Estimation of Additional Downward Radiation from Aerosols over Large Cities

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1. Introduction

Atmospheric pollution over large cities has been increasing rapidly in recent years, and some meteorologists believe that a downward radiation from aerosols plays a rôle of preventing nocturnal cooling in large cities. In fact the presence of an additional downward radiation whose origin is other than H_2O and CO_2 was suggested by Robinson [1] by the analysis of radiation measurements made at Kew Observatory. However, later Yamamoto and Sasamori [2] showed that the observed values of downward radiation at Sendai were in good agreement with the values computed by the Yamamoto radiation chart which treats H_2O and CO_2 as radiator. They also showed that isothermal emissivity can not be determined uniquely from radiation measurements in the atmosphere by the Robinson or Brooks [3] method, that is, higher values of isothermal emissivity were obtained in cases of inversion than in cases of no inversion for the same amount of H_2O and CO_2 . In spite of our investigation, as Sendai is a middle sized city without appreciable sources of air pollution, there still remains a question, whether the atmospheric pollution over large cities is really ineffective to long wave radiation or not. In the present paper the author has examined the problem from an entirely different point of view.

2. Estimation of additional downward radiation from aerosols

First we shall consider the depletion of solar radiation through scattering by air

molecules, water vapour and aerosols. According to Linke [4] the intensity I of the direct solar radiation transmitted through the cloudless atmosphere is given by

$$I = I_0 \exp(-s\tau m), \quad (1)$$

where I_0 is the solar constant, m the air mass measured in terms of the zenith path as unity, s the extinction coefficient averaged over wave length of the pure air of a unit air mass, and τ the turbidity factor. If we write (1) in the following form;

$$I = I_0 \exp\{-(s+s')m\}, \quad (2)$$

then s' means the extinction coefficient of water vapour and aerosols in a unit air mass. Here by the term aerosols we mean all the impurities in the air involving condensation nuclei, smokes and dust particles. From (1) and (2) we have

$$s' = s(\tau - 1). \quad (3)$$

By some investigators the extinction coefficient s' is further separated into two parts, one due to water vapour and the other to dust. For instance, Fowle [5] found that the extinction coefficient of water vapour determined from observations was 64 times larger than that would arise from molecular scattering by H_2O molecules. Later the value 64 was reduced to 5 by Tien Kiu [6]. As regard to the explanation of these results, it is supposed that at least a part of the H_2O molecules occurs in the form of small droplets or clusters of molecules [7]. In the present research we will include such polymerized molecules or small droplets into the category of aerosols and will neglect the extinction due to pure water vapour molecules. In

other words we assume the extinction coefficient s' as due to aerosols alone.

According to Kimball [8] the intensity of direct solar radiation through the dry and dustless air of a unit path length ($m=1$) is given by $I/I_0=\exp(-s)=0.914$. From this we can derive the value of s to be 0.09. Next the value of τ for representative, clear and contaminated atmospheres in Japan are as shown in Table 1. As can be seen in the

Table 1. The annual mean values of τ . [9]

Station	Longitude	Latitude	τ
Tokyo	139°46' E	35°41' N	4.27
Osaka	135°30' E	34°39' N	4.46
Shimizu	133°01' E	32°43' N	2.82
Kakioka	140°11' E	36°14' N	2.83

table, the Osaka atmosphere is most contaminated in Japan. From (3) the value of s' for the Osaka atmosphere can be obtained to be 0.312.

Now let the total quantity of aerosols in a unit air mass be x (g cm^{-2}), then we have

$$s' = tx, \quad (4)$$

where t (g^{-1}cm^2) is the scattering coefficient of aerosols per unit mass with unit cross section. If the value of t is known we can estimate the total quantity of aerosols over Osaka from (4). Unfortunately the value of t for aerosols is not yet known, owing to the difficulty of measuring the amount of aerosols in an open air. Several years ago Yamamoto and Saito [10] measured the transmissivity of visible and infrared radiations through smokes in the laboratory. In this case we were able to measure the amount of smoke particles. The mean value of the mass scattering coefficient for several smokes was $7440 \text{ g}^{-1}\text{cm}^2$ in our experiment. With the use of this value we can derive from (4) the total quantity of aerosols over Osaka to be $x=4.19 \times 10^{-5} \text{ g cm}^{-2}$.

Now the mean absorption coefficient of smokes over infrared region measured in the above quoted experiment by Yamamoto and Saito [10] is $450 \text{ g}^{-1}\text{cm}^{-2}$. Using this value the transmission for slab τ_f by the aerosols

over Osaka is approximately given by

$$\tau_f = \exp(-1.66 \times 450 \times 4.19 \times 10^{-5}) = 0.969, \quad (5)$$

where the numerical factor 1.66 is due to Elsasser [11]. Equation (5) means that the isothermal atmosphere containing aerosols as much as the Osaka atmosphere absorbs only 3 per cent of the incident radiation. Because of the smallness of the absorption by aerosols an approximate estimation will be enough as to the additional downward radiation due to aerosols. If we use the Yamamoto radiation chart [12] the downward radiation D at the surface by H_2O and CO_2 in the atmosphere is given by

$$D = B(T_0) - \int_0^{B(T_0)} \{\tau_f(u, T) - \Delta\tau_f(u, u_{\text{CO}_2})\} dB, \quad (6)$$

where $B(T_0)$ is the black body radiation at the surface temperature T_0 , u , u_{CO_2} the optical paths of H_2O and CO_2 respectively and $\tau_f(u, T) - \Delta\tau_f(u, u_{\text{CO}_2})$ the mean transmission over wave length by H_2O and CO_2 . The downward radiation D' at the surface by H_2O , CO_2 and aerosols is approximately given by

$$D' = B(T_0) - \int_0^{B(T_0)} \frac{\{\tau_f(u, T) - \Delta\tau_f(u, u_{\text{CO}_2})\}}{\times \tau_f(u_{\text{aero}})} dB, \quad (7)$$

where $\tau_f(u_{\text{aero}})$ is the transmission by aerosols whose value is given above.

