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Global temperature distributions from OGO-6 6300 Å airglow measurements

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The OGO-6 6300 Å airglow temperature measurements have been used to develop models of the global temperature distributions under solstice and equinox conditions for the altitude region from 240 to 300 km and for times ranging from dawn in this altitude region to shortly after sunset. The distributions are compared with models derived from satellite orbital decay and incoherent scatter sounding. The seasonal variation of the temperature as a function of latitude is shown to be very different from that derived from static diffusion models with constant boundary conditions.

INTRODUCTION

Current international reference atmospheres such as *CIRA* [1972] are based on satellite orbital decay data. The data on which such models are based are more closely related to the density than the temperature and the temperatures are used as a convenient method of relating densities at different altitudes in a simple manner. *Waldteufel and McClure* [1969] and *Swartz et al.* [1972] have developed models of the neutral temperature based on incoherent scatter sounding data. These data have the advantage that the temperatures are measured relatively directly; however, the number of locations at which such measurements can be made is severely limited and the measurements upon which these models were based did not extend beyond the region from 12° south to 45° north. The measurement of the gradients of both the densities and temperatures is of extreme importance in the calculation of neutral winds which depend upon the pressure gradients. These winds in turn are of critical importance to the understanding of the energy budget of the thermosphere and the behavior of the *F* region of the ionosphere.

The present paper describes global temperature

distributions measured by the 6300 Å airglow photometer on OGO-6. Measurements were made from June 8, 1959 until August 16, 1970 and thus correspond to quite active solar conditions. The details of the experiment have been described by *Blamont and Luton* [1972]. The data on which this work is based were averaged over the height interval 240 to 300 km. The sensitivity of the experiment was such that temperatures could only be measured during daytime and shortly after sunset.

A serious problem with the data used for this study is a discrepancy first noticed by P. Bauer and J. M. Luton between the 6300 Å airglow and incoherent scatter temperatures. The discrepancy appears to have a maximum in the middle of the day when the airglow temperatures are about 100 K lower than the incoherent scatter temperatures. This effect is presently being investigated but the explanation has not yet been found.

MEASUREMENTS

A basic problem with satellite experiments is that the orbit precesses slowly and it takes a considerable period to obtain the diurnal variations. The orbital characteristics of OGO-6 were chosen with great care and it is fortunate that orbits are available in which one half of the orbit was close to the

diurnal maximum and the other close to the diurnal minimum for times which are close to the June solstice and each equinox during the measurement period. Figure 1 shows the variation as a function of latitude for such days. Several things are immediately obvious. It is apparent that the diurnal variation becomes smaller as one approaches the poles and is very small at latitudes greater than about 70° . It is also apparent that at the solstice the maximum temperature is reasonably constant in the summer hemisphere and the minimum temperature reasonably constant in the winter hemisphere. The maximum average temperature does not occur close to the subsolar point but at 60° to 70° N while the minimum average temperature occurs at 40° to 50° S. The equinox curves are quite different and both cases show that the polar temperatures are similar in the north and the south. Again the diurnal variation is small close to the poles and largest close to the equator. The large variation in the ratio of the maximum to the minimum temperature ratio is not due to the diurnal variation in the maximum temperature, which is quite constant over the globe, but to the diurnal minimum. The behavior shown in this figure is quite general for several days of relatively low

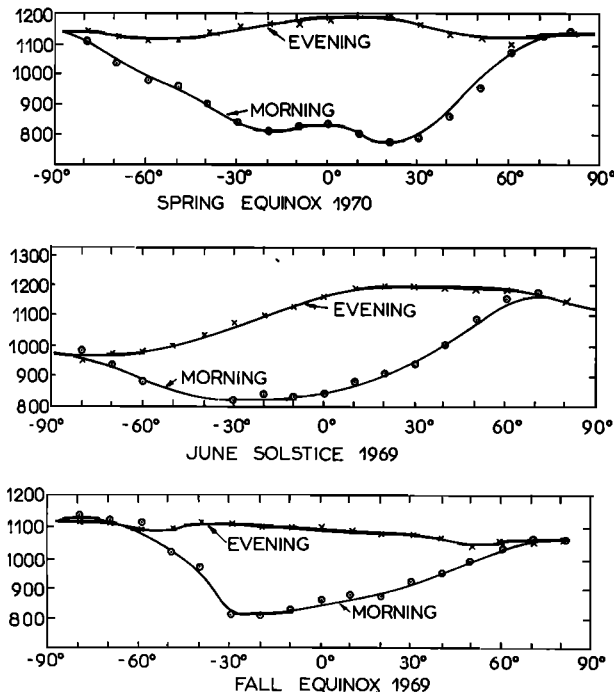


Fig. 1. Temperatures at 270 km at times close to sunrise and sunset as a function of latitude.

