MEASURED AND CALCULATED LONGWAVE RADIATION FLUXES AND THEIR YEAR TO YEAR VARIATION AT MIZUHO STATION, ANTARCTICA

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Abstract: Together with measurements at Mizuho Station during POLEX-South, longwave radiation fluxes are calculated for the same measurement conditions. Comparing the measured and calculated downward longwave fluxes, good agreement is found for most months in 1979 and several months in 1980; however, large disagreements are seen for winter months in 1980.

The variation of longwave radiation between 1979 and 1980 is examined using measured and calculated fluxes. The measured downward longwave flux in the winter of 1980 was extremely large compared with the measured flux in 1979 and the calculated flux in 1980. This large amount cannot be explained only by the difference in temperature and water vapor amount. Calculated longwave fluxes also show differences between 1979 and 1980; but these can be explained by temperature and water vapor amount. Misinterpretation of cloud amount is suspected.

1. Introduction

Longwave radiation is one of the main components of the heat budget at the ground surface, together with shortwave radiation and sensible heat. Longwave radiation is very sensitive to the other meteorological parameters, such as clouds, water vapor amount and temperature. Measurements of longwave radiation were made at Mizuho Station (70°42′S, 44°20′E), Antarctica, during 1979 and 1981 within the radiation observation program of POLEX-South (Japanese POLEX program), and data reports were published by YAMANOUCHI *et al.* (1981a) and ISHIKAWA *et al.* (1982a).

Climatic conditions of 1980 at Mizuho and Syowa Stations (69°00′S, 39°35′E) have been discussed and compared to other years. Some eccentric features are found related to the flow out of the sea ice (KAWAGUCHI, 1983; ISHIKAWA and KOBAYASHI, 1983; ONO, 1984). It is of great interest to examine the variation of longwave radiation in 1980 compared with that in 1979.

In the present paper, calculations of longwave radiation fluxes are done for the same condition of measurements and fluxes are compared to those measured. Calculations are done using the simple wide band model of RAMANATHAN (1976). Then the difference in longwave flux between 1979 and 1980 is discussed using measured and calculated fluxes. From inspection especially of the downward longwave flux, several causes of the difference between 1979 and 1980 are inferred.

2. Measurements and Calculations

Measurements of longwave radiation fluxes were made using Eppley precision infrared radiometers (PIR). One set of two pyrgeometers upfacing and downfacing was mounted 1.5 m above the snow surface and another set was mounted at the top of a 30 m tower. Performance of the apparatus and method of measurements and calibrations have been reported in detail (YAMANOUCHI *et al.*, 1981b). Originally, data were sampled once a minute, but in the present paper, daily or monthly mean data are discussed (YAMANOUCHI *et al.*, 1981a; ISHIKAWA *et al.*, 1982a). Experimental uncertainties are estimated as $\pm 1\ \mathrm{W/m^2}$ for the daily mean.

Simple calculations of longwave flux were carried out using a spectrally integrated scheme derived by RAMANATHAN (1976). The water vapor 6.3 μm band, water vapor rotational band, CO_2 15 μm band, O_3 9.6 μm band, water vapor continuum absorption and overlaps of these bands were taken into account. Absorptions of the water vapor 6.3 μm and rotational ($\lambda > 15.15\ \mu$m) bands are expressed by the flux emissivity of STALEY and JURICA (1970) formulated by RAMANATHAN (1976) as

$$E(\bar{w}, T) = 0.59\left(\frac{T_0}{T}\right)^{1/4}\left(1 - \frac{1}{2}\sum_{i=1}^{2}\frac{1}{1 + A_i \bar{w}^{1/2}}\right), \qquad (1)$$

where

$$\bar{w} = \int \left(\frac{p}{p_0}\right)\left(\frac{T_0}{T}\right)^{1/2} \mathrm{d}w, \qquad (2)$$

is the effective water vapor amount, A_i are constants and w, T and p are water vapor amount, temperature and pressure, respectively. The suffix 0 denotes conditions at STP. Another rotational band ($15.15 > \lambda > 12.5\ \mu$m) is expressed in terms of the transmissivity of RODGERS and WALSHAW (1966). Water vapor continuum absorptions are expressed by flux emissivities based on BIGNELL (1970).

