

Longwave radiation at the ground: IV. Comparison of measurement and calculation of radiation from cloudless skies

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SUMMARY

On cloudless days in summer, longwave irradiance was measured with a radiometer and was calculated from a radiation chart using records from radiosondes released at regular intervals. The results show that the chart underestimated the longwave flux density at the ground by up to 40 W m^{-2} , corresponding to an extra atmospheric emissivity of up to 0.12. The extra emissivity was correlated with turbidity during the day, and on turbid days it showed a diurnal variation with a maximum at noon. It is suggested that emission in the atmospheric window from dust, pollen and spores may account for some of the observed excess.

1. INTRODUCTION

In the absence of cloud, longwave radiation received at the ground from the atmosphere can be accounted for largely by emission from water vapour and polyatomic gases, mainly carbon dioxide and ozone. The importance of emission from suspended solid material has been debated for many years but no clear picture of the magnitude and variability of such a flux has emerged. Consequently aerosol emission is not included in any radiation charts or computational schemes which are important methods for calculating longwave radiation over oceans and remote regions where direct measurements do not exist (Swinbank 1963). It is therefore necessary to know the accuracy with which longwave radiation at the surface can be calculated from measurements in the atmosphere and to identify any systematic deviations. This was an aim of the work described in this paper.

Robinson (1947, 1950) made a detailed study of longwave radiation at Kew, a site close to London, at a time when smoke pollution was much greater than at present. He compared direct measurements of the hemispheric flux density at the ground, L_d , with calculations based on radiosonde records from sites 100 and 140 km from Kew, using charts designed by Elsasser (1942) and by himself (Robinson 1947, 1950). Measurements exceeded calculations by an amount ranging from $+40$ to -10 W m^{-2} (mean $+16 \pm 5 \text{ W m}^{-2}$) for the Elsasser chart and from $+10$ to -20 W m^{-2} (mean $+0.4 \pm 6 \text{ W m}^{-2}$) for the Kew chart. Robinson considered that there were three reasons for the differences. First, emissivity was slightly overestimated in the Kew chart for optical water paths in the range 0.005 to 5 cm. Second, the dependence of emissivity on temperature was ignored in the Kew chart and was probably incorrectly specified by Elsasser. Third, there was an additional component of radiation ranging from about 0 to 25 W m^{-2} which was not allowed for in either chart. Analyses of the angular distribution of longwave radiance revealed that on many occasions the excess radiation may have originated from low-level dust, but there were also times when undetected high cloud appeared responsible.

There appear to have been no other studies on this scale but two reports of longwave radiation when the atmosphere was very turbid gave conflicting results, perhaps because of differences in properties and sizes of particles. Paltridge and Platt (1973) measured shortwave and longwave fluxes from an aircraft in dense haze. Although there was considerable convergence of shortwave flux in the haze layer they were unable to detect differences between longwave fluxes as measured, and those calculated from temperature and humidity

profiles assuming an aerosol-free atmosphere. However, uncertainties in the calculations were large (up to $\pm 40 \text{ W m}^{-2}$). Separate measurements of atmospheric emissivity at $11 \mu\text{m}$ agreed well with measurements made on days when there was no haze. Idso (1972) measured L_d during a dust storm and found increases from normal of about 12–13% which he attributed to emission from the dust.

Rouse *et al.* (1973), who measured L_d in a city and on its outskirts, found that atmospheric emissivity over the city was about 20% larger during the daytime. Aircraft flights showed that the increase was not associated with higher air temperatures over the city and Rouse concluded that there was significant emission from particles.

