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EVALUATION OF TIROS INFRARED DATA

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In order to obtain useful information from the satellite infrared data, it is imperative to know the physical significance of such measurements relative to the problems of climatology and synoptic meteorology. Although from the preliminary analyses of the radiation data collected by Tiros II and III several groups of workers have recently demonstrated the utility of such measurements in meteorology (Nordberg etal, 1962; Bandeen etal, 1961; Wark and Hanel, 1961; etc), several problems are, however, yet to be solved before the wealth of the radiation data now available can be applied to study the atmospheric processes in time and in space. We have attempted to find an answer to two of these questions which are of particular interest to climatology.

It is known that much of the driving force for the meridional circulation of the atmosphere is provided by the net differential heating of the earth as a function of latitude. Theoretical attempts have been made during the last few years to calculate the net radiation balance of the northern hemisphere as a function of latitude and season (Houghton, 1954; London, 1957; Ohring, 1958; Davis, 1961; etc.). Now with the availability of the Tiros radiation data, it has become possible to verify these ideas by the actual measurements of the total outgoing radiation flux for the northern hemisphere and obtain effective values of this flux for the different latitudes of the southern hemisphere as well.

Analysis of Tiros II Data.

The Tiros II contains five infrared censors which respond to radiation in five different spectral regions. They have been described in detail by Bandeen, Hanel, Licht, Stampfl and Stroud (1961). With the view of obtaining the latitudinal distribution of total outgoing radiation for a particular season we have used the data from Channel IV (7-32μ) which corresponds to approximately 80% of the total outgoing radiation from the earth. As Tiros II was launched on November 23, and since the data of Channel IV degraded after the first 650 orbits (Tiros II radiation data, Users Manual, 1961) the analysis was confined to the first 45 days of the observations.

Tiros II satellite has measured the outgoing infrared radiation between 50°N and 50°S. Further, each one of the measurements made by the satellite corresponds to a particular nadir angle, namely the angle between the line joining the satellite to the subsatellite point on the earth and the direction of view of the radiation censor in the satellite. For this analysis only the measurements made at nadir angle $\angle 25^{\circ}$ were utilized. This limitation of the range of nadir angle eliminates the correction due to the limb darkening effect (Wark and Yamamoto, 1961) which is necessary for measurements made at higher nadir angle.

In the present investigation the Final Meteorological Radiation Tapes supplied by the Goddard Space Flight Center were programmed to give an average of all the measurements in each 5° longitude by 5° latitude grid between 50° N and 50° S. Since there are several tens of thousands of satellite observations in each of these grid points it is assumed that the average of these readings yields a good estimate of the mean outgoing radiation for that area. There are, however, some regions on the globe for which no data is available, viz. between 30° to 50° S and 80° to 100° W in the southern hemisphere and its antipode in the northern hemisphere. Surrounding these regions the available satellite coverage is very poor; these areas have therefore been excluded in taking the mean for a 5° latitude belt. The mean outgoing radiation from the first 650 orbits of Tiros II (November 23, 1960 to January 7, 1961) in Channel IV was thus obtained as a function of latitude. In order to obtain the <u>total</u> outgoing radiation from these measurements the correction curve of Wark and Yamamoto (1961) was used. The results are shown as Curve 3 in Figure 3. These results may now be compared with the flux of the outgoing radiation calculated as a function of latitude from a model atmosphere corresponding to this period.

Model Atmosphere

The model atmosphere was constructed for the months of

December and January, which correspond approximately to winter in the northern hemisphere and summer in the southern hemisphere. Owing to the lack of data for the southern hemisphere on the vertical distribution of the various parameters (water vapor, carbon dioxide, ozone, temperature and cloud cover) necessary for the calculation of the flux of the escaping radiation, we have assumed that the distribution of these parameters in the two hemispheres is the same for the same season.

METEOROLOGICAL PARAMETERS

Temperature.

The vertical temperature distribution as a function of latitude was constructed from the meridional temperature cross sections given by London (1957) and Murgatroyd (1957).

Carbon Dioxide.

It was assumed that the carbon dioxide is uniformly mixed in the atmosphere up to 80 km altitude, and that its total amount (240 cm-atm) remains constant with the latitude (Goody, 1954).

Ozone.

Paetzold and Piscaler (1961) have constructed the meridional cross-section of ozone for spring and autumn while London (1962) has given the latitudinal variation of the total ozone amount in the northern hemisphere. In order to obtain the vertical distribution of ozone for winter and summer needed in our calculations, the total ozone amounts given by London were adopted and the vertical ozone profiles were constructed in such a way that the vertical distribution resembled that of Paetzold's, but was consistent with the total ozone variation with latitude as given by London (1962).

Water Vapor.

The water vapor present in the atmosphere is the most important absorber of the earth's infrared radiation. Although the surface mixing ratios are reliably known for the northern hemisphere (London, 1957), its mean vertical distribution in the atmosphere is not very well determined. Roach (1961) has suggested that the mixing ratio of water vapor decreases with altitude as $(p/p_o)^4$ up to an altitude of about 15 km, above which he assumes a constant value of 2μg/g. However, the total amount of water contained in a vertical column per unit area calculated from the surface mixing ratio value and with this model distribution is about 25% smaller than the estimates of London (1957) based on the measured relative humidity values in the troposphere.

