have lower maximum surface pressures, above which they collapse without buckling. If in Eq. 3 we set $\beta$ equal to the limiting surface pressure $\beta_{max}$, however caused, we can solve for $x_{max}$, the maximum length of film. For $u = 23.2$ cm/sec and $\beta_{max} = 73$ dyne/cm, $x_{max}$ is about 1 m.

The Reynolds number $(ux)/\nu$ at $x_{max}$ is $2.32 \times 10^5$, which is low enough so that the entire length of the boundary layer is certain to be laminar (*1*, p. 37). If the boundary layer becomes turbulent, Eq. 3 overestimates $x_{max}$, but this can occur only for longer films with, perhaps surprisingly, lower water speeds. Raising $u$ lowers the Reynolds number at $x_{max}$ because $x_{max}$ varies as $u^{-3}$.

A surface dam is a special case of upwelling and subsidence. Water subsides to get underneath it and then upwells on the downstream side. A stream with a succession of surface dams will be covered with film only for a short distance upstream of each dam. Except for what little material has diffused to the surface since the last upwelling, the rest of the stream will have no covering.

This will be true of upwelling and subsidence of any origin, be it bottom conformation and eddies in streams, density gradient waves in the sea (*3*), or whatever mechanism causes the linear pattern of upwellings and subsidences that go along with wind slicks (*4*). Suppose an array of linear upwellings 100 m apart makes the water flow at 10 cm/sec toward linear subsidences between them. Any film will be compressed into sharp-edged bands over the subsidences. The bands will be only 7 m wide if the collapse surface pressure of the film is 40 dyne/cm.

The sharpness of these edges is suggested by indicators other than the D-line, often when $u$ is too low, or the water is too rough for D-lines to be seen. In wind slicks the film makes its presence known by damping short-wavelength waves. The transition from a ripply to a glassy surface occupies a distance of 30 cm or so, most of which may be the distance that the ripples have to run within the slick to be attenuated to insignificance. A sharper transition is revealed if we spray the water with a garden hose. The spray, splashing into the water, makes many little drops that skid across a clean water surface only to drown within a centimeter or so of the same line at the edge of a slick, a line well on the ripply side of the transition zone defined by the damping of the wavelets.

The sharp edges are predicted by Eq. 3, which applies exactly only if the water which approaches the edge carries no film material on its surface. This is never precisely true, and we can arrange cases where considerable material is carried. Will the film on the approaching water remain undisturbed until it hits and adds to the trapped film at a line of discontinuity; or will the jump in surface pressure occur more gradually? We might expect the latter to occur, unless the film has a very special relation of surface pressure to areal density. Yet if a drop of Mazola oil is added to an expanded Mazola oil slick, an expanding D-line is formed; so if the discontinuity is no longer ideal, the transition is still quite abrupt.

The D-line is not the only demarcation line to be found on water. Lines can often be seen at the meeting of waters of different properties, as when fresh water runs over salt water (*5*). But when we see a thread-narrow line we should suspect the presence of a natural or man-made contaminant film. D-lines will form on flowing Bethesda tap water, and on the cold, clear water of an Adirondack mountain brook.

C. W. McCutchen
*Section on Rheumatic Diseases, Laboratory of Experimental Pathology, National Institute of Arthritis and Metabolic Diseases, Bethesda, Maryland*

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## Planetary Albedo Changes Due to Aerosols

Abstract. *Absorption and scattering by aerosols and reflection of solar radiation from the surface determine the sign of the change in the planetary albedo caused by the presence of aerosols. This change in planetary albedo results in atmospheric heating or cooling. Small changes in the ratio of absorption to scattering over time can reverse such heating or cooling trends.*

In the past few years there has been growing concern over the effects of pollutants on the environment. Although greater attention has been devoted to biological effects, there are also possibly significant meteorological effects. This paper concerns changes in the planetary albedo which, in turn, affect the temperature of the earth. The planetary albedo is the proportion of the incoming solar energy that is reflected back to space by the earth and the atmosphere.

Until recently, it had been believed that an increase in carbon dioxide was contributing to an increase in the mean annual worldwide temperature (*1*). However, the decrease in the worldwide temperature since 1945 has required further hypothesis. McCormick and Ludwig (*2*) suggested that increased aerosol content resulting from pollution increased the backscattered component of incoming solar radiation and thereby caused the cooling. However, Charlson and Pilat (*3*) showed that an aerosol layer, by absorbing solar radiation, could have the opposite effect, that is, warming, and that other hypotheses are required to explain the observed cooling. I shall show here that the ratio of absorption to scattering and the surface (or lower cloud layer) albedo are important parameters in determining the temperature change.

