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DIRECT MEASUREMENT OF THE VERTICAL COMPONENT OF THE TRANSPORT VELOCITY IN THE IONOSPHERIC F REGION

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ABSTRACT

DIRECT MEASUREMENT OF THE VERTICAL COMPONENT OF THE TRANSPORT VELOCITY IN THE IONOSPHERIC F REGION

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A new technique for measuring the vertical component of the ionospheric ambipolar transport velocity V_z in the topside F region is reported. Incoherent backscatter spectra taken on August 7, 9, and 15 (1967) reveal a slight Doppler shift from which the line-of-sight (vertical) velocity of the plasma under observation is calculated. The measurements were taken from 1800 hrs. L.T. to 2200 hrs. L. T. and demonstrate that the velocities during this period were most negative at times corresponding to the rapid cooling of the F region due to ionospheric sunset. The values of V_z range from 40 meters per second downward to 10 meters per second upward.

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I. Introduction

Unlike the lower regions of the ionosphere, the F region and, particularly, the topside F region cannot be studied successfully by consideration of the ionizing radiation alone. The reason for this lies in the slow rate of decay of the ionization in this region. For example, the decay time for the lower lying E region is less than one hour, while the decay time in the F region is on the order of days. This slow rate of decay gives rise to the possibility of a redistribution of the ionization before it disappears. Thus, the simple concept of the ionosphere being controlled by a production-loss mechanism of solar radiation is not applicable.

The determining equation for the electron density in the F region as a function of height and time is the following continuity equation:

$$\frac{\partial N_e}{\partial t} = q - \beta_{eff} N_e - \nabla \cdot (N_e \underset{\sim}{V}) \qquad \text{I-1}$$

where N_e = number density of free electrons

q = rate of electron production

β_{eff} = the effective coefficient of attachment

$\underset{\sim}{V}$ = the transport velocity

The electron density, therefore, is determined by the simultaneous interplay of three processes: 1) electron production; 2) electron loss; and 3) movements of the electrons via the transport velocity. Consequently the transport velocity must be included in any satisfactory theory of F region shape and morphology.

The first requirement of such a theory is, of course, a knowledge of what is to be explained; namely, the electron density as a function of height and time. In the past these electron density profiles have been provided solely by the use of ionospheric sounders. This technique suffers, however, from being able to give information only at heights at and below the F peak. Sounders carried in satellites ("topside" sounders) can and are used to give electron density information at the F peak and above it, but no information concerning the time development of the topside region at a given location can be deduced from this method.

The continuity equation (I-1) has within it the potential to explain most of the features of the F region behavior. Our knowledge of the parameters q, β, and $\underset{\sim}{V}$ is, however, quite uncertain. Without question the transport velocity is the most poorly known of the parameters, for no agreement exists as to either its magnitude or its direction at any given time during the day or night or at any given location. This, perhaps, can be understood if one realizes that the transport velocity is really the resultant of several competing processes: ambipolar diffusion, electrodynamic drifts, rapid temperature changes, and the effect of neutral atmospheric winds.

Probably the first real attempt at solving the continuity equation was made by Ratcliffe et al (1956). They simply neglected the movement term $-\nabla \cdot (N_e \underset{\sim}{V})$, assumed the well known expression for q, first derived by Chapman (1931) and then examined the observed electron densities in the F1 and F2 regions. They demonstrated that these densities were consistent with an attachment-like process above a height of 240 km with the coefficient of attachment β decreasing with increasing height. (Historically this was of significance because it gave quantitative support to the Bradbury (1938) hypothesis of F region formation.)

Yonezawa (1958) found equilibrium solutions, i.e. $\frac{\partial N_e}{\partial t} = 0$, for uniform electrodynamic drift velocities. He found that the altitude of the F2 peak was almost linearly proportional to the drift velocity and that the peak density increased exponentially with peak altitude.

Since Yonezawa, many equilibrium model solutions to the continuity equation have been attempted. Risbeth and Barron (1960) considered the dependence of equilibrium density profiles on uniform drift velocities and on the ratio of loss to diffusion coefficients. Bowhill (1961) proposed a simple model for the formation of the daytime peak of the F2 region in which he ignored transport effects. Nisbet (1963) computed equilibrium model solutions for the effective attachment coefficient during the nighttime allowing the electron temperature to exceed the ion temperature, but also ignoring transport. Tornatore (1964) later extended this work to include uniform electrodynamic drift velocities. Quinn and Nisbet (1965) further extended this work to include both uniform drift velocities and the effects of diffusion.

Risbeth (1964) obtained the first time dependent solutions of the F2 region for both a "fast" diffusion model and a "slow" diffusion model. (The "slow" diffusion model set the diffusion coefficient at one order of magnitude less than the "fast" diffusion model.) He discovered that the "fast" diffusion model best fits observed daytime density profiles, while a slow diffusion model was necessary to explain nighttime density profiles. Risbeth suggested that: a) the diffusion or loss coefficients undergo a diurnal change arising from changes in the neutral composition or b) the night layer is controlled by electrodynamic movements or nocturnal sources of ionization.

Because of its importance in explaining F region behavior the transport velocity and/or its various components have been the subject of much special attention. Chandra et al (1960) tried to extend the analysis of Ratcliffe et al (1956) by including the movement term in the continuity equation. They made assumptions similar to those of Ratcliffe et al about the rate of electron production q and the loss coefficient β, and with these, they wanted to determine the component of the total vertical transport velocity as a function of height. The problem is indeterminate without an additional arbitrary assumption; they chose the initial values of the vertical velocity as between ± 20 m/s at about 240 km. This uncertainty converged at greater heights. The maximum magnitude of their deduced velocities was about 25 m/s. Using several models, Garriott and Thomas (1962) and Degaonkar and Santani (1966) attempted to separate the vertical transport velocity into a diffusion component and an electrodynamic drift component. Both works show that the drift component could be as high as 100 m/s and that it dominated over the diffusion process. Later, Shimazaki (1965) introduced the effect of a rapid temperature change and demonstrated that the vertical transport associated with this rapid temperature change was dominant over both diffusion and electrodynamic drift at times near sunset. Maeda and Kato (1966) extended the work of Martyn (1948) and others to predict the diurnal change of the electrodynamic drifts on the basis of the electric fields produced by the dynamo theory. Kohl and King (1967) and Geisler (1967) have indicated that neutral winds blowing in the F region are also important in determining the total transport velocity. Using radar backscatter measurements at Arecibo, Rao (1968) has estimated nighttime ($q = 0$) solutions of the continuity equation for the vertical transport velocity at heights (400-625 km) where loss is considered negligible. Using backscatter measurement made at Millstone Hill, Evans (1965, a, b, c) has observed a rapid decrease in the electron temperature of the topside F region at sunset and has suggested a strong accompanying downward diffusion.

The present investigation is concerned with the problem of <u>directly</u> measuring the vertical component of the transport velocity at several heights between 300 and 525 km, using a new radar backscatter technique. The technique consists of measuring the Doppler shift of the characteristic backscatter frequency spectrum produced by the moving plasma. The observations were made at Arecibo, primarily at times during and immediately following ionospheric sunset. The Doppler method can be used, of course, equally well at any time of the day or night; sunset was chosen because, extrapolating from the works of Evans and Shimazaki, this was a time when large velocities might be expected.

We shall demonstrate that, in general, the transport velocities become most negative at ionospheric sunset, and that they are strongly correlated with the rapid cooling of the electrons.

II. Description of the Experiment

1. Backscatter Summary

Gordon (1958) proposed the basic theory for the incoherent backscattering experiment based on Thomson scattering of radio waves by ionospheric free electrons. Subsequent investigations by Salpeter (1960), Buneman (1961), Bowles (1961), Fejer (1961), Hagfors (1961), Renau et al (1961), and others have shown that scattering involves both electrons and ions and that the total backscattering cross section appropriate to the F region is

$$\sigma = \sigma_e \left(\frac{1}{1+T_e/T_i}\right) \qquad \text{II-1-1}$$

where σ_e is the classical Thomson electron cross section and T_e and T_i are the electron and ion temperatures. Doppler broadening of the returned signal occurs because of the charged particle thermal motions. The frequency spectrum of the backscattered power depends on T_e, T_i and the masses of the various ionic species present (which is simply 0^+ in the upper F region). Thus, by power and spectral measurements of the backscatter signal the electron density and the electron and ion temperatures can be determined. The present investigation is, however, primarily concerned with directly measuring the large scale vertical motion V_z of the plasma being probed by noting that such a motion will give rise to a uniform Doppler shift of the returned frequency spectrum according to the relation

$$V_z = \frac{c\Delta f}{2\ f_o} \qquad \text{II-1-2}$$

where Δf is the amount the spectrum is shifted, and f_o is the transmitted frequency (430 MHz).

2. Facility and Procedures

The Arecibo Ionospheric Observatory, where these data were obtained, employs a 1000 foot diameter reflector to transmit a 1/6 degree wide beam at 430 MHz and to receive the backscattered signals. The radar equipment, its operation for ionospheric backscatter measurements and the procedures for data gathering and reduction have been described by Gordon and LaLonde (1961); Carlson (1965) and Wand and Perkins (1965).

The measurements used in the present study were obtained by using 500 μs pulse transmissions at a pulse repetition period of 64 ms. Spectra were taken at altitudes of 300 km, 375 km, 450 km, and 525 km. (Note: Since a 500 μs pulse was used these altitudes really reflect an average over a 75 km interval). Typical integration times for the spectra were 15-25 minutes.

The standard ionospheric radar set-up aliases (folds) the returned signal at several different places. Aliasing the signal makes it symmetrical about the fold-over frequency and, thereby, destroys any Doppler shift information which might be contained within the signal. The elimination of aliasing was the only procedural problem encountered and resulted in two slight modifications of the standard set-up.

The Doppler shift of the returned signal Δf was found by the following formula (see Figure 1)

$$\Delta f = \frac{(f_h^{(1/2)} - f_o) - (f_o - f_\ell^{(1/2)})}{2} \quad . \qquad \text{II-2-1}$$

where $f_{\ell}^{(1/2)}$ = the half-power frequency of the low frequency side of the spectrum and $f_h^{(1/2)}$ = the half-power frequency of the high frequency side of the spectrum. The relative power of the spectra as a function of height is printed out in steps of 500 cycles. (The spectra are about 10kc wide.) The half-power frequencies were arrived at by linear interpolation between the two appropriate printed out frequencies. Figure 1 shows that this assumption of linearity of the spectra over the range of interest is an exceedingly good one.

