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Effects of aerosol on the local heat budget of the lower atmosphere

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SUMMARY

The convergence of heat below subsidence inversions was estimated from radiosonde ascents on fine summer days. Sensible heat input from the ground was estimated from measurements of heat fluxes and net radiation over a wheat field. Absorption and back scattering of solar radiation by aerosol throughout the atmosphere were calculated from measurements of solar radiation at the ground. Radiation absorbed by aerosol in the thermal boundary layer heated the lower atmosphere at an average rate of 3·3 degC day^{-1} (60W m^{-2}), about twice as fast as estimated for gaseous constituents.

An equation is derived to show how the net effect of aerosol on the lower atmosphere depends on the reflection coefficient of the surface and on the aerodynamic and surface resistances to vapour transfer. Over the wheat, the net effect was almost zero when the crop was transpiring fast, but when the crop was mature, aerosol caused net cooling.

1. INTRODUCTION

As part of a wider attempt to identify possible causes of climatic change, attention has recently been directed to the absorption and scattering of solar radiation by aerosol, processes which modify the heat balance of the atmosphere and the radiation budget at the earth's surface. Calculations by Mitchell (1971), and by others, demonstrate that aerosol may be responsible either for cooling or for heating the lower atmosphere depending on factors such as the height, thickness and optical properties of the layer and the relative inputs of sensible and latent heat from the ground. The main sources of uncertainty in predicting thermal effects of aerosol in the atmosphere are an incomplete knowledge of (i) optical properties and distribution of aerosol, and (ii) the relationship between net radiation and sensible heat flux at the earth's surface.

We have made direct measurements, during summer anticyclonic conditions in Britain, of heat storage in the lower atmosphere, coupled with measurements of surface radiation and heat fluxes. In the first instance, the measurements were analysed to show the importance of radiant energy absorbed by aerosol in determining the energy budget of the lower atmosphere. The analysis was then extended to estimate the influence of atmospheric aerosol on the input of sensible heat to the atmosphere from the ground. It appears that aerosol may be responsible for a net heating or cooling of the lower atmosphere at a rate which depends on surface reflectivity and on the ratio of aerodynamic and surface resistances to vapour transfer.

2. THEORY

During anticyclonic weather in Britain, inversions formed by subsidence are common at heights between 1 and 2km. Below the main inversion, the atmosphere is heated by radiation and by an input of sensible heat from the ground. Heat may also be entrained from the warmer air above the inversion as discussed by Cattle and Weston (1975). For simplicity, we shall consider the heat budget of a 'thermal boundary layer' defined as that layer of the atmosphere above which there was no measurable change of temperature due

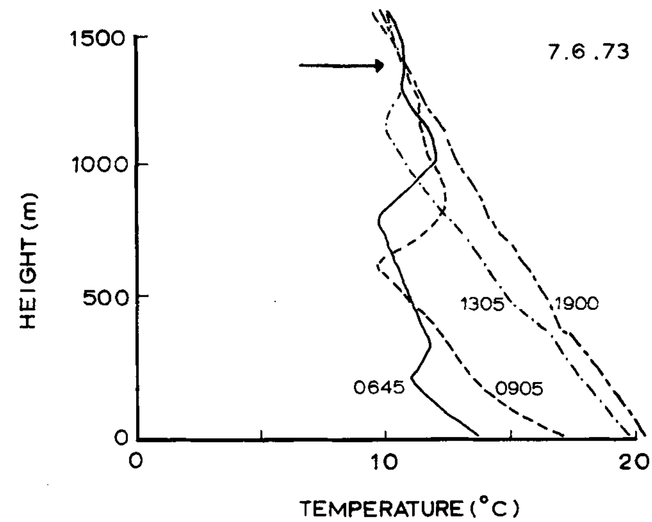

Figure 1. Profiles of temperature in the atmosphere measured with radiosondes on 7 June 1973 when the thermal boundary layer was at 1300m. Times in GMT.

to heating of the surface during the day (see Fig. 1). Entrainment may be regarded as a redistribution of heat within this layer.