On the other hand, the results of Mastenbrook and Dinger (1960) indicate a higher amount of water vapor in the stratosphere. For this reason in the present investigation two types of vertical distribution of water vapor are assumed. In the first type the water vapor mixing ratio was decreased with height as $(p/p_o)^4$ following Roach, while in the other it was decreased as $(p/p_o)^3$. The latter gives a close agreement with the estimates of London for the total amount of water contained in a vertical column. The distribution of water vapor in the stratosphere is constructed with the help of the measurements made by Mastenbrook and Dinger (1961).

Clouds.

London (1957) has compiled the amount and average height of the different types of clouds as a function of latitude and season. The total cloudiness in December and January for each 10^o latitude belt from 50^o N to 50^o S required in our atmospheric model was adopted from the above investigation. The mean effective height of the clouds was determined by giving proper weight to the relative amount of each cloud type. Figure 1 shows the latitudinal distribution of the total cloud cover and its effective height thus obtained.

Absorption coefficients.

The absorption coefficients of water vapor, ozone, and carbon dioxide in the infrared needed for the computation of the flux of outgoing radiation were taken from the most recent theoretical and experimental investigations.

Carbon dioxide.

The absorption coefficients of CO_2 in the 15μ band have been thoroughly investigated by Cloud (1952); Howard, Burch and Williams (1955); Madden (1957); Yamamoto and Sasamory (1958); etc. Kondratiev and Niilisk (1961) have recently summarized the most reliable available data of the absorption functions of CO_2 in the 15μ region and have concluded that the values given by Yamamoto and Sasamory are "nearest to reality." We have therefore adopted these values for our calculation.

According to Kondratiev and Niilisk (1961), however, the total outgoing radiation of the atmosphere does not depend significantly on the amount of CO_2 adopted or on the small variation of the absorption coefficients thereof. Also the 4.3μ band is unimportant.

Ozone.

Ozone has a strong vibration-rotation absorption band centered at 9.6μ and has been investigated by several authors. Elsasser has also tabulated the absorption coefficient of ozone in the 9-10μ region using basically the experimental determinations of Summerfield (1952) and correcting them for the more recent results of Walshaw (1957).

For our investigation we therefore adopted the absorption coefficients for ozone as tabulated by Elsasser.

Water Vapor.

Unlike CO_2 and O_3 the absorption coefficients of water vapor in the region of 7-30μ are very controversial. The 6.3μ fundamental has however been thoroughly studied and the absorption coefficients reliably determined (for example, Howard, Burch and Williams, 1955). Between 8 and 12μ, however, water vapor has a continuum superposed by weak rotation lines which increase in intensity towards longer wavelengths, becoming maximum at approximately 45μ. The absorption coefficients of water vapor in the 8-13μ region are relatively small and this region is known as the "atmospheric window." However, the absolute values of the absorption coefficient differ from one investigator to the other.

Figure 1 shows the absorption coefficient of water vapor as a function of wavelength, after a number of investigators. It is immediately seen that the disagreement between the individual values of the 8-13μ region is highly significant. Kondratiev (1961) has however summarized all the recent reliable data and has given a generalized absorption coefficient for the window which at 9.6μ also agrees with the very accurate determination of Vigroux (1959). Following Kondratiev the values given by the dotted line in Figure 3 were therefore used as the absorption coefficient for water vapor in our calculations.

<u>Temperature and pressure dependence of the absorption</u>.

Due to the variation of the temperature and pressure in the earth's atmosphere it is important to take into account the effect of these parameters on the absorption coefficients. The temperature dependence of the absorption coefficient has been investigated by Elsasser (1960) who gives the following empirical relation for the absorption coefficient L at a temperature T and frequency ν:

$$\log \frac{LT}{L_o T_o} = -a\ (\nu - \nu_o)^2\ \frac{T_o - T}{T}$$

where L_o is the absorption coefficient at temperature T_o, the factor 'a' varies for each band, and ν_o is the frequency of the center of the band. This temperature dependence was used in our computations.

In order to correct for the pressure dependence of the absorption coefficients, we employed the factors given by Kondratiev and Niilisk (1961) -- namely $(p/p_o)^{0.5}$ for CO_2, $(p/p_o)^{0.8}$ for H_2O, and $(p/p_o)^{0.2}$ for O_3.

Method of Calculation.

The model atmosphere, as described in the preceeding paragraphs, was divided into 80 layers of approximately ten millibars each, between 1,010 and one millibar. The temperature of each layer was taken from the vertical profile of temperature for different latitudes and seasons given by London (1957). The H_2O, CO_2, and O_3 concentrations for each layer were adopted as described earlier. The outgoing flux was computed for every 10 cm^{-1} interval in the infrared between 3 and 250μ from the following equation:

$$B_\nu = B_{G\nu}\, \tau_o + \int_{\tau_o}^{1} B_\nu(T)\, d\tau \qquad (1)$$

where B_ν is the total black body radiation in this wavelength interval reaching the top of the atmosphere; $B_{G\nu}$ the ground radiation; τ_o the total transmission of the atmosphere; τ being given by

$$e^{-\int_z^{\infty} (k_1\rho_1 + k_2\rho_2 + k_3\rho_3)\, dz}$$

where k_1, k_2, and k_3 are the absorption coefficients of CO_2, H_2O, and O_3; and ρ_1, ρ_2, and ρ_3 being their densities. The Equation (1) was integrated over all the frequencies between 3 and 250μ by a program of the IBM 7090 computer.