The annual mean temperature and humidity in Osaka are 15.3°C and 73 %, but their vertical distribution is not known, at least, to the author. According to London [13], however, the temperature and humidity at $30^\circ\text{--}40^\circ\text{N}$ belt during March are 13.0°C and 71 % at the surface and the vertical distribution of them is known in detail. Ignoring the minor differences we will assume that the data obtained by London can be applied to the Osaka atmosphere. Then the downward radiation by H_2O and CO_2 computed by the Yamamoto chart is $0.414 \text{ cal cm}^{-2} \text{ min}^{-1}$, and the additional downward radiation by aerosols over Osaka is given by the difference

of (7) and (6) to be $0.004 \text{ cal cm}^{-2} \text{ min}^{-1}$, being less than one per cent of the downward radiation by H_2O and CO_2 . The result of this computation suggests that, even in the contaminated air such as Osaka's the direct detection of the presence of the additional radiation by aerosols from observations of radiation will be difficult, owing to the expected errors of observations of radiation and of humidity in the free atmosphere.

3. Discussion on heat budget of large cities

Owing to a large extinction coefficient for visible light and a small absorption coefficient for infrared radiation of aerosols, the net effect of aerosols to solar and terrestrial radiations is to diminish the amount of radiation received at the surface of large cities and to cool the surface and lower layers in comparison with those of country sides. This is best illustrated in Table 2, in which the mean daily amounts of insolation received by 1 cm^2 of the horizontal surface at several stations in Japan are shown. Obviously the amounts received in large cities are much less than those received at small cities. In comparison with this large difference, the difference of the daily amounts of additional

radiation received at large cities and small cities would be negligible, being probably of the order of a few calories per day per cm^2 .

In spite of the large radiation deficit in large cities it is generally believed that the climate of large cities is warmer than that of suburbs or countries. Taking the circumstance for granted, one of the important causes of it would be, of course, the heat produced from industrial and domestic furnaces in large cities. For example, Osaka and its circumferential cities consume about 10 million tons of coal per year. Assuming one gramme of coal to produced 8000 calories of heat and the area of Osaka and its circumferential cities to be $4 \times 10^{12} \text{ cm}^2$, the mean production of heat by coal is about $55 \text{ cal cm}^{-2} \text{ day}^{-1}$. Although the value seems insufficient to compensate the radiation deficit in large cities, the inhabitants of Osaka use charcoal, petroleum and electric power, etc., other than coal, so that the total amount of heat produced artificially may be sufficient for the warming of air in large cities.

Besides, several meteorological factors favorable to the temperature rise in large cities are considered. One of them is the diminution of evaporation in large cities as already pointed out by Geiger [15]. In large cities a considerable part of precipitation will flow into a drainage system and will be ineffective to cooling of the surface by evaporation. If we assume that a half of the amount of precipitation in large cities is ineffective to evaporation, while all the amount is effective in the country sides, the difference of heat loss due to evaporation between Osaka, whose annual precipitation is 1274 mm, and its suburbs is about $103 \text{ cal cm}^{-2} \text{ day}^{-1}$, which is comparable to the radiation deficit in Osaka.

The difference of the surface condition between large cities and the suburbs further affects the microclimatology through the difference of cooling by long wave radiation. According to Falckenberg [16] the blackness to long wave radiation of the surface of sand

Table 2 The mean daily amounts over five days of insolation during February and August, 1953 [14]

Unit: $\text{cal cm}^{-2} \text{ day}^{-1}$

Station Date	Osaka	Kochi 133°33'E 33°34'N	Tokyo	Utsunomiya 139°52'E 36°33'N
31, Jan. — 4, Feb.	178	310	200	259
5, Feb. — 9, Feb.	291	412	240	278
10, Feb. — 14, Feb.	226	332	164	195
15, Feb. — 19, Feb.	213	330	250	293
20, Feb. — 24, Feb.	218	341	227	295
25, Feb. — 1, March	203	265	179	239
Mean	221	332	210	260
30, July. — 3, Aug.	383	495	327	403
4, Aug. — 8, Aug.	289	446	193	354
9, Aug. — 13, Aug.	314	362	338	333
14, Aug. — 18, Aug.	315	420	209	275
19, Aug. — 23, Aug.	321	515	282	338
24, Aug. — 28, Aug.	221	364	126	216
Mean	307	432	246	320

is about 89 per cent. The blackness of large cities whose surface is mainly made of concrete, brick, tile, etc., will be of the same order as that of sand. Whereas the blackness of the rural districts of Japan, the greater part of which is rice fields, will be about 95 per cent or more. If we assume these values of the surface blackness and the surface temperature of both the city and its suburb to be 15°C , the difference of surface cooling by long wave radiation will amount to about $49 \text{ cal cm}^{-2} \text{ day}^{-1}$.

Lastly the effect of city fog will be mentioned. Although aerosols themselves are nearly ineffective in preventing the nocturnal cooling as described above, they favour the formation of fog which prevents the cooling of the surface and lower layers of the atmosphere most effectively.

These meteorological factors will not necessarily act simultaneously or continuously, but we cannot deny that these will be as important in modifying the city climate as the heat produced artificially.

References

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