Most theoretical studies of effects of aerosols have emphasised effects on transfer of solar radiation and have neglected effects on longwave exchange (Luther 1976). Models developed by Atwater (1971a, b, 1977) and Luther (1976) include longwave exchange and enable aerosol effects on thermal structure, e.g. in urban atmospheres, to be simulated. However, Atwater (1971b) stressed that lack of data on aerosol absorption and scattering coefficients precluded a complete examination of the accuracy of his model in specific situations. Ackerman (1977) showed that effects of aerosols on L_d were significant only for emission in the atmospheric window. Based on aerosol distributions measured over Los Angeles, his calculations show that 'average' aerosol increased L_d by about 14 W m^{-2} but that increases exceeding 60 W m^{-2} occurred when aerosol was dense. The extra radiation attributed to aerosol has two sources: first, emission from particles by virtue of their temperature and emissivity produces a flux of radiation at the ground which is not allowed for in the charts; and second, absorption of shortwave energy by aerosol in the day causes heating of the atmosphere. This source of extra radiation is accounted for in all radiation charts provided that the appropriate temperature profile is measured. Consequently Ackerman's values for the increase in L_d due to aerosol are larger than the excess radiation found by comparing measurements with calculations from charts.

In the first paper of this series we analysed measurements of longwave radiance at Sutton Bonington to show that there were many occasions in this rural area when atmospheric emissivity was larger than Robinson's minimum values at Kew. The origin of the excess radiation could not be investigated because there were no local radiosonde records. In this paper we analyse a series of measurements of L_d at regular intervals on cloudless days in summer when radiosondes were released simultaneously from the same site.

2. THEORY

(a) *Atmospheric emission*

The principles of radiation charts have been discussed by Sellers (1965) and Kondratyev (1969). Essentially all charts make use of empirical and/or theoretical relationships between the isothermal emissivity of a slab of radiating material and the optical thickness of that material. The effect of pressure on the width of the absorption lines of water vapour is allowed for, and the variation of emissivity with temperature is sometimes included. Staley and Jurica (1970) recently calculated the emissivities of homogeneous slabs of water vapour, carbon dioxide and ozone as functions of optical thickness and temperature. Their results correct an arithmetical error in the tables of Elsasser and Culbertson (1960) and they also include tabulations of the correction for overlap of H_2O and CO_2 bands. Yamamoto (1952) based his radiation chart on similar calculations which agree reasonably well with those of Staley and Jurica. Robinson (1947) accepted Elsasser's laboratory values (1942) for water vapour emissivity at small optical paths, but used observed minimum values of atmospheric emissivity at Kew for larger paths. His values of atmospheric emissivity at Kew are

from 2 to 14% smaller than those of Staley and Jurica at optical water paths of 0.1 to 2.0 cm. In the present analysis, values of emissivity taken from the tables of Staley and Jurica were used in a numerical form of the radiation chart described in Robinson (1947).

(b) *Aerosol emission*

Emission from aerosols is likely to influence the radiation budget at the ground only in the atmospheric window, 8 to 13 μm (λ₁ to λ₂). At the ground, the flux density of radiation emitted by a layer of aerosol at temperature *T* is given by

$$L_p = \int_{\lambda_1}^{\lambda_2} \epsilon(\lambda) B(\lambda, T) d\lambda \quad (1)$$

where $\epsilon(\lambda)$ is the emissivity of the layer and $B(\lambda, T)$ is the Planck distribution. For aerosol at low levels, the term $\int_{\lambda_1}^{\lambda_2} B d\lambda$ expressed as a fraction, *f*, of blackbody radiation is almost independent of air temperature, *T*, and so Eq. (1) may be written, integrals being from λ₁ to λ₂,

$$\begin{aligned} L_p &= \frac{\int B d\lambda \int \epsilon(\lambda) d\lambda}{\sigma T^4 \int d\lambda} \sigma T^4 \\ &= f \bar{\epsilon} \sigma T^4 \end{aligned} \quad (2)$$

The value $\bar{\epsilon}$ is the mean emissivity of aerosol averaged over the atmospheric window, and $f\bar{\epsilon} = L_p/\sigma T^4$, is the effective emissivity of the atmosphere for particles. The value of *f* ranges from 0.31 to 0.33 in the temperature range 10 to 30°C.

