

# A Study of Polar Substorm Related Traveling Ionospheric Disturbances Based on Columnar Electron Content Measurements

by

Michael J. Davis

August 1970

Technical Report N° 12

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Prepared under  
National Aeronautics and Space Administration  
Research Grant N° NGR 05-020-001

**RADIOSCIENCE LABORATORY**

**STANFORD ELECTRONICS LABORATORIES**

**STANFORD UNIVERSITY • STANFORD, CALIFORNIA**



A STUDY OF POLAR SUBSTORM RELATED TRAVELING  
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## ABSTRACT

A detailed study of a particular type of traveling ionospheric disturbance (TID) has been carried out using columnar electron content data. This type of disturbance is characterized by an occurrence related to the degree of geomagnetic activity, a high rate of travel (near  $500 \text{ m s}^{-1}$ ) from the polar regions toward the equator, and the retention of its character over distances of the order of 1000 km. The principal purpose of the present work is to present a detailed picture of the morphology of these disturbances and to relate their behavior, in detail, to a separate pattern of geophysical activity, namely, the occurrence of polar substorms.

These disturbances exhibit a diurnal variation in their direction of arrival at a midlatitude observing site which can be explained in terms of an average source location fixed in local dipole time and dipole latitude and which is in the evening sector of the auroral oval. It is possible to relate the occurrence of these disturbances to the occurrence of polar substorms in an event-to-event, as well as a statistical fashion. In addition, it is possible to determine the location of the source of individual disturbances, and these positions agree very well with the average location of the source.

These disturbances are not generally observed to travel across the equator. This lack of transequatorial propagation is explained on the basis of the high rate of damping of the disturbances in the neutral atmosphere which produce them.

The subject of the source of the disturbance is reviewed and the problem of experimentally studying its character is discussed. It is suggested that any study of the physics of the source be directed toward activity in the evening sector of the auroral oval. A possible source related to the westward traveling surges which are observed in auroras is discussed.

In addition to the study of the morphology of these TIDs, a theoretical study of the ionospheric effects of the atmospheric disturbances which are thought to cause them is undertaken. The effects of

these disturbances on production and loss processes and diffusion are considered. Using various models, numerical results are obtained for the height range 200 to 600 km. Inclusion of the effects of perturbations in loss and diffusion are shown to produce considerable changes in the perturbation in the ionization caused by internal atmospheric gravity waves. In particular, the inclusion of the perturbation in diffusion results in a large depression of the perturbed electron concentration at high levels. The limitations in the models used in the calculations are discussed and stressed. The effect of atmospheric gravity waves on electron content is also considered.

## ACKNOWLEDGMENTS

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The collection of the data used for this study owes much to Mr. S. C. Hall, who coordinated the operation of the network of stations used in making the columnar electron content measurements. It is also a pleasure to acknowledge the assistance of Mrs. W. E. Faulkerson in processing much of the data used here. I would also like to thank Miss M. Budney for translating my scrawl into manuscript form.

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## CHAPTER I

### INTRODUCTION

The fact that the ionosphere frequently exhibits an irregular structure has been known at least since the late 1930's [Booker and Wells, 1938]. It has since become evident that many irregularities move horizontally over considerable distances [Beynon, 1948, Munro, 1948] and the expression "traveling ionospheric disturbance" (TID) is used to describe this phenomenon. Extensive summaries of the early work done in this field have been given by Brown [1959].

It is now apparent that various types of traveling irregularities occur in the ionosphere. Georges [1968a] has demonstrated the existence of several distinct classes of naturally occurring TIDs; two of these have considerable dimensions but show quite different behavior. One group consists of large scale (LS) ionospheric irregularities which are characterized by speeds of some  $500 \text{ m s}^{-1}$ , a direction of travel from north to south in the northern hemisphere, little change in character over distances of 100 km or more, and occurrence related to the degree of geomagnetic activity. Medium scale (MS) traveling disturbances constitute a second category; their speeds are considerably slower, they are of much more localized nature, show no apparent correlation with geomagnetic activity or any other geophysical phenomenon and usually consist of a few quasi-sinusoidal oscillations with "periods" commonly near 15 to 30 minutes. An additional type of disturbance which has large dimensions is that studied recently by Titheridge [1969] in the southern hemisphere. These disturbances are only observed at night during local summer and show no correlation with geomagnetic activity. They are generally confined to the topside ionosphere.

Large scale irregularities have been observed by various techniques, including spaced ionosondes [Munro, 1950; Bowman, 1965], fixed frequency backscatter [Valverde, 1958], measurements of the instantaneous frequency of spaced-path stable high-frequency transmissions [Chan and Villard, 1962], incoherent backscatter [Thome, 1964, 1968; Tesud and Vasseur, 1969], and a network of CW doppler sounders [Georges, 1968a]. The present inves-

tigation was limited to large scale TIDs; they were studied by means of the signatures which they leave in columnar electron content records.

This work is divided into two parts. The first will deal with the characteristics of the large scale TIDs which have been observed using columnar electron content measurements. Chapter II describes the experimental technique used and the basic morphology of the TIDs. Using observed velocities of LS TIDs, it is shown in Chapter II that a model for the location of their source can be constructed. Various features of LS TIDs strongly suggest that their source is related to polar substorms. Chapter III presents a detailed substantiation of that suggestion. In addition, it is shown that the location of the source of individual LS TIDs can be found and that it is in good agreement with the average source location found in Chapter II.

Although, superficially, large scale TIDs often appear to be periodic phenomena, they are actually trains composed of a number of individual, often pulse-like, disturbances. (However, on some occasions there is evidence that an individual TID may possibly be part of a damped oscillation.) The periodicity which is apparent in large scale TIDs is caused by the fact that these disturbances are produced by sources which recur at regular intervals and which, as was mentioned above, are related to polar substorms. Large scale TIDs are generally thought to be due to disturbances propagating in the neutral atmosphere. These disturbances are waves in the sense that they transfer energy from one location to another without transferring material between the locations. However, since the word "wave" is often used to describe phenomena of more marked periodicity than that shown by the TIDs, the term "atmospheric disturbance" will generally be employed in this work in preference to the expression "atmospheric (gravity) wave". Nevertheless, the term "wave" is sometimes used and unless otherwise stated, implies only a mechanism for the transfer of energy.

The rate at which the disturbances in the neutral atmosphere which cause LS TIDs are attenuated as they travel is studied in Chapter IV. The attenuation appears to be high enough to explain the lack of observations of transequatorial propagation of the TIDs. Chapter V discusses the nature of the source of the disturbances. Methods for determining the directional character of the source are considered.

The second portion of this investigation deals with the production of large scale TIDs by disturbances propagating in the neutral atmosphere. Chapter VI is devoted to a study of how atmospheric disturbances in the height range 200 to 600 km affect the local ionization. The effects of the disturbances on the motion of ionization and on the production and loss of electrons are included. The principal result of Chapter VI is a partial differential equation describing the perturbation in the electron concentration produced by a disturbance in the neutral atmosphere. These results are for a general disturbance in the neutral atmosphere. However, in order to obtain numerical results, some models must be adopted. It is particularly difficult to obtain a realistic model for the disturbance in the neutral atmosphere. What will be done is to use a very simple model, a sinusoidal, internal atmospheric gravity wave. Such an approach will allow some insight into the problem, without necessitating a more realistic model to be constructed, a project which is beyond the scope of this work. Chapter VII discusses the models necessary to use the theoretical results of Chapter VI. The limitations in the models used are stressed. Chapter VIII presents the results obtained using the theory and models of the preceding two chapters. The effects of internal gravity waves on the local electron concentration and on electron content are shown. Chapter IX is a summary of important conclusions and a list of suggestions for future work.



## **PART I**

### **CHARACTERISTICS OF LARGE SCALE TRAVELING IONOSPHERIC DISTURBANCES**



## CHAPTER II

### MORPHOLOGY OF LARGE SCALE TRAVELING IONOSPHERIC DISTURBANCES

This chapter deals with the behavior of large scale TIDs observed at midlatitudes. The primary goal is to show that a number of simple but important patterns are present in this behavior. The chapter begins with a discussion of the observational technique used, followed by a description of various characteristics of the disturbances. The chapter concludes with a discussion of the morphology of the disturbances, with emphasis on their source.

#### A. Nature of the Observations

The columnar electron content of the ionosphere can be obtained continuously using measurements of the Faraday rotation angle of a signal from a geostationary satellite [Garriott, Smith, and Yuen, 1965]. In converting from Faraday rotation angle to columnar electron content, a factor of proportionality has been used which will be called G and which is defined by the following equation:

$$\Omega = GI \quad (2-1)$$

where  $\Omega$  is the Faraday rotation angle and G is called the layer shape factor [Almeida, Garriott, and da Rosa, 1970]. Using Eq. (2-1)

$$dI/I = d\Omega/\Omega - dG/G \quad (2-2)$$

which may be approximated by

$$\Delta I/I = \Delta\Omega/\Omega - \Delta G/G \quad (2-3)$$

During local nighttime and daytime  $\Delta G/G$  is usually much less than the other terms in Eq. (2-3) [Almeida et al., 1970]; however, during sunrise and sunset changes in G may become important. Since the principal interest here is in relative fluctuations in electron content and since during sunrise and sunset periods it is difficult to obtain good quality data

relating to TIDs, the diurnal variation in G is of little importance to this work. In addition, based on calculations using the "standard" wave model which will be discussed in Chapter VIII, i.e., a period of 90 minutes, a horizontal wavelength of 3000 km, an amplitude of  $20 \text{ m s}^{-1}$  at 200 km and a direction of travel from dipole north to south, the theoretical changes in G (for Stanford) caused by the presence of a TID are due almost entirely to changes in layer height ( $\pm 30 \text{ km}$ ), the changes in shape of the layer being of secondary importance. Such changes in layer height will produce changes in G of roughly  $\pm 1.0$  per cent, whereas, for the model used, changes in I of  $\pm 10$  per cent occur. Waves with smaller amplitudes would produce correspondingly smaller changes in G. Therefore, since  $\Delta I/I \gg \Delta G/G$  seems to be true during passage of a wave, the effect of any variation in G will be neglected here.

A very simple theory predicts that columnar content is inversely related to the solar zenithal angle during the day. Such a relationship would result in a diurnal content curve with a maximum at local noon and symmetrical about noon. The prediction is only roughly fulfilled as can be seen from the fact that, on most days, the curve deviates quite markedly from such a simple shape. The deviation may be a systematic asymmetry of seasonal character [Yuen and Roelofs, 1966] or it may be a gross distortion during geomagnetic storms [Titheridge and Andrews, 1967]. The curve may be smooth or it may have fluctuations. Here we will be concerned only with the study of fluctuations produced by large scale traveling ionospheric disturbances which are occasionally seen superposed on the background diurnal variation. The electron content curves which serve as basic data for this investigation also reveal quite clearly other types of variations, the most common of which belong to the MS category mentioned in the introduction [Titheridge, 1968]. Figures 2-1 and 2-2 contrast slant electron content curves for two different days. In the first figure the major influence on the electron content is the solar zenithal angle; in the second, large fluctuations distort the curve. In this work, the term "disturbed day" will be used to indicate days when such fluctuations are present. These variations are due to large scale TIDs.

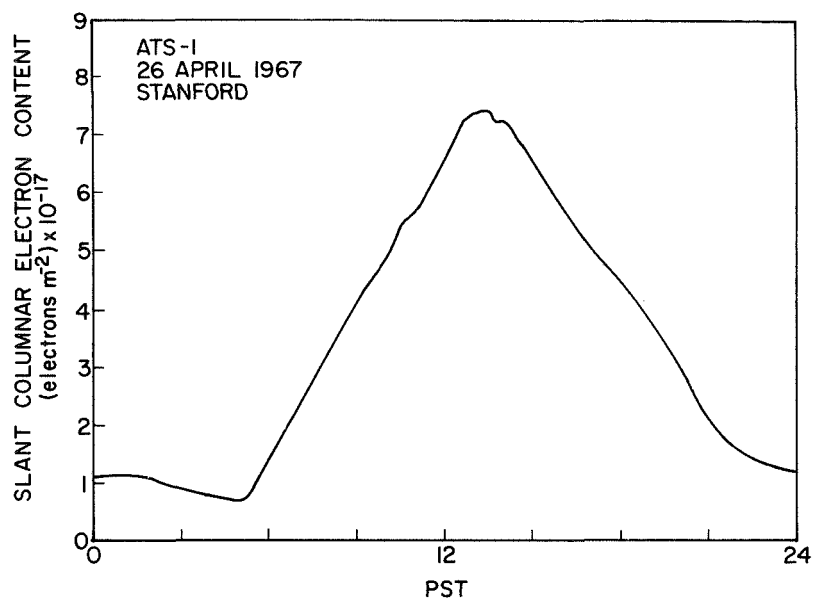


Fig. 2-1. DIURNAL BEHAVIOR OF ELECTRON CONTENT ON A NON-DISTURBED DAY.

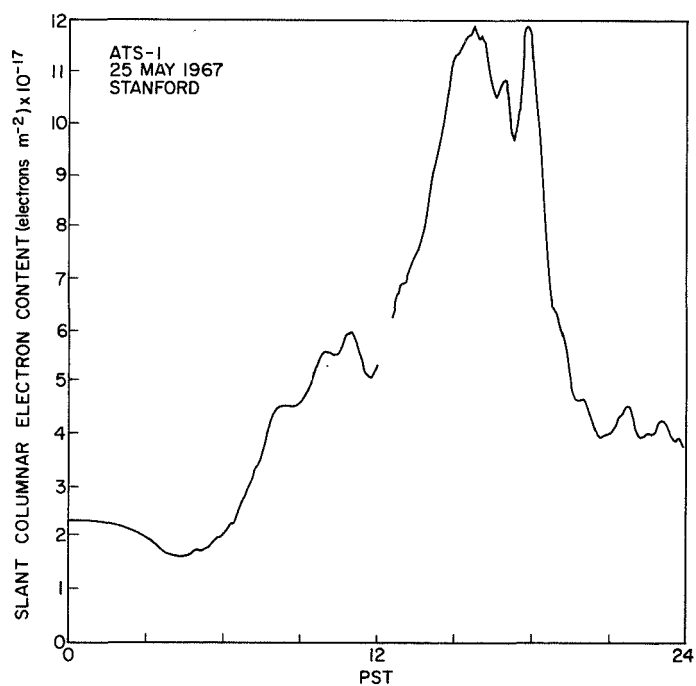


Fig. 2-2. DIURNAL VARIATION OF ELECTRON CONTENT ON A DISTURBED DAY.

In order to study the spatial behavior of LS TIDs, a network of four ground stations observing VHF signals from the geostationary satellite ATS-1 was installed. The location of these observatories, of the sub-satellite point of ATS-I, and of the 300 km sub-ionospheric points for each ray path are shown in Figure 2-3. The resolution in slant electron content of the data recorded is of the order of  $2 \times 10^{15}$  electrons  $m^{-2}$  or better at all the stations.

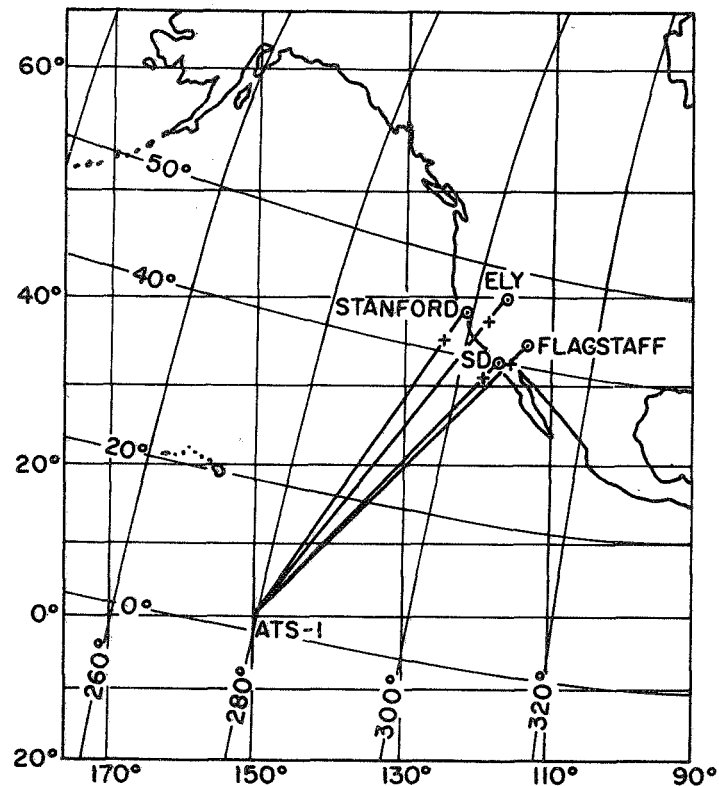


Fig. 2-3. LOCATION OF THE FOUR ORIGINAL GROUND STATIONS. Also shown are the sub-satellite point of ATS-I, and the 300 km sub-ionospheric points (crosses). SD is an abbreviation for San Diego. Both geographic and dipole coordinates are shown.

When one compares the electron content curves obtained at each of the sites on a disturbed day, the similarity of the fluctuations in each of them becomes immediately apparent. Figure 2-4 shows such a comparison in which the curves are arranged in a north-to-south order from the top to the bottom of the figure. A careful examination of the data in

Figure 2-4 would reveal that the fluctuations did not occur in synchronism at all the stations, but that there are definite time shifts between similar features at different stations. These shifts are difficult to see in Figure 2-4 because of the scale used.

In order to facilitate comparisons of the fluctuations, the slow diurnal variation can be eliminated from the data by means of digital filtering; the results of applying this operation to the data in Figure 2-4 are displayed in Figure 2-5. A progressive delay can be observed from station to station suggesting a disturbance traveling southward.

#### B. Characteristics of the Disturbances

Large scale traveling disturbances, as can be seen in filtered records such as those shown in Figure 2-5, produce perturbations in the electron content consisting of a sequence of maximums and minimums. The complete event between two consecutive troughs will be called a "cycle" or, frequently, simply a TID, for reasons which will become apparent later. The duration of a cycle is somewhere between 1 and 3 hours, 1.5 hours being very common. All cycles occurring between two consecutive unperturbed periods constitute a "train of disturbances". We consider as belonging to the same train all cycles separated by quiet periods of relative short duration not exceeding that of about three cycles. A train of disturbances lasts typically 10 to 20 hours, although there are some infrequent examples of much longer trains.