The total outgoing flux leaving the earth's atmosphere, in the absence of clouds, was thus obtained and is shown in Figure 3 as curve 1. It corresponds to the model atmosphere II in which water vapor mixing ratio decreases with height as $(p/p_o)^3$. In order to compare our computations with the TIROS II measurements of the total flux (curve 2), we had however to take into account the presence of clouds. Assuming the clouds to be "black" to the infrared radiation between 3 and 250μ, and using the effective heights as determined earlier, we computed the total outgoing radiation as a function of latitude for an atmosphere completely covered with clouds.

Knowing the percentage of cloud cover in each latitude belt for the season (Figure 2), we determined the actual radiation leaving the atmosphere for the average cloudiness.

The radiation flux leaving the atmosphere thus obtained for each 10^{o} latitudinal belt between 50^{o}N and 50^{o}S is shown by curves 3 and 4 in Figure 3. Curve 3 corresponds to the model atmosphere I (water vapor mixing ratio decreasing as $[p/p_o]^4$) and curve 4 for the rate of decrease as $(p/p_o)^3$ (model atmosphere II).

In the same figure we have plotted the values of the total outgoing flux computed by London (1957).

<u>Comments on Figure 3.</u>

The agreement between the Tiros II measurements for a particular period with our theoretical computations based on thirty-year mean values of the meteorological parameters is striking. The sudden fall in the total outgoing radiation flux at the equator as observed by Tiros II, and also revealed in this study (Curves 3 and 4), and by London's investigation as well, is due to the high percentage of cloud cover in the summer tropics (Figure 2). The subtropical high pressure belts (10^{o}-30^{o} latitude) associated with low cloud amounts (Figure 2) and drier atmosphere leads to higher outgoing flux as apparent from Figure 3. It is difficult to say, however, which of the computed curves (3 or 4) shows better agreement with actual measurements. Although the dependence of the escaping flux on the vertical distribution of water vapor is appreciable, the difference due to the presence of clouds is however highly significant (Curves 1 and 4). As mentioned above, our computations are based on the thirty-year mean values of the meteorological parameters, and the flux measured by Tiros II corresponds to November, December 1960 and January 1961. From the available data of the global meteorological observations (Monthly Weather Records) we have tried to see whether December 1960 was a "normal" or "abnormal" month. The analysis of the latitudinal temperature distribution for this month shows extremely consistent values ($\pm$ 0.5°K) with the thirty-year mean values of each belt. No data, however, is readily available for the cloud cover. It is therefore difficult to conclude whether the small differences between the actual and computed values of the total outgoing flux are due to the discrepancies in the adopted water vapor distributions or the mean cloud cover. Moreover, as mentioned earlier, the meteorological parameters used for the southern hemisphere are those of the northern hemisphere for the same season. In the absence of actual data from the southern hemisphere, it becomes still more difficult to comment on any disagreement between computed and observed radiation fluxes.

Despite these uncertainties it is rather encouraging that the measured flux by TIROS II shows an agreement in the $\pm$ 2% of the computed flux for a mean water vapor distribution decreasing as $(p/p_o)^3$, and an average cloudiness as compiled by London (1957). The agreement of the TIROS data with London's computation of total outgoing flux is also noteworthy. If the mean cloud cover for the same period (November 23, 1960 to January 7, 1961) can somehow be determined from TIROS II cloud photographs, it will perhaps be possible to obtain a better estimate of the water vapor distribution in the atmosphere for different latitudes. In any case the TIROS II data for Channel 4 as corrected by Wark and Yamamoto (1961) can be safely applied to the study of planetary heat budget which is now in progress.

<u>TIROS II Measurements in Channel 2 -- (7 - 13μ).</u>

Apart from measuring the total radiation flux leaving the earth's atmosphere, TIROS II satellite has also a sensor to measure the radiation escaping in the 7-13μ region (Bandeen, Hanel, Stampfl, Licht, and Stroud, 1961). Due to relatively small amount of absorption by water vapor, and almost insignificant absorption by CO_2 and O_3, the observations in this spectral region were made to obtain, in the absence of clouds, an estimate of the ground temperature. A preliminary analysis of the radiation data of TIROS II for this channel (Nordberg, <u>et al</u>., 1962) has however shown that the temperatures measured by this sensor in a cloudless atmosphere are several degrees lower than the actual ground temperature. The discrepancy has been attributed to the water vapor continuum in the 7-13μ region (Figure 2).

From our model atmosphere II we have computed the outward going flux in the 7-13μ region as a function of latitude. The results obtained, both for a clear and a partially cloudy atmosphere, are shown in the Figure 4 (curves 2 and 3). The mean ground temperatures are also plotted in the same figure (curve 1). It is clear that even inthe case of cloudless atmosphere, the temperatures inferred from the satellite measurements in this wavelength region will be several degrees lower than the actual ground temperature. The difference between the two curves varies with the latitude and season, the minimum difference being at sub-tropical latitudes during winter when the water vapor amount in the atmosphere is at its minimum.

In the case of partially cloudy atmosphere (curve 3) the difference between observed temperatures and the ground temperatures is much larger due to the presence of clouds. The curve 3, however, represents average temperature observable from a satellite in the 7-13μ channel under the conditions of average cloudiness. In order to compare our results with the actual observations, we have therefore analyzed the Channel 2 radiation data of TIROS II in the manner already described for

Channel 4. The results are plotted as curve 5 in Figure 4. The reason for a systematic $\sim 10^{\circ}K$ difference between the observed and computed temperatures is difficult to understand, but a small discrepancy of $\sim 3^{\circ}K$ in the TIROS II measurements for this channel due to the following reason can perhaps be rectified.