Evaluation of Eq. (2) requires knowledge of $\epsilon(\lambda)$, which from Kirchoff's Law is identical to the absorptivity, $a(\lambda)$. For the sizes of particle and for the wavelengths considered here, scattering may be neglected so that $\epsilon(\lambda) = a(\lambda) \approx 1 - t(\lambda)$ where $t(\lambda)$, the transmissivity of an aerosol layer depth *z*, is given by

$$t(\lambda) = \exp(-K_a(\lambda)Mz) \quad (3)$$

with $K_a(\lambda)$ the mass absorption coefficient (m²kg⁻¹) and *M* the mass concentration of aerosol (kg m⁻³).

There are few measurements of $K_a(\lambda)$ or $K_a(\lambda)M$ and apparently none in Britain. Volz (1972) and Fischer (1975) studied urban aerosol over Mainz; their mean value of K_a over the atmospheric window was about 10²m²kg⁻¹. Quoted values of *M* in Britain are ~100 μg m⁻³, but large particles with diameters exceeding 1 μm may have been excluded

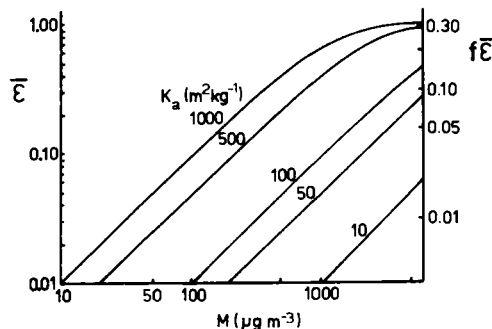


Figure 1. Dependence of the mean emissivity, $\bar{\epsilon}$, of aerosol in the atmospheric window and of the effective emissivity, $f\bar{\epsilon}$, of the atmosphere for particles, on the aerosol mass absorption coefficient, K_a , and on aerosol mass concentration, *M*.

from the measuring systems. Maximum values of M in turbid environments are 500–1000 $\mu\text{g m}^{-3}$.

The distribution of aerosol with height may be very variable (Junge 1963). To make an estimate of the emissivity of aerosol we will assume a uniform distribution to 1 km. Fig. 1 shows the dependence of $\bar{\varepsilon}$ and $f\bar{\varepsilon}$ on M and K_a .

The apparent emissivity of the atmosphere from a radiation chart is $\varepsilon_a = L_{dc}/\sigma T^4$, where L_{dc} is the computed longwave flux density at the ground. The *measured* apparent emissivity when aerosol emission is present would be

$$\varepsilon_m = (L_{dc} + L_p)/\sigma T^4 = \varepsilon_a + f\bar{\varepsilon} \quad (4)$$

The fractional excess emissivity is therefore given by

$$\gamma = f\bar{\varepsilon}/\varepsilon_a \quad (5)$$

3. INSTRUMENTATION AND METHODS

At Sutton Bonington (52.8°N 1.25°W) a Linke–Feussner pyrheliometer was used to measure the incoming longwave radiation from the atmosphere under cloudless skies during the period May–August 1975. The method of measurement and integration to determine L_d was as described by Unsworth and Monteith (1975). During most days the pyrheliometer was also used to determine turbidity by the method of Unsworth and Monteith (1972).

On the few occasions when measurements with the Linke–Feussner were not possible, L_d was measured with the instrument described by Dalrymple and Unsworth (1978). In these cases turbidity was found from contemporary measurements of global and diffuse solar radiation with Kipp solarimeters.

The pyrheliometer was calibrated for shortwave radiation before the measurements against a sub-standard Kipp solarimeter at Sutton Bonington, and after the measurements against an Ångström pyrheliometer at the Meteorological Office, Bracknell. The calibrations agreed within 1%. Unsworth and Monteith (1975) showed that the calibrations for shortwave and longwave radiation agreed within 2%.

Radiosondes (Graw, Model H50 – low-level version) were used to obtain profiles of temperature and relative humidity in the lowest 1.5 km. Before use, the hygrometer elements were conditioned for 48 hours at high humidity and then the manufacturer's calibrations for all sensors were checked in the laboratory against a Fortin barometer and an Assmann psychrometer. Profiles of temperature and humidity above 1.5 km were found from Meteorological Office ascents from Crawley, as experience with simultaneous ascents from both sites showed that above this height variations of profiles with space and time were small and had insignificant effects (<1%) on calculations of L_d .