An expanded system of observatories, established in late 1968, which has replaced the one shown in Figure 2-3, and which is shown in Figure 2-6, has provided evidence that the disturbances retain their character while traveling over north-south distances of nearly 2000 km as well as showing strong similarities across 1500 km of their wavefronts. The system of stations shown in Figure 2-6 is equipped to provide data with better resolution in time and electron content than the original

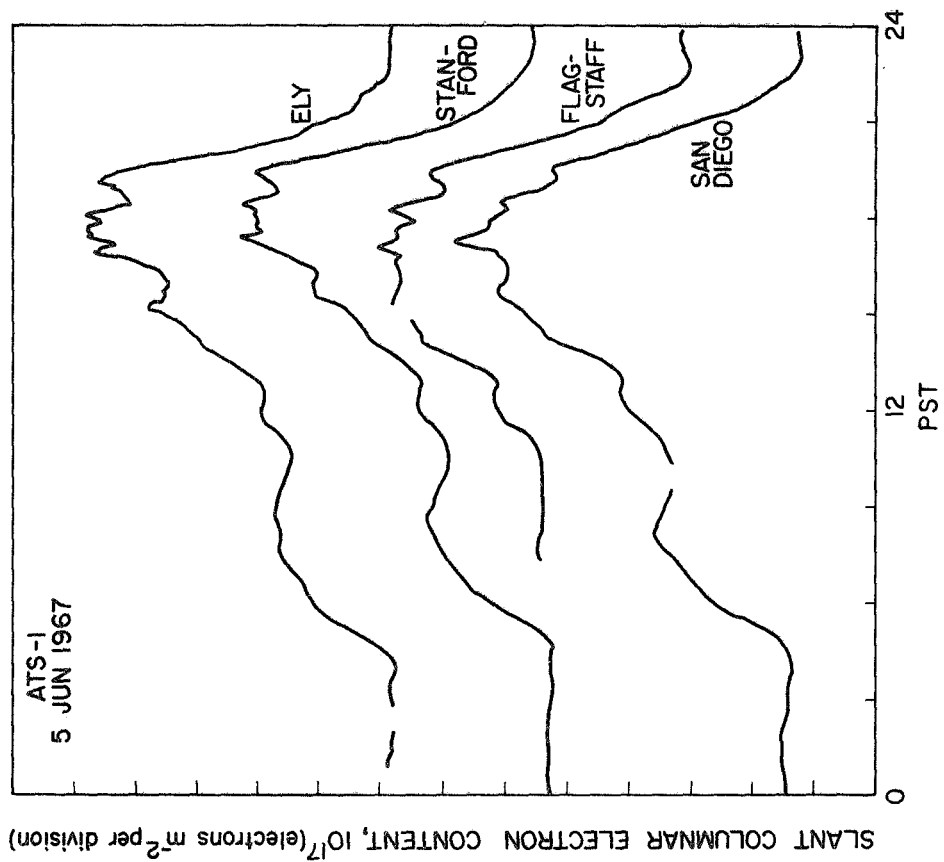


Fig. 2-4. A COMPARISON OF ELECTRON CONTENT OBSERVED AT THE FOUR STATIONS ON A DISTURBED DAY.

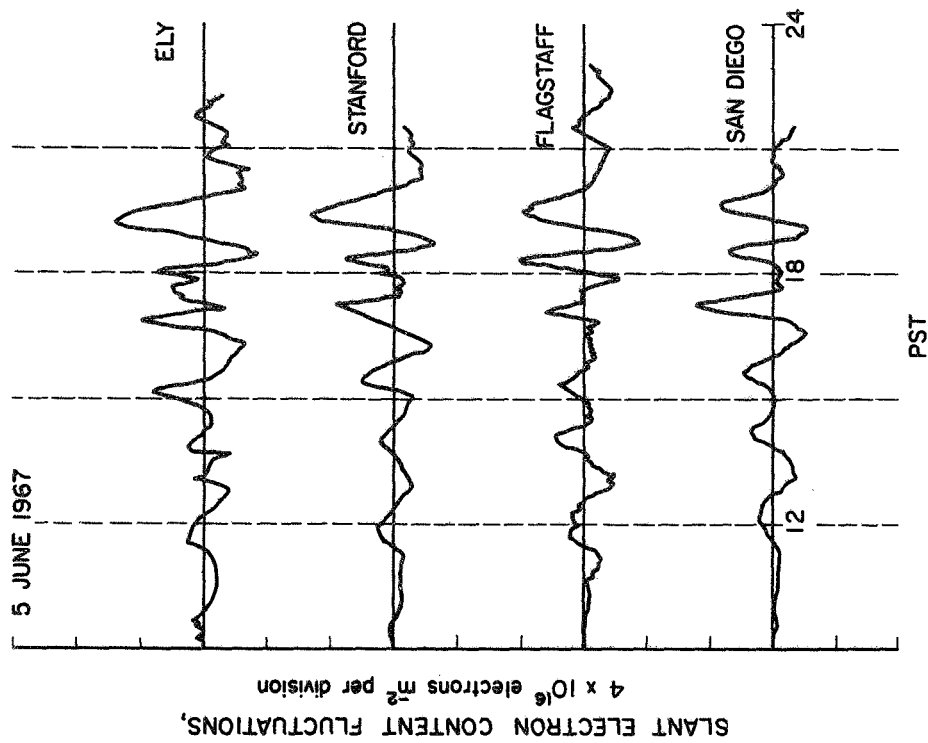


Fig. 2-5. ELECTRON CONTENT FLUCTUATIONS OBTAINED FROM FIGURE 2-4 BY DIGITAL FILTERING.

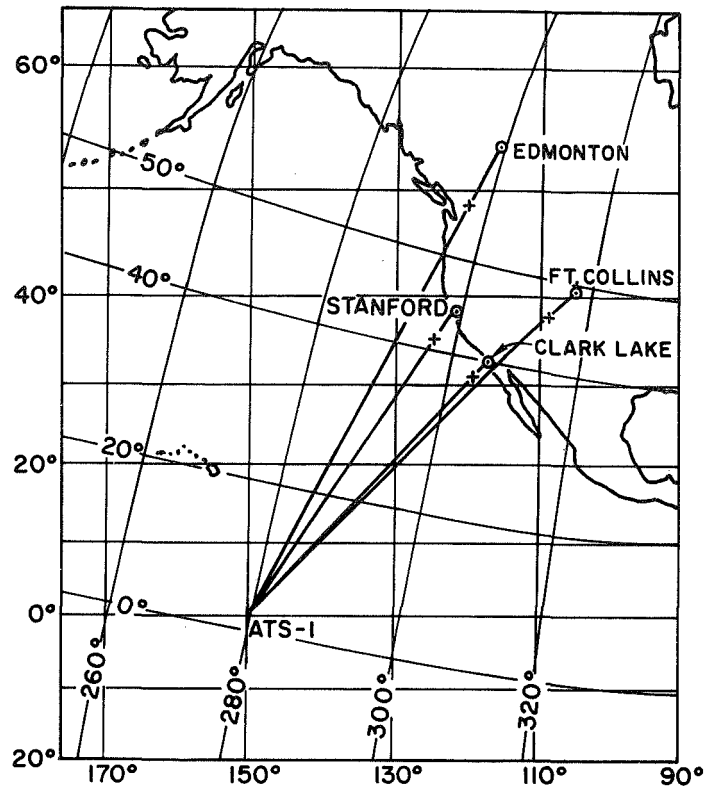


Fig. 2-6. SECOND NETWORK OF STATIONS USED.

system. In addition, it provides greater geographical coverage and perhaps most importantly, allows data to be obtained at a considerably higher latitude than was possible with the first network.

Figure 2-7 shows the times when trains of large scale TIDs were observed during the period from February 1967 through November 1968. Also shown are the principal geomagnetic storms occurring during that period, as reported in Solar-Geophysical Data [passim], and the times when  $K_p$  was 5 or greater. As can be seen, many storms have an accompanying TID and almost all TIDs occur during storms. Disturbances occurred on virtually every occasion on which  $K_p$  was 5 or greater. These results are in good agreement with those of Georges [1968a], who suggested a relationship between  $K_p$  values greater than 5 and occurrence of LS disturbances. However, numerous TIDs were observed during storms when  $K_p$  was below 5. Other workers [Chan and Villard, 1962; Thome, 1968] have also suggested a relationship between enhanced geomagnetic activity and the occurrence of disturbances.

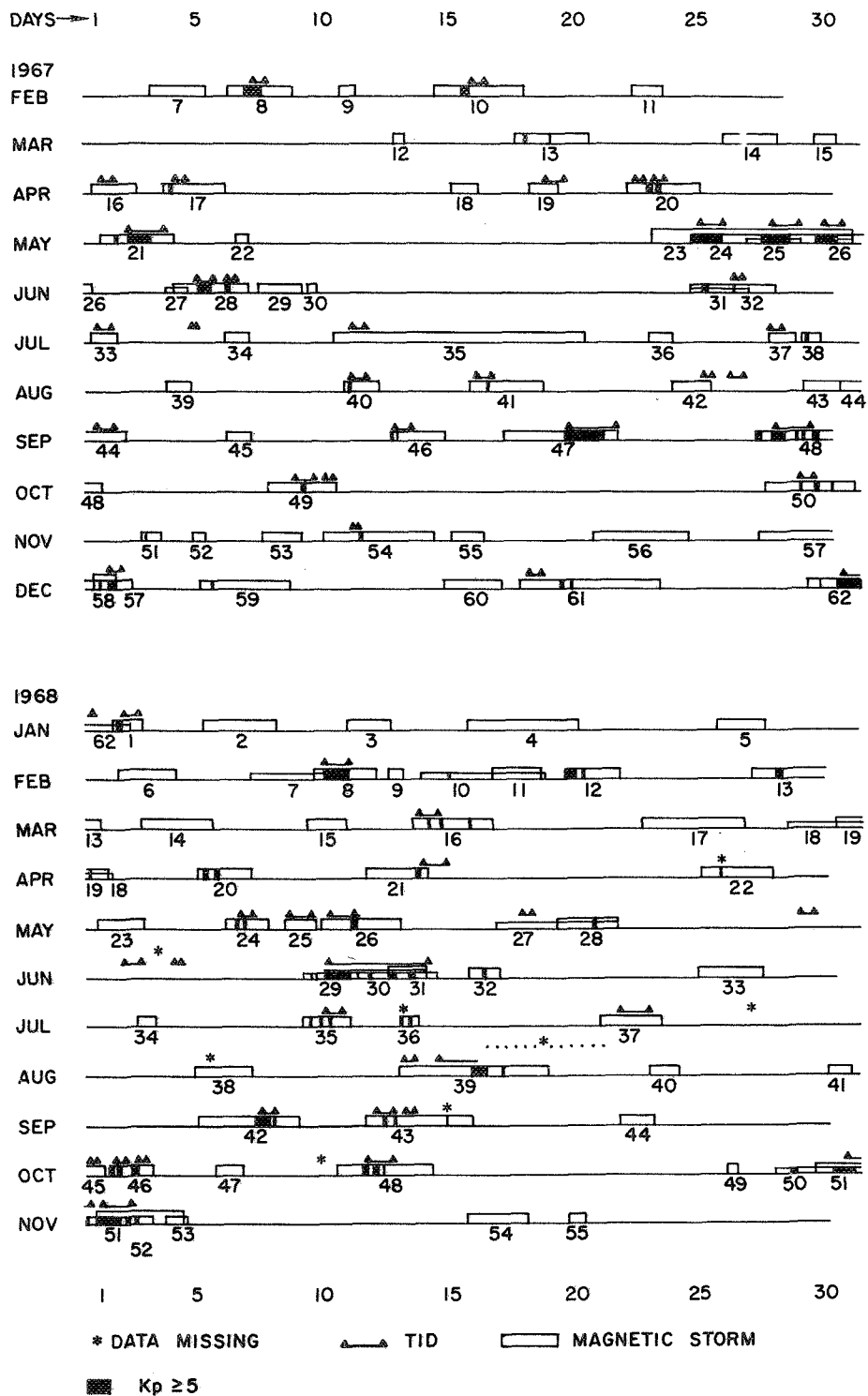


Fig. 2-7. TIMES WHEN LS TIDs WERE OBSERVED AT STANFORD DURING THE PERIOD FEBRUARY 1967 THROUGH NOVEMBER 1968. The numbers under the symbols indicating geomagnetic storms are the numbers of the storms.

TIDs are not observed with equal frequency throughout geomagnetic storms. Each storm during which a disturbance was observed was divided into 20 intervals. The number of times TIDs were observed during each interval was then determined by summing over all the storms. The results are displayed in Figure 2-8. TIDs occur most frequently during the middle of storms with few being observed near the beginning or end.

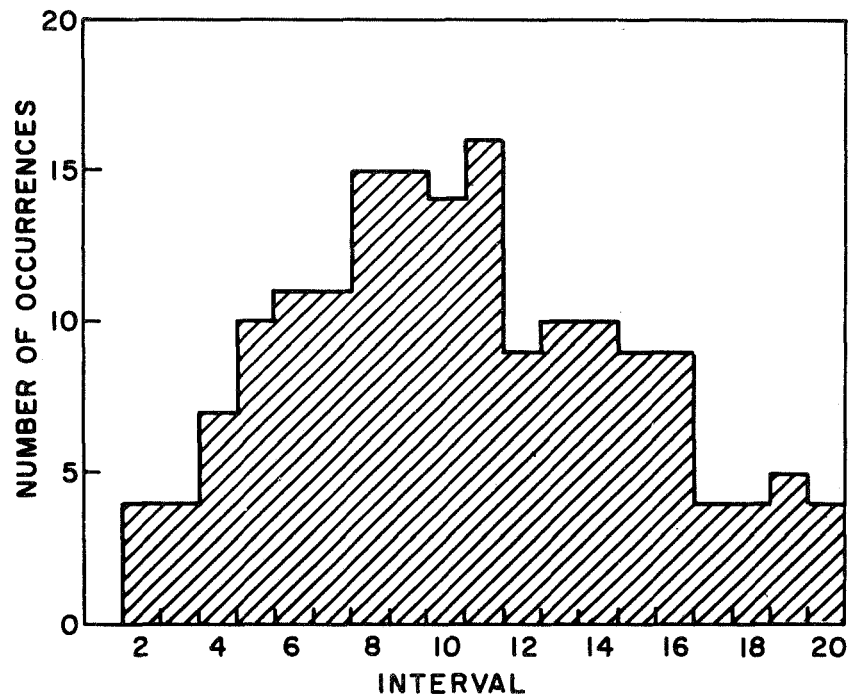


Fig. 2-8. VARIATION IN OCCURRENCE OF LS TIDs AT STANFORD WITH RESPECT TO THE PHASE OF GEOMAGNETIC STORMS.

The disturbances also exhibit a diurnal variation in occurrence. Figure 2-9 shows the number of TIDs observed at Stanford as a function of local time. The distribution is somewhat similar to the diurnal variation curve of the electron content.

The amplitudes of the disturbances vary considerably. On one occasion, during the very intense geomagnetic storm of 25 May 1967, a cycle with a peak-to-trough amplitude of  $30 \times 10^{16}$  electrons  $m^{-2}$  was

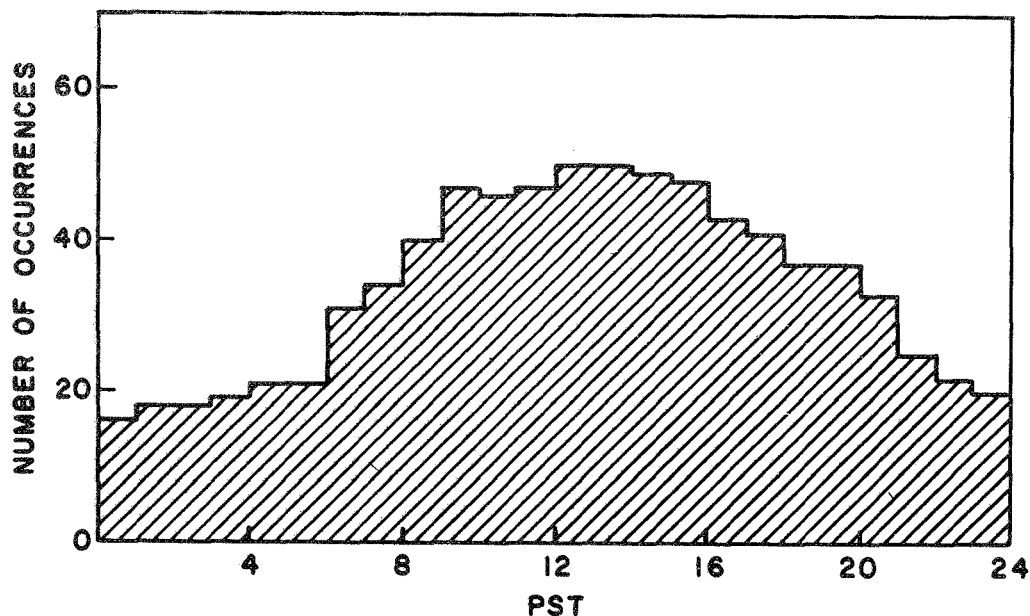


Fig. 2-9. VARIATION IN OCCURRENCE OF LS TIDs AT STANFORD WITH RESPECT TO LOCAL TIME.

observed. However, the amplitudes are usually much smaller. A representative rms amplitude, obtained from the filtered records, for a train of disturbances is  $0.8 \times 10^{16}$  electrons  $m^{-2}$ . Relative amplitude will be defined as the ratio of the rms amplitude of the perturbations to the mean value of electron content for the period of interest. A typical relative amplitude for a train of disturbances is 2%. Assuming that the TIDs are produced by disturbances in the neutral atmosphere, the relative amplitudes of the TIDs are much more meaningful quantities with which to deal than absolute amplitudes, as will be discussed in Chapter VIII.

