A ought to be expressed by:

$$A = \frac{\int_0^\infty a(\theta\lambda) \cdot F(\theta\lambda) d\lambda}{\int_0^\infty F(\theta\lambda) d\lambda}, \tag{3}$$

where $F(\theta\lambda)$ is the vertical component of the incident light corresponding to the wave-length.

Now the relative intensity of the various wave-lengths in sun radiation varies within rather wide limits with the composition of the atmosphere as regards absorbing and scattering constituents, and consequently also with solar height. The albedo of most surfaces of ground can then be expected to vary considerably with the named parameters. In spite of this most values for the albedo of ground and vegetation cover, quoted in the literature, are given without reference either to solar height or to the composition of the incident radiation. It seems reasonable to think that they generally correspond to a solar height of 30°–70° and to the composition of solar radiation characterized by normal turbidity (β about 0.10). They are probably in most cases uncertain within rather wide limits.

As regards the reflection from a turbid medium, like the atmospheric aerosol, it is dependent on the scattering and also on the absorption within the volume element. For the case that there is no absorption but only scattering, the albedo will approach the value 1.0 for great thicknesses of the medium. Practically such an ideal case never occurs, but snow layers and comparatively thin layers of magnesium oxide come very near to it for visible light. From the reflection of such media the ratio between scattering and absorption may be computed. The problem was first treated by SCHUSTER (1905) in a classical paper and the theory has been applied and developed further, in detail by DEIRMENDJIAN & SEKERA (1954) as regards the cloudfree and dustfree atmosphere and by MECKE (1944) for cloud sheets of different density and thickness.

I. *The albedo of the surface of the earth* is highly dependent upon the nature of the surface. It seems appropriate to distinguish between some characteristic groups: (1) woods, grass fields and land covered by other forms of vegetation: $a = 0.05$–0.15; (2) sand and rocks free from vegetation: $a = 0.15$–0.30; (3) snow cover: $a = 0.40$–0.85; (4) water surfaces: $a =$ about 0.06, but highly dependent on the solar height. Approximately, the reflectivity of water surfaces for direct solar radiation may be assumed to follow the laws of Fresnel.

II. *The albedo of the cloudfree atmosphere* is dependent chiefly on the scattering by the aerosol. One must here distinguish between the molecular scattering and the scattering by solid and liquid particles, which generally to a variable extent are polluting the atmosphere. Under average conditions the "back-scattering" by the clear atmosphere and the ground of the total solar radiation is of the order of about 10–20 percent. The practical treatment of this problem will constitute the main object of the present note.

III. *The albedo of the clouds* is highly dependent on cloud form as well as cloud thickness. The measurements are difficult and their interpretation not always free from ambiguity. Especially the attempts to generalize the results must be made with precaution, as the measurements generally are limited to local and occasional conditions, which only seldom can be regarded as fully representative either in time or space. Among recent results we may especially mention those of FRITZ (1954), ROBINSON (1958) and NEIBURGER (1949). Fritz derives on the basis of measurements on the brightness of the dark part of the moon by Danjon, an average albedo of the clouded part of the atmosphere of 49 percent. For *persistent overcast* sky at Kew Observatory ROBINSON (1956) gives the average value 65 percent. NEIBURGER (1949) derives an albedo of homogeneous Str-clouds as function of their thickness and arrives at an albedo of 75 percent for a thickness of 600 m against only 38 percent for a thickness of about 100 m.[1]

The variations of the albedo of the earth

More than a quarter of a century ago (1928) SIR GEORGE SIMPSON (1928, 1929) published detailed studies of the heat balance of the earth, based upon the data then available concerning the solar constant, the absorption of the water vapor and the albedo of the earth. Simpson comes to the conclusion, that there is, as to be expected, a perfect balance (within less than 2 percent) between incoming and outgoing radiation, the earth neither losing nor gaining energy in the long run. Simpson assumed on the basis of measurements of the reflection from clouds from balloon by Aldrich, a value of 0.43 for the albedo of the whole earth which is considerably higher than what later has been derived (Fritz, Robinson, Houghton) namely about 35 percent. The values of Simpson for the solar energy absorbed by the earth and its atmosphere must if the lastnamed value is correct be too low. In order that the outgoing terrestrial radiation shall balance the energy income, we must therefore consider the possibility that it also is higher than was derived by Simpson. Our knowledge of the radiation properties of water vapor and carbon dioxide, which constitute the chief atmospheric radiators, seems, however, not to be so accurate, especially when applied to the highest layers of the atmosphere, that there is any serious difficulty in adjusting the terrestrial radiation values as to give a perfect agreement between income and output of energy. It seems not possible to evaluate either of these quantities with an accuracy greater than about 5 percent. Without being able to prove it, one has therefore at present to accept the perfect equality of income and output, as a matter of probability.

Simpson was by his studies induced to take up for examination the interesting problem of the reaction of the earth to changes in the incoming solar radiation. It was clear to him that in order to produce a change of the outgoing terrestrial radiation in a corresponding degree, it would demand temperature changes which were very large even for rather small changes of the solar constant. The arguments of Simpson in this respect may be supplemented through stating that it, with due regard to the changes in the water vapor content of the atmosphere, which must be a consequence of changes in temperature, seems very doubtful if a change in the incoming radiation is ever followed by any change in the terrestrial radiation from the ground out to space. This being so, another compensating factor must be sought for, and Simpson very ingeniously developed the idea that changes in the incoming radiation, when they occur, give rise to a change in cloudiness through a change in the atmospheric circulation and thus to a change of the albedo, which so to say regulates the income and keeps it comparatively constant. It may easily be derived from (1) that an increase in I of, for instance, one percent, can, if the albedo has a value of about 0.35–0.45, be completely compensated if this value changes with about 0.007. At present, however, it seems rather doubtful whether the solar constant integrated over all wave-lengths may reasonably be as-

[1] See also LINKE, *Meteor. Taschenbuch II*, 1953, p. 522.