A relationship between the planetary albedo, the incoming solar radiation, and the outgoing infrared radiation, which is a function of the mean temperature for a spherical body under the influence of solar radiation at the mean distance of the earth from the sun, is given by

$$(1 - A)\, S = 4\sigma T^4 \qquad (1)$$

where $S$ is the incident solar radiation on a horizontal surface (integrated over wavelength), $A$ is the planetary albedo, $\sigma$ is the Stefan-Boltzmann constant, and $T$ is the mean radiative temperature of the sphere.

It should be emphasized that the mean radiative temperature is the temperature one would obtain by radiometric observation from outside the atmosphere. The surface temperature of the earth is dependent on convection, conduction, evaporation, and the solar and infrared radiation fluxes. Angström (*4*) stated that a change in the mean temperature "can only be taken as a rough estimate" of a change in the surface temperature. This report is concerned with changes in the mean radiative temperature; changes in the surface temperature are merely inferred from the assumption of Angström. Further work to include changes in infrared radiation that affect the mean radiative temperature is necessary to substantiate this assumption itself and to determine the effect that aerosols have on its validity.

The planetary albedo is composed of backscattered radiation from the atmosphere and reflected radiation from the surface and clouds. The amount of radiation returned to space is reduced by absorption within the atmosphere, which heats the atmosphere. The addition of pollutant or natural particles alters both the backscattering and the absorption of solar radiation and can, therefore, alter the albedo.

The changes in planetary albedo due to aerosols, hereafter referred to as aerosol delta-albedo and denoted by $\delta$, may be defined as

$$\delta = A_p - A \qquad (2)$$

where the subscript denotes the presence of the aerosols. Therefore, Eq. 1 can be rewritten as follows to include the presence of aerosols

$$[1 - (A + \delta)]\, S = 4\sigma T^4 \qquad (3)$$

In their hypothesis McCormick and Ludwig assumed that $\delta$ is positive, whereas Charlson and Pilat showed that it is possible for $\delta$ to be negative for an absorbing aerosol layer. A more detailed analysis of the aerosol delta-albedo that includes the surface albedo and the absorption and scattering properties of the aerosols is presented below.

A formula for the aerosol delta-albedo can be derived from the changes in the solar fluxes reflected downward and upward within the atmosphere. The derivation is simplified by normalizing each of the radiation terms in Eq. 3, that is, by dividing each term by the incoming solar radiation, thus making all the terms in the delta-albedo a fraction of the incoming solar radiation.

The net contribution to the delta-albedo from the downward flux within the atmosphere is the upward scattering ($\sigma_p$) caused by the aerosols. It is assumed that the introduction of aerosols does not alter the absorption and scattering characteristics of the original atmospheric constituents.

The change in surface-reflected radiative flux can be represented by the term $a(I_p - I)$, where $a$ is the surface (or lower reflecting cloud layer) albedo, $I_p$ is the normalized flux received at the surface in the presence of atmospheric aerosols, and $I$ is the normalized flux in the absence of particles. This term can be defined further as

$$a(I_p - I) = -\, a(\alpha_p + \sigma_p)$$

where $\alpha_p$ is the fractional absorption due to particles and $\sigma_p$ is the fractional scattering due to particles. This equation would represent the contribution to the delta-albedo in the absence of further atmospheric effects on the beam. However, the reflected radiation is reduced by scattering back to the ground and absorption within the atmosphere. The assumption made for the downward beam will be applied to the reflected beam; that is, the presence of particles does not alter the absorption and scattering due to the other atmospheric constituents. This assumption is valid for the reflected radiation whenever the relation

$$1 - (\alpha_p + \sigma_p) \gg 2(\alpha_m + \sigma_m)$$

is satisfied, where $\sigma_m$ and $\alpha_m$, respectively, are the fractional backscattering and absorption from the atmosphere without the inclusion of particles. It is further assumed here that radiation scattered back to the surface is not reflected a second time. This assumption has the effect of removing terms containing second and higher orders of the albedo. The presence of particles will reduce the reflected radiation by scattering $\sigma_{pr}$ and absorption $\alpha_{pr}$. However, the absorption and scattering of the reflected radiation can be written in terms of the incoming radiation as follows

$$\alpha_{p_r} + \sigma_{p_r} = a\, I_p\, (\alpha_p + \sigma_p)$$

where

$$I_p = 1 - (\alpha_m + \sigma_m) - (\alpha_p + \sigma_p)$$

Therefore, the delta-albedo is given by

$$\delta = \sigma_p - a(\alpha_p + \sigma_p) - a\, I_p\, (\alpha_p + \sigma_p) \quad (4)$$