Since these TIDs are observed primarily during geomagnetically disturbed periods, it is of interest to determine if any relationship exists between the amplitude of the disturbances and the level of geomagnetic activity. Figure 2-10 is a scatter diagram in which the average relative amplitudes of a number of trains of disturbances are plotted

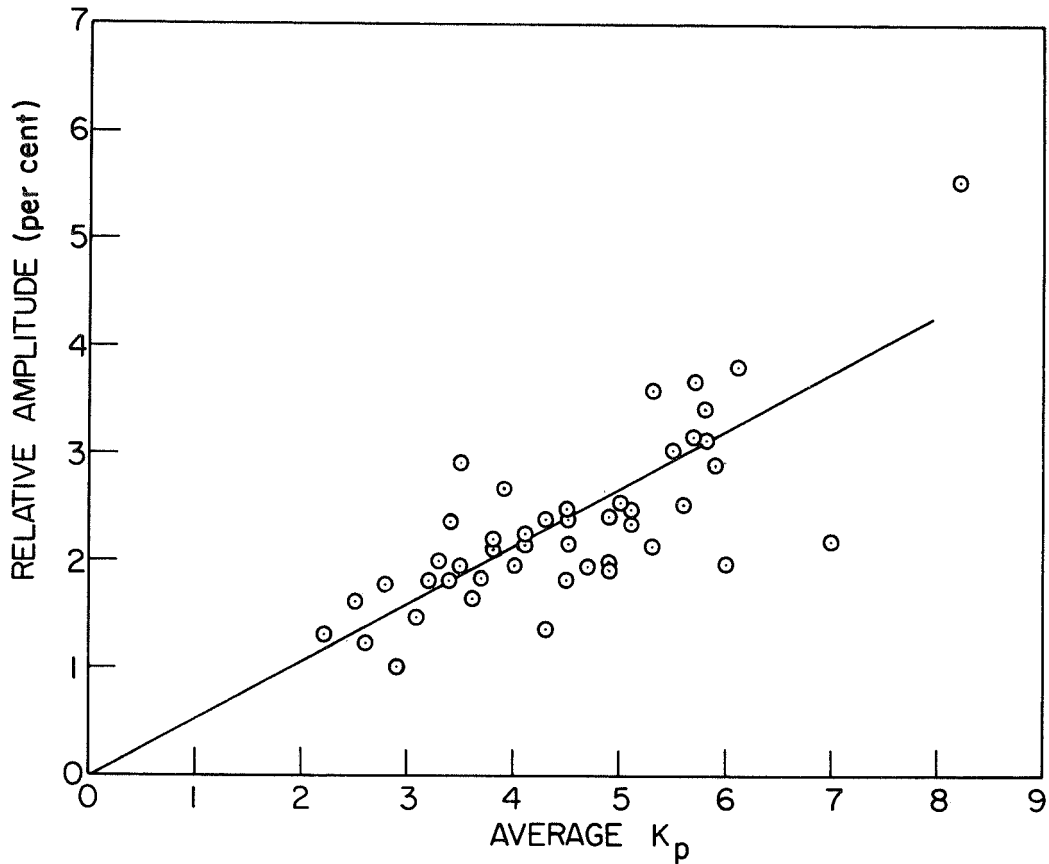


Fig. 2-10. SCATTER DIAGRAM OF AVERAGE RELATIVE AMPLITUDES OF TRAINS OF TIDs VERSUS THE AVERAGE  $K_p$  FOR THE PERIOD WHEN EACH WAS OBSERVED AT STANFORD.

against values of  $K_p$  averaged over the period during which each train was observed. The scatter diagram indicates a strong correlation between the relative amplitudes of the disturbances and the level of geomagnetic activity.

For each disturbance the relative and absolute rms amplitudes were also determined for every two hour interval of the disturbed period. Using these results the average diurnal behavior of relative and absolute amplitudes was determined. Figures 2-11a and 2-11b show the curves obtained. Standard deviations are indicated by the vertical bars.

### C. Speed and Direction of Motion of the Disturbances

Figure 2-5 indicates that the farther south the observer, the later each detail in the electron content fluctuation pattern makes its

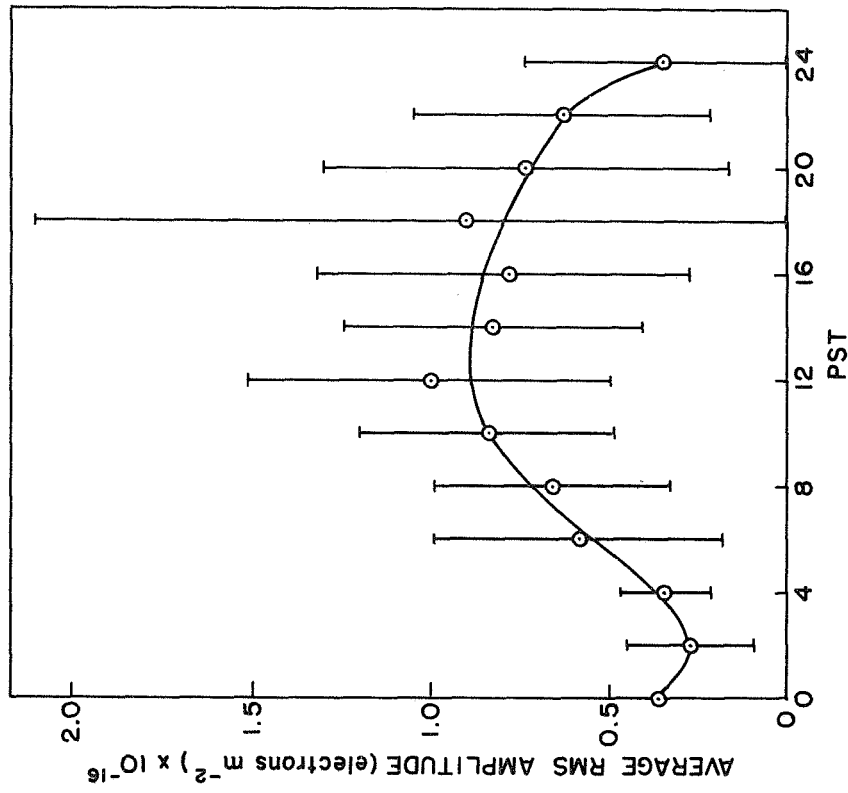


Fig. 2-11(b). DIURNAL VARIATION OF AVERAGE ABSOLUTE RMS AMPLITUDES OF TIDS OBSERVED AT STANFORD.

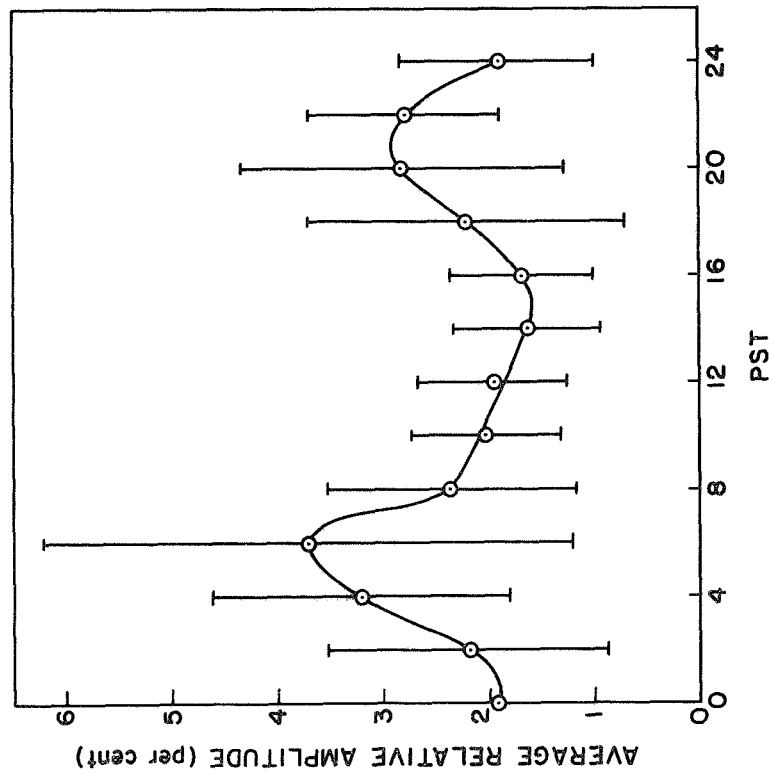


Fig. 2-11(a). DIURNAL VARIATION OF AVERAGE RELATIVE RMS AMPLITUDES OF TIDS OBSERVED AT STANFORD.

appearance. By measuring the time intervals between the occurrence of a given cycle at three different locations, it is possible to determine the direction and the speed of propagation of the disturbance. Since each cycle has its own characteristic shape, differing quite noticeably from that of neighboring cycles, and since this shape is remarkably well preserved from one station to another, the time delays can be determined unambiguously from a simple comparison of similar patterns. Such an approach is found to be easier to apply than determining time delays based on the maximums of the cross-correlation functions of the wave shapes at the different stations. The results obtained using the two methods generally are in good agreement.

In translating time shifts into the magnitude and direction of the velocity of the disturbance, the assumptions of plane wave propagation and a plane earth were made. Note that the last assumption does not mean that the calculated velocities cannot later be used in connection with a spherical earth model when considering distances much larger than those between the stations of the observation network. It is important to point out that the method presented here will only allow the horizontal phase velocity of the disturbances to be determined.

The speeds and directions of travel which will be presented here are averages obtained in a manner which will now be discussed. As can be seen from Figure 2-3, four triangles can be formed from the four-station network. Two of these triangles have their longest leg in the north-south direction (Ely-San Diego) and two have their longest leg in the east-west direction (Flagstaff-Stanford). The former pair give a more accurate measure of speeds of north-to-south traveling disturbances, while the latter pair are better for measurement of their direction of travel. The speeds obtained during 1967 which are presented here are averages of the values obtained from the two "north-south triangles" and the directions of travel are averages obtained from the other two triangles. Since the two speeds and the two directions which are averaged are usually nearly equal, the plane wave assumption made earlier appears to be justified.

No three station coverage was available throughout much of 1968; the network of stations shown in Figure 2-6 was established in October

of that year. The speeds used here for disturbances studied with this system of stations were obtained using the Stanford - Edmonton - Ft. Collins triangle. The direction of travel are averages from that triangle and the Stanford - Ft. Collins - Clark Lake triangle.

The velocities which are calculated are averaged over the time required for an identifiable feature of a disturbance to travel between stations. The values obtained are reasonable provided the spatial and temporal gradients of velocity do not behave in an erratic fashion. There is no evidence that any such erratic behavior occurs.

The data obtained do not provide information on the height of the disturbances. However, Thome [1964, 1968] has studied the same type of disturbances and has shown that they are located as low as 200 km and possibly as high as 800 km. Due to the low electron concentration at the high levels, the relative importance of effects about, say, 500-600 km are probably not very important when compared with those of lower levels. Tesud and Vasseur [1969] have also studied the vertical structure of these disturbances. Although the height range which they studied is not so large as Thome's, they found the disturbances to produce appreciable effects throughout the 200-400 km range. Since the ray paths from the stations to the satellite are nearly parallel, the distance between a given pair of ray paths is very nearly constant over the height range of interest. Hence one does not need accurate information about the heights of the disturbances in order to obtain good estimates of their horizontal velocities. In the calculations carried out in this paper, velocities have been determined assuming heights of 300 km. If, instead, the major influence of these disturbances on our measurements is at 200 km or 400 km, an error of less than about four percent will have been made in the determination of speeds and no appreciable error will have been made in the calculation of direction of travel. (These estimates of error are for the original network.)

The principal inaccuracy in determining the velocities of the disturbances stems from the uncertainty in the measured travel times between stations. Generally the peaks of the disturbances are so broad that timeshifts cannot be estimated any closer than about one minute.

Since our time resolution is generally better than one minute, improving it would not improve the accuracy of the measured timeshifts. Increased station separation seems to be the only method of increasing the accuracy of the measurements. An uncertainty of one minute may be important since travel times between stations, particularly with the original network, are often less than ten minutes.

The accuracy of measurement is a function of the speed and the direction of travel of the TID. For the disturbances discussed here the mean direction of travel is from geomagnetic north to south with a diurnal fluctuation of roughly  $\pm 20^\circ$ . For disturbances coming from such a direction with speeds of the order of  $500 \text{ m s}^{-1}$ , the measured speed is accurate to within about  $\pm 30 \text{ m s}^{-1}$  and the measured bearing is accurate to within about  $\pm 5^\circ$ . These are the errors due to a one minute uncertainty in travel times and are for the Stanford - Ely - Flagstaff - San Diego network. For the enlarged system of stations and for the same disturbances, the speeds are accurate to within about  $\pm 10 \text{ m s}^{-1}$  and the bearings are accurate to within about  $\pm 2^\circ$ .

The measured speeds for a number of TIDs are shown in Figure 2-12

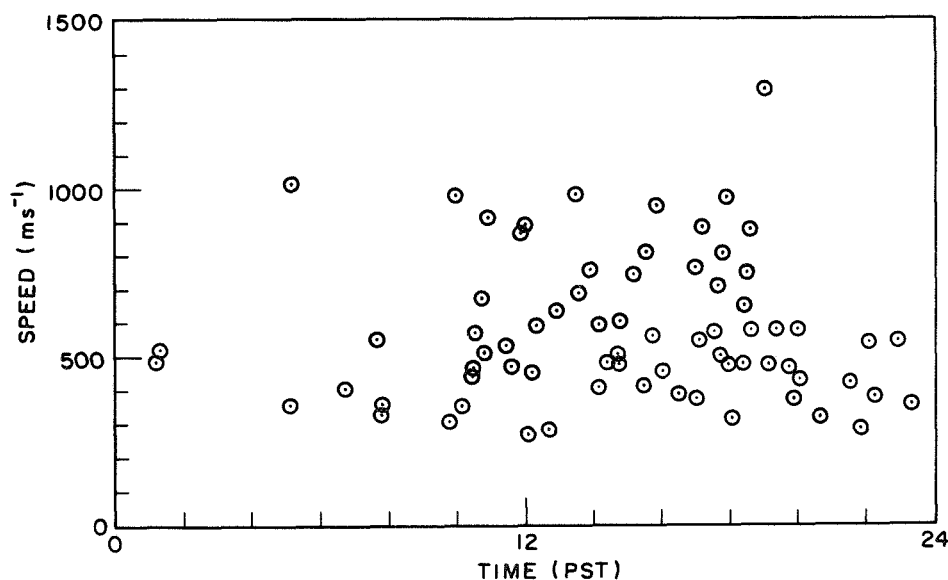


Fig. 2-12. SPEEDS OF A NUMBER OF TIDs AS A FUNCTION OF PACIFIC STANDARD TIME.

as a function of Pacific Standard Time. The speeds show no definite diurnal variation in magnitude; however, a diurnal variation in the number of observations is apparent and is probably due to observational limitations which will be discussed later. Figure 2-13 shows a histogram of observed speeds. Most disturbances have speeds between 300 and 600  $\text{m s}^{-1}$ . Virtually no disturbances have speeds below 300  $\text{m s}^{-1}$ , while an appreciable number have speeds between 600 and about 1000  $\text{m s}^{-1}$ . It has been previously reported [Davis and da Rosa, 1969] that these disturbances exhibit a pronounced diurnal variation in speeds. Indeed, for the disturbances studied in that paper a definite diurnal variation seemed to be present. The diurnal variation which was observed there, and which is now regarded as spurious, seems to be due to the limited sample size and the time averaging method used.

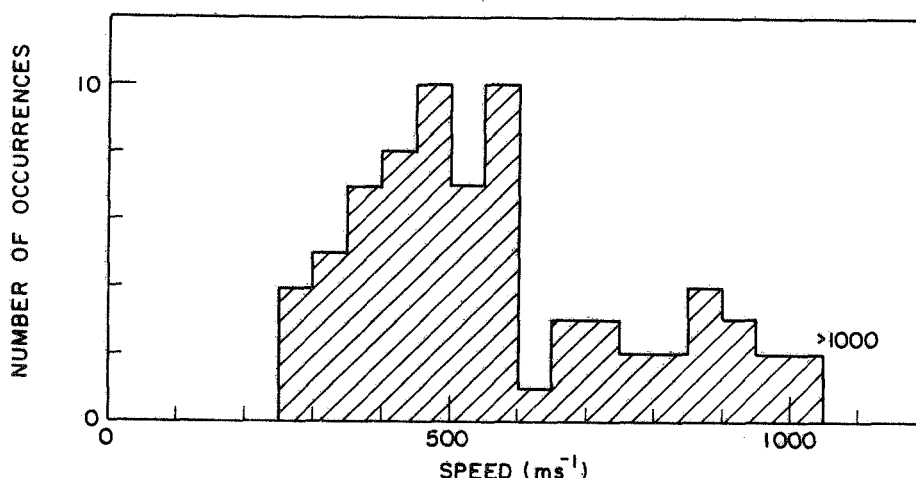


Fig. 2-13. HISTOGRAM OF OBSERVED SPEEDS. The two occurrences indicated between 1000 and 1050  $\text{m s}^{-1}$  are actually the number of occurrences with speeds above 1000  $\text{m s}^{-1}$ .