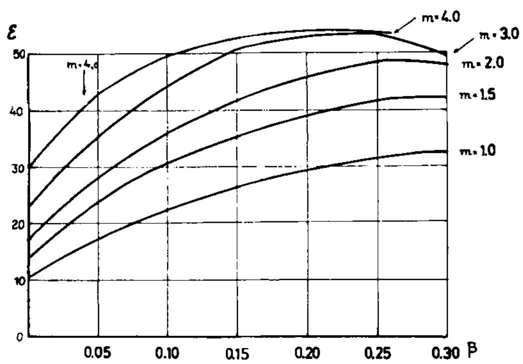

FIG. 1. Photometric loss factor in its dependence on air mass and turbidity.

sumed to have been subjected, during the time when observations have been made, to changes as large as one percent (ALLEN, 1958). But, anyhow, the problem may be reversed from that considered by Simpson, and we may ask *what are the variations of the albedo of the earth* and how do they affect the radiation income and, secondarily, the climate of the earth?

If we consider separately the various albedos, which together give rise to the albedo of the earth, we are led to the following considerations.

The albedo of the snowfree surface of the earth, highly variable as it is from one place to another, must have been comparatively constant with time. The albedo of the oceans has probably a very constant average value, even if local increases may have occurred. As regards the solid surface, it is to large parts covered by vegetation which may naturally change somewhat as well in kind as in extension. But there is no reason to think that the average albedo has been subjected to any but rather small changes from that reason.

Quite another case is presented by the *snow cover*. The albedo is here very high and may perhaps be averaged to about 65 percent. The variation of the extension of the snow cover from one year to another may be considerable, but on the other hand, it is, under present conditions, limited to a rather small part of the earth's surface and to the part where the total energy income constitutes only a small fraction of the total. Further we may expect a change of the average extension of the snow cover to occur as a consequence of a climatic change and not as an ultimate cause of it. The influence of changes in the snow albedo will therefore not be considered here.

The variation of the extension and thickness of clouds probably constitutes the most important cause of changes in the albedo within a given region of the earth. It seems doubtful, however, if such a change can reasonably be assumed to be ultimate cause of changes in the radiation income concerning the earth as a whole. With Simpson one may assume that, if the radiation income is variable from some cause or other, the cloudiness is the natural regulator to counterbalance to some extent the variations. But, if the average temperature remains constant, it seems reasonable to assume that the average cloudiness remains comparatively constant also. The cloudiness may be regarded as an intermediate step between evaporation and precipitation, and they probably both remain rather constant under constant temperature conditions. Whether, however, cosmic factors, e.g. ultraviolet solar radiation or others, may produce variations in the content of condensation nuclei of the upper air layers, and in this way lead to changes in the cloudiness, may still be regarded as an open question. Some observations seem to suggest a relationship between solar activity and the frequency of *Cirrus clouds* without being conclusive as regards generality.

The variations of the albedo in the case of a cloudfree atmosphere

The turbidity of the atmosphere differs from the factors already considered through the fact, that changes occur which are independent of the climatic variations. Such changes are for instance produced by volcanic outbreaks, and by intense forest fires. We have here a factor which undoubtedly may give rise to variations in the albedo and thus in the radiation income. Its existence does not presuppose a climatic change.

An idea of the influence of turbidity on the albedo may now be gained from the measurements or registrations of the illumination corresponding to different values of the turbidity. Such measurements are still comparatively rare. They have, however, been carried out in a systematic way at the Kew Observatory as reported by ROBINSON (1956) and also by Drummond at Pretoria (ÅNGSTRÖM & DRUMMOND, 1962). The lastnamed measurements of

TABLE 2. *Global illumination (kilolux) at sea level to horizontal surface.*

β	m 0	1	1.25	1.50	1.75	2.0	3.0	4.0	6.0
0.00	142	127	100	82	69	59	36	25	15
0.05		118	90	72	60	51	31	19	11
0.10		110	83	66	54	45	26	18	10
0.15		104	77	61	50	41	23	17	11
0.20		100	73	58	46	38	23	17	11
0.25		98	71	55	44	37	22	17	12
0.30		96	71	55	44	38	24	18	14

the illumination, from sun alone, as well as from sky and sun, are supplemented by very extensive filter measurements, which allow an effective control of the illumination values. The measurements are especially valuable, as they refer to conditions where the parameters involved have varied within a large range: the turbidity between $\beta = 0.05$ to $\beta = 0.35$ and absolute air mass (m) penetrated by the sun's rays from 0.85 to 4.0.

The illumination values have been, and will be further treated, in joint papers by the present author and A. Drummond. Here we will limit ourselves to a short summary of the results directly applicable to our present problem.

For the global illumination (by sun + sky) E_{S+D} at the surface of the earth, we have derived the equation (ÅNGSTRÖM & DRUMMOND, 1963)

$$E_{S+D} = \frac{142 e^{-m(0.16+2.4\beta)}}{m} + \frac{8.5 + 84\beta}{\sqrt{m}}, \qquad (4)$$

where the first term refers to sun radiation alone, the second one to sky radiation alone. It may be inferred from this equation, that the selective optical band, filtered out by the human eye from the spectrum of the sun, behaves, as regards its absorption and scattering by the atmosphere, practically, as a homogeneous radiation. The selective absorption and the molecular scattering are expressed by the factor $e^{-0.16m}$, the scattering and absorption by dust by the factor $e^{-2.4m\beta}$. The last-named factor corresponds closely to a wave-length of 555 mμ, if the extinction is supposed to be proportional to $\lambda^{-1.5}$ and β is the scattering coefficient corresponding to the wave-length 1 μ (Ångström, 1929, 1930, 1951).