3. Error Evaluation

The errors indicated in Figures 4-6 are due solely to statistical fluctuations in the returned signal and the noise. Perkins and Wand (1965) indicate that such fluctuations are the overriding source of inaccuracy in the system and give the root-mean-square of deviation of the spectra to be

$$\delta \simeq \left(\frac{4}{N}\right)^{1/2} \qquad \text{II-3-1}$$

for signal-to-noise ratios above unity, where N is the number of pulses in an integration period, i.e. the number of samples being analysed. Equation II-3-1 is the determining equation for the error bars in the figures mentioned above. (Note: we did not analyse the data taken at 525 km precisely because the signal-to-noise ratio feel substantially below unity in most cases and δ is then too large for the data to be interpreted meaningfully).

The basic assumption which allows us to consider statistical fluctuations as the only real source of error is that the incoming noise and the noise of the system across the frequency range is "white" or flat. If the noise is systematically non-flat the present data reduction program would give rise to a false or "equipmental" Doppler shift. Such a shift would have a constant value at all heights and at all times throughout the night. Thus the relative changes in the deduced velocities would remain unchanged, but a correction would have to be applied to their absolute values. Although this is not felt to be a problem, a very useful addition to the present procedure would be to make use of the "idler" channel of the parametric amplifier. The idler channel contains a signal which is the mirror image of the signal channel. Thus a real shift one way in the idler channel corresponds to a shift in the <u>opposite</u> direction in the signal channel. However, the "equipmental" shift would be in the <u>same</u> direction for both signal and idler channel. Thus the spurious shift could be measured and removed. This technique will be tried this coming summer (1968).

III. Theoretical Background

1. Purpose

In the Introduction it was noted that the vertical transport velocity in the F region was a result of several distinct mechanisms: diffusion, electrodynamic drifts, expansion and contraction of the atmosphere due to temperature variations, and neutral winds. In this section the necessary theory for each of these mechanisms is reviewed.

2. The Continuity Equation

The basic equation employed in all models of the F region is the continuity equation for electron or ion density

$$\frac{\partial N_e}{\partial t} = q - L - \nabla \cdot (N_e \underset{\sim}{V}) \qquad \text{III-2-1}$$

where, as before, N_e is the electron or ion number density (charge neutrality is assumed), q and L are the effective production and loss rates of electron-ion pairs, and $\underset{\sim}{V}$ is their transport velocity. In general, electrons and ions obey separate continuity equations, but in the F region these two equations are identical. The reason for this identity is twofold: 1) only one major ionic species is present (O^+), thus q and L refer equally to either electrons or ions; and 2) no currents are believed to flow, thus $\underset{\sim}{V}_i = \underset{\sim}{V}_e = \underset{\sim}{V}$ (Fejer 1965).

3. Production and Loss

Before reviewing transport velocity mechanisms, a brief discussion of production and loss in the F region seems appropriate.

Above 180 km electron density production is due almost entirely to the photoionization of atomic oxygen by solar ultraviolet radiation. (Although molecular nitrogen and molecular oxygen ions are being produced at rates comparable to that of atomic oxygen, they recombine with electrons so rapidly that their average density is very small (Donahue 1965)). At this height the solar flux is essentially unattenuated. The production rate q is, then, the product of the solar flux, the ionization cross sections, and the atomic oxygen number density integrated over all wavelengths. It is usually assumed that q is a constant times the atomic oxygen density.

Atomic oxygen-electron pairs are lost primarily according to the following:

$$O^+ + N_2 \rightarrow NO^+ + N \qquad \text{(rate } \gamma_1\text{)}$$

$$O^+ + O_2 \rightarrow O_2^+ + O \qquad \text{(rate } \gamma_2\text{)}$$

followed by dissociative recombination

$$NO^+ + e \rightarrow N + O \qquad \text{(rate } \alpha_1\text{)}$$

$$O_2^+ + e \rightarrow O + O \qquad \text{(rate } \alpha_2\text{)}$$

The rate of change of O^+, NO^+ and O_2^+ are governed by

$$\frac{dn(O^+)}{dt} = q - \gamma_1 n(O^+)n(N_2) - \gamma_2 n(O^+)n(O_2)$$

$$\frac{dn(NO^+)}{dt} = \gamma_1 n(O^+)n(N_2)-\alpha_1 n(NO^+)N_e$$

$$\frac{dn(O_2{}^+)}{dt} = \gamma_2 n(O^+)n(O_2)-\alpha_2 n(O_2{}^+)N_e$$

since

$n(O^+) + n(O_2{}^+) + n(NO^+) = N_e$ and assuming equilibrium conditions, these governing equations can be solved to give an effective loss rate L:

$$L = \frac{\alpha_1\alpha_2[\gamma_1 n(N_2)+\gamma_2 n(O_2)]N_e{}^2}{\alpha_1\alpha_2 N_e+\alpha_2\gamma_1 n(N_2)+\alpha_1\gamma_2 n(O_2)} \quad . \qquad \text{III-3-1}$$

In the F region the denominator of equation III-3-1 is dominated by $\alpha_1\alpha_2 N_e$, thus L can be simplified

$$L = [\gamma_1 n(N_2) + \gamma_2 n(O_2)]N_e = \beta_{eff} N_e \qquad \text{III-3-2}$$

where $\beta_{eff} = \gamma_1 n(N_2) + \gamma_2 n(O_2)$

We now turn our attention to the theory of transport velocities.

4. Transport Velocities

Following the development of Doupnik (1967), Holt and Haskell (1965, chapters 6 and 7) and Dougherty (1961), we can derive an expression for the transport velocity. The treatment is one in which a continuum model is used. The plasma is considered as an interacting mixture of electron, ion and neutral-molecule fluids in a constant magnetic field. Using such a model, the momentum-conservation law can be written as

for electrons:
$$M_e \frac{\partial}{dt}(\underset{\sim}{V}_e N_e) + N_e \mu_{en} \nu_{en} (\underset{\sim}{V}_e - \underset{\sim}{V}_n) + N_e \mu_{ei} \nu_{ei} (\underset{\sim}{V}_e - \underset{\sim}{V}_i) + M_e N_e (\underset{\sim}{V}_e \cdot \nabla) \underset{\sim}{V}_e = -N_e q [\underset{\sim}{E} + \underset{\sim}{V}_e X \underset{\sim}{B}] - M_e N_e \underset{\sim}{g} - \nabla (N_e \; kT_e) + \underset{\sim}{V}_e M_e (q-L) \qquad \text{III-4-1}$$

for ions:
$$M_i \frac{\partial}{dt}(\underset{\sim}{V}_i N_e) + N_e \mu_{in} (\underset{\sim}{V}_i - \underset{\sim}{V}_n) + N_e \mu_{ei} \nu_{ei} (\underset{\sim}{V}_i - \underset{\sim}{V}_e) + M_i N_e (\underset{\sim}{V}_i \cdot \nabla) V_i = +N_e q [\underset{\sim}{E} + \underset{\sim}{V}_i \; X \underset{\sim}{B}] - M_i N_e \underset{\sim}{g} - \nabla (N_e \; kT_i) + \underset{\sim}{V}_i M_i (q-L) \qquad \text{III-4-2}$$

where the electron and ion pressures have been replaced by $N_e kT_e$ and $N_e kT_i$ respectively; and where M_e and M_i are the electron and ion mass, ν is the indicated collision frequency, μ is the indicated reduced mass, and q is the electronic charge. $\underset{\sim}{E}$ is an assumed electrostatic field due to both external sources and to slight charge separation, and $\underset{\sim}{B}$ is the geomagnetic field.

In the F region it is known that

a) the ion-neutral collision frequency $\nu_{in} <<$ ion gyro-frequency, Ω_i,

b) $\mu_{en}\nu_{en} << \mu_{in}\nu_{in}$.

In addition we will assume that

a) the time scales for bulk changes are much longer than the collision intervals, so that accelerations are negligible

b) the geomagnetic field lines are linear, have a dip angle I below the horizontal and lie in a north-south plane.

c) the neutral wind $\underset{\sim}{V}_n$ is horizontal

d) there are no horizontal gradients

e) as stated before, no currents flow ($\underset{\sim}{V}_e = \underset{\sim}{V}_i$).

Under these conditions, equations III-4-1 and III-4-2 can be added together and, choosing a coordinate system such that the x-axis is to the south, the y-axis is to the east and the z-axis is upward, the vertical component of the ambipolar transport velocity can be found. The result of the calculation is

$$V_z = V_{dz} + W_{nz} + W_{dz} \qquad \text{III-4-3}$$

where $$V_{dz} = -\sin^2 I \; \frac{k(T_e + T_i)}{M_{in}\,\nu_{in}} \left[\frac{1}{N_e}\frac{\partial N_e}{\partial z} + \frac{M_i g}{k(T_e+T_i)}\right] \qquad \text{(a)}$$

$$W_{nz} = V_{nx} \sin I \cos I \qquad \text{(b)}$$

and $$W_{dz} = \frac{E_y}{B} \cos I \qquad \text{(c)}$$

This has a useful physical interpretation: V_{dz} is the vertical component of the ambipolar diffusion velocity through a stationary neutral atmosphere - since $\nu_{in} << \Omega_i$, the diffusion is solely along field lines; W_{nz} is the vertical component of the plasma being pushed up or down the field lines by the meridional (north-south) component of the neutral wind (V_{nx}); and W_{dz} is the vertical component of the $\underset{\sim}{E} \times \underset{\sim}{B}/B^2$ drift velocity.