The measured directions of the velocity vectors of a number of TIDs are shown in Figure 2-14a as a function of Pacific Standard Time. The direction of arrival has a diurnal variation which is perhaps more apparent in Figure 2-14b in which the direction of arrival for TIDs on several selected days is shown. The direction of arrival has a definite diurnal variation; the disturbances come from an easterly direction

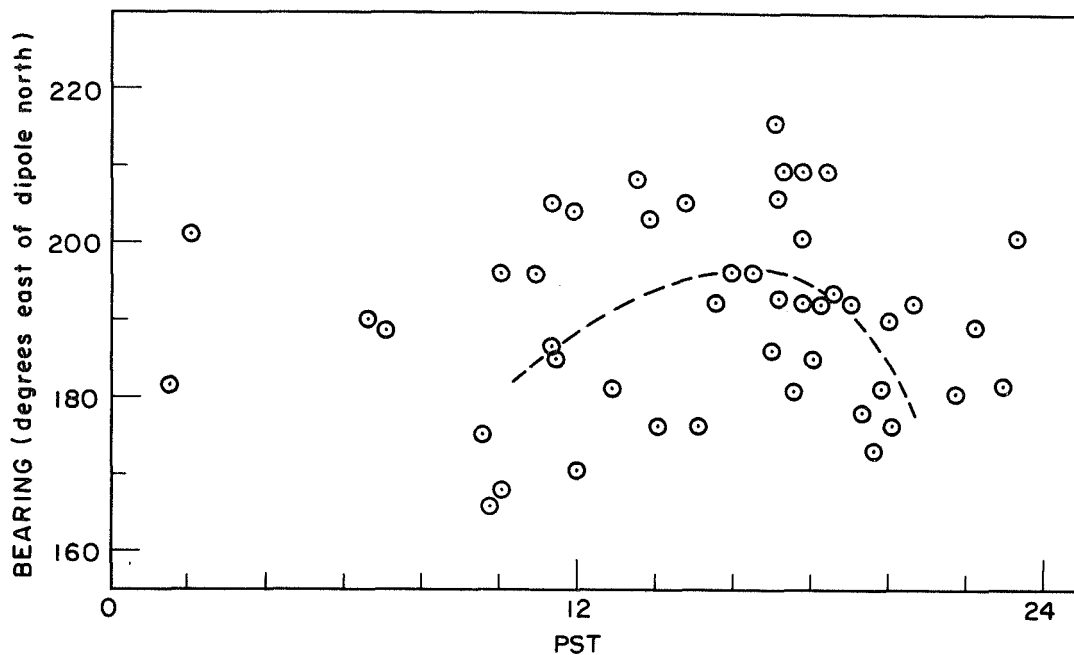


Fig. 2-14(a). MEASURED AZIMUTHS OF A NUMBER OF LS TIDs AS A FUNCTION OF PACIFIC STANDARD TIME. The dotted line is drawn through averages made at three hour intervals. 180° means propagation from dipole north to south.

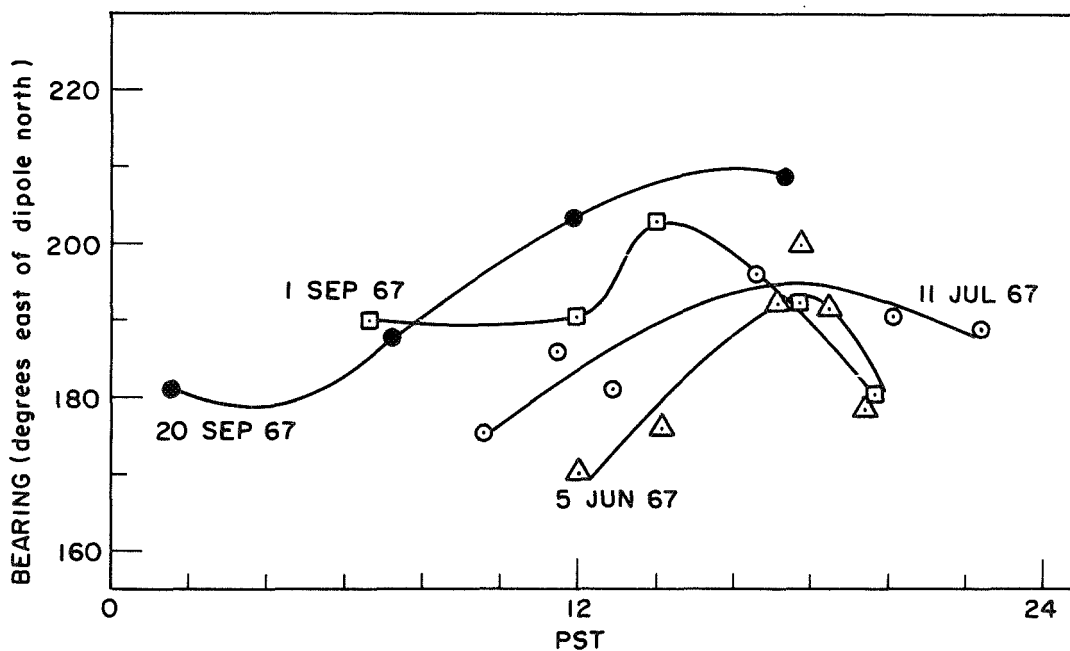


Fig. 2-14(b). MEASURED AZIMUTHS OF A NUMBER OF LS TIDs ON SEVERAL DAYS.

during local afternoon and appear to come from a westerly direction during local morning. The average direction of travel is from roughly geomagnetic north to south. Again, the diurnal variation in number of observations is probably due to observational limitations.

#### D. Discussion

As has been mentioned previously, a large diurnal variation in the number of observations occurs. This variation is due at least in large measure to observational limitations. One cannot detect fluctuations in electron content when the absolute value of their amplitude falls below a minimum level,  $I'_{\min}$ . Therefore, there is a minimum value of relative amplitude which can be observed

$$R_{\min} = I'_{\min} / I_0$$

As will be discussed in Chapter VIII,  $R$  is a measure of the amplitude of the disturbance in the neutral atmosphere. Due to the reduced electron content at night, the nocturnal value of  $R_{\min}$  becomes large enough to prevent the effects of many disturbances in the neutral atmosphere from being detected, while during the day, with a larger electron content and consequently a lower detectability threshold, the effects of a substantially greater number can be observed. This explains qualitatively the shape of the histogram of TID occurrences in Figure 2-9.

Assuming that the average amplitude of the disturbances in the neutral atmosphere is constant and that the coupling between them and the ionization does not vary diurnally, the average diurnal variation of the absolute amplitude,  $I'$ , of the TIDs should be similar to that of the electron content. This is approximately so as can be seen from Figure 2-15 where normalized average values of  $I'$  and  $I_0$  for the period 1967-1968 are compared. The discrepancies between the two curves can be attributed not only to the approximate nature of the assumptions made above but also to the fact that our data filtering process tends to consider the fast content variations near sunrise and sunset as part of irregularities, although care was taken to avoid this problem. The

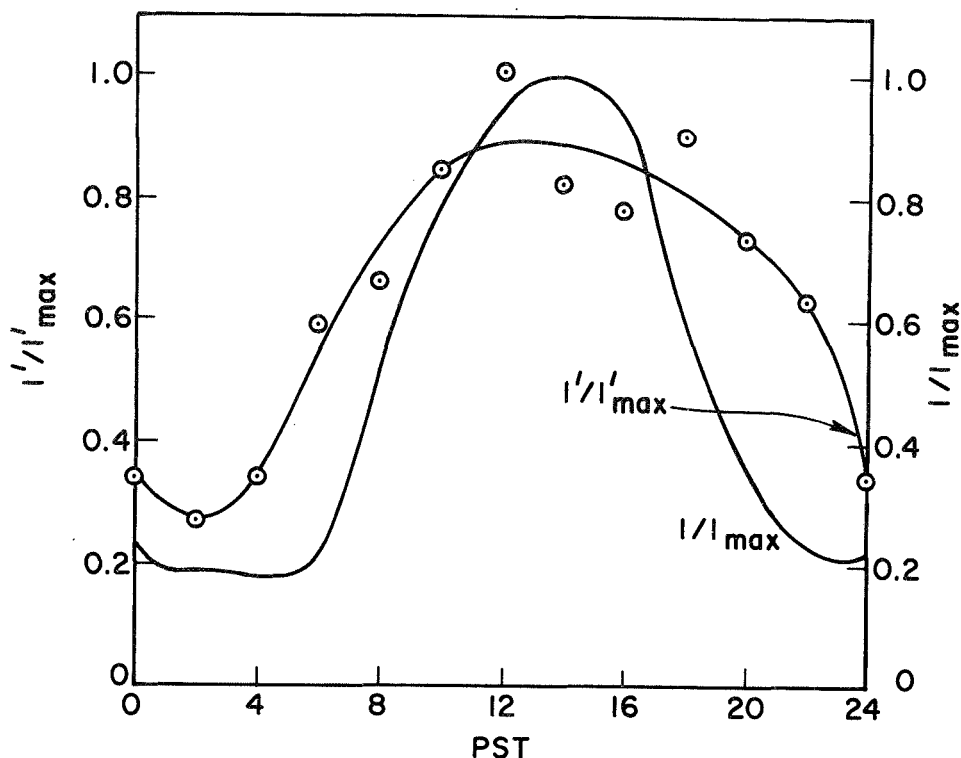


Fig. 2-15. A COMPARISON OF THE NORMALIZED DIURNAL VARIATION IN AVERAGE RMS AMPLITUDES OF TIDs AND THE NORMALIZED DIURNAL VARIATION IN AVERAGE ELECTRON CONTENT.

diurnal variation in amplitude will be considered in detail in Chapter V.

It will be shown in Chapter VIII that the relative amplitude,  $R$ , of electron content fluctuations associated with a TID is proportional to the amplitude of the disturbance in the neutral atmosphere causing them. On the other hand, it was found experimentally that these TIDs tend to occur during geomagnetically disturbed periods [cf. Figure 2-7] and that their amplitude,  $R$ , is proportional to the geomagnetic activity index,  $K_p$ . These facts suggest that the generation of the atmospheric disturbances which cause the TIDs is probably related to enhanced geomagnetic activity. This conclusion, taken together with the general north-to-south direction of travel of the TIDs, suggests that the source of the disturbances which we have observed lies in the northern auroral region. Such a location for the source has been previously suggested by Valverde [1958], Bowman [1965], Georges [1968a], Thome [1968], and others.

The concept of geomagnetic (dipole) time will be used throughout the following discussion. Geomagnetic time is determined by the position of the sun with respect to the geomagnetic meridian at the observing site; it is useful for studies dealing with time variations in geomagnetic phenomena. In this study dipole time will be calculated using a centered dipole and is defined by the spherical angle at the dipole pole formed by the local dipole meridian and the dipole meridian passing through the subsolar point. Dipole midnight is defined as the time of the transit of the sun over the dipole meridian opposite to that at the observing site. All calculations of dipole time were done using a solar declination of zero. The reader who is interested in an extended discussion of geomagnetic time is referred to Simonow [1963].

Since visual auroras, known to be related to geomagnetic activity, have a maximum frequency of occurrence near geomagnetic midnight in the auroral zone [Hultqvist, 1967, and references therein], it is not unreasonable to imagine that atmospheric disturbances might originate from this same location. Viewed from a fixed point on earth, at midlatitudes, a source located at geomagnetic midnight in the northern auroral zone appears to rotate, once every 24 hours, in a westerly direction around the north geomagnetic pole, causing a diurnal variation in the direction of arrival of any disturbances produced by that source when observed at the midlatitude location. Such a variation is in qualitative agreement with the behavior depicted in Figure 2-14 where it is seen that the mean direction of propagation of the TIDs is from geomagnetic north-to-south with a diurnal swing of  $\pm 20^\circ$ .

To further test the hypothesis described, we considered an isotropic source of disturbances and assumed that the resulting irregularities travel in the direction in which they are launched with a speed independent of latitude. The expected direction of arrival of disturbances can then be calculated as a function of time for any observing location. This was done assuming an observer located at the center of the polygon formed by the 300 km ionospheric points of the satellite-to-receiver ray paths for the network shown in Figure 2-3. The results obtained shows a diurnal variation in the direction of arrival which lagged appreciably the observed ones and which was, in addition, somewhat larger. In order to

obtain a better match between theory and experiment, other possible locations for the source were examined. By trial and error it was found that a source located at 1900 dipole time and dipole latitude  $78^{\circ}$  N gives quite good agreement with our observations. The length of the path traveled by a disturbance from such a source is shown in Figure 2-16 as a function of the universal time of origin. The calcu-

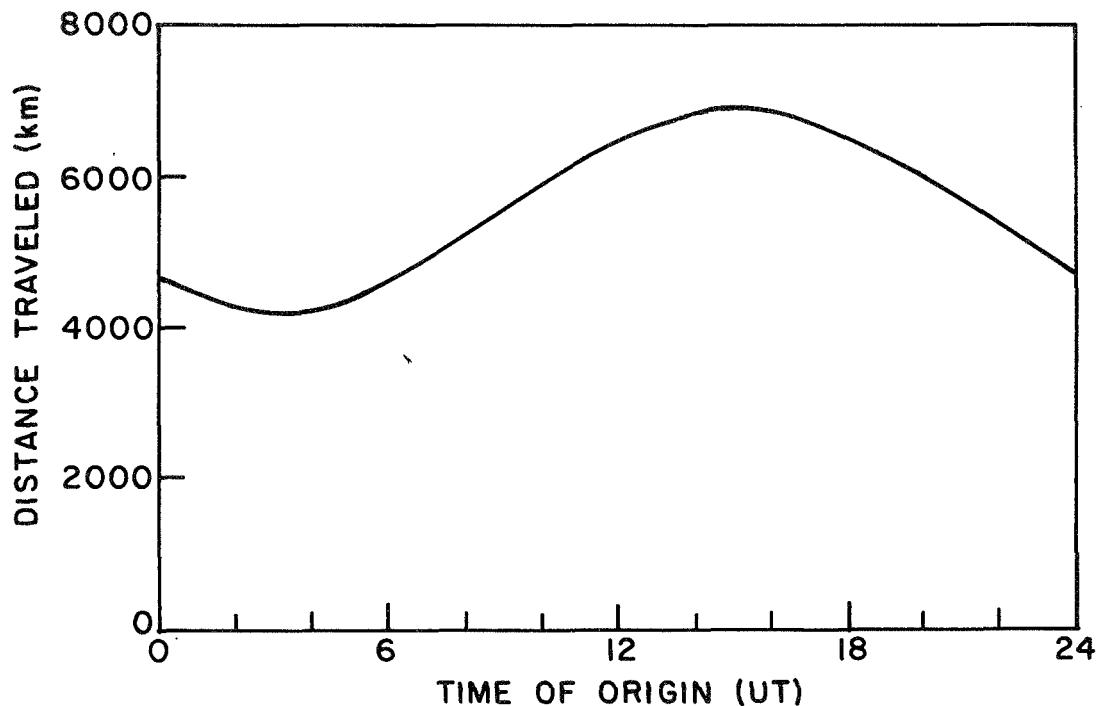


Fig. 2-16. THE DISTANCE TRAVELED BY A DISTURBANCE PRODUCED AT 1900 DIPOLE TIME AND  $78^{\circ}$  NORTH DIPOLE LATITUDE BETWEEN ORIGIN AND DETECTION VERSUS TIME OF ORIGIN.

lated directions of arrival versus Pacific Standard Time are shown in Figure 2-17 along with the averages of the observed values. The vertical bars indicate the standard deviation of the observations. Calculations were done using several constant speeds. The calculated direction of arrival versus time curve is fairly insensitive to the actual assumptions made about the speeds of the disturbances.

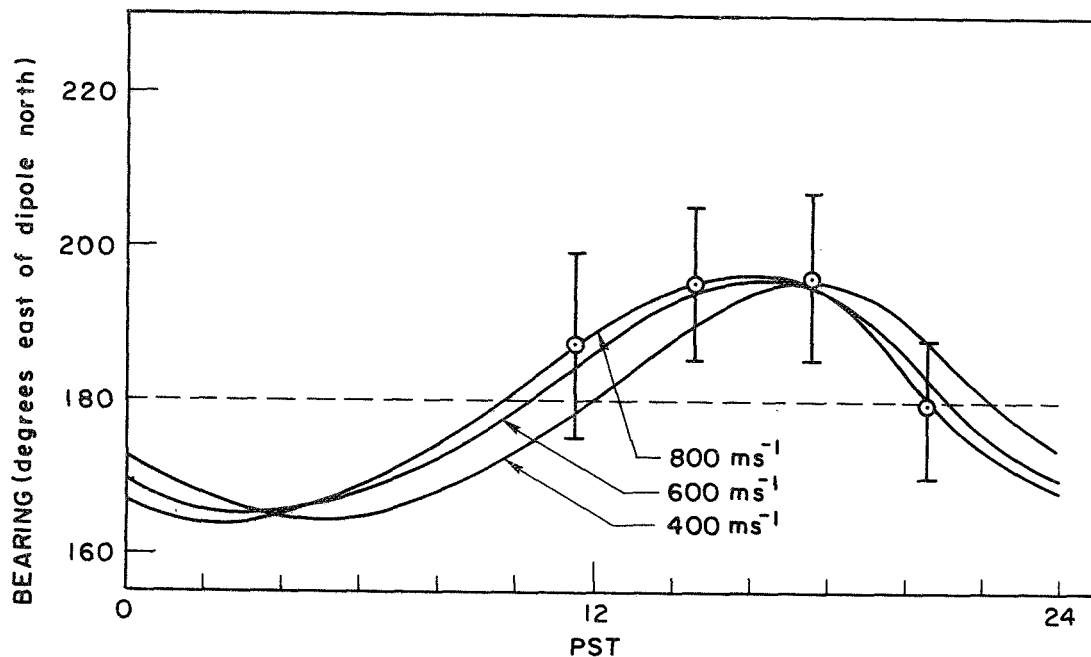


Fig. 2-17. A COMPARISON OF THE OBSERVED AVERAGE DIURNAL VARIATION IN DIRECTION OF ARRIVAL OF TIDs AND THE EXPECTED DIURNAL VARIATION.

The data in Figures 2-12 and 2-14 were obtained using two separate arrays of stations. It is assumed that speed and azimuth are only functions of the local time of observation. Using this assumption the data obtained using the second network were translated 0.3 hours later in time in order to plot them with the data obtained using the original network. This small time shift is the difference in local time between the centers of the two networks. Such a translation allows the two sets of data to be compared meaningfully. Local time at the center of the original network is UT-7.8 hours. Since the north-south separation between the centers of the two networks is small, it is neglected. All subsequent discussion refers to the centroid of the original array of stations.

A source of disturbances at the location determined above would be very near the auroral oval determined by Feldstein [1963] in which auroras are seen in the zenith with a frequency of occurrence of 75% or more. However, the actual location of aurora at a particular instant

depends upon the degree of geomagnetic activity [Akasofu, 1964 and references therein]. If the source of the disturbances is actually located in a region characterized by auroral activity at the time the disturbance is produced, then, since such a region moves further toward the equator with increased geomagnetic activity, the extent of diurnal variation in angle of arrival should depend upon the degree of geomagnetic activity. This point will be discussed further in Chapter III.