For the "loss of illumination" ΔE through the intervention of the atmosphere we get $\underset{m=0}{E} \sin h$ being the extraterrestrial luminous flux to a horizontal surface:

$$\Delta E = \underset{m=0}{E} \sin h - E_{S+D} \qquad (5)$$

or

$$\Delta E = \frac{142(1 - e^{-m(0.16+2.4\beta)})}{m} - \frac{8.5 + 84\beta}{\sqrt{m}}. \qquad (6)$$

The values of E_{S+D}, according to (4) are given also in Table 2. Subtracting these values

TABLE 3. *Photometric loss factor of cloudfree atmosphere at sea level.*

β	m 1	1.25	1.50	1.75	2.0	3.0	4.0	6.0
0.00	0.11	0.12	0.14	0.15	0.17	0.23	0.30	0.36
0.05	0.17	0.21	0.24	0.26	0.28	0.34	0.46	0.53
0.10	0.23	0.27	0.31	0.33	0.37	0.44	0.49	0.59
0.15	0.27	0.33	0.36	0.38	0.42	0.51	0.52	0.55
0.20	0.30	0.36	0.39	0.43	0.47	0.52	0.52	0.53
0.25	0.31	0.38	0.42	0.46	0.48	0.53	0.52	0.51
0.30	0.33	0.38	0.42	0.46	0.47	0.49	0.50	0.43

TABLE 4. *Photometric loss factor ($\bar{\varepsilon}$) in its dependence on turbidity (β) declination of the sun (δ) and latitude (ψ).*

ψ	0°			30°			60°		
δ	−23°.5	0°	+23°.5	−23°.5	0°	+23°.5	−23°.5	0°	+23°.5
$\bar{\varepsilon}_1$	0.315	0.302	0.315	0.340	0.312	0.312	0.50	0.370	0.333
for β	0.125	0.125	0.125	0.10	0.10	0.10	0.03	0.075	0.10
$\bar{\varepsilon}_2$	0.393	0.380	0.393	0.455	0.398	0.385	(0.65)	0.480	0.420
for β	0.25	0.25	0.25	0.20	0.20	0.20	0.06	0.15	0.20
$\bar{\varepsilon}_2/\bar{\varepsilon}_1$	1.25	1.26	1.25	1.34	1.28	1.23	(1.30)	1.30	1.28

ε_1 = photometric loss factor under normal conditions of turbidity.
ε_2 = photometric loss factor after increase to the double of turbidity.

from $142/m$ we get the illumination loss. Dividing these differences by $142/m$ we get what we may call the *loss factor of illumination:* ε. This loss factor is given in Table 3 in its dependence on air mass and turbidity.

If nothing of the radiation within the visual optical band $\int_0^\infty \varphi(\lambda) F(\lambda) d\lambda$ [$\varphi(\lambda)$ denotes the sensitivity function of the human eye, $F(\lambda)$ the energy of the extraterrestrial solar radiation], was absorbed in the atmosphere, and if the earth surface was totally black, viz. did not reflect anything of the incident radiation, ε would be what is often called the *photometric albedo* of the system earth and atmosphere in the case of a clear sky. This albedo can as we will see later be assumed to closely coincide with the energy albedo for the total visible and ultraviolet spectrum $\int_0^{700} F(\lambda) d\lambda$ (ÅNGSTRÖM & DRUMMOND, 1962) *under the same presumptions.* Now these presumptions generally fail to be valid, because: (1) the atmospheric ozone absorbes a couple of percent within the named spectral region, (2) the atmospheric dust generally not only scatters but also absorbs a certain fraction, (3) the part of the energy reflected from the ground, *which is not absorbed* in the atmosphere, must be added to the value of ε.

We will consider separately the influence of these factors on the energy albedo. First, however, some words shall be said as regards the information already gained concerning the variation of ε with air mass and turbidity. It is clear that if we do not consider the *energy* which remains within the system earth-ground after the process of absorption and reflection, but simply wish to find the changes which are produced by the cloudfree atmosphere as regards the *visible* radiation, the value of ε and its behaviour are of prime importance. It is the corresponding region of the spectrum, which is responsible for most photobiological and germicidal effects. It is consequently of considerable interest to investigate how a change in turbidity affects it. With this problem in view we are not concerned with the corresponding amount of energy turned to heat through absorption. *This energy is simply lost, from the region of wave-lengths, which matters as regards photobiological and similar processes at the earth's surface.*

From the values given in Table 2 we may now easily compute graphically, what may appropriately be called the effective *photometric loss* factor $\bar{\varepsilon}$ at various latitudes at different times of the year. The daily illumination income at the upper limit of the atmosphere at a given locality is a function of latitude ψ and declination δ and given by

$$\bar{E} = 142 \int_{\tau_1}^{\tau_2} \sin h \, d\tau, \tag{7}$$

where τ_1 and τ_2 are the times of sunrise and sunset, and

$$\sin h = \sin\psi \sin\delta + \cos\psi \cos\delta \cos\tau. \tag{8}$$

From (7) and (8) in combination with (6), the value of $\bar{\varepsilon}$ is derived through graphical integration. Results corresponding to latitudes 0°, 30° and 60°, and to declinations 0°, +23.5° and −23.5° are given in Table 4. The turbidity values are highly approximative and suggested by the results of a preliminary investigation. The uncertainty as regards the regional distribution of β prohibits us from trying to press the results of the computation above what can be regarded as a reasonable approximation. The average value of $\bar{\varepsilon}$ at the equator and also at 30° lat. comes very close to 0.32 with a slight increase to about 0.34 at higher latitudes. The winter value at latitudes above 60° is evidently of small consequence for the illumination integral over the year, as the illumination itself is then rather weak. The measured and computed winter values of the illumination at corresponding low solar heights are uncertain and must in reality be strongly influenced by the albedo of the snow surfaces.