The electric field $\underset{\sim}{E}$ in the F region is associated with the electric field produced in the dynamo region of the atmosphere (100 to 150 km) and is essentially the same (Farley, 1959; Martyn 1953). $\underset{\sim}{E}$ in the dynamo region is found from the geomagnetic daily variations on the ground. The central idea of the dynamo theory is that $\underset{\sim}{V}_n$, the neutral wind in the dynamo region (assumed to arise from horizontal pressure variations due mainly to solar heating) pushes ions across the field lines. Electrons are largely "tied" to the field lines because their gyro-frequency is much higher than their collision frequency with the neutral gas. An electric current $\underset{\sim}{J}$ responsible for the variations in $\underset{\sim}{B}$ then results. (Note: it is the neutral winds in the dynamo region that cause the electrostatic field in the F region, not the neutral winds in the F region itself.)

The method of computing the dynamo's electric field can be quickly outlined following Kato and Maeda (1966). The basic equations are Ohm's law and Maxwell's equations

$$\underset{\sim}{J} = \underset{=}{\sigma} (\underset{\sim}{E} + \underset{\sim}{V}_n \times \underset{\sim}{B})$$

$$\nabla \cdot \underset{\sim}{J} = 0$$

$$\nabla \times \underset{\sim}{E} = 0$$

which enable us to know the relation between $\underset{\sim}{V}_n$ and $\underset{\sim}{E}$ or $\underset{\sim}{V}_n$ and $\underset{\sim}{J}$

$$\nabla . \{\underline{\underline{\sigma}} \, (E + \underset{\sim}{V} \times \underset{\sim}{B}\} = 0$$

$$\nabla \times \{\underline{\underline{\sigma}}^{-1} . \underset{\sim}{J}\} = \nabla \times (\underset{\sim}{V}_n \times \underset{\sim}{B})$$

where $\underline{\underline{\sigma}}$ is the conductivity tensor, $\underset{\sim}{V}_n$ is the dynamo neutral wind and $\underset{\sim}{B}$ is the total magnetic field.

The usual procedure is to infer a global distribution of $\underset{\sim}{J}$ from the many magnetogram stations throughout the world that record daily geomagnetic variation $\Delta\underset{\sim}{B}$; the complicated relationship between $\Delta\underset{\sim}{B}$ and $\underset{\sim}{J}$ has been worked out be several workers (Chapman and Bartels, 1940; Maeda, 1955; and Kato, 1956). Assuming one knows $\underline{\underline{\sigma}}$, then a global distribution of $\underset{\sim}{J}$ determines a global distribution of $\underset{\sim}{E}$. The vertical component of the induced drifts in the F region W_{dz} has been computed by Maeda and Kato (1966) and is shown in Figure 2.

An alternative to the wind driven dynamo theory is the suggestion that electric fields are generated in the magnetosphere and transmitted down along geomagnetic field lines to produce the dynamo currents (Fejer, 1965). The necessary theory, however, has not been developed.

In a very interesting article, Dougherty (1961) has shown that collisions between drifting charged particles and the neutral gas in the F region greatly alter the simple picture of $\underset{\sim}{E} \times \underset{\sim}{B}/B^2$ drifts. If one assumes that the F region neutral wind is initially zero, then within 20 minutes to one hour the neutral wind would be set in motion by the horizontal component of the plasma drift.

If one additionally assumes that the final result of the collisions is to make the horizontal components of $\underset{\sim}{V}_i$ and $\underset{\sim}{V}_n$ equal, then both W_{dz} and W_{nz} disappear as does the $\sin^2 I$ term in front of V_{dz}. The physical picture becomes one of ambipolar diffusion without a geomagnetic field.

Until recently very little was known about the neutral winds in the F region and, consequently, they were usually set equal to zero (as we have just seen in the previous paragraph). In 1967, Kohl and King calculated a model wind system for this region. The driving force was again the horizontal variation of atmospheric pressure due to solar heating. Molecular viscosity, the coriolis force and ion-drag were included in the equations of motion. Unfortunately, the effects of the electrostatic electric field and the $\underset{\sim}{E} \times \underset{\sim}{B}/B^2$ drifts were neglected. The result of the calculation imply an average neutral wind that is horizontal and has a magnitude of about 100 m/s. The vertical component of the plasma transport velocity arising from this wind is shown in Figure 3.

Our discussion of transport velocities is not yet complete. Equations III-4-1 and III-4-2 make no allowance for the effect of a sudden temperature change (i.e. they are independent of $\frac{\partial T}{\partial t}$). T. Shimazaki (1959, 1966) has pointed out that the motion of expansion and contraction of the atmosphere due to temporal temperature variations should be an important mode of the transport velocity in the F region at times near sunset and sunrise. We will quickly sketch Shimazaki's development.

The basic equation is

$$\frac{d \log p}{dh} = \frac{-1}{H} \qquad \text{III-4-4}$$

where p is atmospheric pressure, H is the scale height, and H is the altitude (and equals z in our coordinate frame). Integration of Equation III-4-4 gives

$$p = p_a \, e^{-\int_a^h \frac{dh}{H}} = p_a \, e^{-Z} \qquad \text{III-4-5}$$

with

$$Z = \int_a^h \frac{dh}{H}$$

where the subscript a refers to a reference altitude, in this case the altitude at which T (temperature) is taken to be almost constant throughout the day (about 100 km). From Equation III-4-5 we see that Z is a constant on an isobaric surface. The motion of the isobaric surface is thus

$$Z = \text{constant or } \frac{dZ}{dt} = 0$$

which gives the velocity of the surface as

$$V_T = -\overline{H} \int_a^h \frac{\partial}{\partial t} \left(\frac{1}{H}\right) dh = -\frac{\overline{H}k}{Mg} \int_a^h \frac{\partial}{\partial t} \left(\frac{1}{T}\right) dh \qquad \text{III-4-6}$$

where $\overline{H}$ is the average scale height over the integration range. From Equation III-4-6 it can readily be seen that V_T has a non-uniform height variation and becomes largest at times of sudden temperature change. For completeness we will also write down the divergence of V_T with respect to z (or h).

$$\frac{\partial V_T}{\partial z} = \frac{1}{H} \left(\frac{\partial H}{\partial t} + V_T \frac{\partial H}{\partial t}\right) = \frac{1}{T} \left(\frac{\partial T}{\partial t} + V_T \frac{\partial T}{\partial z}\right) \qquad \text{III-4-7}$$

This derivation has been for neutral particles; it amounts, in fact to a <u>vertical</u> neutral wind which "blows" charged particles up or down the field lines. The vertical component of this motion is $V_T \sin I$ which we will call W_{Tz}.

This completes the discussion on the relevant processes capable of causing plasma transport. To summarize, then, the vertical component of the ambipolar transport velocity V_z is the result of one or more of the following: V_{dz}, the component due to ambipolar diffusion; W_{nz}, the component due to horizontal neutral winds; W_{dz}, the component due to electrodynamic drift; and W_{Tz}, the component due to sudden temperature changes.

IV. Observations

On August 7, 9 and 15 (1967) the ionospheric transport velocities at heights of 300 km, 375 km, and 450 km were measured by the Doppler shift technique. The velocities are shown in Figures 4, 5, 6, 7, 9, and 11. The following features can be observed:

1. The transport velocity becomes most negative at times coinciding closely to ionospheric sunset.

2. After ionospheric sunset the velocities become less negative and are approximately zero by 21:30 hours L.T.

3. The higher the altitude, the more negative the transport velocities appear to be (during the sunset period).

The ion temperatures T_i (=T_e during the present velocity measurements) were also deduced from the back-scatter spectra and are shown in Figures 8, 10, and 12. The striking feature of these plots is the extremely rapid cooling of the ionosphere during sunset (1900 hours to 2030 hours L.T.) At the altitudes under observation, the ionosphere has undergone the majority of its temperature change by 2030 hrs. L.T. By 2130 hrs. it is approximately isothermal at a temperature of $1000^{\circ}K$.

V. Discussion and Comparison with Other Workers

Examination of Figures 4-12 demonstrate clearly that the vertical transport velocities are most negative when the ionosphere is cooling the most rapidly (also see Table I). The maximum observed velocities occurred on August 9, and it was on this day that the ionospheric temperatures underwent their greatest temporal change. This result is in agreement with the work of Shimazaki, who, as mentioned before, demonstrated that the contraction of the ionosphere should be an important mode of plasma transport at times near ionospheric sunset. It should also be pointed out that, due to a more rapid cooling at the higher heights and a longer path of integration, Equation III-4-6 for V_{Tz} predicts qualitatively the apparent velocity-height variations of the observations.

Our results are also consistent with the strong negative diffusion velocities predicted by Evans (1965) for times near sunset. This is due to a rapid temperature change lowering the scale height $H \left(= \frac{k(T_e+T_i)}{M_i g}\right)$ while the electron density remains (temporarily) unchanged. Evans' prediction can be shown from Equation III-4-3a if one realizes that the term in brackets is almost identically zero and is therefore very sensitive to a decrease in electron temperature T_e or ion temperature T_i if the electron number density and its height gradient $\frac{\partial N_e}{\partial z}$ remain constant. (In our experiment relative values of N_e and $\frac{\partial N_e}{\partial z}$ were available and were found to vary by less than 5% during sunset, while T_e and T_i changed typically by more than 20%.)

In an experiment performed in August of 1965 at Arecibo, P. B. Rao (1968) deduced transport velocities at 400 km from the temporal changes in the measured electron density-height profiles and the continuity equation. He found the vertical transport velocity V_z to be most negative immediately following sunset (with maximum velocities at -15 to -25 m/s) becoming approximately zero by 2200 hrs. L.T. Thus, his results are also in excellent agreement with the present experiment.

The vertical component of the $\underset{\sim}{E} \times \underset{\sim}{B}/B^2$ drifts arising from the dynamo electric field as developed by Maeda and Kato and shown in Figure 2 can be seen to be inadequate to explain the sunset velocities for two reasons: 1) in contrast to the observed velocities, the $\underset{\sim}{E} \times \underset{\sim}{B}/B^2$ drifts change magnitude only slightly throughout the sunset period; and 2) the electrostatic field $\underset{\sim}{E}$ is believed to be constant in the F region and, therefore, electrodynamic drift velocities are independent of height and cannot explain any velocity height variations.

The vertical component of the neutral winds of Kohl and King (1967) can be seen (Fig. 3) to become steadily less negative during the sunset period and, therefore, cannot be used to explain the sudden increase of the downward transport velocity usually observed at that time. It is very interesting to note, however, that later in the evening W_{nz} becomes zero at almost precisely the same time as does the observed transport velocity, i.e. at about 2130 hrs.