If the intensities of auroral and geomagnetic activity are good indices of the rate at which energy is being deposited in the polar atmosphere during geomagnetically disturbed periods, then such energy is not delivered uniformly, but is, in part, deposited explosively during polar substorms, primarily in the vicinity of the midnight sector. Assuming that the likelihood of producing an atmospheric disturbance is enhanced during periods when the amount of energy deposited in the atmosphere is suddenly increased, then it is probable that TIDs are produced by a number of isolated sources associated with separate substorms. Such a source mechanism would probably produce disturbances at a time near geomagnetic midnight in the form of a series of separate pulses, or damped oscillations, rather than a single periodic disturbance, a suggestion consistent with our observations. During geomagnetically disturbed periods, substorms occur roughly every one or two hours [Akasofu, 1965], times which are near the "periods" of observed LS TIDs. It is known that substorms provide the main contribution to  $K_p$  during storms and also during periods of relative quiet [Piddington, 1968], hence the good correlation between the relative amplitudes of disturbances and  $K_p$  observed by us. The relationship between the origin of large scale TIDs and substorms will be discussed in detail in the following chapter.

One would expect that disturbances of the type discussed here should also be generated at high latitudes in the southern hemisphere during geomagnetic storms and should therefore be detectable in the northern hemisphere propagating northwards, provided the long distances involved did not cause them to be too strongly attenuated. In fact, these disturbances have been observed in the southern hemisphere [Munro, 1950; Bowman, 1965; French, 1968]. We have, however, observed no

northward moving disturbances. This problem has been discussed by Thome [1968] who observed a southward moving disturbance by means of ionograms taken at stations located along the 75th meridian. The amplitude of the ionospheric disturbance increased as the wave moved southward towards midlatitudes and then decayed as it approached the equator, disappearing completely at low latitudes and not reappearing, as expected, in the southern hemisphere. The explanation given for this behavior is that the disturbance in the neutral atmosphere is progressively attenuated by viscous losses as it moves south. Its ionospheric signature is, however, enhanced at midlatitudes by the favorable inclination of the geomagnetic field lines, little perturbation in the ionosphere being expected when the field is either vertical or horizontal. By the time the atmospheric disturbance reaches the midlatitudes of the southern hemisphere, it has been sufficiently attenuated to be undetectable. However, Bowman [1965] has reported evidence of round-the-world propagation of these disturbances. The rate at which these disturbances are damped as the travel will be studied in Chapter IV. The problem of transequatorial propagation will then be considered again.

The observations presented in this chapter illustrate the well-known characteristics of large scale TIDs. These features are the following:

1. Occurrence highly correlated with geomagnetic activity
2. Propagation from the polar regions toward the equator
3. Speeds of the order of  $500 \text{ m s}^{-1}$
4. Retention of character over large distances.

In addition, several other features have been suggested.

5. A mechanism of production the output of which is related to the level of geomagnetic activity
6. An origin at high dipole latitudes at a location in the evening sector of the auroral oval in the vicinity of 1900 dipole time
7. A source mechanism related to the occurrence of polar substorms.

Many of the problems introduced here will be considered in more detail in subsequent chapters.

## CHAPTER III

### LARGE SCALE TIDS AND POLAR SUBSTORMS

It was suggested in Chapter II that large scale TIDs originate during polar substorms. This suggestion was based on the observed character of the TIDs and the known character of polar substorms. The present chapter pursues the possible relationship between large scale TIDs and substorms in more detail. This is done by comparing substorms and TIDs on a one-to-one basis as well as comparing them statistically; both approaches give convincing evidence that the TIDs and polar substorms are related. In addition, it is shown that it is possible to determine the location of the source of an individual TID if it can be associated with a substorm and if its velocity is known.

#### A. Background

Experimental verification of the fact that large scale TIDs occur during geomagnetically disturbed periods has been available for nearly a decade. However, a convincing experimental demonstration of a connection between these disturbances and substorms is not available in the literature. A detailed study relating infrasonic waves to polar substorms has been carried out by Wilson [1969]. In some respects the morphology of these waves is similar to that of large scale TIDs. For a single case French [1968] has shown evidence suggesting a connection between a supersonic auroral motion (which undoubtedly occurred during a substorm) and a subsequently occurring TID.

A detailed substantiation of the suggestion that LS TIDs originate during polar substorms in a fashion consistent with the morphology of the disturbances presented in Chapter II is not available. The purpose of this chapter will be to establish a firm connection between LS TIDs and polar substorms.

#### B. Polar Substorms

Since much of the following discussion involves the use of the concept of a polar substorm, a very brief discussion of the phenomenon will

be presented. For a detailed discussion of the subject, the reader is referred to the work of Akasofu [1965, 1968], and to the references mentioned therein.

Probably under the influence of the solar wind, the earth's magnetosphere appears to become intermittently unstable and at such times large amounts of energy are suddenly released into the polar upper atmosphere. This sudden release of energy is apparent in a number of diverse phenomena, for example, intense auroral activity, the precipitation of energetic particles and x-ray bursts. This activity, the polar substorm, may occur repeatedly during active periods and has a typical lifetime of one to three hours. The concept of the polar substorm unifies a number of superficially separate phenomena. Each of the phenomena comprising a polar substorm may itself be labeled a substorm, as for example, the auroral substorm.

In order to relate polar substorms and LS TIDs some means of detecting the occurrence of the former must be used. Since the occurrence of a substorm is apparent in many different phenomena, numerous methods are available. The most practical approach is to use geomagnetic field records from high-latitude observatories. During polar substorms an enhanced electric current, the polar electrojet, flows in the region of the auroral oval. The magnetic disturbance produced by this current is called the polar magnetic substorm. The magnetic disturbance is particularly intense in the midnight sector during a substorm, although little effect may be seen at stations located near noon. In addition to a strong dependence on local time, the character of the magnetic disturbance observed depends upon the latitude of the observer. The maximum effects are seen at a station in the auroral oval (at about  $68^\circ$  N dipole (dp) latitude) near midnight. In the midnight sector polar magnetic substorms appear as negative bays in the horizontal component of the geomagnetic records of an observatory located in the auroral oval. At other observatories negative or positive bays or more complicated structures may be seen depending upon their location with respect to the oval. In order to study the occurrence of polar magnetic substorms throughout the day a number of high latitude geomagnetic observatories must be used.

Figure 3-2, which will be discussed in more detail later, provides an example of how the same substorm leaves different types of signatures at different stations. The records from four observatories are shown. The first magnetic substorm is apparent near 1800 PST. Since College is in the late afternoon sector, no major effects are seen there. A second substorm is seen at 2300 PST. By this time Sodankyla has moved well into the morning sector. A positive bay is seen at College. The record from Great Whale illustrates the impulsive nature of the substorm.

It is important to note that polar magnetic substorms will be used to indicate the occurrence of a more widespread activity, i.e., the other manifestations of the polar substorm. Any correlation which may exist between the occurrence of polar magnetic substorms and LS TIDs does not necessarily imply a relationship between the cause of the magnetic disturbance, the enhanced current flow, and the source of the TID, although that may be the case. Unless additional evidence is available, a correlation between LS TIDs and polar magnetic substorms should be thought to imply only a relationship between some unspecified process occurring in the polar substorm and the source of the TIDs.

### C. Approach Used to Connect TIDs and Substorms

Four high-latitude, northern hemisphere observatories were chosen for this study, namely College (Alaska), Sodankyla (Finland), Leirvogur (Iceland) and Great Whale River (Canada). The locations of the observatories are shown in Figure 3-1 along with the 300 km sub-ionospheric points of the network of stations used in measuring columnar content. Dipole coordinates are used. For the period of interest, 1967-1969, high latitude Soviet geomagnetic data are not yet available. The stations selected provide the best coverage possible with only four stations, when Soviet data are not included. The use of four stations allows auroral electrojet (AE) indices to be constructed; use will be made of these indices later in this chapter. (For a discussion of AE indices see Davis and Sugiura [1966].)

The approach to be used in the following section will be to present a number of progressively more complicated examples to illustrate

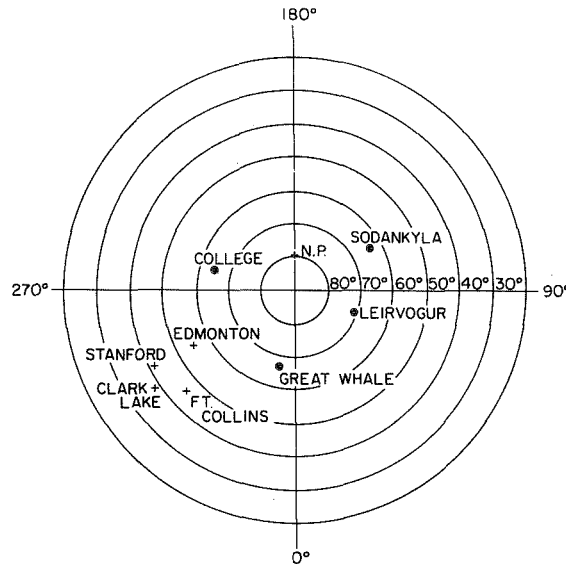


Fig. 3-1. LOCATION OF THE GEOMAGNETIC OBSERVATORIES USED. Dipole coordinates are used. Also shown are the 300 km ionospheric points for Stanford, Edmonton, Ft. Collins, and Clark Lake

the possibilities and problems of attempting to correlate TIDs and substorms. In addition, a statistical approach will be employed.

Two principal observational limitations are present in the method used. The first is the gap in coverage between College and Sodankyla. The difficulty which this gap presents is that it is possible to miss small substorms which might occur when the midnight sector is between College and Sodankyla. The observational difficulties associated with the electron content data are discussed in detail in Chapters II, V, and VIII. Briefly, it is difficult to detect TIDs during local night because of the small electron content observed then and, in addition, if TIDs of sufficiently small amplitude occur at any time, they are not detectable.

When isolated LS TIDs and substorms occur, it is found to be possible to associate a TID with a substorm in a reasonably unambiguous fashion if the speed with which the TID travels is known, since the distance traveled by a TID can then be estimated. Based on the theoretical distances to the average source of LS TIDs calculated in Chapter II, a TID originating in the auroral oval during a substorm should appear on the

Stanford records one to five hours after the substorm, assuming typical TID speeds. Therefore, even if the speed is not known, a convincing association can still be made if the substorm is truly isolated and the time separation between substorm and TID is reasonable. An isolated substorm preceding an isolated TID observed at Stanford by one to five hours would be convincing evidence for a connection between the two events.

Much of the energy of the atmospheric disturbances which cause large scale TIDs is associated with very low frequencies. Based on the dispersion relations for atmospheric waves (cf. Chapter VII), it is theoretically expected that group and phase speeds should be nearly equal at such frequencies. Hence the measured speeds can be treated as either group or phase speeds. These speeds will be used in calculating distances traveled.

In order to calculate the distance traveled by a TID, one must determine first a travel time and then find the distance using the measured velocity. It may sometimes be difficult to estimate the travel times accurately. The travel times used will be the separation between the (negative) peak of the substorm and the peak of the TID associated "bump" in the content curve for Stanford. It will be assumed that the TID speeds are constant and do not vary with position. By making these assumptions, one can calculate the distance traveled by the TID. If that distance agrees reasonably well with the average distance determined in Chapter II, then it would appear justifiable to claim a connection between the two events. If the two distances do not agree, then either the two events are not related or one or more of the assumptions, including the assumption that the average model is correct, is in error. The problem of which substorm to attempt to relate to a given TID becomes very difficult during large geomagnetic storms when many substorms occur. Unless the substorm separation is sufficiently large, it is not possible to associate substorms and TIDs given only the velocity and time of origin. Additional information related to the location of the origin must be provided. However, for the case of well isolated TIDs and substorms the problem is greatly simplified. The first examples studied will be for such well isolated cases.

#### D. Examples

Figure 3-2 shows data for March 16, 1969. The four content curves are displayed in such a fashion that only relative changes have any meaning. Two TIDs are apparent near 1900 PST and are indicated by the arrows. The speed of the first TID is  $1300 \text{ m s}^{-1}$ . That of the second is  $585 \text{ m s}^{-1}$ . A substorm occurs at Sodankyla and Leirvogur near 1750 PST. At Great Whale the substorm has a more complex structure and in fact it is possible that two separate substorms occurred. However, because of the closeness of the bays--a separation of about 20 minutes--it is assumed that only one substorm did occur. No other substorms are apparent for more than twelve hours preceding the substorm at 1750. A second substorm occurs near 2200 PST, after the TIDs are observed at midlatitudes. Assuming that the two TIDs originated in a single substorm at 1750 PST, the first required 50 minutes to reach Stanford, the second 115 minutes. These times imply distances traveled of 3900 and 4000 km for the two disturbances. The theoretical value based on the results of Chapter II is 4200 km for a substorm at 1800 PST. This close agreement suggests strongly that the substorm and TIDs are related.

The substorm observed at 2200 PST did not produce a detectable TID. If a TID had been produced, it would have arrived at Stanford near 2400, assuming a speed of  $500 \text{ m s}^{-1}$ . An inspection of records near that time revealed no signature which could be related to a TID. This lack of observation of a TID may be due to the fact that the substorm produced no TID or that the content was so depressed at that time that it was not detectable.

Figure 3-3 shows a second simple case. On January 25, 1969, a single TID, indicated by the arrows, was observed traveling southwards with a speed of  $725 \text{ m s}^{-1}$ . Because of the occurrence of what appears to be medium scale TIDs at Stanford, numerous other "bumps" are apparent; however, only one gives a reasonable agreement with the observations at the other stations. From the geomagnetic records, a substorm occurs at 1650 PST, following a period during which the records are reasonably quiet for more than four hours. Note that the substorm is visible only as a small positive bay at College. Assuming a substorm at 1650 PST, a distance traveled of 3650 km is calculated, whereas the simple theory of Chapter II suggests 4400 km. Again the agreement is reasonable. If the spike apparent near 1200 PST on the Leirvogur record were assumed

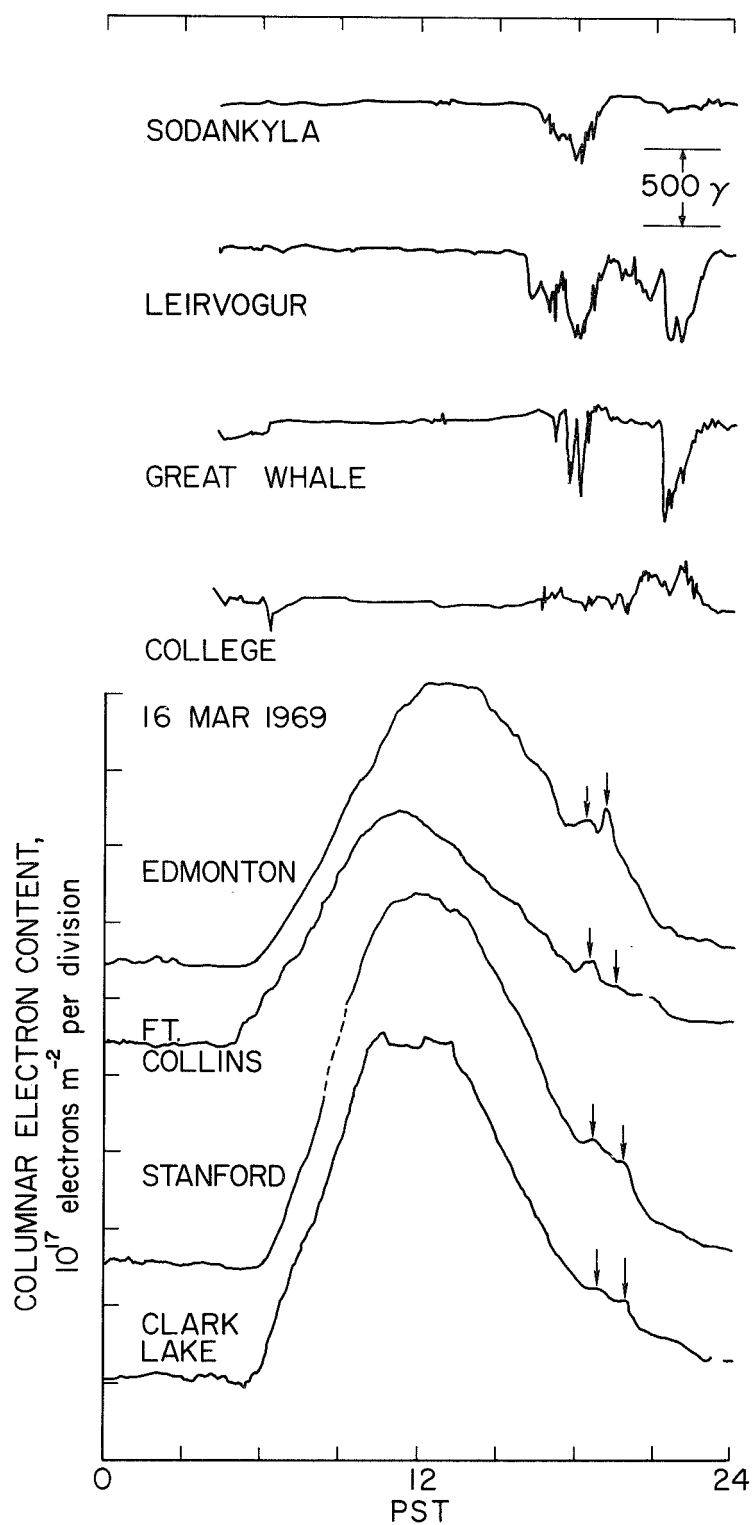


Fig. 3-2. COMPARISON OF GEOMAGNETIC RECORDS WITH ELECTRON CONTENT DATA FOR 16 MARCH 1969. The horizontal component of the geomagnetic field is shown. The arrows indicate large scale TIDs.