What here interests us especially is the change in the photometric loss factor produced by a change in the turbidity. Let us assume the β-values to be increased with 100 percent, viz. to double the normal value. The new values for the photometric loss factor, $\bar{\varepsilon}$ may then be computed in the same way as previously. In Table 3 these values are given under the heading $\bar{\varepsilon}_2$. We find the loss factor to have increased in the proportion 1.25–1.30, viz. with about 25 to 30 percent, almost independently of latitude and declination. Now an increase of turbidity to the double frequently occurs in connection with violent volcanic eruptions or even following large forest fires. An example of the former kind thoroughly investigated, is given by the outbreak at Mount Katmai in Alaska in June 1912. For several months the turbidity over large parts of northern America, Europe and northern Africa was increased by more than 100 percent.

The photometric loss factor may for values of β between 0.05 and 0.25, be regarded as being, practically, a linear function of the turbidity, and in general an increase of β of 10 percent corresponds to an increase in the loss factor of about 2–3 percent. The climatic importance of the variations of turbidity is clear from these considerations.

Application to the problem of the energy albedo of the earth

We have dealt up to now with values which have been derived from direct observations. So far as the measurements of the illumination and also the value of extraterrestrial luminous flux of solar radiation are correct, the values for the photometric loss may be regarded as comparatively correct also. Some uncertainty as regards the regional distribution of turbidity —especially the conditions over the oceans are not well known—may, however, have introduced an uncertainty also in the generalizations and the absolute values of $\bar{\varepsilon}$ perhaps be in error within a couple of percent.

If we try to use our results concerning the illumination for forming an idea of the *energy albedo* of the earth and the atmosphere, we are met with serious difficulties on account of the deficiency of our knowledge of some important parameters, like the absorption of dust in the atmosphere, the reflection power of clouds and the absorption of water vapor in the atmosphere and of water in the clouds.

As has been mentioned already the photometric loss factor does not coincide with the energy albedo of earth and atmosphere within the photometric band $\int_0^\infty \varphi(\lambda)F(\lambda)d\lambda$, on account of the absorption of the ozone and the dust within this region, and also on account of the reflection of the ground, which adds a certain amount to the apparent albedo. The absorption of the dust and the ozone appears as a loss of illumination but is actually converted into energy by the atmosphere. This apparent loss ought consequently to be subtracted from the photometric loss factor, when we try to derive the energy albedo.

Both the quantity of ozone present in the atmosphere and its absorption are comparatively well known from investigations by various authors. If the average thickness of the ozone layer is taken to be 0.26 cm, the effective absorption within the band $\int_0^\infty \varphi(\lambda)F(\lambda)d\lambda$ is 1.6 percent at vertical incidence of the sun's rays. From this a graphical integration gives 3.0 percent as the average absorption of solar radiation within the named band for the whole earth.

As regards the absorption by dust an evaluation is more difficult. Very different opinions have been expressed on this question.

Houghton, in the absence of quantitative information, counts one-half of the dust depletion as absorption. Robinson assumes the dust of the atmosphere above Kew Observatory to be about 9 percent, which would mean that about 20 percent of the depletion through dust is caused by absorption. It seems that both these evaluations, if we are concerned with the ordinary pollutions of the atmosphere, are too high.

A mathematical treatment based on considerations similar to those of Schuster in his classical paper on "Radiation through a foggy atmosphere", permits us to calculate the "photometric loss factor", defined above, corresponding to different values of the ratio between absorption and scattering. The calculation suggest that only about a tenth of the extinction by "dust" is due to absorption. DEIRMENDJIAN & SEKERA (1954) have, on the basis of theoretical considerations, computed values for the diffuse sky radiation in the case of a pure and dustfree atmosphere. No absorption is here combined with the scattering and the molecules scatter equal amounts of energy forwards and backwards. Presumed that the dust scattering takes place according to the same laws, which is certainly not strictly true, we may compute the diffuse sky illumination in the case of a turbid atmosphere, with application of the results of DEIRMENDJIAN & SEKERA (1954) (see especially Tables II and V of their paper). We arrive at a value of 15 kilolux under the assumption of the turbidity $\beta = 0.10$, a ground reflection of 0.15, and vertical incidence of the sun's rays. The corresponding value of equation (4) derived from actual observations, is 16.9 kilolux. If the dust scattering was combined with an absorption, we should expect the observed value to be smaller than the theoretical one. Now just the opposite is the case. This may be due to one or more of the following causes: (1) the forward scattering of an elementary particle is in excess of the backward scattering in the way that the effect of an absorption is overcompensated as far as the emergent radiation is concerned; (2) the ground albedo is larger than 15 percent; (3) the measurements of the diffuse sky radiation are too high, for instance on account of occasional tendencies to cloud formation.

Probably all these factors, with exception perhaps of (2) have to some extent influenced the experimental results. Neither of them, however, to a degree, which makes it justified to assume an absorption by dust, much exceeding 0.1 of the extinction.

The comparatively low value for the absorption, which corresponds to a limit value for the atmospheric albedo of about 50 percent for an extremely turbid atmosphere, suggests conclusions as regards the nature of the scattering particles. We will, however, not enter upon this question here.