In the absence of cloud, the rate of heating of the thermal boundary layer can be expressed as $\partial T/\partial t = gH/(c_p \delta p)$ where H is the heat flux convergence (W m^{-2}) between pressure levels differing by δp(N m^{-2}), g is the acceleration of gravity (m s^{-2}), and c_p is the specific heat of air at constant pressure (J g^{-1}K^{-1}). When $\partial T/\partial t$ is expressed in units of K day^{-1} and δp is in mb, $\partial T/\partial t = 8{\cdot}4H/\delta p$.

The convergence of the total heat flux can be expressed as the sum of three components: the input of sensible heat from the ground H_S; the absorption of radiant energy by atmospheric gases H_G; and the absorption of radiation by aerosol H_A. The term H_G has a shortwave component $H_{G,S}$ and a longwave component $H_{G,L}$.

The absorption and scattering of shortwave radiation by particles in the size range 0·1 to 1μm are much more important energetically than the corresponding processes for longwave radiation, which we neglect in this analysis. Following nomenclature and approximations used by Mitchell, the amount of shortwave energy absorbed by aerosol before and after reflection at the ground is written $a(1+A)S_0$ where S_0 is the radiant flux density on a horizontal surface at the ground in the absence of aerosol, a is a linear absorption coefficient, and A is the reflection coefficient of the ground.

If the absorption of radiation by aerosol in the thermal boundary layer is a fraction D of absorption in the whole atmosphere, the convergence of radiative flux attributable to aerosol absorption can be written

$$H_A = Da(1+A)S_0 \quad . \quad . \quad . \quad (1)$$

and the total convergence of flux is

$$H = H_S + H_R + Da(1+A)S_0 \quad . \quad . \quad . \quad (2)$$

The value of D can be estimated from this equation when all other terms have been estimated or measured. Direct measurements of D are rare but from evidence given by Roach (1961a) and Robinson (1966) a value between 0·3 and 0·6 seems appropriate for the lowest kilometre of the atmosphere during anticyclonic conditions in England. Mitchell proposed a value of $D = 0{\cdot}6$ for continental regions in summer, and Paltridge and Platt (1973) found $D = 0{\cdot}8$ in a dense haze of smoke particles in Australia.

The net loss of radiant energy at the ground surface depends on scattering by aerosol as well as absorption. If b is a back-scattering coefficient for incoming radiation and b' is the corresponding coefficient for radiation reflected from the ground, the net loss will be

$$\delta S_0 = S_0(a+b-Ab') \qquad (3)$$

To investigate the net effect of aerosol on the heat balance of the thermal boundary layer, two distinct processes must be accounted for: the additional absorption of radiation shown explicitly by Eq. (1) and the decrease in the sensible heat flux, δH_S, caused by a decrease in the net radiation flux density R_N absorbed by the ground. The net convergence of heat attributable to aerosol is

$$\delta H = Da(1+A)S_0 - \delta H_S \qquad (4)$$

The two terms on the right-hand side of this equation are both positive by definition so aerosol heats the atmosphere when $Da(1+A)S_0$ exceeds δH_S and vice versa.

In previous analyses which explored the conditions for net heating or cooling, the calculation of δH_S failed to take proper account of the partitioning of sensible and latent heat at the ground. For example, Neumann and Cohen (1972) assumed that

$$\delta H_S = \delta S_0\, \partial R_N/\partial S_0 = (1-A)\delta S_0$$

where $R_N = (1-A)S_0+L$ is the net flux density of radiant energy at the ground surface and L is the net longwave flux density. This value of δH_S is valid only when: (i) the surface temperature (and hence the net longwave exchange) is independent of S_0; (ii) the conduction of heat below the surface is negligible; (iii) latent heat used for evaporation at the ground is released locally by condensation in the thermal boundary layer; or (iv) the evaporative rate is zero.