The Graw sondes ascending at about 3 m s^{-1} transmitted measurements of pressure and humidity at about 4 mb intervals. Thus there were about 50 sets of readings from the surface to about 800 mb and these were augmented with about 10 sets of readings from Crawley. Mixing ratios for water vapour were calculated for each set and the optical water path was calculated, making a square root pressure correction (Robinson 1947; Kondratyev 1969). Values of water vapour emissivity were determined from tables (Staley and Jurica 1970) – emissivities for carbon dioxide and ozone being taken as fixed quantities (Staley and Jurica 1972). Longwave radiation was then calculated by a computer program based on the scheme described by Robinson (1947, Table IV) which divided the atmosphere into layers and calculated the product of the effective emissivity for each layer with the blackbody radiation at mean layer temperature.

Robinson concluded that maximum errors in L_d at Kew computed from the measured

temperature and humidity distributions at rather distant sites, were about $\pm 3 \text{ W m}^{-2}$. This figure is unlikely to be exceeded in the present measurements because the sondes were released during the radiation measurements and from the same site. Moreover, they rose more slowly than standard Meteorological Office sondes and so gave more detailed information and they had sensitive humidity elements and pressure sensors designed for low-level operation.

On a few occasions during the day, and at night when radiosondes could not be released at Sutton Bonington, the optical water path was estimated from the Crawley sonde and from surface observations at both sites using an empirical method established by comparing previous simultaneous sondes from the two sites (Dalrymple 1977). Experience showed that, provided surface vapour pressure differed by less than 3 mb between sites, errors in L_d introduced by this method were less than $\pm 5 \text{ W m}^{-2}$.

A few measurements of L_d were made with the pyrliometer by one of us (G.D.) in September 1974 at Victoria, British Columbia, in conditions of very low turbidity. Radiosonde records from nearby stations in Canada and the United States were used to calculate incoming radiation.

4. RESULTS

Twenty-five sets of observations of L_d were made below cloudless skies at Sutton Bonington. Three observations were made at Victoria, British Columbia. On almost all occasions the measured apparent emissivity, ϵ_m (Eq. (4)), exceeded ϵ_a calculated from the appropriate radiosonde data. The excess emissivity was generally largest when ϵ_a was large, so it is appropriate to use the fractional excess emissivity, γ (Eq. (5)). Fig. 2 shows the

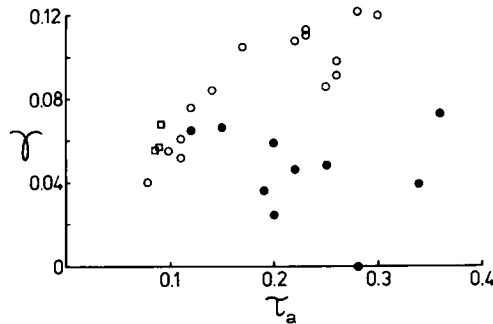


Figure 2. Variation of the fractional excess emissivity, γ , with aerosol turbidity, τ_a . Observations at Sutton Bonington (day - open circles; night - solid circles) and Victoria, B.C. (day - open squares).

dependence of γ on turbidity τ_a , assuming that the turbidity at night was the mean of the last measurement of τ_a on the previous evening and the first value on the following morning. The data are divided into measurements between 0700 and 1815 GMT; and measurements for the remainder of the day.

The relationship between γ and τ_a in the daytime is described by the linear equation $\gamma = c\tau_a + d$, where $c = 0.30 \pm 0.04$, $d = 0.03 \pm 0.01$ ($r^2 = 0.75$). The results from Victoria appear consistent with measurements in low turbidity at Sutton Bonington and so the regression excluding the observations in Victoria is not significantly different. Figure 2 shows that at night γ was small and independent of turbidity.