In the lack of more accurate knowledge we will in the following assume the absorption of regular atmospheric dust—volcanic dust and industrial combustion products excluded—to be 0.1 of the dust extinction. The absorption by ozone α_1, as well as by dust α_2 being in general small, we obtain, with reference to equation (4) for the added effective absorption α, within the photometric band:

$$\alpha = \alpha_1 + \alpha_2 = 0.016m + 0.24\beta m. \qquad (9)$$

The absorption at vertical incidence of the solar radiation and for a turbidity $\beta = 0.10$ is accordingly about 4 percent. For the earth as a whole we get through graphical integration an effective absorption of about 7 percent (± 1 percent). Consequently this percentage amount must be subtracted from the photometric loss factor 32 percent in order that we may obtain the loss of energy within the photometric band. It is the rest, viz. 25 percent, which is available for reflection. To these 25 percent must be added the part of the radiation reflected from the ground, which is not absorbed in the atmosphere. Houghton evaluates the average reflected radiation under clear sky conditions to about 10 percent, a value which is highly influenced by the comparatively low reflection from the oceans. With due regard to the absorption by dust and ozone we evaluate the reflected energy reaching outer space to 3.5 percent of the extraterrestrial illumination. We thus get for the real albedo a_p within the photometric band, at clear sky:

$$a_p = 0.32 - 0.07 + 0.035 = 0.285. \qquad (10)$$

To this value a small correction ought to be applied on the following reasons.

Our determination of the photometric loss factor ε was determined through subtracting the global illumination corresponding to sea level conditions, from the extraterrestrial illumination. The global radiation was derived from direct measurements at Pretoria, where the ground reflection may be assumed to have had a value of about 0.15. Now, over the oceans, where the reflection is only about 0.065 (Houghton), the back scattering from the atmosphere must be smaller than at Pretoria. From this reason and on the basis of computations by DEIRMENDJIAN & SEKERA (1954), we derive a slightly higher value for a_p than is given by (10), namely:

$$a_p = 0.295. \tag{11}$$

This is thus the true energy albedo for the whole earth, corresponding to the wave-length integral $\int_0^\infty \varphi(\lambda) F(\lambda) d\lambda$. With good approximation this albedo can be assumed equal to the albedo corresponding to the wave-length of maximum visual sensitivity, namely 0.555 μ.

This albedo value represents the conditions at a *perfectly clear sky*. For the actual conditions of cloudiness at the period of his observations, Danjon derived a value of 0.39 for the visual albedo. Danjon's measurements extend over a period of 9 years, and we may tentatively assume, like Fritz, that the average cloudiness of the reflecting area of the earth and its atmosphere at Danjon's experiments was approximately equal to the average cloudiness C_m as computed from long series of meteorological observations. The cloudiness is expressed in fractions of the whole sky which are covered by clouds and as a weighted average mean a value $C_m = 0.52$ is derived by Houghton. If we assume the average visual albedo of a cloud-covered portion of the system earth-atmosphere to be a_{pc} we get for the visual albedo A_p of the average, partly cloud-covered earth:

$$A_p = (1 - C_m) 0.295 + C_m a_{pc} = 0.39, \tag{12}$$

which, if we introduce: $C_m = 0.52$, gives:

$$a_{pc} = 0.48. \tag{12a}$$

There are reasons to assume that this value for the *visual albedo* of the cloudy atmosphere comes close to its albedo for the total solar radiation integrated over all wave-lengths. Fritz has in this respect drawn attention to the work of Hewson, who has measured the reflection of clouds for different wave-lengths. The reflection differs highly for different types of clouds but it is common for them all that the reflection is practically constant up to a wave-length of 1.4 μ. For longer wave-lengths the reflection is lower with 10 to 20 percent but as only about 10 percent of the total extraterrestrial solar energy falls within this region, the albedo is only slightly reduced by these conditions. If we accept the albedo value of (12a) for the wave-length 555 mμ, an integration from Hewson's curves gives us the slightly lower value

$$a_c = 0.46. \tag{13}$$

for the total energy albedo of the cloud-covered part of the atmosphere.

Albedo of the earth as a planet

We may use the result of the previous considerations to derive the albedo for the whole spectrum. We will first consider the albedo for the wave-length region 0–0.7 μ, viz. for the visible and ultraviolet regions of the solar spectrum. Danjon has published measurements of the reflected light from the moon for three different wave-lengths, namely 0.467, 0.545 and 0.606 μ and has given the results expressed in stellar magnitudes. Applying the standard equation:

$$n - m = 2.5 \log \frac{B_m}{B_n}. \tag{14}$$

where n and m are the stellar magnitudes of two given cases and C_n and C_m are the corresponding brightnesses, we arrive at the following albedo values:

$$\lambda = 0.467 \quad 0.545 \quad 0.606,$$
$$a_\lambda = 0.48 \quad 0.39 \quad 0.315.$$

Within the corresponding spectral region there is between albedo and wave-length an almost linear relationship, which can be expressed by:

$$a_\lambda = 1.03 - 1.19\lambda, \tag{15}$$

where λ is expressed in microns.

Assuming, which naturally must be rather a coarse approximation, that this linear relationship holds between 0.3 and 0.7 μ and that there is a total absorption below 0.295 μ (what is assumed as regards this absorption is of small consequence in this connection) we may compute the effective albedo $a_{0-0.7}$ from the relationship:

$$a_{0-0.7} = \frac{\int_{0.3}^{0.7} a_\lambda \cdot F(\lambda) d\lambda}{\int_0^\infty F(\lambda) d\lambda}. \tag{16}$$