Mitchell removed the last two restrictions by confining his analysis to a cloud-free boundary layer and by using the ratio of sensible to latent heat flux, the Bowen ratio B, as a parameter specifying the wetness of the surface. The ratio of sensible to total heat transfer is then $B/(B+1)$ so that $\delta H_S = (1-A)[B/(B+1)]\delta S_0$. This formula is valid when the partition of radiant energy into sensible and latent heat occurs at a surface with a fixed temperature but it fails to allow for the change of Bowen ratio that would occur if the surface became cooler as a result of the decrease in irradiance.

A more rigorous expression for δH_S can be obtained from the formula which Monteith (1973) derived for the flux of sensible heat from an extensive surface of vegetation, viz.

$$H_S = \frac{\gamma(1+r_s/r_a)(R_N-G)-\rho c_p \delta e/r_a}{\Delta+\gamma(1+r_s/r_a)} \qquad (5)$$

where γ = psychrometer constant (mb K^{-1})
Δ = rate of change of saturation vapour pressure with temperature (mb K^{-1})
r_a = resistance to transfer of heat and water vapour between the surface and the atmosphere at height z (s m^{-1})
r_s = resistance to transfer of water vapour within the surface – a measure of surface wetness (s m^{-1})
G = downward soil heat flux (W m^{-2})
δe = saturation deficit of the atmosphere at height z (mb)

Assuming that the soil heat flux is constant, or is a negligibly small fraction of R_N, Eq. (5) may be differentiated to give

$$\delta H_S = \frac{\partial H_S}{\partial R_N} \delta R_N = \frac{\gamma(1+r_s/r_a)\delta R_N}{\Delta+\gamma(1+r_s/r_a)} \quad . \quad . \quad (6)$$

To evaluate δR_N, the dependence of net radiation on solar radiation is written as $\partial R_N/\partial S_0 = (1-A)$, an approximation which neglects the influence of surface temperature on longwave exchange.

From Eqs. (3) and (6) it can now be shown that

$$\delta H_S = \frac{\partial H_S \partial R_N}{\partial R_N \partial S_0} \delta S_0 = \frac{\gamma(1-A)(a+b-Ab')(1+r_s/r_a)S_0}{\Delta+(1+r_s/r_a)} \quad . \quad (7)$$

and this value of δH_S can be used in Eq. (4) to predict the net thermal effect of aerosol, δH.

For comparison with previous work on the radiative effects of aerosol at the same site (Unsworth and Monteith 1972), it was convenient to use a turbidity coefficient, τ_a. By using an empirical expression for relating the ratio of diffuse to total solar radiation from cloudless skies as a function of τ_a, it may be shown that τ_a is related to the linear absorption and back-scattering coefficients a and b by $a+b = \tau_a(m-k)$ where m is the air mass number and k is a constant (0·75) derived from the empirical expression which was valid for $1 \cdot 1 < m < 2 \cdot 0$. The separate portions of radiation absorbed and back-scattered were estimated by a method originally described by Robinson (1962) and b' was assumed equal to b. There is one important difference between this treatment and Mitchell's. Values of τ_a for anticyclonic weather in Britain are usually larger than 0·2 and consequently values of $(a+b)$ on most days were an order of magnitude larger than those assumed by Mitchell (see Table 1). Because the analysis was confined to values of m less than 2, the errors introduced by the linear coefficients a and b were less than 5%. For $m > 2$, it would be necessary to use more rigorous exponential functions for absorption and scattering.

TABLE 1. MEAN FLUXES OF SENSIBLE HEAT, RADIATION AND LATENT HEAT IN THE THERMAL BOUNDARY LAYER AND AT THE SURFACE ON FIVE DAYS ($W\,m^{-2}$). MEAN TURBIDITY, ABSORPTION AND BACKSCATTER COEFFICIENTS AND BOWEN RATIO ARE ALSO SHOWN. SYMBOLS CORRESPOND TO SECTIONS 2 AND 4 IN THE TEXT