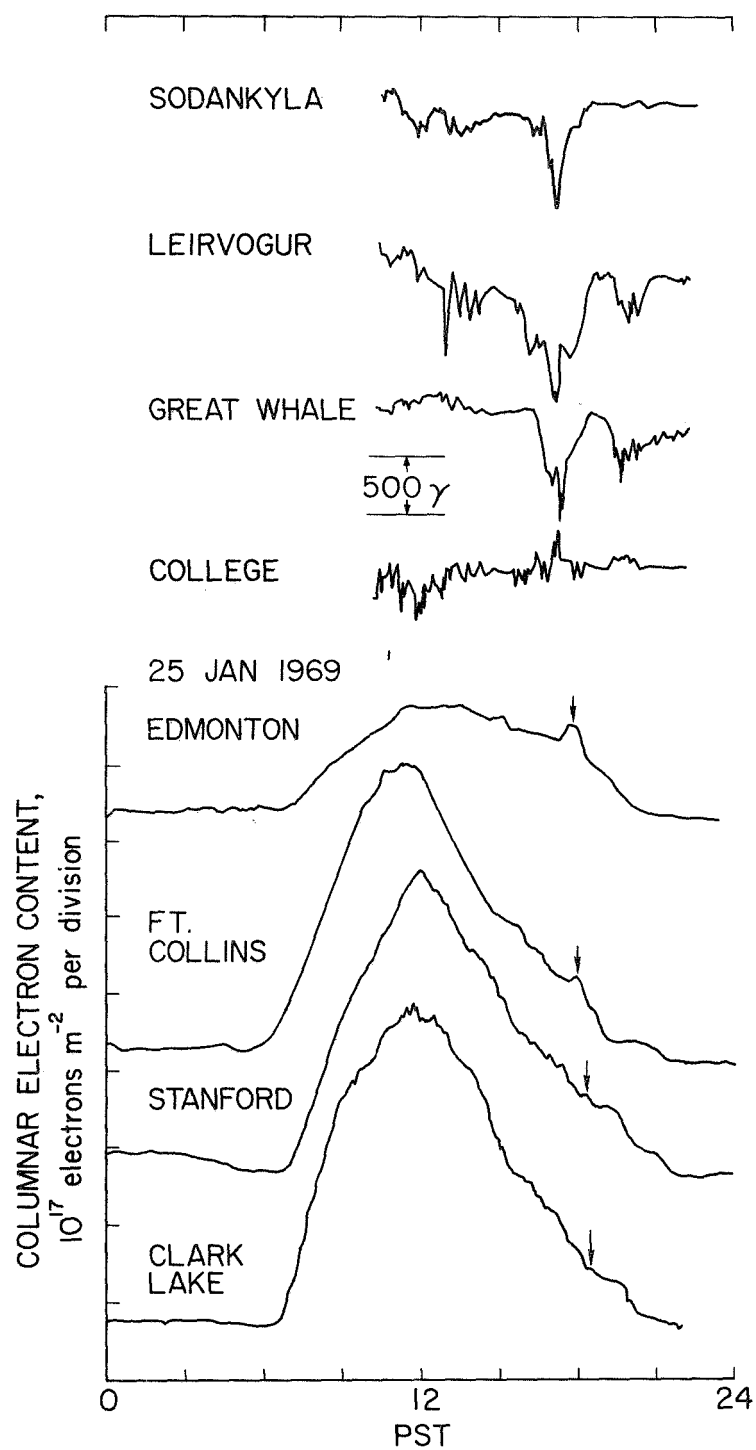


Fig. 3-3. A COMPARISON FOR 25 JANUARY 1969.

to be related to the source, a distance of about 14,000 km would be suggested. This value is much too large and is therefore rejected.

The two cases which have been presented show that a reasonable connection can be established between TIDs and substorms for simple cases. A somewhat more complicated case is shown in Figure 3-4. Eight TIDs were seen at Stanford on September 1, 1967. They are numbered on the content curve. The substorm to which they are apparently related are labeled with corresponding numbers. No obvious substorm can be associated with TID number 2, although it may be related to one of the small bays between substorms 1 and 3. TIDs 4 and 4b appear to have originated in substorm 4, which occurs near the time when the coverage is poorest, i.e., midnight between College and Sodankyla. This probably explains the small signature in the magnetometer record. TID number 6 matches best with substorm number 6, but the substorm labeled 6? is a possible candidate. This last case illustrates the difficulty which occurs when substorms occur closely together: The calculated distances for more than one substorm may appear reasonable and no clear-cut choice can be made.

Finally, Figure 3-5 illustrates a very complex case. On May 2, 1967, numerous but fairly widely spaced TIDs were observed. During the same period, a number of small substorms also occurred. On May 3, 1967, many TIDs and many substorms are apparent. Because of the very close spacing of substorms, it is not possible to relate particular substorms to particular TIDs. However, it can be seen that the frequency of occurrence of TIDs increases dramatically with the large increase in the rate of occurrence of substorms. Few TIDs are seen during the night probably due to the low electron content. That many of the substorms appear to occur only at College is due to the location of the observatories with respect to the auroral oval. College is the most suitably located for detecting substorms during this period.

The examples in this section have illustrated the fact that a convincing correlation between individual TIDs and substorms can be made only for relatively quiet conditions. In order to examine the relationship between TIDs and substorms more carefully for disturbed conditions, a statistical approach will be used in the following section.

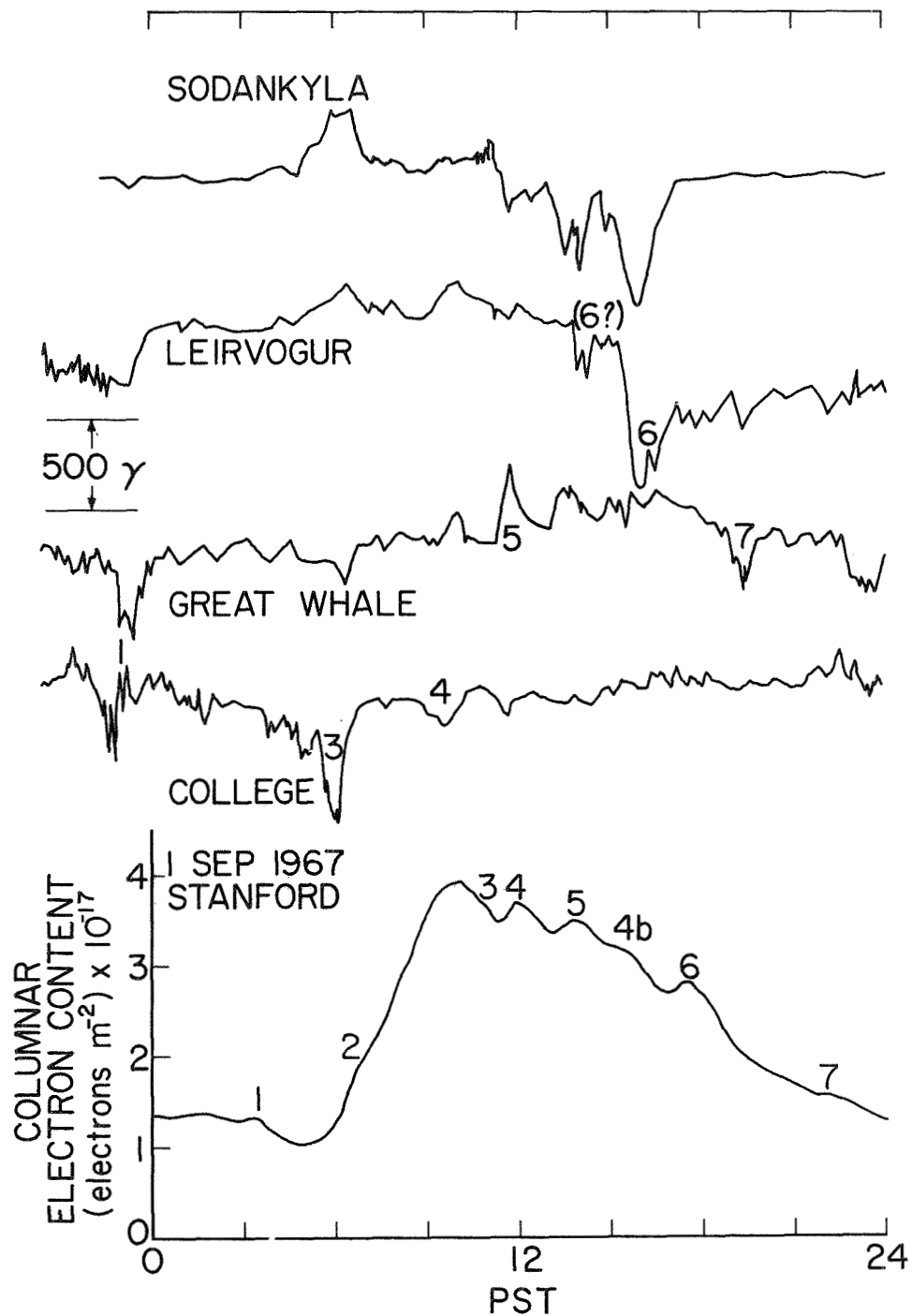


Fig. 3-4. A COMPARISON FOR 1 SEPTEMBER 1969.

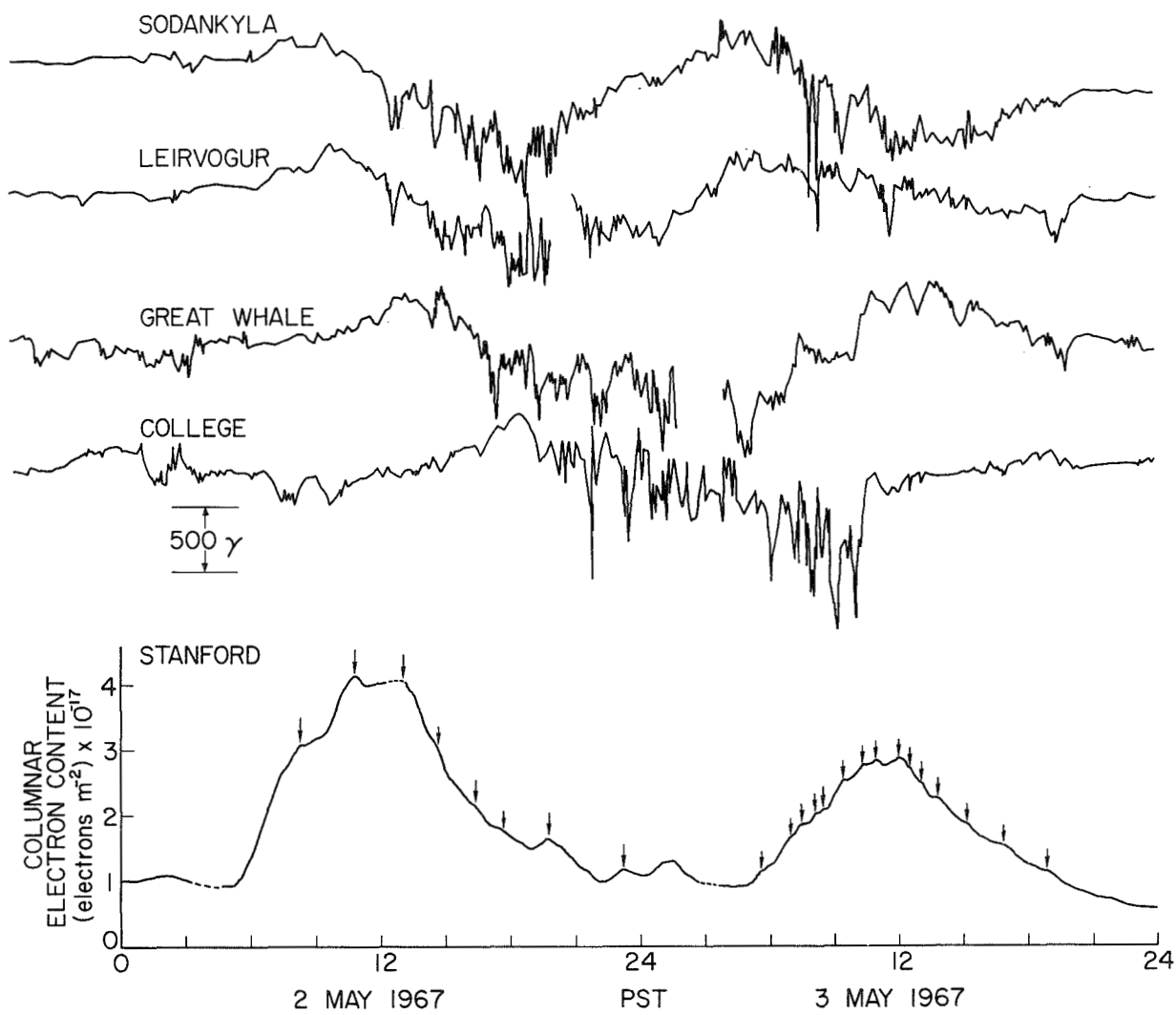


Fig. 3-5. A COMPARISON FOR 2 AND 3 MAY 1967.

### E. Statistical Approach

Since it is not always possible to pair TIDs and substorms unambiguously, an alternate approach will be used to study their relationship. A fairly large number of periods (a total of 28) during which TIDs were detected at Stanford were selected for study. The time interval within which TIDs occurred in each case was established and the number of TIDs counted. Then assuming that the TIDs required at least one but less than five hours to travel from their source, a time interval was established for which the number of substorms was counted. The following definition was used to define a substorm: any bay (positive or negative) with a magnitude greater than  $250 \gamma$  and with a 20 minute or larger separation from another bay. This definition is quite arbitrary and application of it requires a certain amount of judgement. In a sense, then, the number of substorms counted is somewhat subjective. Figure 3-6 shows the results for the 28 intervals. A good correlation between

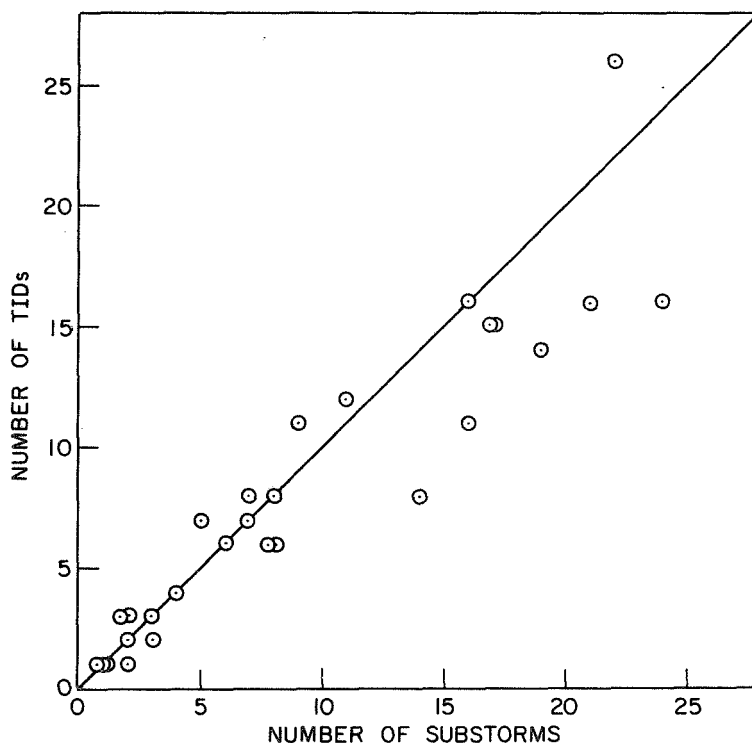


Fig. 3-6. COMPARISON OF THE NUMBER OF TIDS OBSERVED WITHIN A NUMBER OF INTERVALS WITH THE NUMBER OF SUBSTORMS OCCURRING WITHIN THOSE INTERVALS.

the number of substorms and TIDs is apparent. The fact that the number of substorms is generally larger than the number of TIDs within an interval is probably due to an inability to detect TIDs with very small amplitudes and also to the possibility that not all substorms produce TIDs. The frequency of occurrence of substorms, their "periodicity", varies appreciably from case to case for the data in Figure 3-6. Since the same time intervals were used, the "periodicity" of the LS TIDs varies in a similar manner. The good agreement evidenced in Figure 3-6 for such a sizeable sample of substorms strongly suggests an important connection between the fundamental mechanisms involved in producing the atmospheric waves related to the LS TIDs and the occurrence of polar substorms.

#### F. Location of the Source of Large Scale TIDs

It has been shown that it is possible to associate individual substorms with individual LS TIDs. In addition it is possible, by making a number of assumptions, to estimate the distance traveled by a TID before it is detected. Using such a distance, along with the measured direction of travel, it is possible to calculate the apparent origin of the atmospheric disturbance which produces the TID. With this approach the source location may be examined in more detail than could be done in Chapter II, when only the measured direction of travel of the disturbances was used.

Figure 3-7 is a plot of the calculated distances traveled by a number of TIDs before detection at Stanford versus the times of the peaks of the substorms which appear to be the sources of the disturbances. Each point is for a single TID. The poor coverage from 0600 to 1400 UT can be explained by the fact that any disturbances originating during those hours would arrive over the western United States at the most unfavorable time for detection. The distance which a disturbance originating at dipole midnight at  $68^{\circ}$  N dp latitude would travel (dashed line) is plotted along with the distance for a source at 1900 dp time and  $78^{\circ}$  N dp latitude. The latter is the average location calculated in Chapter II. It is readily apparent that the agreement is much better for the previously calculated average location than it is for a source at dipole midnight.

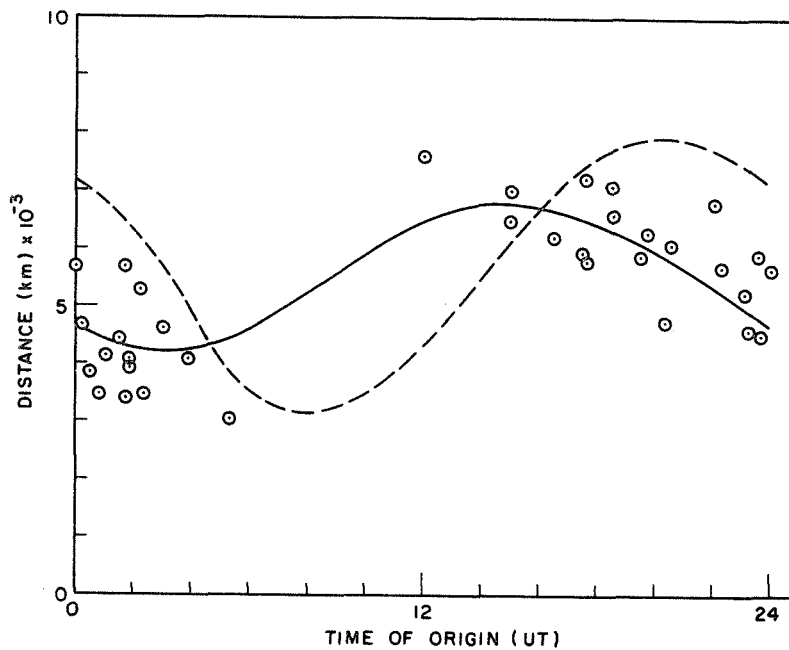


Fig. 3-7. CALCULATED DISTANCES TRAVELED BY A NUMBER OF TIDs VERSUS TIME OF ORIGIN. The curves are theoretical distances: the dashed line for a source at 2400 dipole time and  $68^{\circ}$  N dp latitude; the solid line for a source at 1900 dipole time and  $78^{\circ}$  N dp latitude.