The computation is conveniently made graphically and gives:

$$a_{0-0.7} = 0.41. \tag{17}$$

Evidently the albedo for the whole region ultraviolet-visible, is slightly higher than the photometric albedo 0.39 obtained by Danjon. As the reflection from clouds is practically constant with wave-length within the region in question (Hewson), the increase must be due mainly to differences in the albedo by the cloudfree atmosphere. The albedo 0.295 must in order to be applicable to the whole ultraviolet-visible region be slightly increased by application of a factor ν defined by:

$$(1 - C_m)\,0.295\nu + C_m \cdot a_c = 0.41. \tag{18}$$

Introducing $C_m = 0.52$, $a_c = 0.46$ we obtain

$$\nu = 1.20, \tag{19}$$

which thus means that the energy albedo for short wave radiation ($\lambda < 0.7\ \mu$) in the case of a clear sky is 0.355. The increase of the albedo in the case of the wider region is confirmed to some extent by the relationship found by ÅNGSTRÖM & DRUMMOND (1962) to exist between the illumination on the one hand and the direct solar radiation within the region 0–0.7 μ on the other. The authors found that the illumination (kilolux) could be expressed by the relationship:

$$E = 475\,\varrho \cdot w, \tag{20}$$

where w is the radiation expressed by the integral $\int_0^{0.7} F_1(\lambda)\,d\lambda$ which can be derived from simple pyrheliometric measurements with aid of the filter Scott RG8. For the value of the factor ϱ, we found:

$$\varrho = 0.315(1 + 0.032\,m), \tag{21}$$

where m is the absolute air mass (vertical air mass at sea level equal to unity). From (20) and (21) we get:

$$w = \frac{E}{150(1 + 0.032\,m)}. \tag{22}$$

The equation (22) implies that for growing values of m we have to reduce the value of the illumination with greater and greater numbers, if we wish to derive from it a value for w. This means that the solar radiation is through the scattering of the atmosphere robbed of a part which has a lower luminous efficiency than the rest. The scattered radiation must thus have a smaller luminous efficiency than the direct solar radiation. If we wish to derive the albedo for the whole short-wave region ($\lambda < 0.7$) from the albedo at the effective wave-length of the illumination (λ = about 0.555) we must multiply the latter albedo value with a factor $[1 + f(m)]$, where $f(m)$ is a function of the air mass. The value of ν given through (19) seems with regard to these considerations and apart from what can be derived from Danjon's measurements to be a reasonable one.

The next step for computing the total albedo ($0 < \lambda < \infty$) must consist in an evaluation of the reflected energy for waves longer than 0.7 μ. From measurements by Hess and also from theoretical computations by DEIRMENDJIAN & SEKERA (1954) on the scattered sky radiation, it is evident that the percentage amount of reflected energy of wave-length longer than 0.7 μ must to some extent be dependent on the turbidity, being larger the greater the turbidity. An attempt to evaluate the ratio between the energy amount corresponding to wave-lengths longer than 0.7 μ, to the energy amount below this limit, from the curves given by Hess, leads to values between 0.10 and 0.25, the highest value corresponding to an exceptionally high turbidity. There is not sufficient justification for trying to derive a very accurate value for the named ratio and we will assume the total scattered radiation to be obtained through multiplying the energy within the interval $0 < \lambda < 0.7\ \mu$ with the factor 1.15. This is approximately true for the average sky radiation and we assume it to hold also concerning the energy reflected out to space. As the total energy of the extraterrestrial solar radiation within the interval $0 < \lambda < 0.7\ \mu$ is 0.940 (from tables of Nicolet) and the solar constant has a value 1.98 (Nicolet on the basis of measurements by Smithsonian Institution), we get for total albedo at clear sky:

$$a_0 = \frac{0.355 \cdot 0.940 \cdot 1.15}{1.98} = 0.195 \tag{23}$$

and for the planetary albedo under average conditions of cloudiness:

$$A = 0.48 \cdot 0.195 + 0.52 \cdot 0.46 = 0.333. \tag{24}$$

TABLE 5.

$\beta =$	0.00	0.05	0.10	0.15	0.20	0.25
$A =$	0.29	0.31	0.335	0.36	0.38	0.40

Application to the influence of changes in the turbidity on the albedo

We have shown earlier in this paper, that a change of the turbidity coefficient of 10 percent causes a change in the photometric loss factor of about 2.5 percent. The same percentic effect can be expected to occur in the energy albedo value at clear sky. We may then, with some approximation, put:

$$A = (1 - C_m)(0.105 + 0.90\beta) + C_m a_c, \quad (25)$$

or with introduction of $C_m = 0.52$, $a_c = 0.46$

$$A = 0.293 + 0.43\beta. \quad (26)$$

Table 5 gives the albedo to be expected for different values of the turbidity.

Here the average cloudiness is assumed to be constant and independent of the changes in the turbidity. A change of the turbidity from 0.05 to 0.15 may evidently be expected to cause a change in the energy available for warming the earth from 0.69 to 0.64, viz. with about 7 percent of 0.69, a change which by far surpasses any change in the solar constant as far as our present knowledge goes. On the other hand a change in the solar constant, if and when it occurs, must in general affect larger regions than does a change in the turbidity and the cases are therefore not quite comparable. The table given above for the influence of turbidity on the albedo is highly approximate, as is evident from a number of rather uncertain assumptions behind it. But the order of size of the computed *changes* may in spite of a justified doubt as regards absolute values be expected to be correct within about ± 20 percent, of their own values.

Summary; historical and critical comments

In the present paper we have considered some problems closely connected with the albedo of the earth.

1. On the basis of illumination measurements by A. J. Drummond at Pretoria values for what we have called the *photometric loss factor* have been derived, corresponding to clear sky conditions and to various values of the turbidity coefficient and of the air mass. Corrections have been applied (in earlier paper) to the South African illumination values in order that they may represent sea level conditions. Consequently the same holds concerning the photometric loss factor here derived.

2. Values of the photometric loss factor integrated over various seasons at various latitudes have been computed. The change of the named factor with changes of the turbidity has been investigated. Generally the loss factor changes with about 2–3 percent with a change of 10 percent in the turbidity coefficient.