If one examines the transparency of the filter for Channel 2 (Figure 5) used for the sensor in the TIROS II satellite, one finds that, apart from being transparent in the 7-13μ region, it has another small window between 15-21μ. Although the transmission in the 15-21μ region is only 3% compared to 35% in the "window," this may be a significant source of error in the estimation of the effective temperature of the 7-13μ "window."

From Figure 4 we know that the average effective temperature for the 7-13μ region should be in the neighborhood of $275^{\circ}K$. But similar computations in the 15-21μ region give a mean effective temperature of approximately $220^{\circ}K$. This is due to the high opacity of CO_2 and water vapor in this spectral region -- implying that most of the radiation is coming from near the tropopause.

The total energy reaching the sensor of Channel 2 in the TIROS II satellite is therefore partially weighted by this low temperature radiation and therefore gives systematically lower temperatures than what one would obtain if transmission of the filter was rigorously limited between 7-13μ.

From these considerations a crude estimate in the correction necessary due to this extra window in the filter has been obtained. They range between + 2.5°K and + 5.5°K for different latitudes, being higher at the equator and lower at the middle latitudes. These estimates correspond to a cloudless atmosphere.

In the presence of clouds, however, the amount of correction necessary decreases. We have therefore adopted a mean value of + 3°K to correct the TIROS II observations for all the latitudes. Curve 4 in Figure 4 shows the corrected effective temperatures as measured by TIROS II in this channel. This brings the observed curve nearer to curve 3, but still a difference of ~ 5-10°K exists between the two. The explanation probably lies in the calibration of the sensor or the difference between the assumed and actual cloud cover to which the Channel 2 data should be very sensitive.

<u>10-11μ Window for Nimbus</u>.

Due to the failure of TIROS II in looking at the ground through the 7-13μ Channel, the Nimbus satellite will carry a sensor which is strictly confined to the 10-11μ region (Stampfl and Press 1962). In this wavelength interval the water vapor absorption is very small, and this spectral region lies outside the absorption band of ozone. We have used our model atmosphere to compute the outgoing flux in 10-11μ interval and hence the temperature which will be measured by this sensor. As usual we have corrected our values for the average cloud cover for different latitudes, and the effective temperature thus obtained as a function of latitude are plotted in Figure 6, Curve 2. Curve 1 shows the actual ground temperature. It is obvious that the agreement between the two is much better than in the case of Channel 2 of the TIROS satellite, but still the temperatures measured are lower than the ground temperature. This is mainly due to the inclusion of clouds in our model. In the cloudless atmosphere, however, the temperatures measured are within 2^{o}K of the ground temperature, thus indicating the usefulness of such an experiment.

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LEGENDS

Fig. 1 - Global distribution of mean cloud amount and its effective height (after London, 1957).

Fig. 2 - Absorption coefficients of water vapor in the near infrared.

Fig. 3 - Total outgoing flux in the infrared as a function of latitude (computed and observed).

Fig. 4 - Mean ground temperature, and effective observed and computed temperatures for the 7-13μ region as a function of latitude.

Fig. 5 - Effective spectral response of Channel 2 vs. wavelength.

Fig. 6 - Mean ground temperature and effective emission temperature computed for the 10-11μ window as a function of latitude.