Finally, it should be mentioned that the tendency toward zero of V_z later in the evening is a necessary condition if one is to have any chance of explaining the puzzling maintenance of the nighttime F region. The problem is simply that one cannot explain the relatively high nighttime electron concentration and its slow decay from the simple concept of loss and diffusion. The difficulty can be overcome either by assuming a nighttime source of ionization or by introducing a mechanism which is capable of decreasing the decay rate. From Equation III-3-2 we see that the effective loss rate L ($=\beta N_e$) is proportional to the number density of O_2 and N_2, both of which fall off exponentially with height. Thus the decay rate depends on the height of the F region; any mechanism which will shift the F region upward will decrease its decay rate. Diffusion, however, will tend to move the layer downwards and <u>increase</u> the decay rate. Thus we need an overriding process to move the ionization upward. This may be induced by an electric field W_{dz} or caused by a neutral wind W_{nz}. Unless we have a source of nighttime ionization, a negative vertical transport velocity maintained throughout the night is impossible. Hanson and Patterson (1964) have, however, discussed just such an additional ionization source, namely an ionization flux from the exosphere during the night. Yet, their calculations show that the exosphere cannot be replenished during the day in an amount equal to that needed to maintain the F region at night. An all night negative transport velocity would merely aggravate that situation.

Despite the arguments against it, several authors argue in favor of the exospheric flux of ionization:

Yonezawa (1965) and Evans (1965). In short, the problem is still not settled. The present method of measuring the plasma transport velocity should be able to greatly aid in this debate as our Doppler shift technique alone can distinguish between a true ionization transport and an apparent motion caused by the effects of production and loss.

Before concluding let us remark that we have analyzed in detail the vertical component of the transport velocity <u>at sunset during the summer</u> and have been able to say with some certainty which of the various mechanisms available for ionospheric movements were important. This is really, however, a prototype for a future experiment in which the transport velocity will be monitored throughout several 24 hour days during the summer and, perhaps, the winter. By a study of these velocities in a way similar to that done here at sunset we should be able to: 1) add significantly to the present understanding of which mechanisms at what times are important for ionospheric movements and; 2) apply our measured transport velocities to the continuity equation in order to make better estimates of the production and loss parameters q and β_{eff} than those now available.

Conclusions

The Doppler shift technique described and used in this experiment is a new approach to the problem of determining the transport velocities in the ionosphere. It has provided us with <u>direct</u> measurements of these velocities in the F region at times of approximately 1800 hrs. L.T. to 2200 hrs. L.T. The transport velocities appear to be caused by the rapid cooling of the ionosphere. The more rapid the cooling, the more negative the vertical transport velocities. Ionization movements induced by electric fields or driven by neutral winds seem unimportant at times near sunset.

Although the present results are interesting and important in their own right, the most significant aspect of thid work is felt to be the establishment of the Doppler shift technique itself. It seems certain that this technique has the potential to clear away much of the confusion which at present envelopes our knowledge of the F region.

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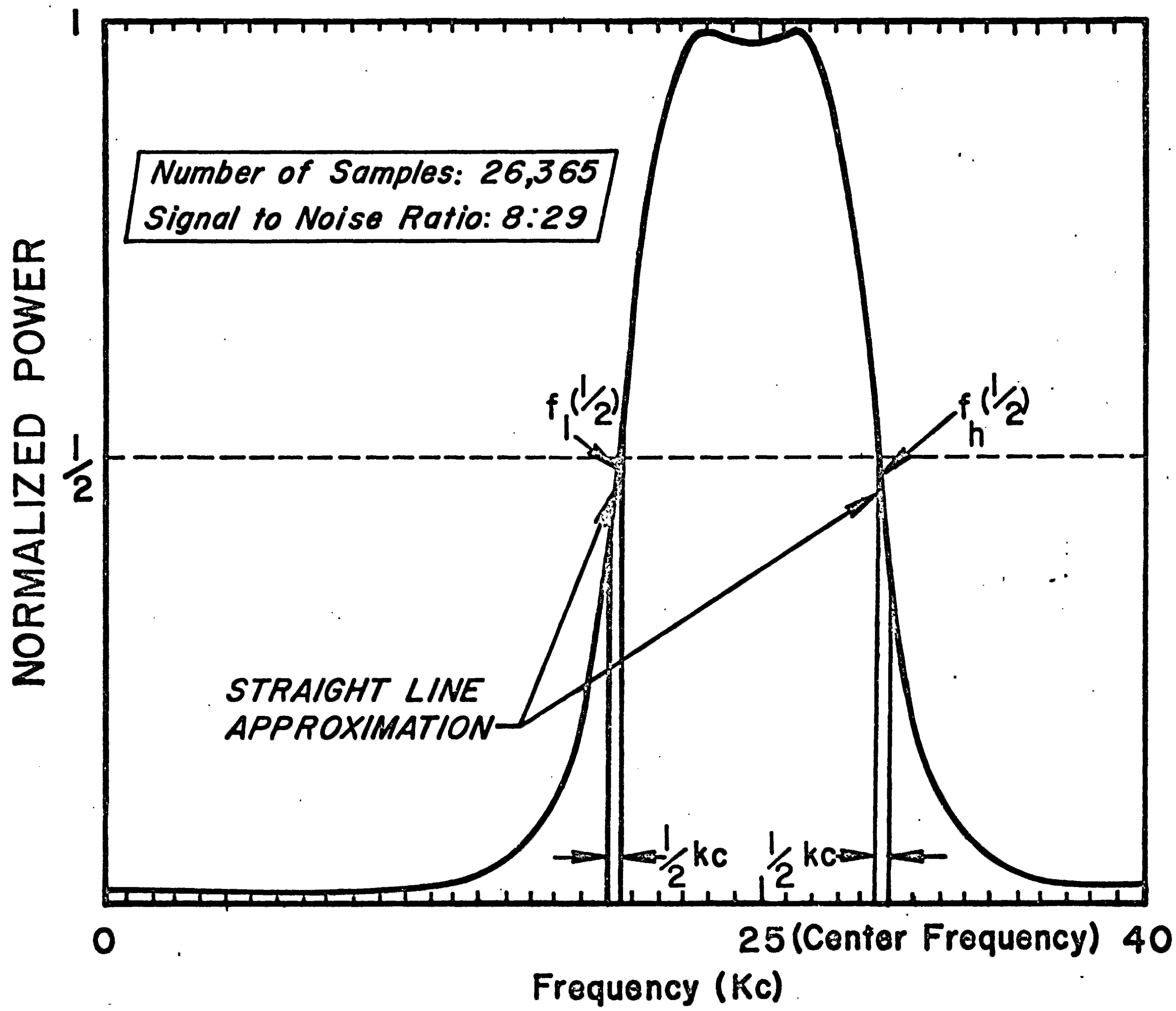
TABLE I

RATE OF TEMPERATURE CHANGE
(°K/min.)

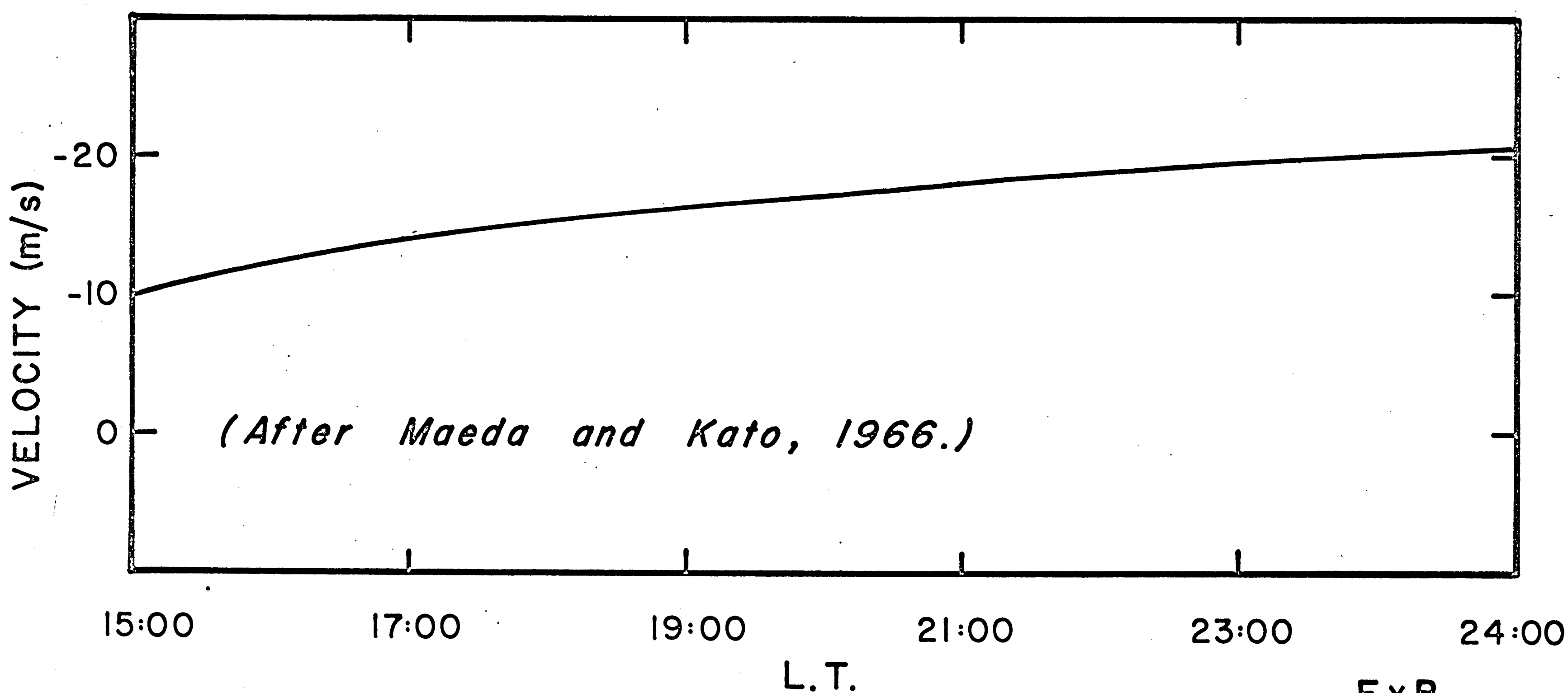
AUGUST 7 TIME	AT 300 km	375 km	450 km
5:45	0.0	+0.5	+0.8
6:33	+1.5	-1.5	-1.3
7:09	-2.2	-3.0	-3.5
7:51	0.0	-0.5	-0.5
8:33	-3.7	-1.4	-2.4
9:09	0.0	0.5	-1.2
AUGUST 9			
5:33	-2.8	-0.3	-2.1
6:09	-0.3	-0.0	-0.7
6:51	-0.3	-3.0	-1.8
7:36	-6.6	-6.6	-6.5
AUGUST 15			
6:36	-1.3	-3.0	+2.5
7:33	-2.4	-1.6	-3.2
8:18	-1.0	-1.0	-2.3
8:51	-0.1	-0.1	+0.1
9:21	-0.3	-0.4	-0.5