3. From the values of the loss factor the energy albedo:

$$\frac{\int_0^\infty a_\lambda \varphi(\lambda) F(\lambda) d\lambda}{\int_0^\infty \varphi(\lambda) F(\lambda) d\lambda}$$

is derived. A combination of this albedo value with the photometric albedo obtained by Danjon from measurements of the reflected light from the moon, gives possibility to determine the photometric energy albedo of the average cloud-cover. A value of 0.48 is derived for the lastnamed albedo.

4. The photometric energy albedo has with application of assumptions which seems reasonable in view of measurements and computations, been corrected to represent the whole region ultraviolet + visible ($\lambda < 0.7\ \mu$). A value of 0.41 is derived for this albedo.

5. The total planetary albedo of the earth is computed and a value 0.333 derived, corresponding to average conditions of turbidity and cloudiness.

6. On the assumption that the cloud-amount remains constant but the turbidity changes we find that a change of 10 percent in the turbidity produces a change of about 1.5 percent in the albedo value, or about 0.8 percent in the energy available to warm the earth.

The considerations applied in the previous study are in many respects similar to those already used by Fritz, Drummond and Deventer, Houghton, and Robinson for deriving the albedo, or in treating the problem of the heat balance of the planet earth and its atmosphere in general. For us, however, the question of the

TABLE 6. *Results of different authors as regards the albedo of the earth and its atmosphere.*

Author	Year	Clouds a_c	Ground + atmosphere	Planetary albedo	Photometric albedo	Remarks
Abbot & Fowle	1908	0.65		0.37		Based on general meteorological considerations and radiometric measurements.
Aldrich	1919	0.78	0.17	0.43		
Simpson	1928	0.74	0.17	0.43		
Baur & Philips	1934	0.72	0.085	0.415		Based on detailed meteorological computations and results of radiometric measurements.
Fritz	1948	0.47–0.52	0.17	0.347	0.39	Based on meteorological computations and acceptance of the value 0.39 of Danjon for the photometric albedo.
Houghton	1954	0.55	0.135	0.34		Based on detailed meteorological computations.
Ångström I	1962	0.46	0.195	0.33	0.39	Based on measurements of illumination in dependence of turbidity, general meteorological considerations and accepting the value 0.39 of Danjon for the photometric albedo.
Ångström II	1962	0.55	0.195	0.38	0.43	Based on own derivation of albedo of ground + atmosphere and on Houghton's value for the albedo of clouds.
Danjon	1928				0.29	Based on astrophysical measurements of the brightness of moon light.
Danjon	1936				0.39	

absolute value of the albedo has been a secondary one. Our chief aim has been to investigate how changes in the turbidity can be expected to influence the albedo within different regions of the solar spectrum. For that purpose we have tried to obtain, also, absolute albedo values.

At the derivation of such values, all the authors, the present one not excepted, are met with serious difficulties, apparent from the comparatively great spreading from author to author of the values tabulated in Table 6. If one, like Simpson, Bauer and Philips and in recent times Houghton, tries to arrive at a result from purely meteorological parameters, the difficulty arises chiefly from the deficiency of our knowledge concerning the absorption by water vapor, dust and clouds. The different assumptions concerning cloud absorption are, as Houghton very adequately expresses it, "little more than an educated guess". For the evaluation of the albedo this creates an uncertainty which is only slightly reduced through the very accurate computation of the albedo of the earth's surface—ground and oceans—presented in a comprehensive paper by HOUGHTON (1954). Through overrating the absorption as well of water vapor and of dust, and underrating the scattering by dust, Houghton has, according to our opinion, obtained too low a value for the albedo of the cloudfree atmosphere. On the other hand, it seems probable that he has slightly underrated the absorption by clouds, and thus obtained a planetary albedo closely coinciding with that obtained by Fritz and also with that given above by the present author.

It is clear that with these difficulties in mind the possibility to use the photometric albedo derived by Danjon through astrophysical measurements of the earth light reflected from the moon, compared with the reflected sun light, should be greeted with great hope and expectation.

Fritz, as well as the present author, has tried this way out of some of the difficulties. The procedure thus has been to extrapolate on the basis of what has seemed to be reasonable assumption, a value of the total albedo from Danjon's values of the albedo within the visual band within the solar spectrum. This extrapolation has its hazardous elements. An assumption concerning the absorption of dust must be introduced and on this point evidently,

TABLE 7.

Year	1926	1927	1928	1929	1930	1931	1932	1933	1934
Photometric albedo	0.41	0.44	0.40	0.39	0.39	0.40	0.38	0.37	0.38

the opinions differ within wide limits. Apart from this it must be realized that systematic as well as accidental errors may to some extent adhere to the results of Danjon. From one of his tables (VII) we may derive values for the photometric albedo for the years 1926–34. They are given in Table 7.

The variations from year to year may depend on several influences. (1) The average cloudiness can have been different; (2) the average density and albedo of the clouds may have changed; and finally (3) the average turbidity may have been subjected to changes from different reasons.

From our equation (25) we may easily get a general idea about how these different factors are liable to influence the albedo. Introducing the average values found through observations for:

$$A,\ C_m,\ \beta \ \text{ and } \ a_c$$

we obtain:

$$\frac{dA}{A} = 0.4\frac{dC_m}{C_m} + 0.1\frac{d\beta}{\beta} + 0.6\frac{da_c}{a_c}. \tag{26}$$

We see that a small percentic change in C_m affects the albedo about 4 times as much as the same percentic change in β, and about $\frac{2}{3}$ as much as the change in the cloud albedo. The amount and albedo of clouds are the dominating factors. With regard to these conditions the variations from year to year found by Danjon are not surprising.