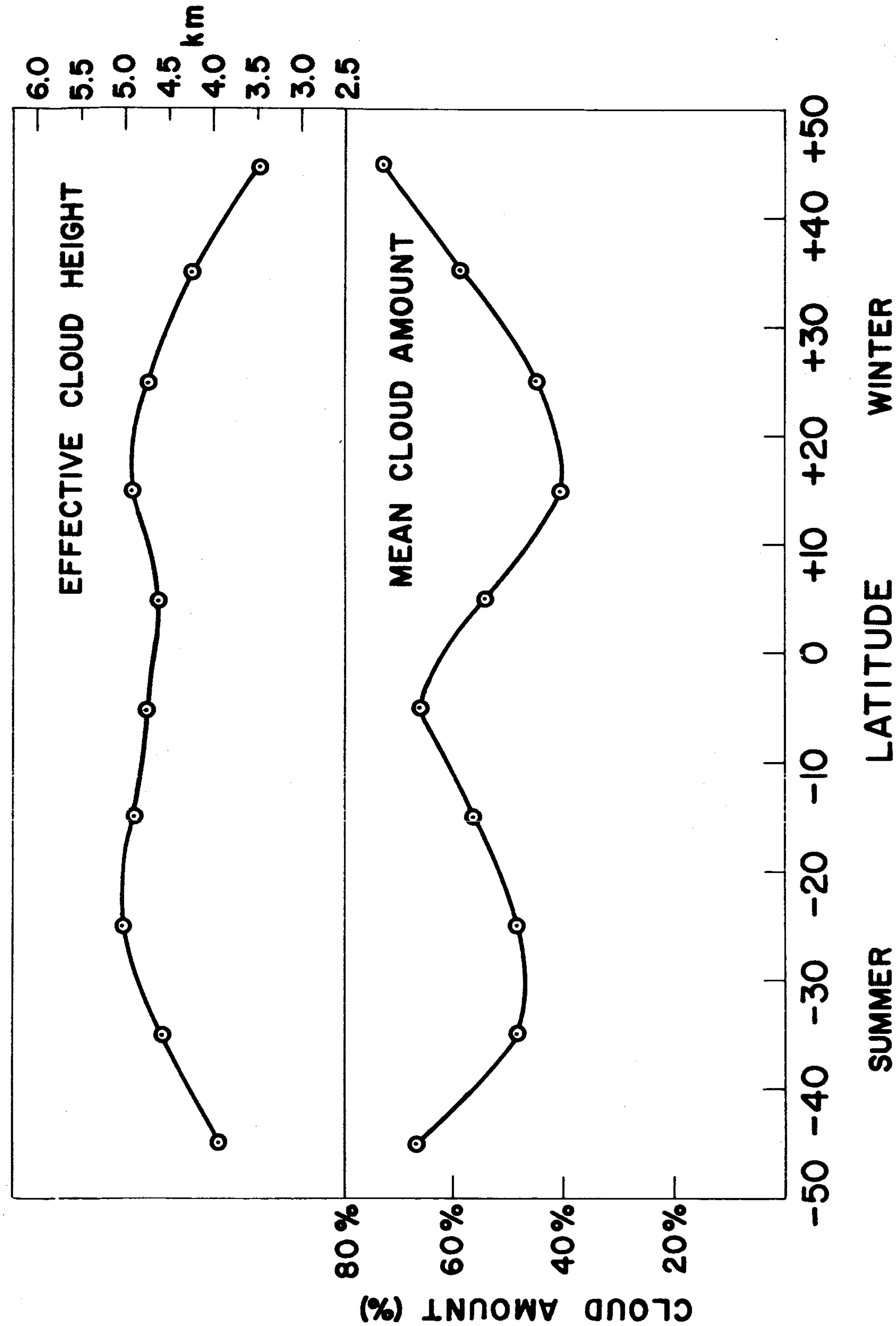

Fig. 1. Global distribution of mean cloud amount and its effective height (after London, 1957).

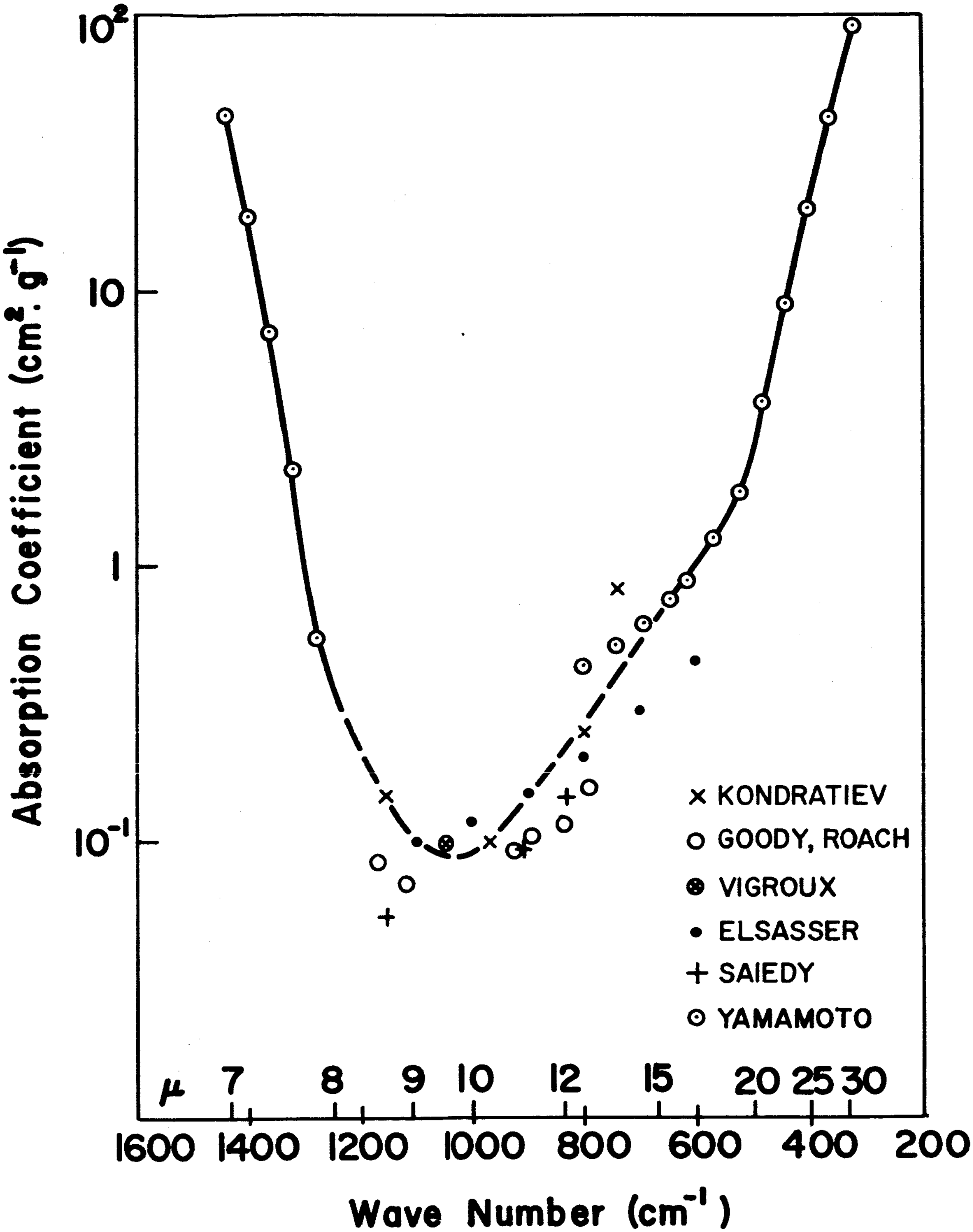

Fig. 2. Absorption coefficients of water vapor in the near infrared.