Date	Period (GMT)	H	$H_{G,S}$	$H_{G,L}$	H_A	$H_{S,M}$	λE_M	τ_a	a	b	B
6 June 1973	0625–1615	130	20	−25	125	50	215	0·40	0·20	0·07	0·23
7 June 1973	0645–1910	120	20	−25	105	70	240	0·30	0·16	0·05	0·29
15 June 1973	0645–1910	140	20	−20	110	95	220	0·25	0·14	0·04	0·43
16 June 1973	0615–1615	175	25	−20	115	90	240	0·40	0·20	0·07	0·38
1 Aug. 1973	0615–1915	180	15	−30	135	160	120	0·35	0·22	0·06	1·33

3. EXPERIMENTAL

At Sutton Bonington (52·8°N 1·25°W), the temperature and relative humidity of the atmosphere from ground level to the top of the thermal boundary layer were measured with radiosondes on cloudless or almost cloudless days in 1972 and 1973. The sondes were released at intervals of about 3 hours from dawn to dusk and rose at about $3\,m\,s^{-1}$. The mean flux convergence between flights was calculated from the radiosonde records. The input of sensible and latent heat at the ground was estimated from measurements of net radiation and of the Bowen ratio over an extensive field of wheat representative of the cereal crops which occupy about 30% of the surrounding area (Williams 1971). A further 60% of the area is under grass but measurements of changes in soil water content in previous years showed that the difference in evaporation rate from the two types of cover was unlikely to exceed 20% except for a few weeks after harvest of the cereals when the stubble was dry. Assuming a mean Bowen ratio of 0·5, the uncertainty in estimating a mean areal value for the sensible heat input from measurements over wheat is of the order of $\pm 20\%$. The turbidity factor, τ_a, was estimated by comparing the direct beam irradiance measured with a Linke-Feussner radiometer with the value calculated for a model aerosol-free atmosphere. Values of $H_{G,S}$ were estimated from the results of Roach (1961b) and the longwave convergence, $H_{G,L}$, was derived from a radiation chart (Robinson 1950) using appropriate additional radiosonde records of temperature and humidity from the *Daily Aerological Record.*

4. Heat budget analysis

(*a*) *Radiation*

Table 1 shows values of the heat budget components for five days, chosen because there was no systematic change of temperature, during the day, above the thermal boundary layer. When this condition was satisfied, it was assumed that the air mass was uniform and that horizontal advection was negligible. The values of τ_a are characteristic of anticyclonic air with a long track over the continent (Unsworth and Monteith 1972). Values of H were determined from the temperature profiles over the periods shown. Solar zenith angles during these periods did not exceed 60° ($m = 2$), the limit of validity of the empirical radiation formulae. On 16 June there was scattered cloud for part of the day so H_A was weighted by the direct radiation recorded at the surface expressed as a fraction of the direct radiation calculated for a cloudless atmosphere with the same turbidity. This procedure is justified if absorption by aerosol is confined to the direct solar beam, an assumption supported by observations of Paltridge and Platt. The columns $H_{S,M}$ and λE_M show the measured inputs of sensible and latent heat from the wheat crop and B is the measured Bowen ratio $H_{S,M}/\lambda E_M$. Values of H_S determined as a residual from Eq. (3) depend on the value assumed for D and Fig. 2 shows the dependence of the ratio $(H_S - H_{S,M})/H_{S,M}$ on D for the five days.

The figure shows that $H_S = H_{S,M}$ for values of D between 0·25 and 0·7 consistent with values already quoted. Because of variability in the partitioning of sensible and latent heat over different land surfaces, H_S may be expected to differ from $H_{S,M}$ by at least $\pm 20\%$.

Adopting the mean value $\overline{D} = 0{\cdot}5$, the daytime heating rates in the thermal boundary layer correspondingly obtained by averaging the data in Table 1 are

		degC/day
Convergence of shortwave flux:	gases	1·1
	aerosol	3·3
Convergence of longwave flux		−1·3
Input of sensible heat from the ground		5·2
Net heating		8·3

According to these figures, radiative heating of the lower atmosphere by aerosol can account for 40% of net heating during the day.