It is now possible to determine a combination of parameter values (that is, values of the surface albedo and of the ratio of absorption to backscattering) that produces an aerosol delta-albedo equal to zero. The ratio at zero aerosol delta-albedo, $R_0$, hereafter called the equilibrium ratio, is

$$R_0 \equiv \left(\frac{\alpha_p}{\sigma_p}\right)_{\delta = 0} = \frac{1 - a(1 + I_p)}{a(1 + I_p)} \quad (5)$$

The limits on $I_p$ are 0 and 1. Therefore, the limits on the ratio $R_0$ are

$$\frac{1-a}{a} \geq R_0 \geq \frac{1-2a}{2a}$$

where the first (or upper) limit is approached in strong aerosol extinction and the second (or lower) limit is approached in weak extinction. Values of the ratio of absorption to backscattering that exceed the equilibrium ratio produce a negative delta-albedo which, according to Eq. 1, results in a higher mean temperature. Values of the absorption-to-backscattering ratio that are less than the equilibrium ratio result in lower mean temperatures. Strong extinction invalidates the assumptions used in the derivation of the aerosol delta-albedo, so the upper limit is not strictly applicable. Because a negative ratio of absorption to backscattering is physically meaningless, a negative equilibrium ratio indicates that equilibrium is not possible.

An interesting speculation can be made from the above analysis concerning the warming and cooling of the atmosphere. If it is assumed at a given time that the earth is in equilibrium, then several factors can alter the equilibrium. These factors include (i) a change in the surface (or lower cloud layer) albedo; (ii) a change in the extinction (absorption plus backscattering) of solar flux reaching the surface; or (iii) a change in the absorption-to-backscattering ratio of the aerosols. The first two factors change the equilibrium ratio. The third is a function of the composition of the aerosols.

The largest change in surface albedo results from changes in snow cover. Changes also may result from urbanization. When the aerosols are above cloud layers, as, for example, in the stratosphere, the albedo of the lower surface also depends on the amount of clouds. Changes in the amount and composition of aerosol in the atmosphere can result from sedimentation or from volcanic activity as well as from man's activities.

If it is assumed that the equilibrium ratio is unchanged over short periods of time, then small changes in the ratio of absorption to backscattering determine whether the atmosphere is heating or cooling. For instance, in some American cities pollution controls and a changeover from soft coal to oil and gas have changed the composition of the particles. This change in the character of the pollution emissions, even if the total increases, may have reduced the absorption-to-backscattering ratio during this period. Thus the new value of the ratio may be less than the equilibrium value, which, in itself, would result in lower temperatures. In the future the ratio may be further reduced by the transition to a nuclear (or other "clean") technology. One might further hypothesize that during transition periods of this type (reduced absorption