Measurements like those of Danjon, however, put a rather high demand on accuracy, as two sources of light which differ highly in intensity must be compared. A number of necessary corrections must be introduced. Earlier measurements in the year 1928 by the same author gave as low a value as 0.29 for the photometric albedo, a value obtained on the basis of an inaccurate formula for the brightness of the moon. The new method is stated to eliminate presumptions as regards this factor and is based totally on direct observations. This means that the author replaces the observations, on which the inaccurate formula on the moon's brightness was based, by his own observations, and the question naturally arises what accuracy he has been able to obtain. A systematic error of let us say 0.05, for instance, from 0.44 to 0.39, would mean that the cloud albedo would be 0.55 and the planetary albedo 0.39 instead of about 0.34 obtained from the supplementary assumptions of Fritz or from those given in previous parts of the present paper.

To these remarks may be added the following ones. It seems highly doubtful if we are justified in assuming that the average cloudiness effective at the measurements of Danjon is the same as the average cloudiness of the whole earth. The measurements were carried out at a given locality (Observatory of Strasbourg) and generally at a clear or almost clear sky. Decisive for the scattered light from the part of the moon not illuminated by sun radiation is then the reflection from a part of the earth's surface, where Strasbourg has a special position. As a perfectly clear sky often occurs in connection with high pressure systems of considerable extension, it seems worth while to consider, whether a clear sky at the place of observation does not involve a cloudiness lower than the average one within the reflecting area of the earth, effective at the experiments. The correlation between the cloudiness at a given station and the average one within such a wide area around it, is probably rather small, but may yet be sufficient to produce a marked deviation from average conditions. The influence goes in the direction to make the albedo value of Danjon too small.

If we modify the computations by Houghton, by introducing a value for the albedo of the cloudfree atmosphere of 0.195 instead of his value 0.135—which latter we, on reasons given above, regard to be too low—we obtain, with application of the cloud albedo value 0.55, which coincides with that of Houghton, a value for the planetary albedo of 0.380. It therefore seems as if the purely meteorological considerations would be in favour of a value for the planetary albedo of about 0.38, which would imply that the photometric mean albedo 0.39

TABLE 8.

Albedo of cloudfree atmosphere and earth	0.195
Albedo of average totally cloud covered atmosphere	0.55
Photometric albedo	0.43
Planetary albedo	0.38

derived by Danjon from 9 years measurements is too low to be climatologically representative and ought to be about 0.43 instead. It is evident from what has been said, that this correction of Danjon's value, need not necessarily be due to a remaining systematic error but equally well to the inability of his short series to represent a climatological average for the whole earth. On the basis of the last considerations the present author is in favour of suggesting as *probable* values for the various albedos which collaborate in producing the mean planetary albedo of the earth, the values given in Table 8.

In summarizing the conditions in this way, we may also say that we have used the meteorological parameters for determining the photometric albedo, instead of going the reversed way. The value obtained for the planetary albedo does very closely coincide with the mean value of those given by Abbot and Fowle, Aldrich, and Baur and Philips on the one hand and Fritz and Houghton on the other. However, as Robinson has pointed out, the uncertainty in the result of all these computations must amount to about ± 0.03.

It seems appropriate to finish this paper by emphasizing some points to which especial attention must be paid in future works on the same subject. The importance of a continuation of the work inaugurated by Danjon and of attacking the albedo problem by satellite observations has already been stressed by Fritz. It must take a rather long time, however, before the satellite observations may be able to furnish mean values for the albedo factor as a climatological parameter. Everyone familiar with the technical difficulties of radiation measurements, already at the more easily controllable conditions at the ground, shares the hope attached to satellite observations, but regards it probable that much experience is yet necessary before the needed accuracy can be gained. Observations which seem desirable are extended measurements of both cloudiness and turbidity within an expanded network, especially over the oceans and also of water vapor absorption through radiometric techniques. It seems highly important that more clear ideas should be gained concerning the extinction by dust and that especially the controversial problem of the relation between scattering and true absorption—where the energy is converted into heat by this agent—is seriously attacked. Here an intimate collaboration between observers in the field and experimenters in the laboratory seems desirable.

Symbols

I = extraterrestrial solar radiation to surface perpendicular to sun's rays at the limit of the atmosphere. $I = 1.98$ gram calories cm^{-2}, min^{-1}.

T = absolute temperature °K.

t = temperature °C.

A = planetary albedo of earth and atmosphere.

θ = angle of incidence.

λ = wave-length. Unit microns (μ).

E = luminous flux through horizontal surface.

E_{S+D} = luminous flux from sun and sky to horizontal surface.

m = absolute air mass. Unit: vertical air mass at sea level of standard atmosphere.

h = solar height.

ε = photometric loss factor.

$\bar{\varepsilon}$ = photometric loss factor corresponding to integral loss from sunrise to sunset.

τ_1 = time of sunrise.

τ_2 = time of sunset.

β = turbidity coefficient (Ångström).

$\int_{\lambda_1}^{\lambda_2} F(\lambda)\,d\lambda$ = extraterrestrial solar radiation at the limit of the atmosphere to horizontal surface, integrated over wave-length interval from λ_1 to λ_2.

$\int_{\lambda_1}^{\lambda_2} F_1(\lambda)\,d\lambda$ = solar radiation (direct and scattered) to horizontal surface integrated over wave-length interval from λ_1 to λ_2.

a_p = planetary photometric albedo = $\int_{\tau_1}^{\tau_2}\int_0^\infty \varphi(\lambda) F(\lambda) d\lambda d\tau / \int_{\tau_1}^{\tau_2}\int_0^\infty F(\lambda) d\lambda d\tau$ at clear sky.

$\varphi(\lambda)$ = standard sensitivity function of human eye.

A_p = planetary photometric albedo of actual atmosphere and earth (average cloudiness).

a_{pc} = planetary photometric albedo of homogeneous average cloud layer.

a_λ = planetary photometric albedo at clear sky corresponding to wave-length λ.

a_c = planetary energy albedo of cloudy atmosphere and earth.

ϱ = luminous efficiency factor.

C_m = average planetary cloudiness.

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