CO_2 15 μm and O_3 9.6 μm band absorptions are expressed by total band absorptances of RAMANATHAN (1976) and KIEHL and RAMANATHAN (1983) as

$$A = 2A_0(T)\sum_j \sum_i \tau_j \ln\left\{1 + \sum_k \frac{\bar{u}_k}{\sqrt{4 + \bar{u}_k(1 + 1/\beta_k)}}\right\}. \qquad (3)$$

The sum in the logarithm argument accounts for closely neighboring, totally overlapped sub bands. The sum over j accounts for bands which are separated but may be overlapped by the wings of the stronger bands. τ_j is the transmission function of the wings of the strongest bands in this region. A_0 is an effective bandwidth parameter, evaluated for a pressure weighted temperature along the path. The dimensionless optical depth $\bar{u}_k$ and line width parameter β_k for the k-th sub band are given as

$$\bar{u}_k = \int \frac{S_k/Q_v}{A_0(T)} \exp\left\{1.439\, W_k\left(\frac{1}{T_0} - \frac{1}{T}\right)\right\} \mathrm{d}u, \qquad (4)$$

$$\beta_k = \frac{4}{d_k \bar{u}_k} \int \gamma_k^0(T) \frac{p}{p_0} \mathrm{d}\bar{u}_k, \qquad (5)$$

where S_k, Q_v, W_k, d_k, γ_k^0 and u are the band strength, vibrational partition function, energy of the lower level of transition, mean line spacing, mean line half width and absorber thickness, respectively.

Three regions of band overlap are considered: water vapor continuum and rotational band; CO_2 15 μm, water vapor rotational and continuum; and O_3 9.6 μm band and water vapor continuum.

Errors of the calculation scheme are discussed in detail by RAMANATHAN (1976) and KIEHL and RAMANATHAN (1983). The maximum error is expected in the treatment of band overlap, and that is about 4% at most for the downward flux by the CO_2 15 μm band overlapped with the water vapor rotational band. So, it is small enough for the present sensitivity study.

As for the model atmosphere, the temperature and humidity profiles above the 700 mb level are derived from the monthly mean aerological data at Syowa Station (JAPAN METEOROLOGICAL AGENCY, 1981, 1982). Though Syowa Station is about 250 km away from Mizuho Station, the upper free atmosphere is the same at both stations on an average (KAWAGUCHI, 1982). This assumption is also supported from a comparison of low level sondes flown at Mizuho in 1980 (KAWAGUCHI *et al.*, 1984) with monthly mean aerological data at Syowa Station as shown in Fig. 1. Except for the inversion layer, the temperature of the upper layer at Mizuho lies around the mean temperature at Syowa Station. The temperature profile below the 700 mb level is determined from the surface temperature at Mizuho Station, using the typical shape of the inversion layer assumed from the low level sondes. The relative humidity

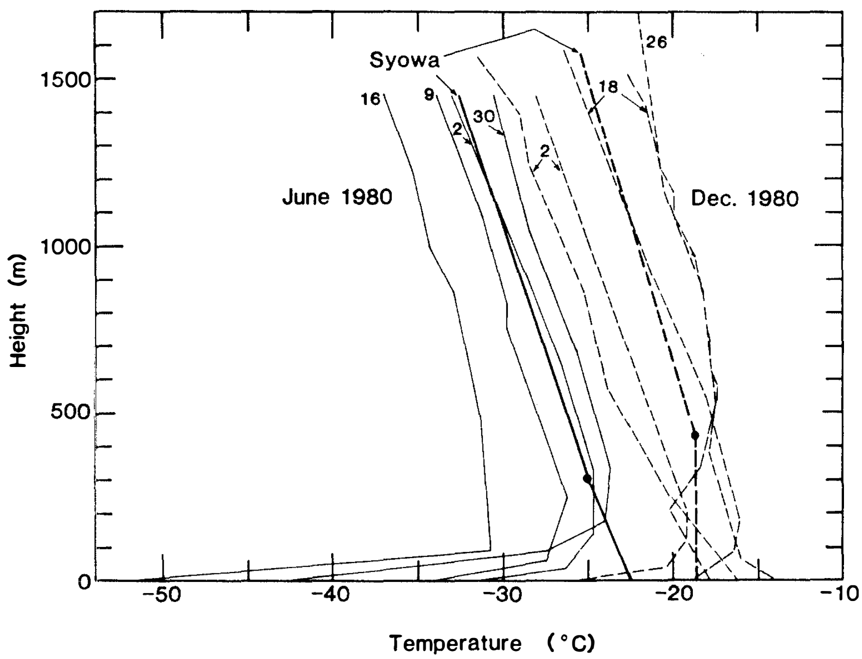

Fig. 1. Examples of temperature distribution at Mizuho (thin lines) and Syowa Stations in June and December 1980. At Mizuho Station observations are from low level sondes and at Syowa Station from monthly mean aerological data. Height is based on the surface elevation of Mizuho Station, 2230 m.