TYPICAL SPECTRUM

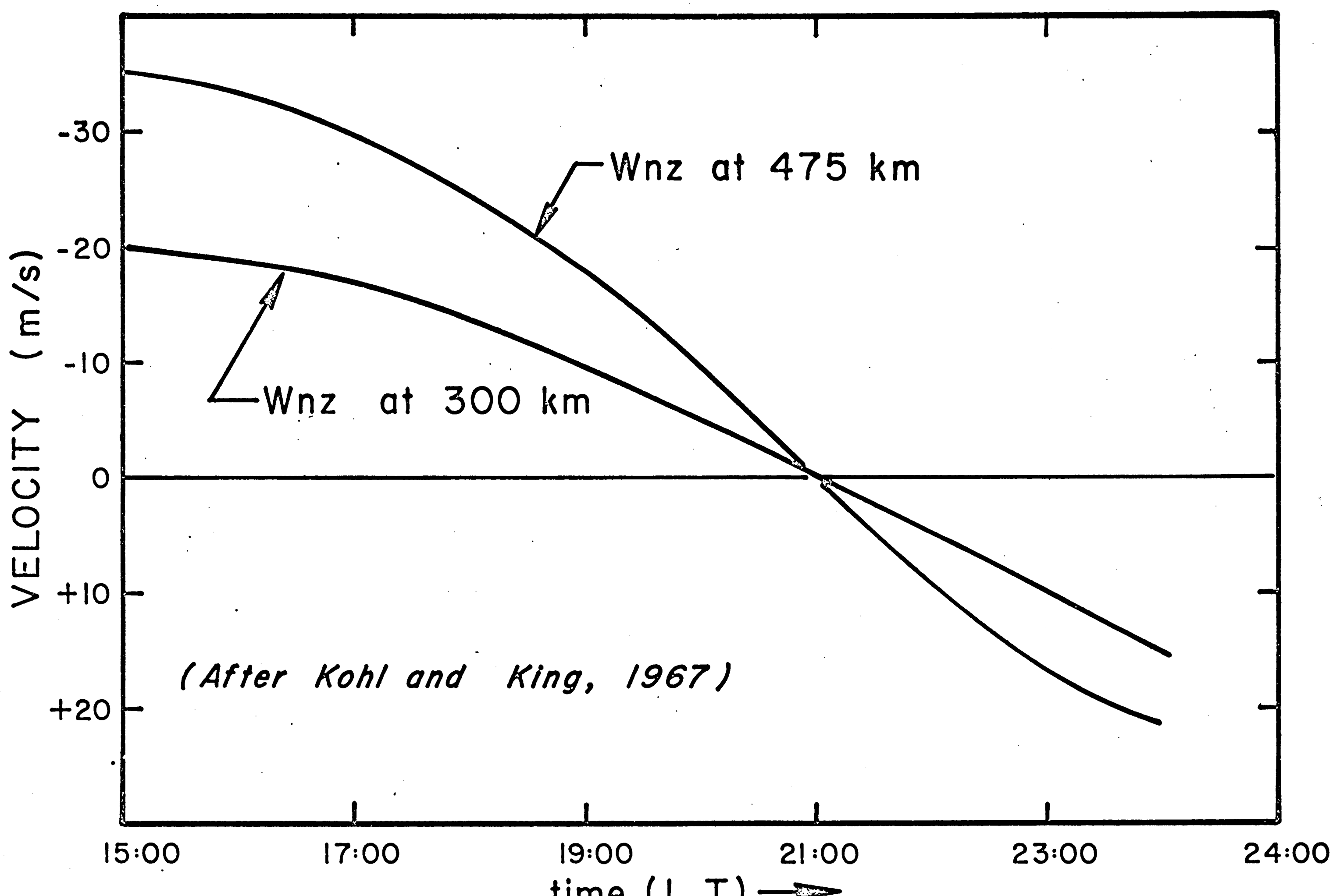
(taken on Aug. 7, 1967, at 20:18 hrs., at 300 km)

Figure 1



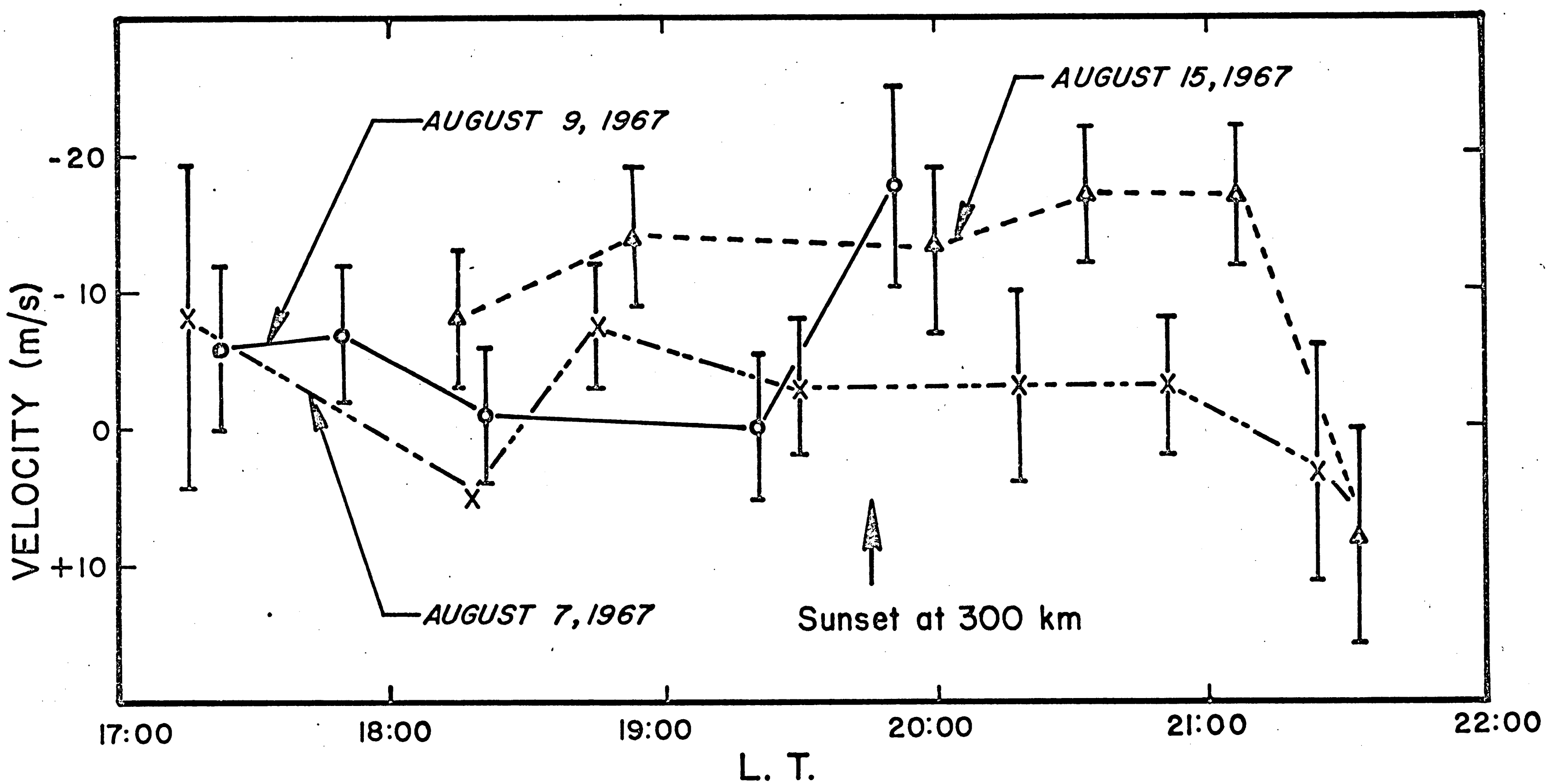
VERTICAL COMPONENT OF TRANSPORT VELOCITY PRODUCED BY $\frac{\underset{\sim}{E} \times \underset{\sim}{B}}{B^2}$ DRIFTS