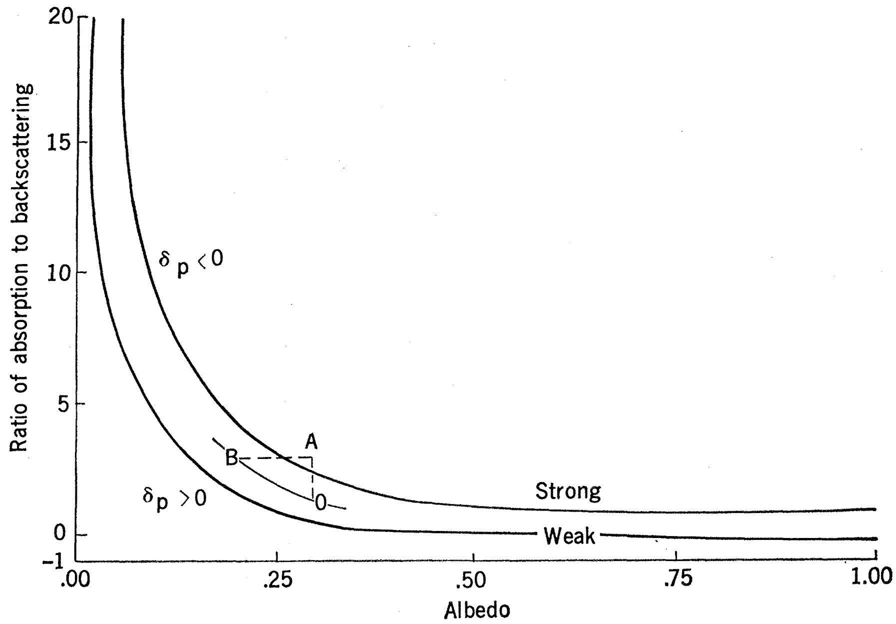


Fig. 1. Equilibrium ratio as a function of the surface albedo and the ratio of absorption to backscattering for aerosols. Both strong extinction and weak extinction are indicated.

to backscattering) the surface albedo response lags sufficiently, so that it reflects, for example, equilibrium conditions for the earlier technology. Then, any further reduction in the ratio of absorption to backscattering would result in further cooling.

Feedback effects from the warming or cooling and the interaction of the three factors introduce many complexities. For instance, a small increase in the absorption-to-backscattering ratio resulting in atmospheric heating would probably reduce the snow coverage and the surface albedo. Thus, the equilibrium ratio would change. Figure 1, which exhibits values of the equilibrium ratio as a function of surface albedo for weak and strong extinction, shows the hypothetical change in albedo (from *A* to *B*) that would result from such a change in the absorption-to-backscattering ratio (from 0 to *A*). In practice, the transient stages indicated in Fig. 1 are probably replaced by more complex transition states.

At this point, we may now consider numerical values of the parameters described above. Values of albedo of the lower reflecting surface have been tabulated by many authors. These values include 0.05 to 0.20 for vegetation, 0.15 to 0.30 for sand and rocks, 0.40 to 0.85 for snow, and 0.06 (but dependent on solar zenith angles) for water (*4*). The albedo for clouds ranges from nearly 0 for thin clouds to 0.7 for thick stratus clouds (*5*). The average planetary albedo without clouds is 0.195 and with clouds is 0.33 to 0.38 (*4*). Representative values of the various parameters needed to determine the delta-radiation are needed. For purposes of demonstration, values used were given by or inferred from Lettau and Lettau (*6*) for urban (Kew, England), desert (La Joya, Peru), and prairie (O'Neill, Nebraska) conditions. These values are based on the model of Lettau and Lettau and on values of radiant flux observed by Robinson (*7*), Stearns (*8*), and Lettau and Davidson (*9*). Near noontime, cloudless values for a city indicate that the absorption-to-backscattering ratio has a value of 4.10 while the equilibrium ratio is 2.85. The inferred value of the absorption-to-backscattering ratio in the desert case is 0.700 (2.100 is the value of the equilibrium ratio) and in the prairie case is 0.585 (1.520 is the equilibrium value). These values result when one assumes that one-third of the scattered radiation is returned to space (as assumed by Lettau and Lettau).

Using these values, we see that the present urban aerosol–surface albedo environment would produce a warming trend if solar radiation were the only thermal process acting. On the other hand, the aerosol-albedo combination characteristic of the desert environment and prairie environment appears to produce cooling trends (again in the absence of other thermal processes).

Additional data, both in time and in space, are needed to compute representative values of the equilibrium ratio and the absorption-to-backscattering ratio for different types of aerosols. Although changes in these parameters are important, there appear to be no data concerning such changes. Further work will be necessary to remove the assumptions made in the present derivation and to include other thermal effects, particularly the infrared effects. Then the hypothesis concerning the temperature change in the atmosphere can be examined in greater detail.

MARSHALL A. ATWATER
*Center for the Environment and Man, Inc., Hartford, Connecticut 06103*

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# Boconó Fault, Venezuelan Andes: Evidence of Postglacial Movement

Abstract. *Postglacial, right-lateral, strike-slip movement along the Boconó fault, measured on detailed topographic maps, averages 66 meters. The rate of movement was approximately 0.66 centimeter per year, as indicated by carbon-14 dating of associated soil. This evidence suggests that postglacial movement between the Caribbean and Americas plates occurred mainly along the Boconó fault and the north coast of Venezuela.*

There is increasing evidence (*1*) that the northern boundary of South America lies within the zone of contact between two plates of lithosphere, the Caribbean and the Americas plates (Fig. 1A). Right-lateral, strike-slip dis-