below the 700 mb level is interpolated between 100% at the surface and the value at the 700 mb level. These temperature and humidity profiles are not confirmed to be averages; however, since a minor difference in profiles will not make any noticeable difference in longwave fluxes as shown by YAMANOUCHI and KAWAGUCHI (1984), these profiles are regarded as typical for the purpose of the present paper. One of the examples of the temperature and humidity distributions are shown in Fig. 2a with the solid and dashed lines, respectively. Calculations are made only for the average, since we have no evidence that the temperature and humidity of the upper air at Mizuho coincide with those at Syowa in daily range. Daily values deviate greatly as seen in Fig. 1 owing to the synoptic scale disturbance. 30 levels are chosen for grid points, the top at 0.1 mb and the bottom at 730 mb assumed to be the surface level at Mizuho Station (2230 m). The thickness of layers is 50 mb in the middle part of the atmosphere from 100 to 700 mb; the near-surface part of the atmosphere between 700 and 730 mb is divided into 7 layers.

The downward flux L_d and upward flux L_u at level p for a particular band are calculated for example with total band absorptance A as

$$L_d = -\int_{p_t}^{p} B[T(p')]\,\mathrm{d}A(p' \to p), \tag{6}$$

$$L_u = \sigma T_s^{\,0} - \bar{B}(T_s)A(p_s \to p) - \int_{p_s}^{p} \bar{B}[T(p')]\,\mathrm{d}A(p' \to p), \tag{7}$$

where σ is the Stefan-Boltzmann constant, $\bar{B}(T)$ is the mean Planck function for the particular band at temperature T and p_t, p_s and T_s are the pressure at the top and at the surface and the temperature at the surface, respectively. The emissivity of the snow surface is assumed to be unity in this calculation for convenience (DOZIER and WARREN, 1982).

3. Measured and Calculated Fluxes

General descriptions of measured longwave radiation and discussions related to atmospheric temperature, water vapor amount, clouds and surface inversion were already done by YAMANOUCHI and KAWAGUCHI (1984) and by ISHIKAWA *et al.* (1982b). Since upward radiation directly and downward flux vaguely depend on surface temperature, net longwave radiation flux depends linearly on the strength of the surface inversion (YAMANOUCHI and KAWAGUCHI, 1984).

Vertical distributions of calculated fluxes for June 1979 for example are shown in Fig. 2b with the temperature and water vapor profiles used in the calculation in (a). In Fig. 2c, the heating rate of the atmospheric layer derived from the net longwave flux calculated (b) is shown as a reference. In the inversion layer, owing to the variation of the slope of the net flux, extremely large cooling of more than 10 K/day is seen in the upper part and large heating of about 6–7 K/day is seen near the surface. Though the peak values are not representative since they vary according to the grid selection, a general trend is well expressed.

Calculated downward fluxes are compared with measured fluxes for the clear sky in Fig. 3 for 1979 (a) and 1980 (b), respectively. Measured fluxes are monthly