The diurnal variation of γ is shown in Fig. 3 where the data have been divided into two classes, depending on turbidity. When τ_a exceeded 0.15 there was a diurnal variation in γ , with a maximum between 11 and 15 GMT. Also shown in Fig. 3 are measurements made

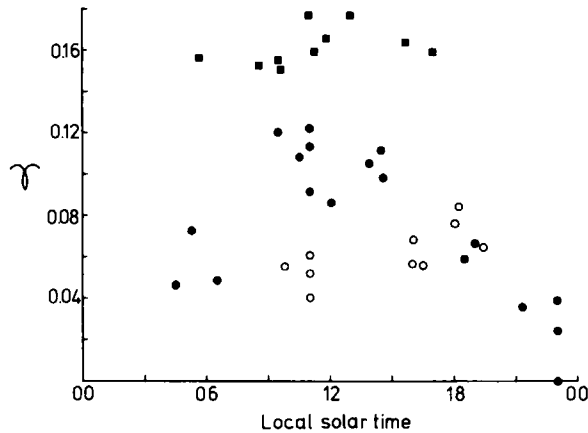


Figure 3. Diurnal variation of the fractional excess emissivity for two turbidity classes, $\tau_a < 0.15$ (open circles) and $\tau_a > 0.15$ (solid circles). Observations below thin cirrus are also shown (solid squares).

when there was cirrus present, showing that γ was then larger than on days of high turbidity, and exhibited no diurnal variation.

5. DISCUSSION

(a) *The origin of the excess radiation*

Figure 2 shows a consistent underestimate in L_d calculated from radiosonde records, and the magnitude of the discrepancy appears to be linearly related to turbidity during the day, but not at night. A number of possible sources of this difference will be considered, some due to systematic error and some to extra radiation sources.

(i) *Instrumental errors*: Different radiosondes were used to derive each point on Fig. 2, so a systematic trend in error is improbable. The pyrliometer maintained its calibration throughout the measurements.

(ii) *Tenuous high cloud*: This could account for the variation in Fig. 2, but examination of the angular dependence of longwave radiation by the method of Robinson (1947) provided no clear evidence of cloud. The diurnal variation of γ observed under cirrus was quite different from the usual form (Fig. 3), and variations in cloud density to explain the day/night differences in γ seem unlikely.

(iii) *Errors in the assumed relationship between emissivity and optical water path*: As discussed in section 2, there are considerable uncertainties in the emissivity of water vapour. Use of Robinson's emissivity data, which are consistently smaller than theoretical values, would alter the magnitude of γ in our results and would give a positive value of γ when $\tau_a = 0$ in Fig. 2; however, the apparent dependence of γ on turbidity would persist. The intercept in Fig. 2, which does not differ significantly from zero, is physically realistic and supports the use of Staley and Jurica's data.

Sasamori (1968) compared calculations similar to ours but based on the radiation chart of Yamamoto (1952), with more rigorous calculations of radiative transfer by the method of Rodgers and Walshaw (1966) and found good agreement in derived values of atmospheric cooling rates. It is therefore unlikely that errors in the radiation chart method account for the present discrepancy.

(iv) *Emission by water vapour self-broadening or by water dimer molecules*: Bignell (1970) measured in the laboratory the absorption of water vapour in the atmospheric window and found a component of absorption which increased with partial water vapour

pressure. The mechanism may involve water dimers (Burroughs *et al.* 1969) or self-broadening (Bignell *et al.* 1963). A positive correlation between turbidity and humidity would allow water vapour emission in the atmospheric window to explain our results partially. However, such an explanation does not seem consistent with the way the excess emission persists when Robinson's chart is used, because his chart was based on atmospheric measurements which must include the water vapour continuum.

Vittori *et al.* (1974) measured monochromatic extinction of solar radiation in the atmospheric window. After they had corrected their results for absorption by water vapour (Tomasi *et al.* 1974) there remained a component of absorption which varied diurnally; they attributed the effect to droplets or particles.