Assuming that the direction of travel for these disturbances does not vary as they move toward the south, it is possible to determine the location of the origin of LS TIDs using the calculated distances and measured azimuths. Figure 3-8 shows, in dipole coordinates, the calculated point of origin for a number of disturbances. The coordinates are local dipole time and dipole latitude. The hatched area is the auroral oval, a region in which aurora are seen 75% of the time or more in the zenith [Feldstein, 1963]. The arrow attached to each point shows the direction toward Stanford at the time of origin. The number beside the arrow is the auroral electrojet index, in 100's of gammas, corresponding to the peak of the substorm during which the disturbance originated. There does not seem to be a correlation between the size of the substorm and the location of the source. The cross with a circle around it is the average location of the source determined in Chapter II, 1900 dp time and  $78^{\circ}$  N dp latitude. The second cross at 1840 dp time  $74^{\circ}$  N dp latitude is the average of the points plotted. The agreement between the two average locations is excellent.

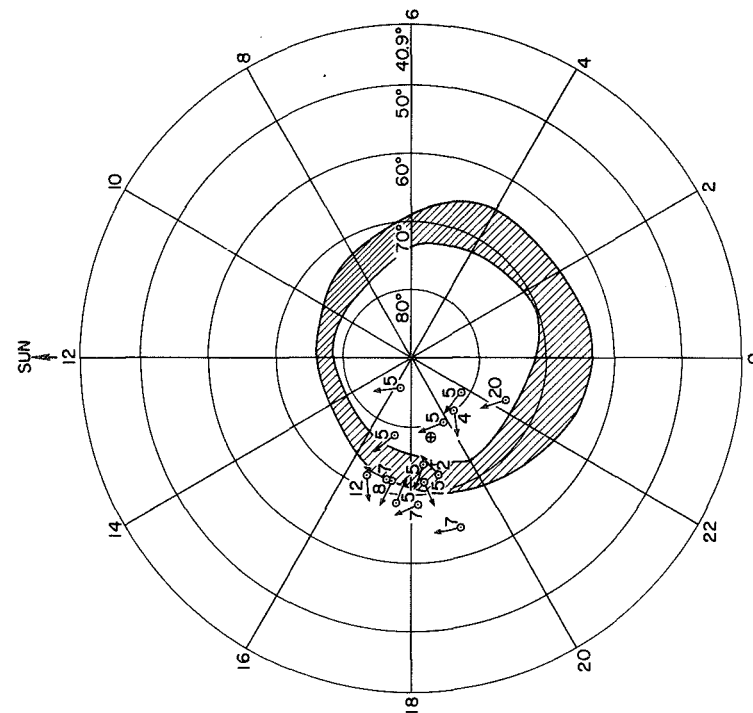


Fig. 3-8. LOCATION OF THE SOURCE OF A NUMBER OF LS TIDS. The hatched area is the auroral oval.

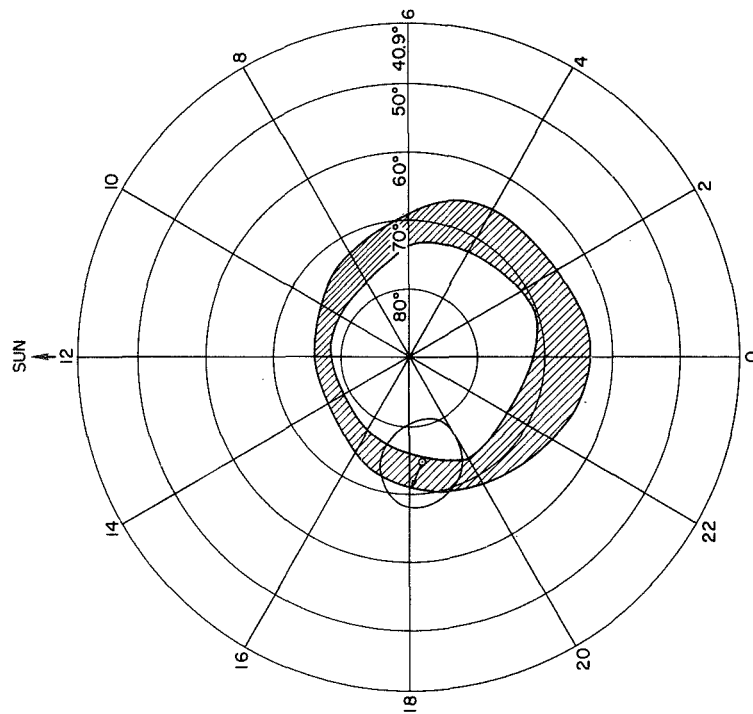


Fig. 3-9. AN ESTIMATE OF THE ERROR IN THE SOURCE LOCATION.

An estimate of the error in the calculated source locations is illustrated in Figure 3-9. The coordinate system is the same as was used in Figure 3-8; however, only a single point is plotted. The "circle" around the point is an estimate of the error in its position based on a  $\pm 5^\circ$  accuracy in measured azimuths and  $\pm 10$  per cent accuracy in calculating the distance traveled. These seem to be reasonable uncertainties provided the assumptions made are valid. The important point to note concerning Figure 3-9 is that the error size is quite small compared to the size of the auroral oval. Hence, it can be said with reasonable confidence that the actual source of the disturbances lies in the evening sector of the oval. Although the "point" of origin of a number of disturbances has been found, it is understood that the source of the disturbances is most likely not a point, but instead probably has considerable size. Since the error circle shown has fairly large dimensions, slightly over a 1000 km diameter, it might be conceptually useful to consider the source to be inside or primarily inside, that circle, rather than at a single point.

#### G. Discussion

This chapter presents several important results. First, it is possible to associate individual substorms and LS TIDs for simple cases. Secondly, it is statistically possible to relate substorms and LS TIDs for both fairly quiet and highly disturbed conditions. The results obtained provide strong support for the contention that LS TIDs originate during polar substorms. Furthermore, it has been shown that it is possible to calculate the location of the source of LS TIDs on an event-to-event basis. The method used is considerably more powerful than that used in Chapter II, but the results are in good agreement with those obtained there. They provide very strong evidence that the source of these disturbances is in the evening sector of the auroral oval, a factor which should be considered in attempting to construct a model for the mechanism by which the disturbances are produced.

## CHAPTER IV

### EXPERIMENTAL INVESTIGATION OF THE DAMPING RATE OF THE ATMOSPHERIC DISTURBANCES WHICH CAUSE LARGE SCALE TIDS

It is of interest to learn the rate at which the atmospheric disturbances which produce large scale TIDs are damped. Since the data used in the present study are columnar electron content measurements, it is logical to attempt to determine the rate of damping of these disturbances from a study of the amplitudes of the perturbations they produce in columnar electron content at two locations.

In determining the rate at which a disturbance is damped it is best to make measurements at points which are located such that the line separating them is parallel to the direction of travel. The disturbances causing LS TIDs travel very nearly from geomagnetic north to south in the northern hemisphere. As can be seen in Figure 2-6, the 300 km ionospheric points on the raypaths from both Stanford and Edmonton to ATS-I lie very nearly on the same dipole meridian. Hence, the line connecting these two points is approximately parallel to the direction of travel of the disturbances at all times. This provides a very nearly ideal geometry for studying the rate at which the atmospheric disturbances which cause large scale TIDs are damped.

The analysis required to investigate the rate of damping of the atmospheric disturbances in terms of content perturbations observed at two points will be developed in the following section for the case of small damping of a sinusoidal atmospheric wave. Then this analysis will be used along with observations of TIDs made at Stanford and Edmonton to study the rate of damping of the atmospheric disturbances which cause the ionospheric disturbances which have been observed.

#### A. Analysis

The rate of damping of an atmospheric wave will be found by first determining the ratio between values of the energy flux associated with the wave observed at two points along the path traveled by the wave in terms of the distances of the points from the source of the wave and

the rate of attenuation of the wave. Secondly, the ratio between the fluxes observed at the two points will be found in terms of electron content perturbations. Using the two relations, the attenuation rate may be found directly in terms of the content perturbations observed at two stations.

The instantaneous energy flux associated with an atmospheric wave in a nonviscous medium is given by

$$\vec{F} = p_n \vec{U}_n = (p_{no} + p'_n) \vec{U}_n \quad (4-1)$$

where  $U_n$  is the velocity of the neutral gas associated with the wave and  $p_n$  is the total pressure.  $p_{no}$  is the unperturbed pressure and  $p'_n$  is the pressure perturbation due to the wave. When this flux is averaged over a cycle, the contribution due to  $p_{no} \vec{U}_n$  vanishes. For the waves of interest, the vertical velocity is much less than the horizontal velocity so that only the horizontal component of Eq. (4-1) will be considered.  $p'_n$  can be related to  $U_{rn}$ , the horizontal velocity, by a polarization relation;

$$p'_n = \epsilon U_{rn} e^{i\varphi} \quad (4-2)$$

$\epsilon$  and  $\varphi$  are functions of height. The ratio,  $\epsilon e^{i\varphi}$ , between  $p'_n$  and  $U_{rn}$  is obtained from a model of wave propagation.  $p'_n$  and  $U_{rn}$  are assumed to have the form  $e^{i\omega t}$ . When the effects of viscosity are included,  $p'_n$  and  $U_{rn}$  will be modified somewhat as compared to the nonviscous case. However, even using these modified values, Eq. (4-1) will not give the complete energy flux, since there is now a flow of energy associated with the viscous forces [cf. Pitteway and Hines, 1963]. For the case of small damping, i.e., energy flux associated with viscous forces much less than the energy flux of the wave, Eq. (4-1) may still be used. The approach which will be used here is to first assume a small damping rate, carry through the analysis and then determine whether the damping rate obtained, using the analysis along with the observations, is consistent with the original assumption of small damping.

A coordinate system will be used in which  $z$  is the distance from the surface of the earth and  $r$  is the radial distance from the source of the wave. It will be assumed that  $U_{rn}$  has the following form:

$$U_{rn} = A_o \exp i(\omega t - \int K_r dr - \int K_z dz) \quad (4-3)$$

where

$$K_r = k_r - i\alpha_r$$

and

$$K_z = k_z - i(\alpha_{z1} - \alpha_{z2})$$

$\alpha_r$  is the rate at which  $U_{rn}$  is attenuated in the horizontal direction;  $\alpha_{z1}$  is the rate of damping in the vertical direction. This damping is due to the effects of viscosity and heat conduction.  $\alpha_{z2}$  is the rate of growth of wave amplitude with height due to the decreasing density.

Therefore,

$$U_{rn} = A_o \exp(-\int \alpha_r dr) \exp[\int (\alpha_{z2} - \alpha_{z1}) dz] \exp i(\omega t - \int k_r dr - \int k_z dz) \quad (4-4)$$

$U_{rn}$  may be written as

$$U_{rn} = U_o(r, z) \exp i(\omega t - \int k_r dr - \int k_z dz) \quad (4-5)$$

The time averaged energy flux is then

$$\bar{F}_r = \frac{1}{2} \epsilon U_o^2(r, z) \cos \varphi \quad (4-6)$$

Consider now the flow of power away from the source. Figure 4-1 shows a small tube through the atmosphere at some height  $h$ . The tube is of angular width  $\Delta\theta$ , vertical width  $\Delta h$  and originates at the source. It

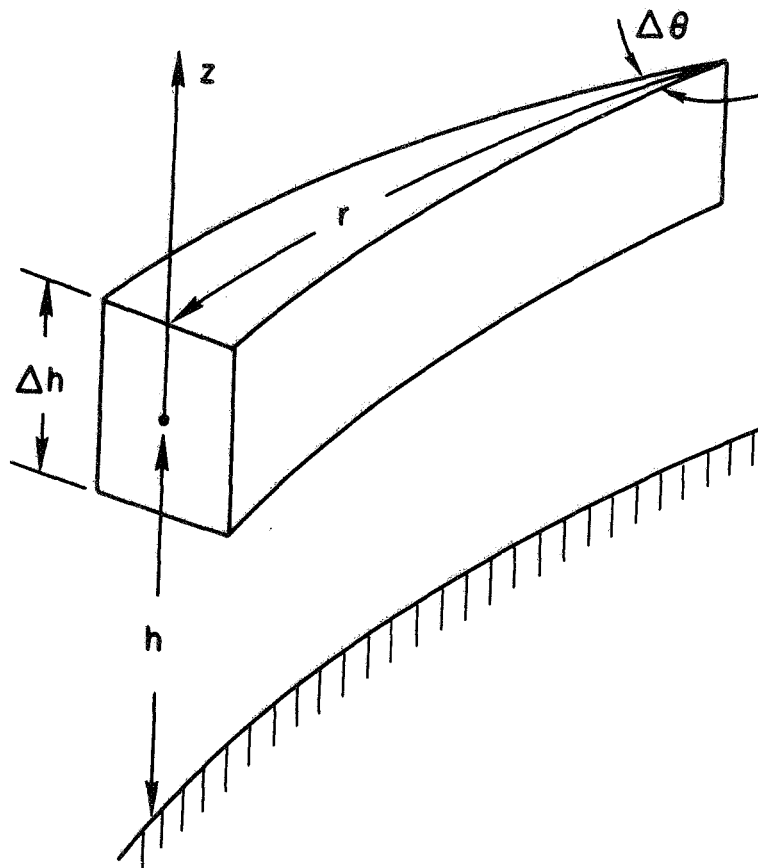


Fig. 4-1. DIAGRAM OF THE SMALL TUBE THROUGH THE ATMOSPHERE USED IN THE CALCULATIONS.

is assumed that the cross section of the tube is so small that  $\bar{F}_r$  is constant over it. Then the flow of power through the tube is given by

$$\bar{P}_r = \iint \bar{F}_r r d\theta dz = r \bar{F}_r \Delta\theta \Delta h \quad (4-7)$$

Hence

$$\bar{F}_r = \frac{\bar{P}_r}{r \Delta\theta \Delta h} \quad (4-8)$$

Since  $\bar{F}_r$  is proportional to  $\exp(-2 \int \alpha_r dr)$ , so is  $\bar{P}_r$ , and  $\bar{P}_r$  may be written as

$$\bar{P}_r = P_o e^{-2 \int \alpha_r dr} \quad (4-9)$$

where  $P_o$  is the power flow into the tube at the source.

Therefore

$$\bar{F}_r = \frac{P_o e^{-2 \int \alpha_r dr}}{r \Delta\theta \Delta h} \quad (4-10)$$

The ratio between fluxes at two points located at distances  $r_1$  and  $r_2$  from the source is given by

$$\frac{\bar{F}_{r_2}}{\bar{F}_{r_1}} = \frac{r_1}{r_2} \exp \left[ -2 \left( \int_0^{r_2} \alpha_r dr - \int_0^{r_1} \alpha_r dr \right) \right] = \frac{r_1}{r_2} \exp \left[ -2 \int_{r_1}^{r_2} \alpha_r dr \right] \quad (4-11)$$

The ratio  $r_1/r_2$  is due to the changing distance from the source, which results in changes in the cross-sectional area of the tube. The exponential term results from the attenuation of the wave as it travels.

It will be shown in Chapter VIII that the perturbation in electron content caused by an atmospheric wave may be written as

$$I' = m_h U_{rnh} I_o \quad (4-12)$$

where  $I_0$  is the unperturbed content,  $U_{rn h}$  is the amplitude of the horizontal gas velocity at height  $h$  and  $m_h$  is a constant of proportionality which is evaluated numerically using the approach outlined in Chapter VIII.  $m_h$  depends upon the distribution of ionization, the parameters of the wave and the path along which  $I$  is determined. The amplitude of the velocity at height  $h$  may be expressed in terms of the relative perturbation in electron content,  $R$ .

$$U_{rn h} = R/m_h \quad (4-13)$$

Hence

$$U_{rn}(r_1, z, t) = (R/m_h) \exp\left[\int_h^z (\alpha_{z2} - \alpha_{z1}) dz\right] \exp i(\omega t - \int_0^{r_1} k_r dr - \int_h^z k_z dz) \quad (4-14)$$

and

$$\bar{F}_{r_1} = \frac{1}{2} \epsilon (R/m_h)^2 \exp\left[2 \int_h^z (\alpha_{z2} - \alpha_{z1}) dz\right] \cos \varphi \quad (4-15)$$

Assuming now that the shape of the vertical variation in amplitude does not change with position along the path of travel, the ratio between the fluxes at two points located at distances  $r_1$  and  $r_2$  from the source may be found.

$$\frac{\bar{F}_{r_2}}{\bar{F}_{r_1}} = \left( \frac{R_2 m_{h1}}{R_1 m_{h2}} \right)^2 \quad (4-16)$$

Using Eq. 4-11 the fluxes may be eliminated to give

$$\left( \frac{R_2 m_{h1}}{R_1 m_{h2}} \right)^2 = \frac{r_1}{r_2} \exp\left[-2 \int_{r_1}^{r_2} \alpha_r dr\right] \quad (4-17)$$

whence

$$\int_{r_1}^{r_2} \alpha_r dr = - \ln \left[ \frac{R_2 m_{h1}}{R_1 m_{h2}} \left( \frac{r_2}{r_1} \right)^{1/2} \right] \quad (4-18)$$

Assuming that the rate of attenuation does not vary between the two locations (or that  $\alpha_r$  in the following equation is an average value), Eq. (4-18) may be simplified to give

$$\alpha_r = - \frac{1}{(r_2 - r_1)} \ln \left[ \frac{R_2 m_{h1}}{R_1 m_{h2}} \left( \frac{r_2}{r_1} \right)^{1/2} \right] \quad (4-19)$$

This expresses the rate at which the amplitude of the atmospheric wave is damped in terms of the content perturbations observed at two points at distances  $r_2$  and  $r_1$  from the source.