Figure 2



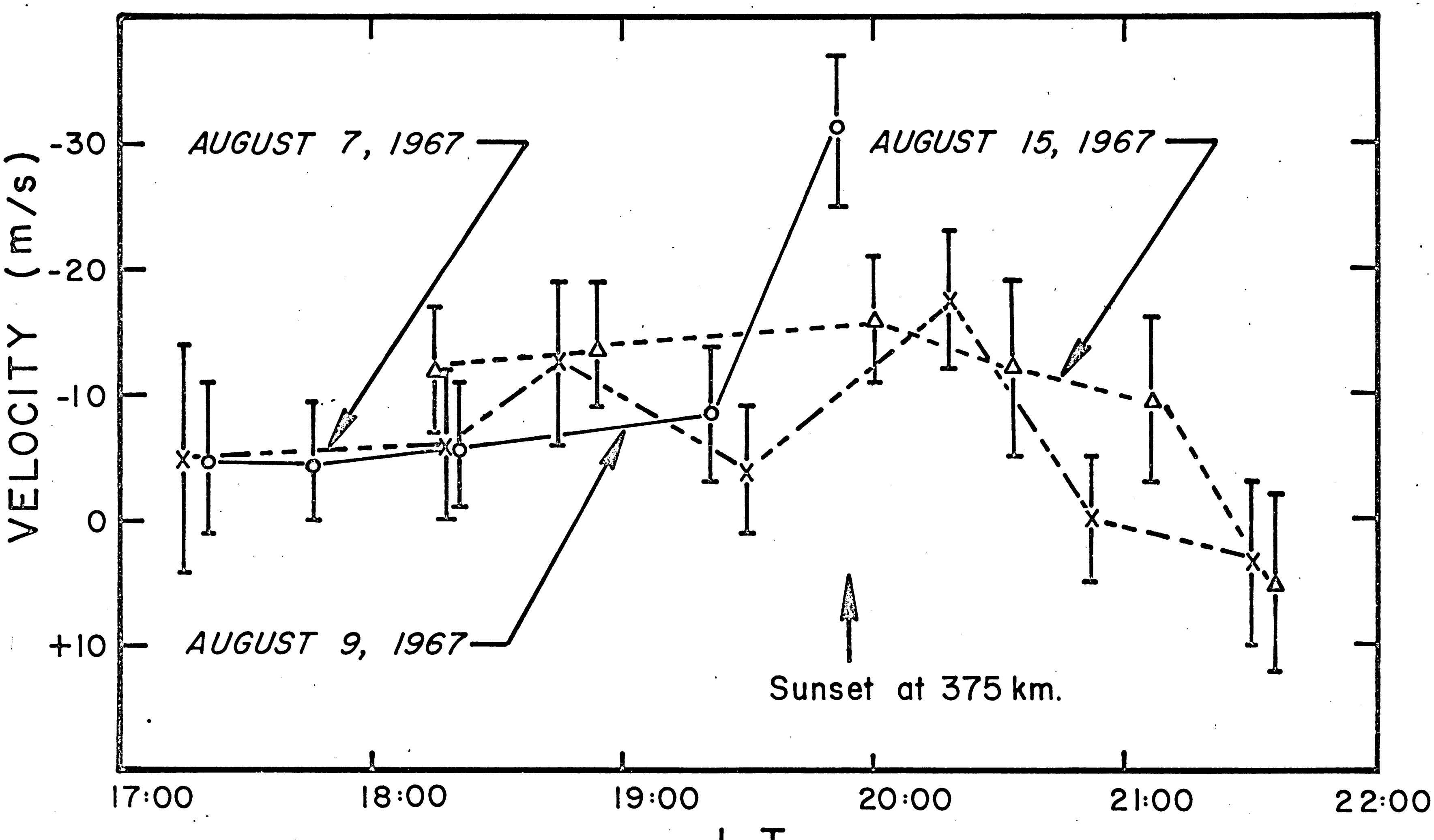
VERTICAL COMPONENT OF THE TRANSPORT VELOCITY
PRODUCED BY NEUTRAL WINDS

Figure 3



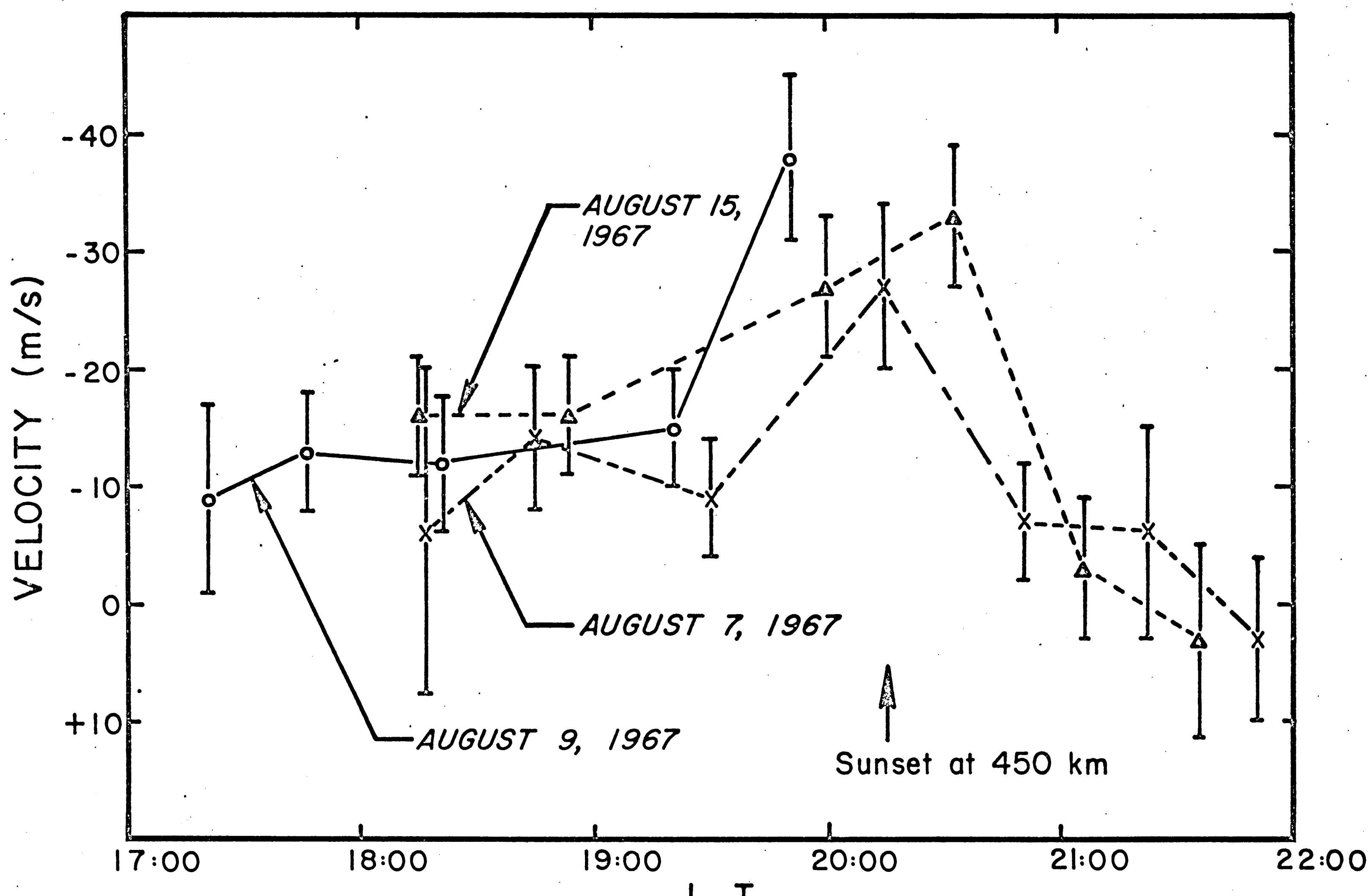
TRANSPORT VELOCITIES AT 300 km

Figure 4