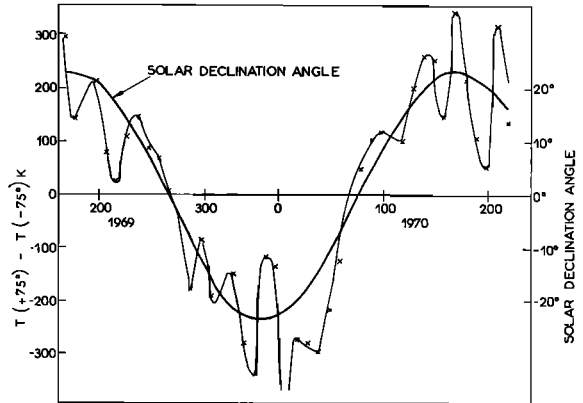


Fig. 2. Difference in temperature at 270 km between 75° N and 75° S.

magnetic activity for which the diurnal maximum and minimum conditions were measured. It must be concluded that for these equinox cases the maximum in the diurnal average temperature did not occur close to the subsolar point but that there were two maxima, one at either pole.

The way in which the atmosphere changes over from the summer to winter conditions is of considerable interest because of the influence it has on global circulation patterns and on the information it can give about thermal delays in the thermosphere. Figure 2 shows the variation of the difference between the temperatures in the northern and southern regions at latitudes whose magnitudes exceeded 70° . In each case measurements were averaged over a 24-hr period. The procedure is not sensitive to large local time perturbations because of the relatively small local time effects at high latitudes. Taking the differences in temperatures a first-order correction is provided for the quite large variations which occur at high latitudes related to magnetic activity. Despite the scatter the overall annual variation is sinusoidal with an amplitude of about 300 degrees. The changeover from positive to negative occurs within 10 days of the equinox in either case. Further data on other years would be required before more categorical statements could be made about the phase of the component but it does not appear that large delays relative to the solar declination are involved. The changeover at the equinoxes is very rapid with a slope of about 7 K per day. It is thus apparent that in attempting to develop global temperature distributions characteristic of equinox conditions, data separated from the equinox by more than a few days must be treated with

caution, and the period over which it is possible to allow the satellite to precess is much smaller than for solstice conditions.

GLOBAL TEMPERATURE DISTRIBUTIONS

It took OGO-6 approximately 3 months to precess through 24 hr in local time. It is apparent that this period is a major fraction of a year and that during this period major temperature variations occur correlated with seasonal changes. Because of the location of the orbits it was possible to establish the major portion of the diurnal variation without using data separated by more than 21 days from the solstices and this appeared to be tolerable.

Within these periods the variations correlated with solar and magnetic activity were comparable with the diurnal and latitudinal changes it was desired to determine. As has been shown by *Waldteufel* [1970] and *Rohrbaugh et al.* [1973], though a good correlation exists between the mean monthly 2800 MHz solar flux and the thermospheric temperature, the correlation with the component varying with the solar rotation is not nearly so good. Large changes occur from one day to the next that are not well related to ground measurements of solar activity. A similar conclusion was reached by *Knight et al.* [1973] for the 304 Å solar flux measured by OSO-6. It was therefore concluded that some other method of normalizing the measurements should be used than one based on solar-geophysical indices. In the method employed a day was selected close to the solstice for which measurements were available at the equator for times close to sunrise and sunset. As has been shown by *McClure* [1969] the diurnal temperature minimum occurs at Jicamarca (12°S) at about 5 hr local time and the diurnal maximum at about 18 hr. These two points were then used with normalized diurnal variations from Jicamarca measurements to establish a mean diurnal variation at the equator. All available equatorial temperatures were compared with this variation to check that the distribution appeared consistent with the measurements. Despite the expected scatter the curve appeared to provide a reasonable fit to the data. This diurnal variation was then used to normalize the variation along the orbit by choosing the quiet day in a five-day period when the equatorial temperature was closest to the mean diurnal variation, to reject data that was too far from these values, or to scale the data up or down by the difference between the measured and mean temperatures at the