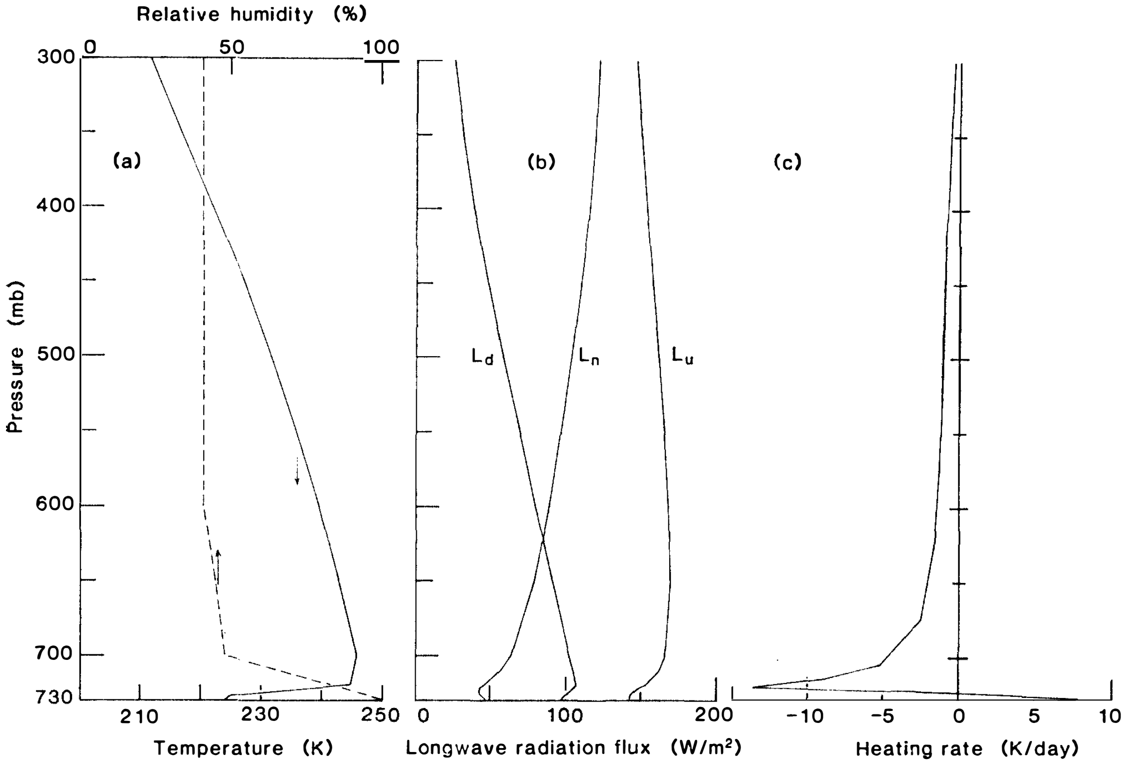

Fig. 2. Temperature (solid line) and humidity (dashed line) distributions of the model atmosphere in June 1979, for example, used in calculation (a), calculated longwave radiation fluxes (b) (L_d: downward, L_u: upward, L_n: net) and heating rate (c).

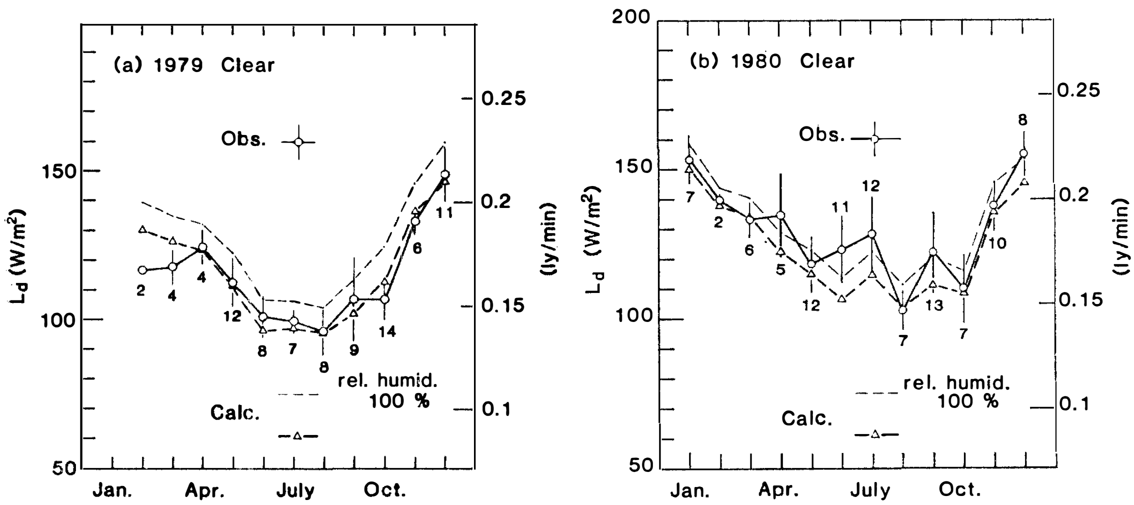

Fig. 3. Comparison of calculated and measured downward longwave radiation fluxes for a clear sky at Mizuho Station in 1979 (a) and 1980 (b). The thick broken line with triangles shows calculations for the same conditions as the measurements, and the thin broken line represents calculations for 100% relative humidity in the troposhere with other parameters remaining constant. Vertical bars for measured points are standard deviations of measurements. Numbers in the figure indicate the number of clear days ($n<1/10$) in each month.