(*b*) *Total heat*

The net thermal effect of aerosol on the lower atmosphere given by δH may now be estimated by applying Eqs. (4) and (7) to the records for the five days analysed in Table 1 and Fig. 2. Values of D for each day were taken from Fig. 2 when $H_S = H_{S,M}$. Assuming

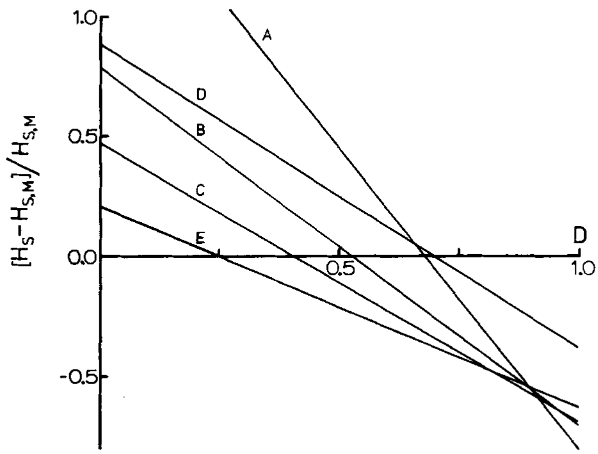

Figure 2. Dependence of the relative difference $(H_S - H_{S,M})/H_{S,M}$ between calculated (H_S) and measured $(H_{S,M})$ sensible heat fluxes at the ground on the fraction D of flux convergence due to aerosol in the thermal boundary layer.
A: 6 June 1973 B: 7 June 1973 C: 15 June 1973 D: 16 June 1973 E: 1 August 1973

a mean value $r_a = 0{\cdot}4$ s cm^{-1} appropriate for a mean wind speed of 2m s^{-1} over a cereal crop about 1m tall, the mean value of r_s for each day was calculated from the Bowen ratio (Monteith 1964). Values of r_s were consistent with values from Monteith, Szeicz and Waggoner (1965). The appropriate value of Δ was calculated from the mean air temperature; the reflection coefficient was taken as 0·20, and $\gamma = 0{\cdot}66$mb degC^{-1}. Table 2 shows the values of parameters used, and the corresponding calculated values of δH based either on the measured value of D or on the mean value $\bar{D} = 0{\cdot}5$.

TABLE 2. QUANTITIES USED IN EQS. (4) AND (7) TO CALCULATE δH FOR VALUES OF D SHOWN IN THE TABLE AND FOR A MEAN VALUE $\bar{D} = 0{\cdot}5$

Date	D	r_s (s cm^{-1})	H_A (W m^{-2})	$\delta H(D)$ (W m^{-2})	$\delta H(\bar{D})$ (W m^{-2})
6 June 1973	0·68	0·5	125	35	12
7 June 1973	0·53	0·5	105	9	7
15 June 1973	0·41	1·1	110	−11	−1
16 June 1973	0·70	0·9	115	24	0
1 Aug. 1973	0·26	2·6	135	−51	26

Table 2 demonstrates that, during daylight, aerosol may cause net heating or net cooling of the lower atmosphere at rates comparable with the cooling caused by longwave flux divergence (Table 1). Variation in $\delta H(D)$ between days is caused by variation in D and r_s. The uncertainty in D is about $\pm 0{\cdot}2$, and the column $\delta H(\bar{D})$ shows more clearly the influence of r_s on δH when a value $\bar{D} = 0{\cdot}5$ is used. As the crop matured, r_s increased and δH changed from a heating term to an appreciable net cooling term. The influence of r_s on δH for a single day is shown in Fig. 3 which illustrates the dependence of δH on r_s for 7 June 1973 taking $a = 0{\cdot}16$, $b = 0{\cdot}05$ (Table 1) and $D = 0{\cdot}53$ (Table 2). The point corresponding to the measured value over wheat ($r_s = 0{\cdot}5$ s cm^{-1}) is shown on the figure. Mitchell emphasized the importance of the ratio of aerosol absorption to back-scattering, a/b, in determining δH. The value of τ_a on 7 June corresponded to $a/b = 3{\cdot}2$ (Table 1), and to illustrate the influence of back-scattering, Fig. 3 also shows δH calculated for $a/b = 1{\cdot}6$ and $6{\cdot}4$, assuming that