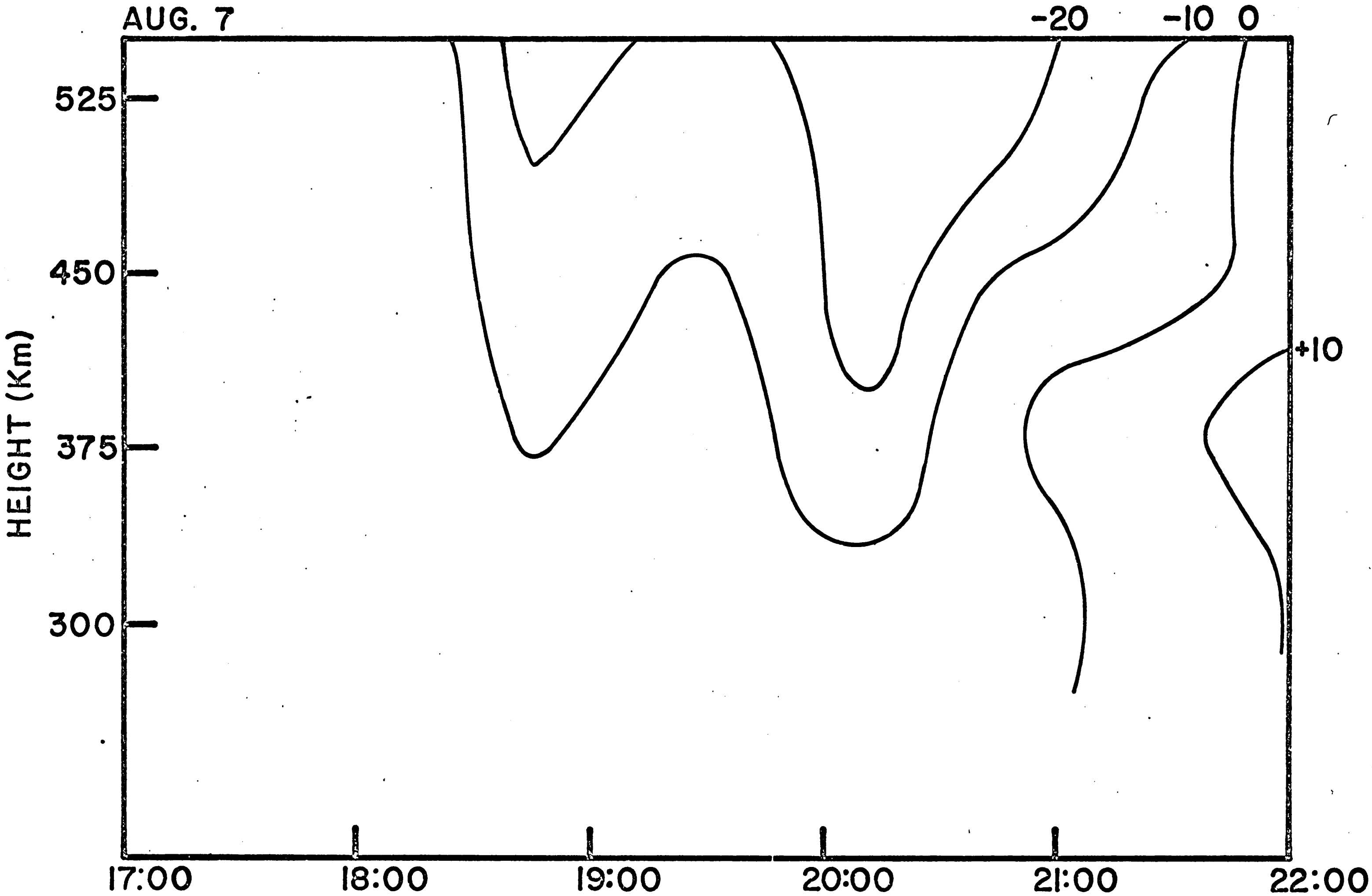
TRANSPORT VELOCITIES AT 375 km

Figure 5



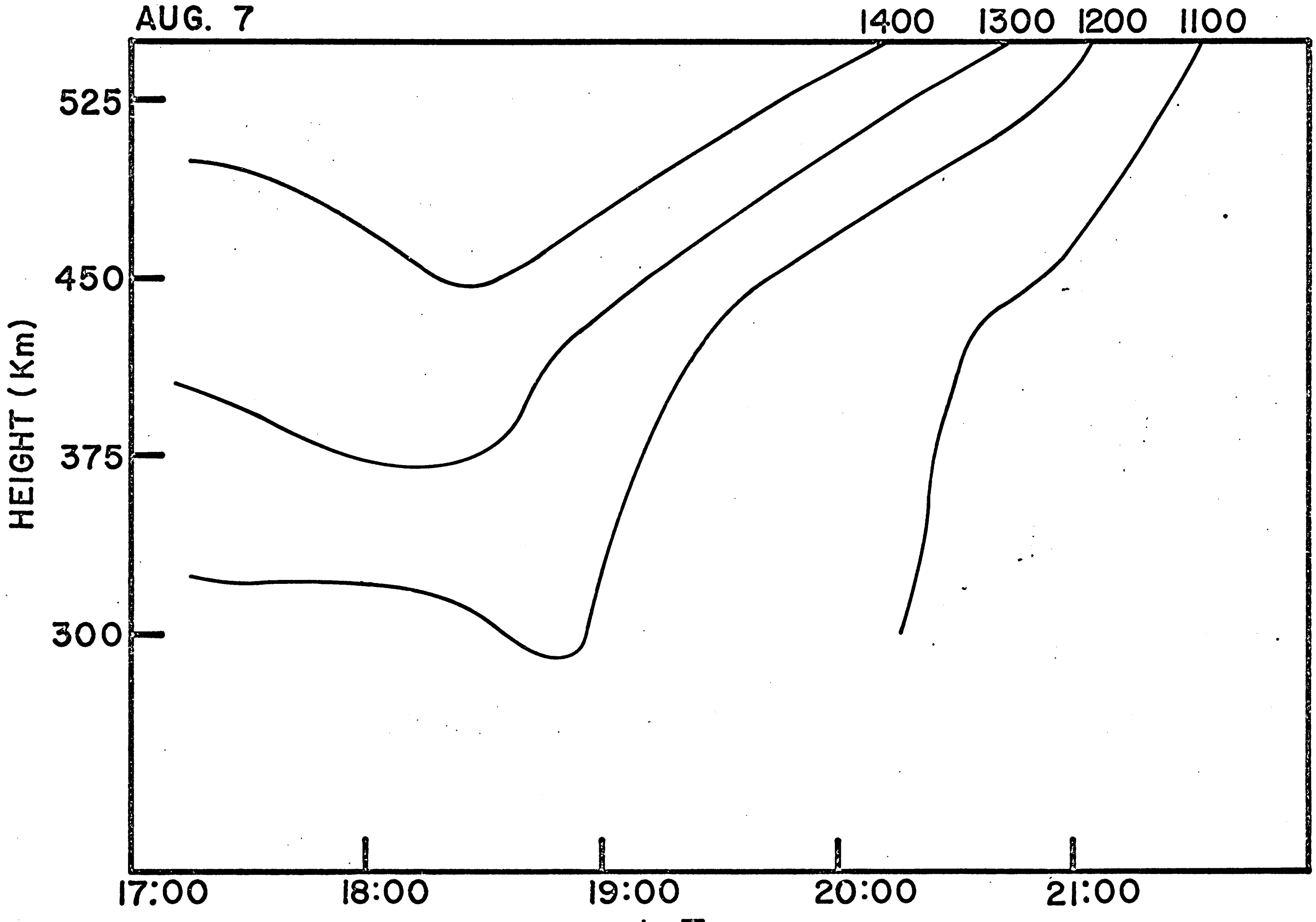
TRANSPORT VELOCITIES AT 450 km

Figure 6

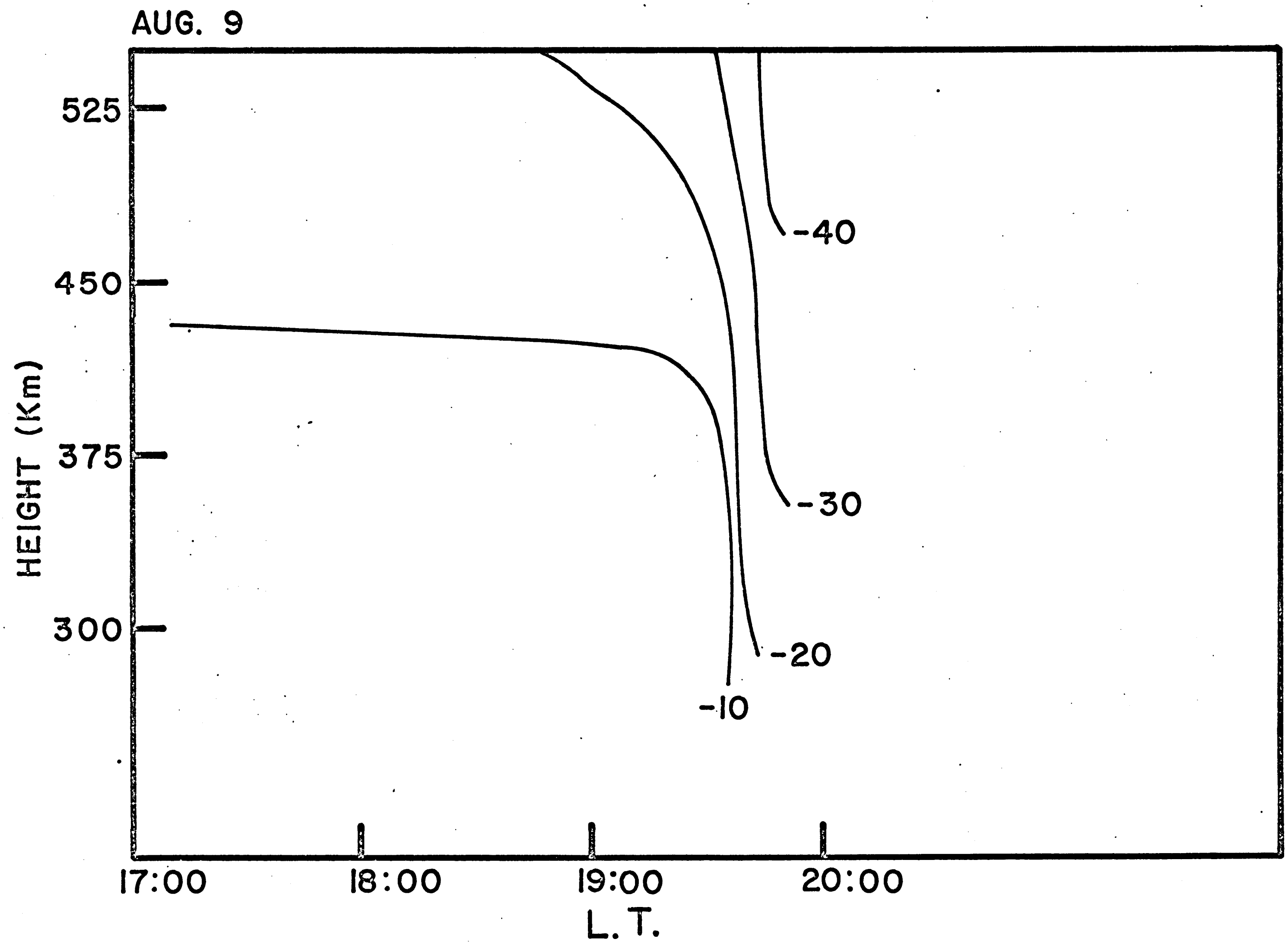


TRANSPORT VELOCITY CONTOURS (m/s)