averages only for clear days (cloud amount $n<1/10$). Numbers of clear days vary between 2 and 14 per month in both years as shown in Fig. 3. Fluxes are calculated for monthly average temperature and humidity distributions for whole days in each month, including clear, partly cloudy and overcast skies. However, only molecular absorption/emission is considered and results are regarded as for the clear condition. In the figure, calculated fluxes for 100% relative humidity in the whole troposphere are also shown as the maximum estimated fluxes, since some uncertainties in humidity cannot be excluded. Good agreements within standard deviation of measured amounts are seen after April in 1979 and January to March and after October in 1980. Calculated fluxes for February and March in 1979 are larger than measured fluxes. In February, since measurements were made only for two days at the end of the month, the measurement is not representative of the monthly average. In March, the measured amount is small assuming a normal seasonal variation. Another disagreement of calculated and measured fluxes is noticeable in winter in 1980. Calculated amounts in April, June, July and September are much smaller than measured, sometimes exceeding the standard deviation of measured values. Sometimes calculations for 100% relative humidity are yet smaller than measurements. Also characteristic is that standard deviations of measured fluxes are remarkably large from April to October in 1980.

4. Comparison of Longwave Radiation Flux between 1979 and 1980

Measured downward, upward and net longwave fluxes in 1979 and 1980 for a clear sky are compared in Fig. 4. In general, downward and upward fluxes for 1980 are larger than those for 1979, especially in winter. Within a clear sky the largest difference, about 30 W/m^2, is seen for the downward flux in July between 1979 and 1980. Net longwave fluxes have not so large difference between two years, because of compensations of both fluxes.

As the cause for this difference, atmospheric temperature is examined in Fig. 5. At Syowa Station, surface temperature in 1980 is extraordinarily high from March to September compared to 1979. It is also high compared to the long-term mean (ISHIKAWA and KOBAYASHI, 1983). This abnormally high surface temperature is owing partly to the flow out of sea ice around Syowa Station (KAWAGUCHI, 1983; ONO, 1984). At the surface at Mizuho Station, temperature does not show a large difference between the two years except in April and June to August. Temperatures in the upper layer are of great importance, however, we don't have monthly mean temperatures in the upper layer at Mizuho. The temperature at the 700 mb level at Syowa Station is examined. In contrast to that at the surface, temperature at 700 mb level shows quite a similar tendency with that at the surface of Mizuho. This is also the case for the higher layer up to 300 mb. Temperatures in June, July and August 1980 are higher than in 1979, which is said to be caused by the macro-scale circulation of air (KAWAGUCHI, 1983; ISHIKAWA and KOBAYASHI, 1983). This temperature difference in both years seems to support the difference in the downward longwave flux.

An effective emissivity defined as

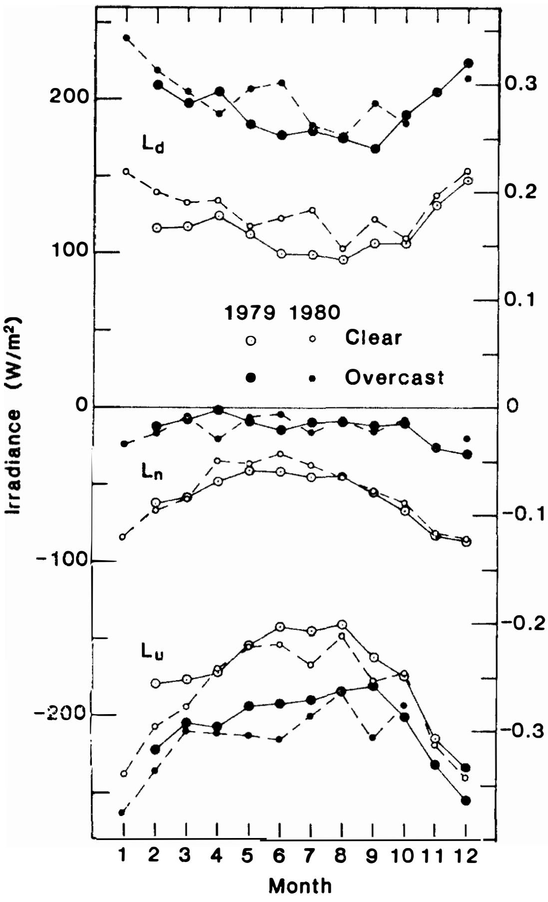

Fig. 4. Comparisons of downward (L_d), upward (L_u) and net (L_n) longwave radition fluxes of 1979 and 1980 for clear (cloud amount $n<1/10$) and overcast days ($n=10/10$, middle or low clouds).