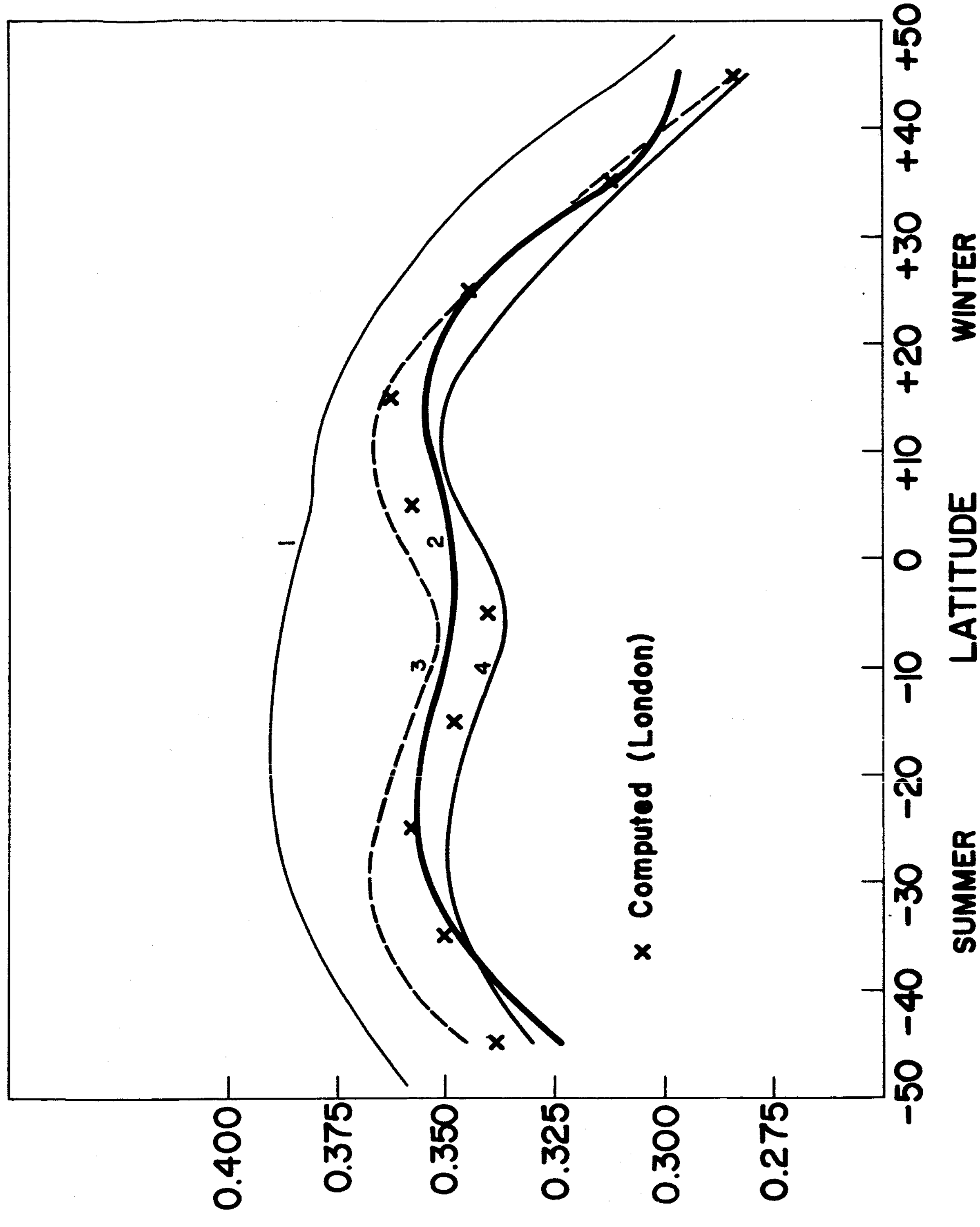

Fig. 3. Total outgoing flux in the infrared as a function of latitude (computed and observed).