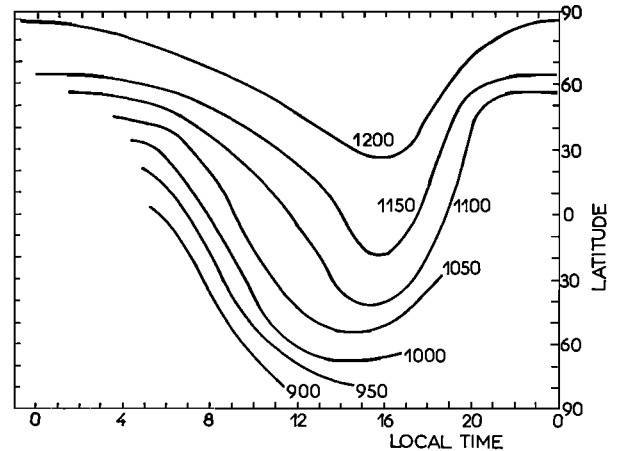


Fig. 3. Global temperature distribution at 270 km, June solstices 1969 and 1970.

equator at each latitude. The effect of this scaling was to remove data that was very far from the average for the period and to use the measurements of the temperature gradients along the orbital path for those remaining.

As has been shown by *Blamont and Luton* [1972] very large changes occur in the temperatures at high latitudes under disturbed magnetic conditions. At the winter pole data were only obtained for a short period when the region was sunlit and the average of the quiet data was used to establish the temperatures in this region. At the summer pole which was continually sunlit enough measurements were available to establish that the diurnal variation was very small. The asymptotic temperature determined for the day selected to normalize the equatorial temperatures was then used for the pole. Days were selected with asymptotic temperatures close to this value to establish the high latitude temperature gradients.

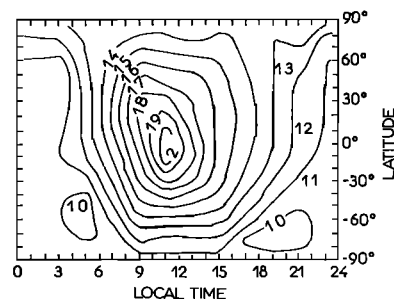


Fig. 4. Ratio of density to global minimum (June solstice): *Barlier et al.* [1972], $\delta = 20^\circ$.

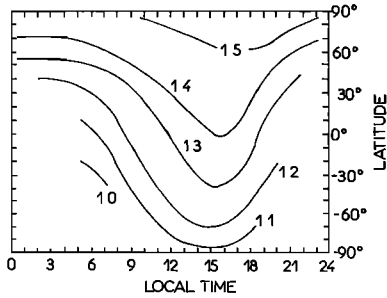


Fig. 5. Ratio of temperature to global minimum (June solstice).

Figure 3 shows the calculated global temperature distribution for the June solstice. This was obtained using data from both the 1969 and 1970 solstices. The data were normalized to conditions corresponding to a 2500 MHz solar flux of about $150 \times 10^{-22} \text{ w m}^{-2}\text{Hz}^{-1}$ and an A_p value of about 4.

The basic features of the behavior are quite obvious. It is apparent that at all local times during the day the temperatures increase monotonically from the winter to the summer pole. There is a very marked effect of the day length causing the temperature gradients to be much larger at the morning increase and sunset decrease in the winter hemisphere. The gradient in the maximum temperature is small in the summer hemisphere and the gradient in the minimum temperature is small in the winter hemisphere.

This model is compared with the 280-km densities of *Barlier et al.* [1972] in Figures 4 and 5. It is apparent that there are very large differences in the location of the global maximum and in the diurnal variation at all latitudes. In Figure 6, the temperature variation for the December 1969 solstice derived in the same manner as for the June

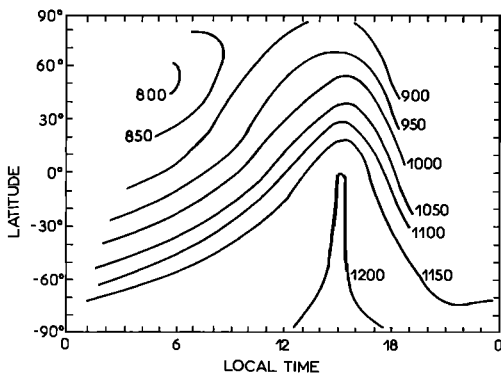


Fig. 6. Global temperature distribution at 270 km, December 1969 solstice.