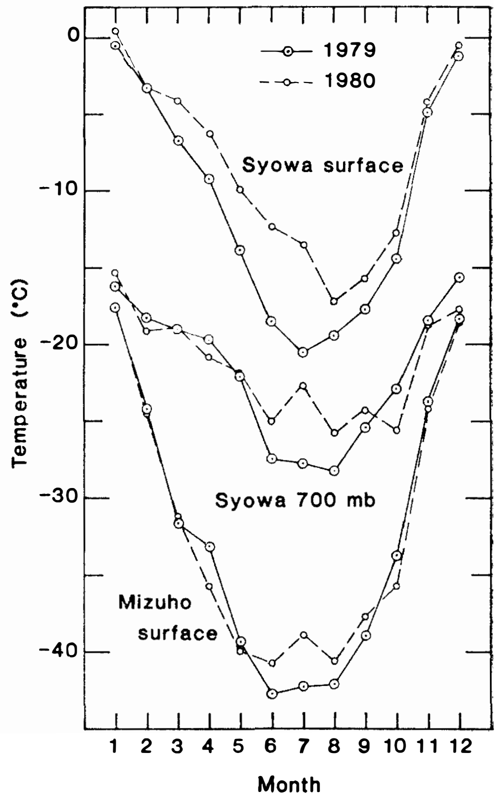

Fig. 5. Comparisons of monthly mean atmospheric temperature of 1979 and 1980 at the surface and 700 mb level at Syowa Station and at the surface of Mizuho Station.

$$\varepsilon^* = L_d/\sigma T_x^4 , \tag{8}$$

is introduced to separate the effect of the atmospheric temperature. L_d is the downward flux at the surface and T_x is the atmospheric temperature at the top of the inversion layer, for which the monthly mean temperature at the 700 mb level at Syowa Station was used. In Fig. 6, effective emissivities under a clear sky for 1979 and 1980 are plotted against T_x. Effective emissivities for the two years differ systematically about 0.05 (10%) on an average. Then, the difference in the downward longwave radiation flux between 1979 and 1980 cannot be explained only by the temperature difference.

Other meteorological parameters for the two years are compared in Fig. 7: the relative humidity at the 700 mb level at Syowa Station (a), column water vapor amount estimated for the condition at Mizuho Station (b) and cloud amount at Mizuho and Syowa Stations (c). It is apparent that the relative humidity in 1980 is 10% higher than that in 1979, and the column water vapor amount in 1980 is also

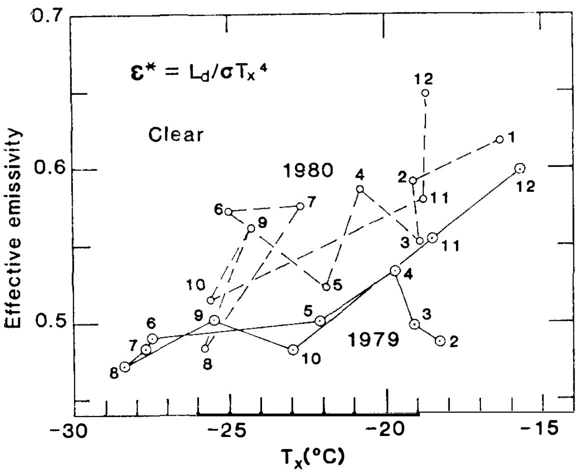

Fig. 6. Effective emissivity ε^* against 300 m level temperature T_x, taken to be the monthly mean temperature at 700 mb level of Syowa Station. Monthly means are for clear days ($n<1/10$) in 1979 and 1980. Small numbers indicate months.