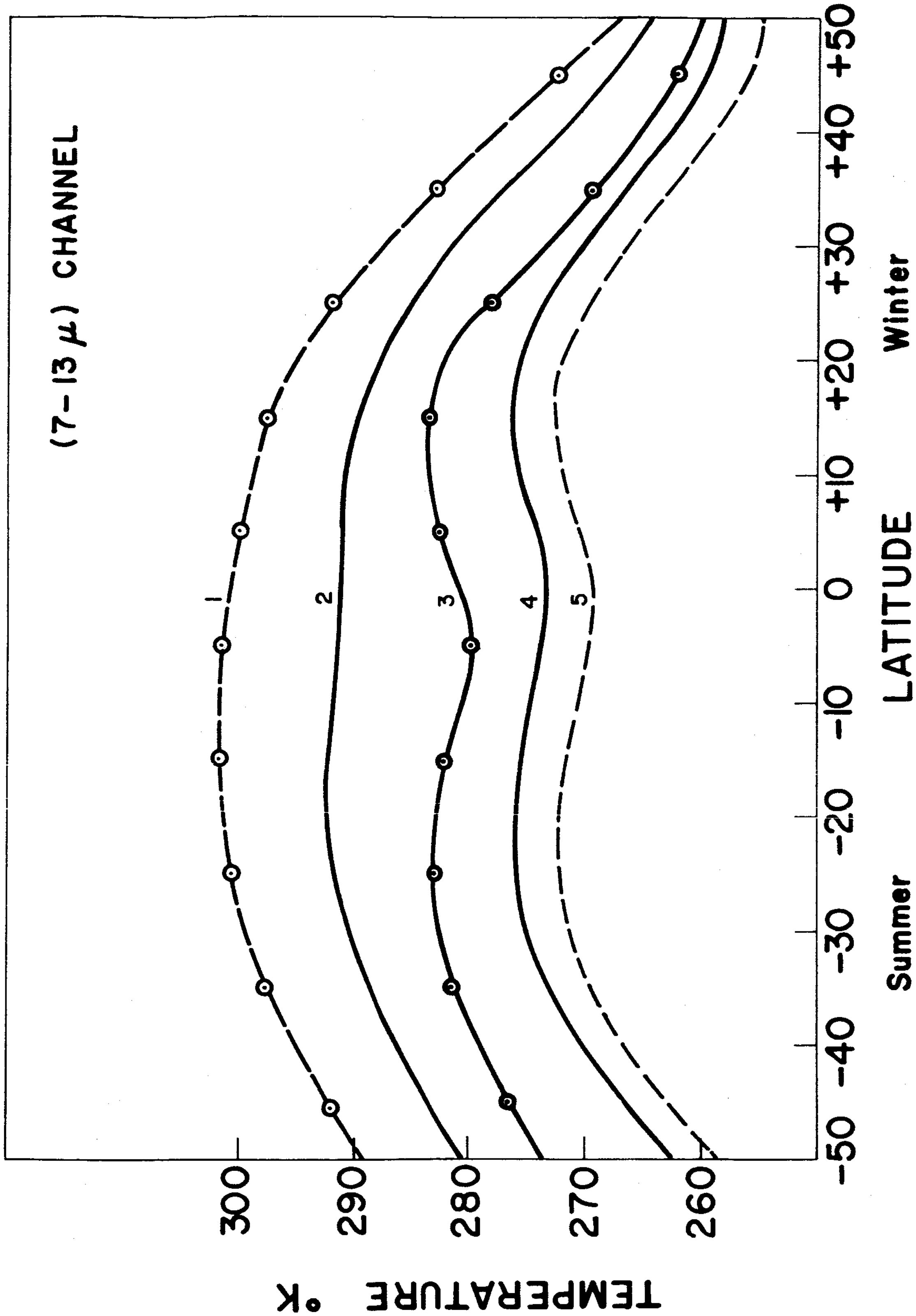

Fig. 4. Mean ground temperature, and effective observed and computed temperatures for the 7 - 13μ region as a function of latitude.