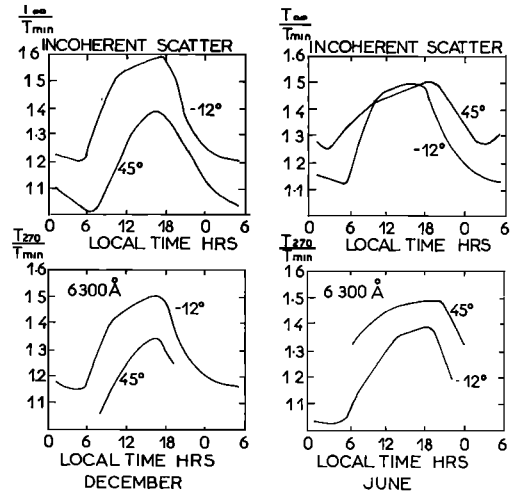


Fig. 7. Comparison of ratio of temperature to diurnal minimum for present model compared with data of *Waldteufel and McClure* [1969].

solstice is presented. Major differences in the diurnal behavior of the densities and temperatures have been known for some time as has been shown for example by *Nisbet* [1967] using Arecibo incoherent scatter data. *Waldteufel and McClure* [1969] attempted to use the data from Jicamarca (12°S) and St. Santin (45°N) to develop a preliminary model. The extrapolation to other latitudes does not agree well with our measurements but in Figure 7 we have plotted the diurnal variation at 12°S and 45°N in summer and in winter normalized to the minimum value in winter at 45°N. This is compared with *Waldteufel and McClure* [1969] values. Based on an examination of many days of data it would seem that the agreement is well within the day-to-day variation.

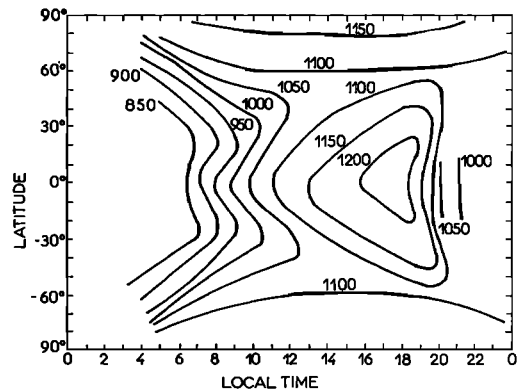


Fig. 8. Global temperature distribution at 270 km, Spring 1969 equinox.

The equinox behavior is more difficult to determine than that for the solstices because of the rapidity of the change. The method employed to calculate the distribution shown in Figure 8 used the data for the spring equinox in 1970. These measurements corresponded to times close to the diurnal maximum and minimum at most latitudes and the data at intermediate times were interpolated using typical equinox diurnal variations at Jicamarca, Arecibo, and St. Santin at latitudes up to 45° and by assuming symmetry from north to south at higher latitudes.

CONCLUSIONS

It has been shown that at altitudes in the vicinity of 270 km the maximum temperature occurs close to the summer pole at the solstices and that the changeover occurs within a period of about 20 days close to the equinoxes. The differences between the diurnal variations in the temperatures measured by incoherent scatter and the equivalent temperatures derived from satellite density measurements are confirmed. At high latitudes the effect of day length becomes very pronounced and the day-to-night temperature ratio becomes small at latitudes greater than 70° . Based on the measurements described here, the day-to-night temperature ratio at any latitude and season can be approximated by

$$T_{\max}/T_{\min} = 1 + 0.21 \cos \theta [1 - \cos (2\pi D/24)]$$

where θ is the geographic latitude and D is the day length angle at 200 km given by

$$D = \cos^{-1} [(-0.1738 - \sin \theta \sin \delta) / \cos \theta \cos \delta]$$

where δ is the solar declination.

While the techniques used to analyze the data

would have the effect of compensating for discrepancies that are simple functions of local time, uncertainties must remain until the cause of the divergences with incoherent scatter temperatures is found.

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