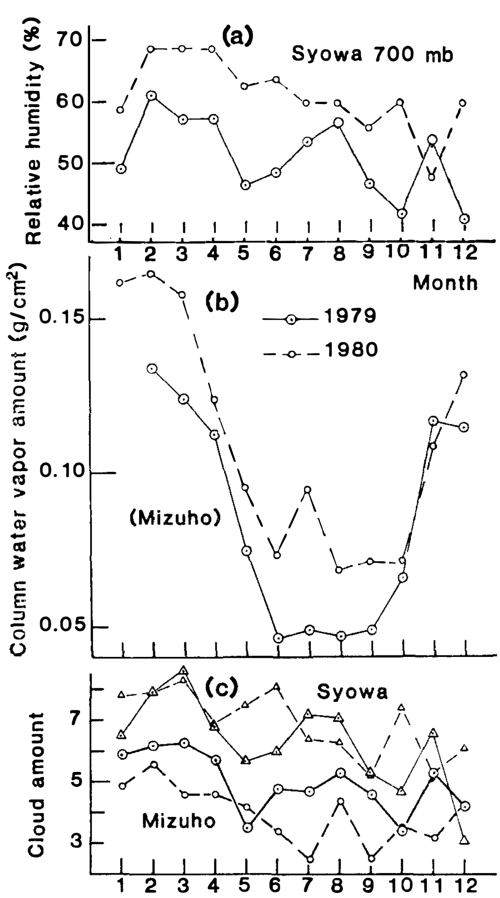

Fig. 7. Comparison of monthly mean relative humidity at 700 mb level at Syowa Station (a), column water vapor amount estimated for the model atmosphere at Mizuho Station (b) and cloud amount at Syowa and Mizuho Stations between 1979 and 1980 (c).

0.02 to 0.05 higher than in 1979 from June to September. The maximum difference in the column amount is seen in July when the amount in 1980 is about twice as that in 1979. As for the cloud amount, slightly larger values are seen for 1980 at Syowa Station; however, at Mizuho Station, values for 1980 are smaller than for 1979 for all months except May, October and December. These small cloud amounts in 1980 disagree with the large water vapor amounts in (a) and (b) of Fig. 7.

From the calculated fluxes, an effective emissivity at 700 mb defined as

$$\varepsilon^{*\prime} = \frac{L_{d700}}{\sigma T_x^4}, \tag{9}$$

is plotted against the column water vapor amount above the 700 mb level in Fig. 8. $L_{d\,700}$ is the downward longwave flux at 700 mb, assumed to be the downward flux above the inversion layer. In Fig. 8, $\varepsilon^{*\prime}$ can be approximated by a single curve within very small deviations (1%). No essential difference is seen between points

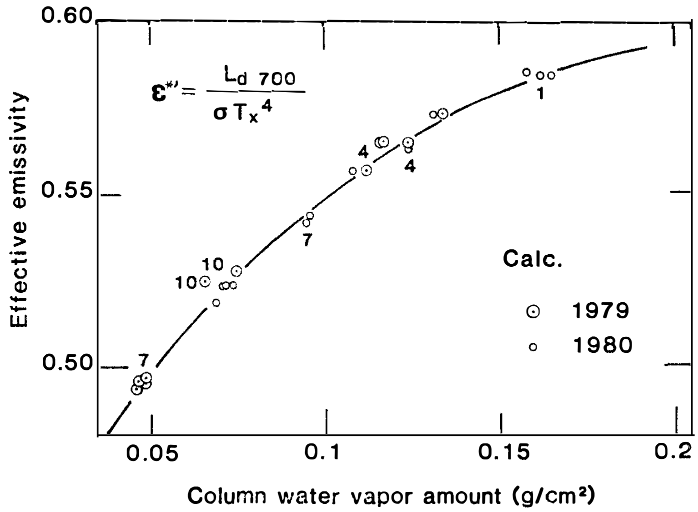

Fig. 8. Effective emissivity estimated at the 700 mb level against column water vapor amount. Monthly means are for clear days ($n<1/10$) in 1979 and 1980. Numbers indicate months of the year.

for 1979 and 1980. Without the effect of surface inversion, calculated downward fluxes can be uniquely explained by the temperature and column water vapor amount.