Figure 7

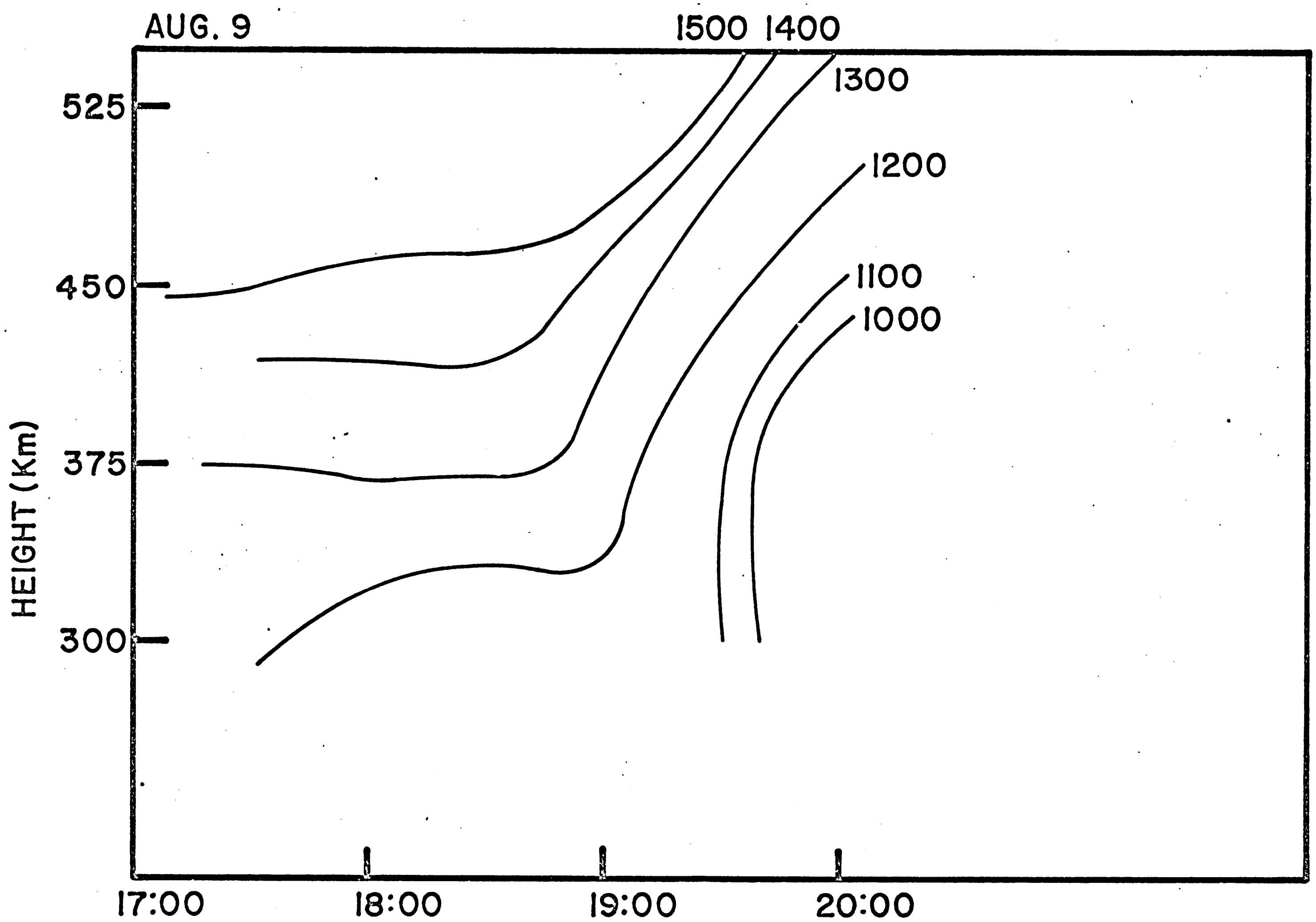


TEMPERATURE CONTOURS
(°K) Figure 8

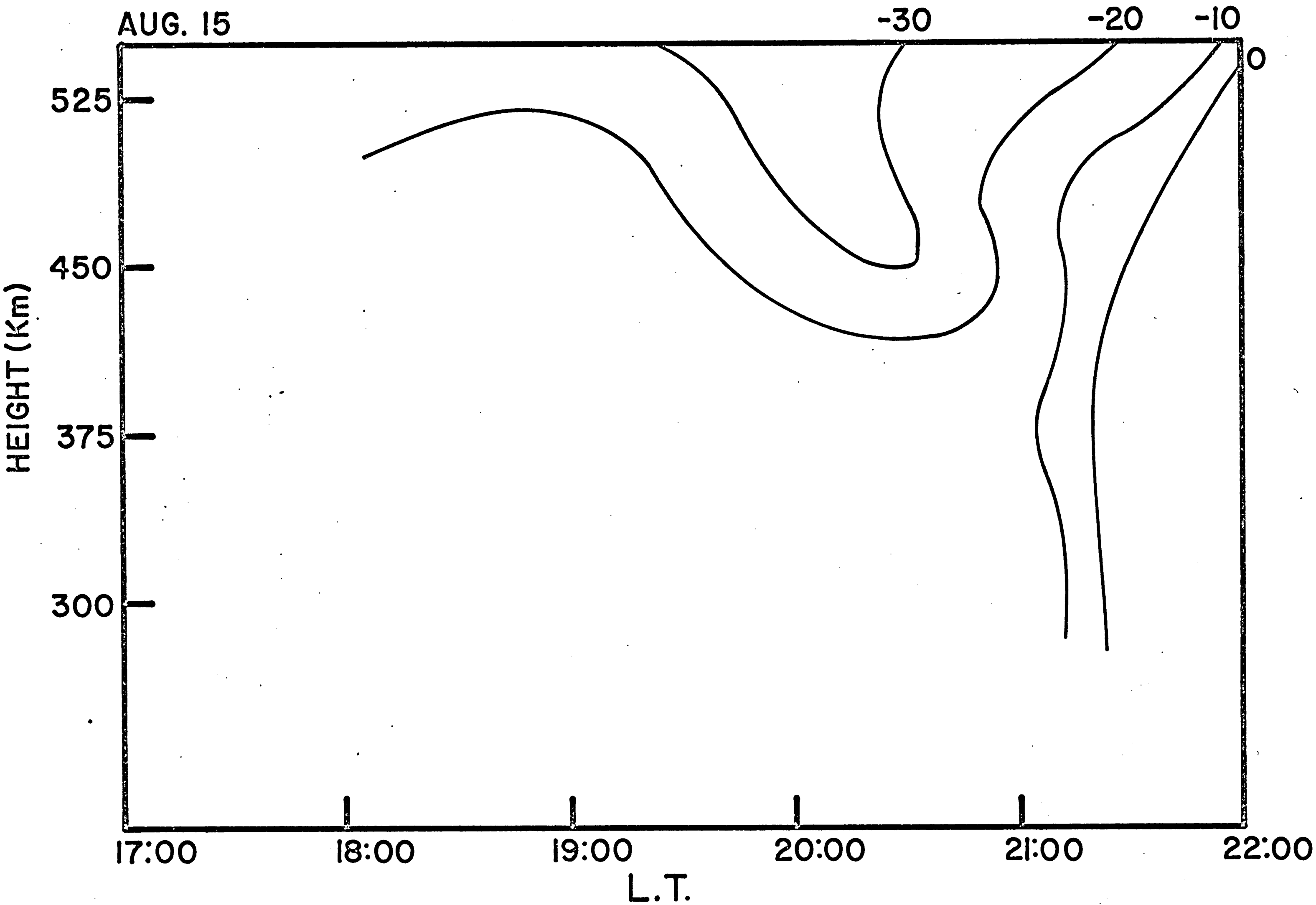


TRANSPORT VELOCITY CONTOURS (m/s)

Figure 9

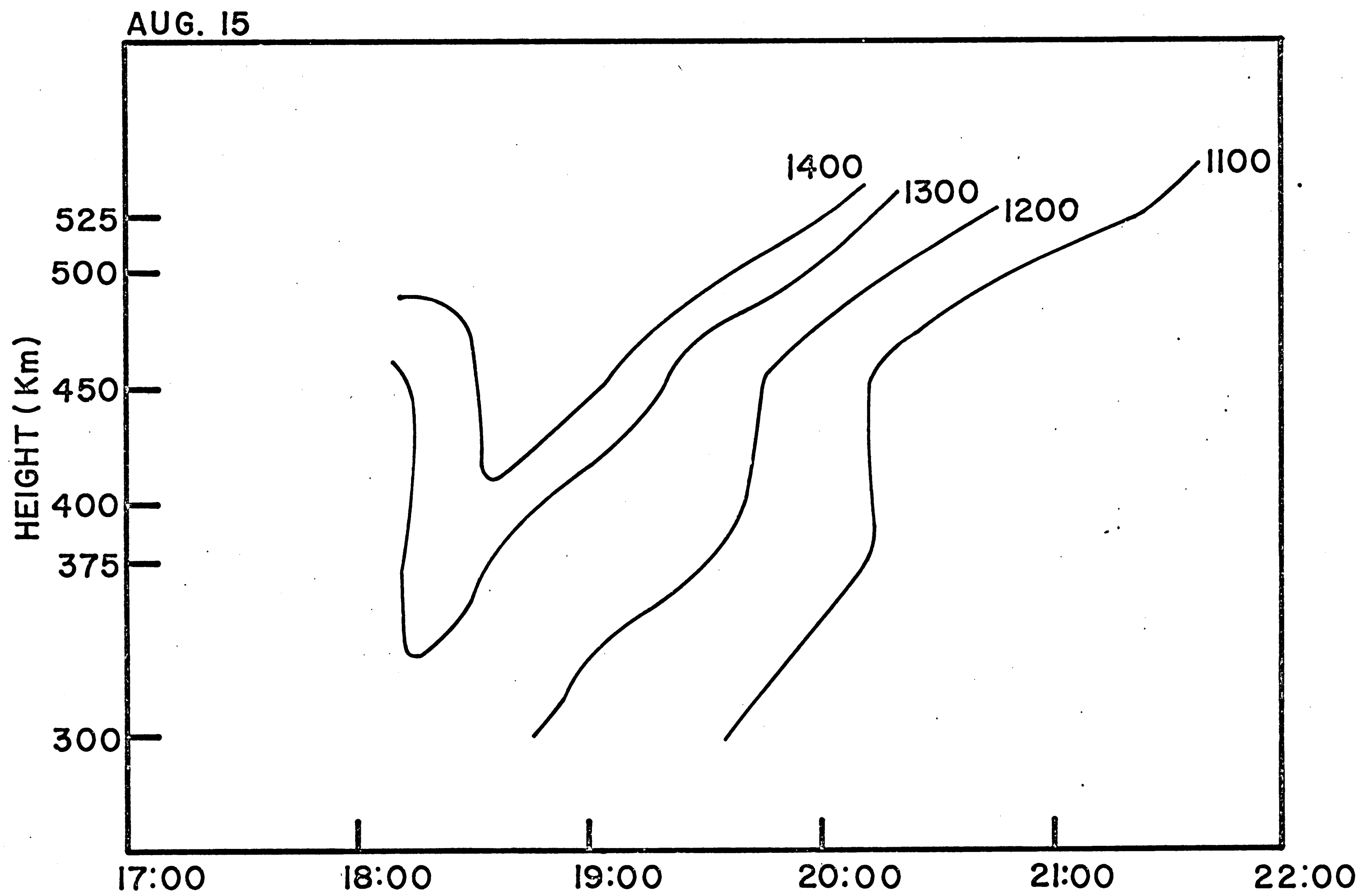


TEMPERATURE CONTOURS
(°K) Figure 10



TRANSPORT VELOCITY CONTOURS (m/s)

Figure 11



TEMPERATURE CONTOURS
(°K) Figure 12