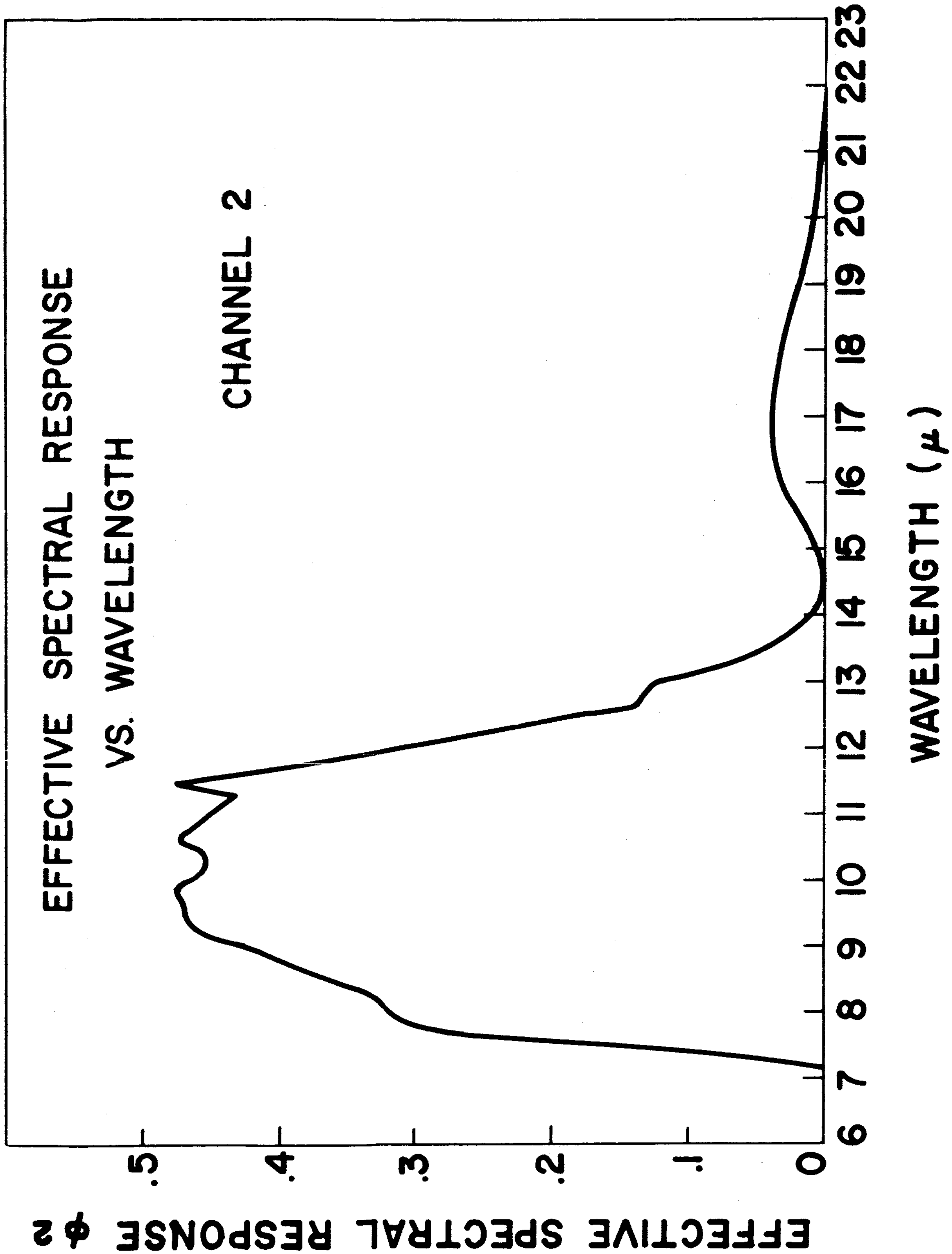

Fig. 5. Effective spectral response of Channel 2 vs. wavelength.

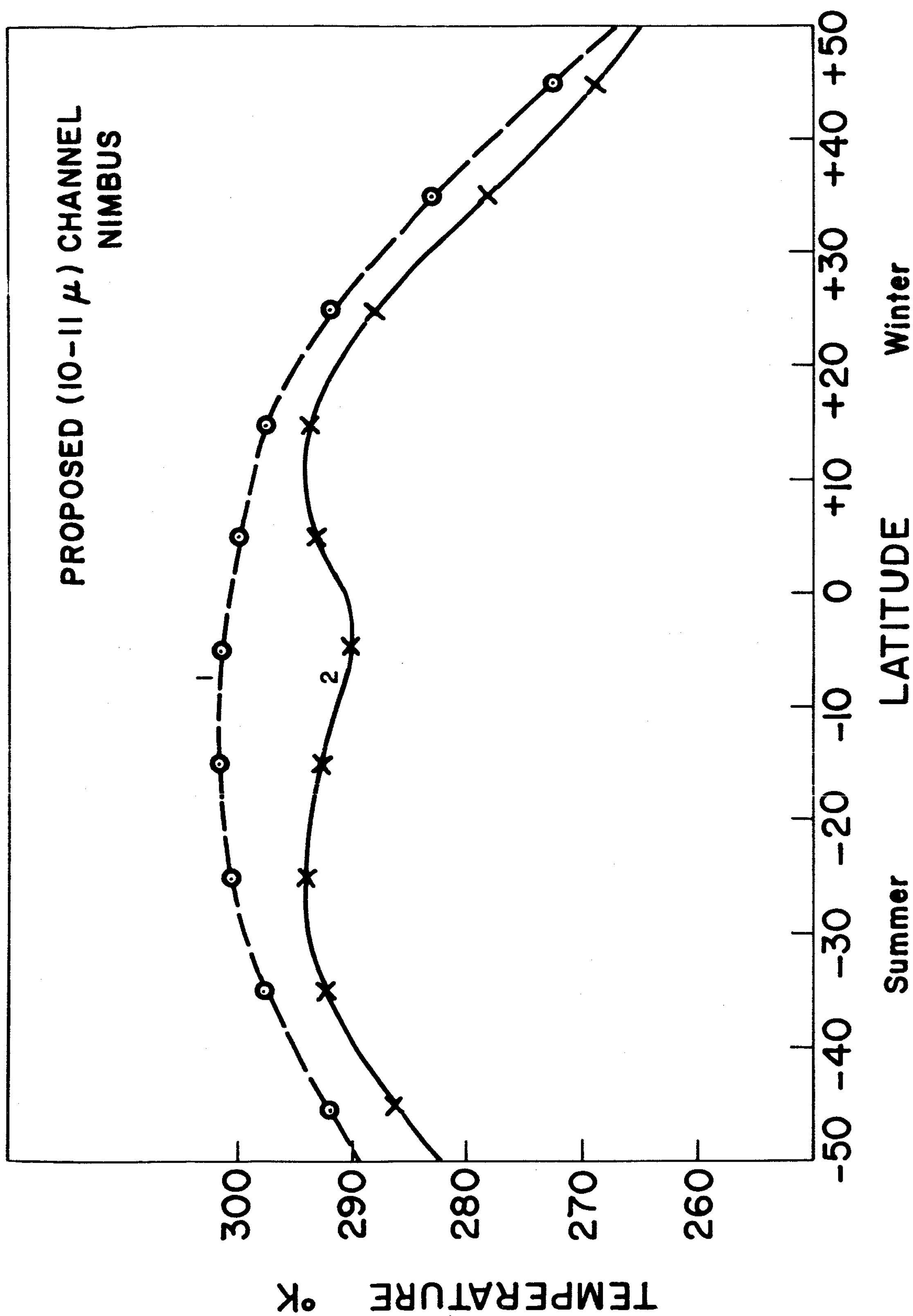

Fig. 6. Mean ground temperature and effective emission temperature computed for the 10-11 μ window as a function of latitude