(v) *Emission from trace gases*: Increases in the optical depths for carbon dioxide and ozone from the mean values assumed in our calculations would increase γ . However the diurnal trend of γ is inconsistent with observed diurnal variations in carbon dioxide concentration at Sutton Bonington in summer. Natural variations in the optical path for ozone may change atmospheric emissivity by ± 0.02 only (Staley and Jurica 1972). The maximum ozone concentration from air pollution in the lower atmosphere in rural areas of England is about 150 p.p.b. (Atkins *et al.* 1972) and this would increase emissivity by only about 0.002.

(vi) *Particulate emission*: Maximum values of γ observed correspond to $f\bar{\epsilon} \approx 0.1$, and Fig. 1 shows that these would require denser concentrations of aerosol, greater mixing depths and/or larger absorptivities than have been reported elsewhere. Dense concentrations of aerosol seem initially inconsistent with observed values of turbidity, which were typical of settled weather in central England (Unsworth and Monteith 1972). However, turbidity is determined mainly by small particles ($< 1 \mu\text{m}$) whereas it is large particles ($\sim 10 \mu\text{m}$) that interact most effectively with the longwave radiation field. The day/night differences in Fig. 2 and diurnal variation of γ in Fig. 3 could be ascribed to emission from large particles kept airborne by convection during the day but removed from the atmosphere by sedimentation at night. For example, concentrations of spores (typical diameter $10 \mu\text{m}$) and pollen ($30 \mu\text{m}$) often vary diurnally in the necessary manner (Gregory 1973). Daily spore and pollen counts from a nearby urban area, Derby (Midlands Asthma and Allergy Research Association, Derby, pers. comm.), during the period of radiation measurements, combined with Gregory's observed mean diurnal variation, suggest that peak concentrations in Derby may have been $400 \mu\text{g m}^{-3}$ and presumably concentrations are larger in rural areas. There is apparently no information regarding the absorptivity of pollen in the atmospheric window, but the ratio of particle circumference to wavelength is relatively large so such particles may be optically active.

(b) Comparisons with other measurements

The typical scale of the excess radiation with $\gamma \approx 0.08$ is about 18 W m^{-2} . Robinson (1947, 1950) found that the measured radiation exceeded calculations from the Kew chart by amounts up to about 30 W m^{-2} but this maximum would be smaller if the calculations were repeated with the emissivities used here. There may be at least two reasons why we measured greater excess radiation than Robinson. First, his results were taken thirty years ago, near London, and it is probable that aerosols and trace gases are different in present-day rural atmospheres. Second, very few of his observations were made in the period May–August, when aerosol turbidity is largest, whereas all our measurements were in this period.

The large range of excess radiation in the present study is consistent with the range of slopes and intercepts in the analysis of the angular distribution of longwave radiation reported in the first paper of this series (Unsworth and Monteith 1975).

In their comparison of measured and calculated longwave radiation, Staley and Jurica (1972) concluded that there was an excess emissivity of about 0.03 which corresponds to $\gamma \approx 0.04$, but the measurements, which were made with a net radiometer over bare soil, assuming a surface emissivity of 0.90, may have had large errors.

6. CONCLUSIONS

Measurements and calculations show that longwave radiation below cloudless skies in summer often exceeded that predicted from a radiation chart (or similar computational scheme) by about 20 W m^{-2} . Emission from aerosol seems the most likely source of the excess. Current knowledge of the emissivity of aerosols is meagre, but emission from large concentrations of spores and pollen in rural areas in summer could explain the observed variations of excess emission with turbidity and with time.

Robinson (1977), reviewing a recent book on radiation in the atmosphere, remarked on the proliferation of results from mathematical models rather than from direct measurements. Models have made little progress in explaining the magnitude of aerosol emission mainly because insufficient is known of the optical properties of gases and solids, particularly in the atmospheric window. The origins of the excess radiation, first identified by Robinson thirty years ago, seem most likely to be revealed ultimately by experimental programmes combining measurements of longwave radiation at the ground with detailed studies of the composition and temperature of the lower atmosphere.

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