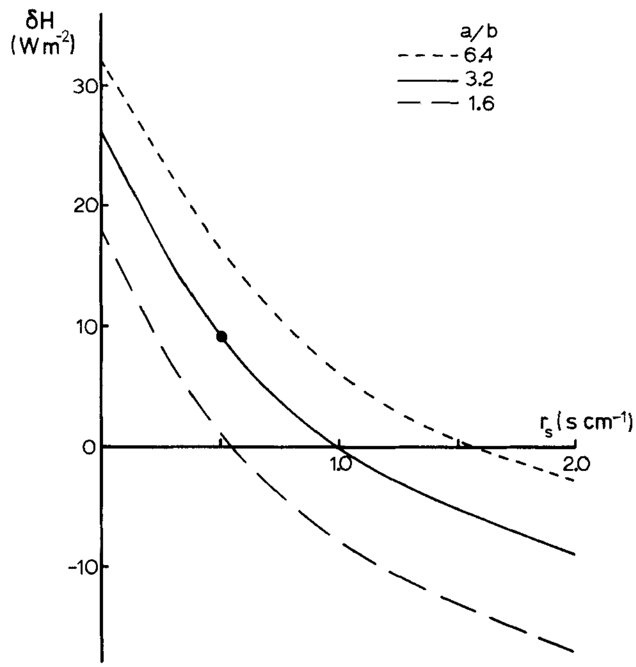

Figure 3. Dependence of the net heat flux δH ascribed to aerosol on the surface resistance r_s and on the ratio of aerosol absorption and backscatter coefficients, a/b, for 7 June 1973. The point corresponding to the measured value over a wheat crop, $r_s = 0{\cdot}5\mathrm{s\ cm^{-1}}$ ($a/b = 3{\cdot}2$) is shown in the figure.

τ_a was unchanged. Mitchell's calculations covered a wider range of a/b but he concluded that typical values were probably of the order of unity.

5. Conclusions

Although a number of approximations were needed to assess the radiative and net thermal effects of aerosol, none of these affect our general conclusions, and refinements could readily be introduced. The salient feature of the analysis is that the surface resistance is treated as a conservative quantity, independent of the income of radiant energy at the ground. Mitchell's assumption that the Bowen ratio is conservative will be satisfied only in exceptional cases.

When aerosol is present in stable air masses, absorption of radiation will often be a significant, and may sometimes be a dominant, term in the radiative balance of the lower atmosphere. For cloudless summer days in southern England, a mean rate of absorption of $60\mathrm{W\ m^{-2}}$ implies that the maximum figure in very turbid air may exceed $100\mathrm{W\ m^{-2}}$. The convergence of heat attributable to aerosol is therefore comparable in importance with the entrainment of heat through the inversion as estimated, for example, by Cattle and Weston (1973, 1975) and by Rayment and Readings (1974). As Cattle and Weston neglected the absorption of shortwave radiation, the values which they quote for the input of sensible heat at the ground are probably too large by at least 30% and possibly by as much as 50%. The corresponding value of the Bowen ratio, 2·5 for the midday period of 25 March 1972, seems to have been overestimated too. From experience and by calculation from Eq. (5), we should not expect the ratio to exceed unity.

The wider implications of this analysis are that over the oceans and other extensive water surfaces the presence of aerosol in the lower atmosphere will increase the net input of heat below an inversion. Farmland and forest will behave in the same way provided the foliage is green and there is adequate water for transpiration (say $r_s < 1$ s cm^{-1}). Over a very dry area such as a desert or an extensive conurbation, aerosol will decrease the net input of heat. Vegetation which is short of water or covered with dead leaves will tend to behave in the same way.

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