In Fig. 9, the effective emissivity defined at the surface ε^* (eq. (8)) for calculated and measured longwave radiation fluxes is plotted against the column water vapor amount. Different from Fig. 8, these ε^* include the effect of surface inversion. ε^* for the calculated flux shows hysteris-like curves, with no essential difference between 1979 and 1980, though ε^* has two values for one particular water vapor amount—small in autumn to winter and large in spring to summer. The difference in ε^* for the same column water vapor amount is due to the difference in the strength of the inversion. When the inversion becomes stronger, ε^* becomes smaller. This tendency is already known from Fig. 2 of YAMANOUCHI and KAWAGUCHI (1982), which shows the Ångström ratio. The Ångström ratio is directly related to the atmospheric emissivity ε by

$$A_0 = 1 - \frac{L_d}{L_u} \approx 1 - \frac{L_d}{\sigma T_s^4} = 1 - \varepsilon. \tag{10}$$

For the calculated flux, the difference between 1979 and 1980 can thus be explained by the difference in temperature and water vapor amount.

The measured results sometimes show extraordinary values, small values for February and March of 1979 and large values for June, July and September of 1980. Since the column water vapor amounts are monthly means for all days and ε^* is only for clear days, the data source is not strictly the same. The water vapor amount might be smaller for clear days. Then the small ε^* for February and March in 1979 can be partly explained. No reasonable explanation for the large ε^* for June, July and September in 1980 is obvious. One possibility is the drifting snow, which will increase the downward flux. It is certain that the mean wind speed is stronger in 1980 than in 1979 and stronger wind will cause a stronger drifting snow. However,

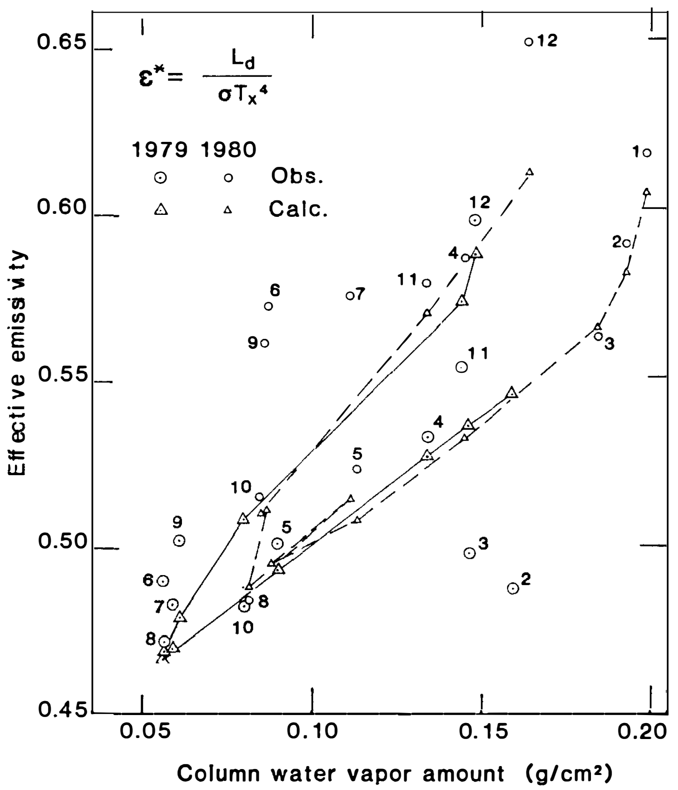

Fig. 9. Effective emissivity ε for measured and calculated radiation against column water vapor amount. Monthly means for clear days (n<1/10) in 1979 and 1980. Numbers indicate months.*

there is no apparent correlation between the average wind speed for the clear sky and the difference of the measured ε^* and measured downward longwave flux from the calculated flux. Another possibility is that there were some clouds in the sky which was taken for to be clear in the surface observation because of the darkness of the polar night. If the lowest measured value, namely, (the monthly average)—(standard deviation), for these months in 1980 is considered as the monthly value, the effective emissivity would coincide with the emissivity in 1979 and also with the calculated emissivity. This misinterpretation is suspected because the observed cloud amount in 1980 is extremely small judging from the large water vapor amount in the atmosphere. This is also assumed from the large standard deviation for the measured downward flux under a clear sky in winter 1980 (Fig. 3). Clouds affect the downward longwave flux greatly. A thick cloud of amount 10/10 will increase the downward flux about 80 W/m^2 (YAMANOUCHI and KAWAGUCHI, 1984). So, even thin clouds or cloud-like matter (fog and so on) have the possibility to increase the downward flux.

On account of many assumptions adopted in the calculation, for instance, temperature and humidity distribution of the upper atmosphere, the comparison between the measured and calculated longwave fluxes is accompanied by some uncertainties. It is difficult to draw a definite conclusion. More precise observations of clouds, which are important sources of downward radiation, should be made in the future.

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