The most important assumptions involved in obtaining Eq. (4-19) will now be discussed in order to make very clear the meaning of any  $\alpha_r$ 's found using this expression. It was assumed that the shape of the vertical variation of amplitude does not vary with position along the path of travel. Since the vertical amplitude variation has been assumed to have the same shape at both observing points,  $m_{h1}/m_{h2}$  is independent of  $h$ , because

$$\frac{m_{h1}}{m_{h2}} = \frac{R_1/U_{h1}}{R_2/U_{h2}} \propto \frac{R_1}{R_2}$$

Hence,  $\alpha_r$  determined by Eq. (4-19) is independent of height, and is independent of the height for which the  $m$ 's are evaluated. Actually it is expected that  $\alpha_r$  should increase with height.

In order to determine  $m$ , some model for the height variation of the wave amplitude must be adopted, in the absence of any additional information about the vertical structure of the wave. The accuracy of  $\alpha_r$  is limited by the accuracy with which the height variation of amplitude can be specified. In addition, unless the actual electron concentration profile is known, the accuracy of  $\alpha_r$  is further decreased. It seems

probable that the greatest weakness associated with Eq. (4-19) lies in the determination of the  $m$ 's. The value of  $\alpha_r$  determined can most reasonably be considered as only a crude average value of the attenuation rate.

Other important limitations also apply. To use the method which has been outlined, columnar content must be measured at two points along the raypath of the wave. If it is not, the additional assumption must be made that the wave amplitude does not vary across the wavefront. Also, the method will give misleading results if there is a large vertical component of power flow. The possibility of interference effects due to the finite size of the source has been neglected. In the preceding calculation the possibility of a decrease in wave amplitude due to dispersive propagation was not discussed since a monochromatic wave was assumed. For the atmospheric disturbances causing large scale TIDs, much of the energy is associated with low frequencies (periods  $> 1$  hour). Based on the theoretical dispersion relations for atmospheric waves, little dispersion is expected at such low frequencies and it can probably be neglected.

Figure 4-1 implies that a peculiarly shaped source has been assumed, i.e., a vertical line. It has not been intended to suggest such a shape. The method outlined should only be used at points well away from the source; at such locations, the actual shape of the source region is probably not too important if the observing points are on the raypath of the waves and interference effects are negligible.

## B. Experimental Results

Figures 3-2 and 3-3 show examples of isolated TIDs which can be seen at a number of stations as they travel southward. In both cases the perturbation in the content seems to be smaller at Stanford than at Edmonton. Numerous examples of LS TIDs which are visible at both Stanford and Edmonton have been obtained. Using these data, it is possible with Eq. (4-19) to obtain an estimate of the rate at which the disturbances causing them are damped. The same electron concentration profiles are used for both Stanford and Edmonton in obtaining values for  $m$ . However, the different geometries and dip angles result in different values for  $m$  at the two stations as is shown in Chapter VIII.

Determination of the distance traveled by LS TIDs involves a number of assumptions, as was discussed in Chapter III. It will be assumed that

the distance traveled is given by

$$r = v\Delta T \quad (4-20)$$

where  $v$  is the measured speed and  $\Delta T$  is the time between the peak of the substorm which is thought to be the source and the peak of the associated perturbation in the content. Eq. (4-19) may be rewritten to give

$$\alpha_r = - \frac{1}{\Delta r} \ln \left[ \frac{R_{S^m_E}}{R_{E^m_S}} \left( \frac{\Delta T_S}{\Delta T_E} \right)^{1/2} \right] \quad (4-21)$$

where  $\Delta r$  is the distance between the 300 km ionospheric points, along the geomagnetic meridian, and the subscripts "S" and "E" refer to Stanford and Edmonton respectively. In cases in which the time of origin of a TID cannot be estimated, the distance from the average source will be used to evaluate Eq. (4-19).

The most difficult task involving the data is the estimation of the content perturbation. The value which will be used is the maximum deviation from what appears to be the unperturbed curve. In essence, a high pass filter is used to remove the diurnal variation. As the Edmonton data shown in Figure 3-3 strongly suggest, the disturbances are often more pulselike than sinusoidal.

The production of TIDs by internal atmospheric gravity waves will be considered in Part II. Results are presented in Chapter VIII for two models for the coupling between motions in the neutral atmosphere and the resulting perturbations in the ionization. The simpler model assumes that the atmospheric wave only moves ionization along the field lines via collisions. This model is labeled the collision geomagnetic equilibrium (C.G.E.) case. The other model used is more complete and considers the effect of the wave on losses and on plasma diffusion; for simplicity it will be called the "full" or "complete" model. The results obtained in Chapter VIII will be used here to provide values for the  $m$ 's in Eq. (4-21). In determining the values for the  $m$ 's, an atmospheric wave with a 3000 km wavelength, a period of 90 minutes, an amplitude of  $20 \text{ m s}^{-1}$  at 200 km, and propagating from dipole north to south was used. The approximations and limitations of the models used in obtaining these  $m$ 's will be discussed in Part II.

For the C.G.E. case and the "standard" wave described above,  $m_E/m_S = 0.54$ ; for the complete model used the ratio is 0.62.  $\Delta r$  is 1530 km. Using these values and a sample size of 12, the mean value of  $\alpha$  for the C.G.E. case is  $7.2 \times 10^{-4} \text{ km}^{-1}$  with a standard deviation of  $3.0 \times 10^{-4}$ . For the full model  $\alpha$  is  $6.4 \times 10^{-4} \text{ km}^{-1}$  with a standard deviation of  $3.0 \times 10^{-4}$ . In terms of db/1000 km, these become 6.3 and 5.5 respectively. The large standard deviation may be due to a variation in parameters from case to case or to the fact that the models used may be poor approximations for certain disturbances. All of the disturbances in the group studied occurred during local daytime and only isolated events were included, i.e., no TIDs occurring during large storms were considered, since at those times the analysis of Part II is least valid. This restriction, plus a limited amount of data, resulted in the small sample size.

### C. Discussion

In the course of our observations no large scale TIDs have been detected traveling from south to north. However, it will be shown in Chapter VIII (cf. Figure 8-5) that the sensitivity of our observing technique is such that disturbances coming from the south should be detectable. As was discussed in Chapter II, the lack of observations of disturbances traveling across the equator has been attributed to a damping of the atmospheric disturbance to such an extent that it is undetectable when it reaches the midlatitudes of the hemisphere opposite to that in which it originated. It is interesting to examine this point further.

As Figure 2-11a shows, disturbances with a relative rms amplitude as large as six per cent seldom occur at Stanford. However, relative amplitudes below one per cent are difficult to detect. (It is worth pointing out that the detectability threshold for disturbances coming from the south is probably highest when they are most likely to occur, that is, during geomagnetically disturbed conditions when many disturbances arrive from the north and possibly mask any smaller disturbances arriving from the south.) Stanford is at about  $40^\circ \text{ N}$  dp latitude. Assume that the statistics of disturbances at  $40^\circ \text{ S}$  dp latitude are the

same as at 40° N. Then a disturbance with a six per cent amplitude would be an unusually large disturbance at 40° S. In addition, the sensitivity to disturbances coming from the south is about 0.6 that for disturbances coming from the north (cf. Figure 8-5). Therefore, assuming a source at the south dipole pole which produces a disturbance which has a six per cent amplitude at 40° S dp latitude and a one per cent amplitude at 40° N dp latitude, Eq. (4-19) implies that the rate of damping is  $10^{-4} \text{ km}^{-1}$ . This is the maximum damping rate a disturbance originating in the southern hemisphere could have and still be detectable at Stanford (assuming, of course, that a six per cent amplitude at 40° S dp latitude is unusually large). This calculation includes the effect of angular spreading, but neglects possible refocusing at the antipode of the source. If the rate of damping is higher than this, disturbances should not normally be detectable in the midlatitudes of the hemisphere opposite that in which they originate.

An atmospheric wave which had  $\alpha_r = 10^{-4} \text{ km}^{-1}$  and a wavelength of 3000 km would have its amplitude attenuated by about 25 per cent in one wavelength. This would seem to be a case of "small" damping and the analysis of this chapter should be applicable to waves with such an  $\alpha_r$ . Note that the values of  $\alpha_r$  calculated in the preceding section are much longer than  $10^{-4} \text{ km}^{-1}$ . In fact, for the full model the average  $\alpha_r$  minus a standard deviation is still  $3.4 \times 10^{-4} \text{ km}^{-1}$ . These results indicate that  $\alpha_r$  is considerably larger than is necessary to prevent observation of transequatorial propagation. However, the fact that the calculated  $\alpha_r$ 's are so large implies that it is really not justifiable to use the simple theory of this chapter to calculate the damping rate of the atmospheric disturbances which cause large scale TIDs.

## CHAPTER V

### THE SOURCE OF LARGE SCALE TIDS

The preceding chapters have discussed the morphology of large scale TIDs and have shown that it is possible to relate them to polar substorms. The physics of the source of the disturbances has been neglected. This chapter will first review various suggestions for the mechanism involved in producing the atmospheric disturbances which cause the TIDs. Next the problem of using experimental observations to determine a source mechanism will be discussed. Following this discussion some speculations concerning the source physics are made. The principal conclusion is that the source may possibly be related to the westward traveling surges which are observed in aurora.

#### A. Background

A number of workers have discussed the physics involved in producing the atmospheric disturbances which cause large scale TIDs. Most of these discussions are based upon mechanisms related to the flow of electric currents in the auroral regions.

T. Gold in a personal communication to G. Thome in 1963 [Thome, 1968] suggested that the disturbances may be produced in the following manner: since most auroral energy is eventually deposited in the polar atmosphere as heat, the atmosphere will expand, with the thermal energy being converted to potential energy. The "bulge" in the polar atmosphere decays by radiating atmospheric disturbances. Cole [1962] has estimated that due to Joule heating the scale height of the polar atmosphere at 150 km may increase by a factor of five during moderate geomagnetic storms. Thome [1968] estimated that the energy available from Joule heating is sufficient to power the disturbances he observed.

Piddington [1963, 1964] has suggested that the disturbances may be produced by motions in the neutral atmosphere caused by charged particles moving in the direction of the electric field which exists during periods of enhanced activity.

The only quantitative studies of the problem appear to have been attempted by Chimonas [1968], Blumen and Hendl [1969], and Chimonas and Hines [1970a]. Chimonas considered the production of disturbances by the Lorentz force,  $\vec{J} \times \vec{B}_0$ , on the auroral ionization, which is transmitted to the neutrals by collisions. The current  $\vec{J}$  is the enhanced current flowing during periods of auroral activity and  $\vec{B}_0$  is the geomagnetic field. Using a simple model, he was able to suggest that such a mechanism seems capable of producing large scale TIDs. Blumen and Hendl have examined the problem of coupling energy into atmospheric wave motions through Joule heating. Again, the currents considered were those occurring during geomagnetically disturbed periods. Chimonas and Hines considered both the Lorentz force and Joule heating. However, they were not able to conclude which is the more important due to lack of information concerning the nature of auroral currents.

French [1968] has suggested that large scale TIDs might be produced by atmospheric disturbances generated by supersonic auroral forms. He discussed an example of a supersonic auroral form and a subsequently occurring TID to support his thesis. Supersonic auroral forms have been shown to produce infrasonic waves [Wilson, 1967, 1969].

Although a number of convincing suggestions have been advanced to explain the source of the disturbances of interest, the subject is still not fully resolved; much of this chapter will be devoted to a discussion of methods of obtaining experimental support of an explanation for a source mechanism.

## B. Experimental Determination of a Source Mechanism

Since in situ measurements in the auroral oval of a type which would allow one to observe the physics of the source are not available, this discussion will be restricted to indirect, midlatitude measurements. The problem can be approached in the following fashion. Assume the source of the disturbances to be inside a large black box of sufficient size to fully enclose the region of the source. Measurements of disturbances produced by the source will be made outside the boundaries of the box. The problem is that of making measurements of a type which will provide information on the physics of the source. The following characteristics would seem to be pertinent: location of the source in space

and time, the directional character of the source, i.e., its "radiation pattern", and the properties of the atmospheric disturbances produced by the source. The location of the source in space and time is the evening sector of the auroral oval during polar substorms. The directional character of the source is not known; this problem will be discussed in detail later. However, any directional character of the source will be due to a combination of the physics of the source and its geometry. The two effects would have to be separated in order for the physics to be studied. The properties of the disturbance produced by the source which seem to be most relevant are its temporal and spatial scales and its amplitude. As has been discussed previously, the time scales of the disturbances observed are consistent with a source related to substorms; in fact they are quite suggestive of this (cf. Chapter II). The spatial scales of the observed disturbances are likely to be misleading if used to infer information about the source, since they are primarily due to the time separation between substorms and the speed of travel of the disturbances.

If the energy input to the source could be estimated, along with the energy output, important evidence relating to the source physics would be available. It was shown in Chapter II that the amplitudes of the disturbances seem to be approximately linearly related to  $K_p$ , which is roughly logarithmically related to the amplitudes of geomagnetic disturbances. Therefore, the output seems to be fairly weakly related to the input, assuming the input is proportional to the amplitude of the geomagnetic disturbances.

The evidence obtained points to a source in the evening sector of the auroral oval. The source locations which have been determined should not be understood to imply a point source. However, it would seem reasonable to assume that they represent something like an "effective" source location and are probably meaningful in terms of establishing a location at which much of the energy is coupled into the atmospheric disturbances.

Intuitively, one would expect the disturbances to be produced near geomagnetic midnight in the auroral oval. In addition, with the

exception of French's [1968] suggestion regarding supersonic auroral forms, the mechanisms proposed would also seem to predict a source near midnight, since the current flow is largest in that region. The Lorentz force and Joule heating are probably also maximized there. The fact that the evidence indicates that the disturbances are not produced near midnight is somewhat surprising. The present study strongly suggests that an investigation of polar substorm related phenomena in the evening sector should be carried out to determine possible sources of the TIDs. This subject will be pursued further, later in this chapter.

Of the important characteristics of the source listed above, only the radiation pattern has not been studied. The other properties mentioned have been treated earlier in this study. Therefore, the following section will consider the problem of the experimental determination of the directional character of the source.

#### C. Radiation Pattern of the Source

There are two fairly obvious techniques for determining the directional character of the source. First, sensors could be positioned around the source and their measurements could then be compared. Such an approach would allow a study of the characteristics of the source both for a single event and, in a statistical sense, for many events. The second approach involves the use of only a single sensor. Since the source of the disturbances rotates with respect to an observer located outside the auroral zone, this single sensor method allows the directional character of the source to be studied in a statistical manner, although it would not be useful for determining the radiation pattern for a single event. A single sensor can essentially obtain the same results as many sensors, although it lacks their redundancy, if only information on the statistical behavior of the source is desired. The single sensor method will be discussed here.

The nature of the "sensor" to be used to make these measurements has not been specified. Ideally some measurement technique which would be sensitive to density and/or velocity variations in the neutral atmosphere should be used. Although atmospheric waves have been detected directly in the neutral atmosphere at the heights of interest [Newton,

Pelz, and Volland, 1969], there does not seem to be a suitable method available for the present study. The method which will be used here to study the radiation pattern of the source will be based on the signatures produced in columnar electron content records by disturbances in the neutral atmosphere. Such an approach involves the use of ionization as a tracer for motions in the neutral atmosphere. As has been pointed out [Chapter VIII, Hooke, 1968b; Georges and Hooke, 1970] such an approach has unavoidable observational biases. In addition, and far more important for the present discussion, the fact that the ionization undergoes a large diurnal variation introduces considerable difficulties. This problem is equally important for either a single or multiple sensor experiment. The large diurnal change in ionization results in a sensitivity to disturbances in the neutral atmosphere which varies greatly throughout the day.

As will be shown in Chapter VIII, it is possible to relate the amplitude of perturbations in electron content to the disturbances in the neutral atmosphere causing them. The day may be divided into a number of time intervals and the average relative amplitude of disturbances within each interval may be determined. Using the analysis of Chapter VIII, this average relative amplitude can be related to the average amplitude of the disturbances in the neutral atmosphere. Using an average source location such as was determined in Chapters II and III, the average distance traveled by disturbances for each of these time intervals may be determined. The results of Chapter IV may then be applied to estimate the amplitude of the disturbances at the source. The average amplitude of the disturbances at the source, in the direction of the sensor, as a function of time, is then known if an average speed of travel for the disturbances is assumed. This average amplitude as a function of time can be converted directly into average amplitude as a function of direction away from the source; i.e., its radiation pattern.

A more accurate approach, which requires more information, is the determination of the amplitude of each disturbance at the source and then the averaging of the amplitudes at the source. Such an approach would require that the time of origin of each disturbance and its speed be known. However, the first, and simpler, method is adequate to show

any large anisotropy in the radiation pattern. All of the tools required to apply these methods have been developed. What is now needed to use the first approach is the average amplitude of the disturbances as a function of local time at the observer.

The diurnal variation in the average measured relative amplitude of LS TIDs has already been determined (cf. Figure 2-11a). This is not the same as knowing the average relative amplitude, since not all TIDs are detected. The following expression sets a lower limit on the relative amplitude,  $R$ , which can be observed

$$R_{\min} = \frac{I'_{\min}}{I_0} \quad (5-1)$$

where  $I'_{\min}$  is the minimum detectable perturbation in electron content. Since disturbances with relative amplitude below  $R_{\min}$  will not be detected, the values of  $R$  which are obtained are samples from a singly truncated distribution with a terminus at  $R_{\min}$ . What is desired is the mean of the actual distribution of relative amplitudes; what are available are means of samples from a distribution having the same shape but truncated at  $R_{\min}$ . Cohen [1961] has given tables which allow the maximum likelihood estimate of the mean of the original distribution to be determined given the mean and variance of a sample taken from the truncated distribution for the case in which the original distribution is normal and there is a sharp, known truncation point. Values for the truncation point,  $R_{\min}$ , must be determined as a function of local time.

$I'_{\min}$  depends on the shape of the content curve; it is more difficult to detect small perturbations when the content is changing rapidly than when it is reasonably constant. However,  $I'_{\min}$  does not depend upon the magnitude of  $I_0$ . Since the shape of the content curve is a function of local time,  $I'_{\min}$  is a function of local time also. Of course,  $I_0$  depends upon local time; hence  $R_{\min}$  also depends upon local time.

Figure 5-1a shows the average electron content for 1967-1969 observed at Stanford: the average curve for 1967-1968 is virtually indistinguishable from the curve shown. Values of  $I'_{\min}$  will be determined using the average curve. This is equivalent to assuming that the

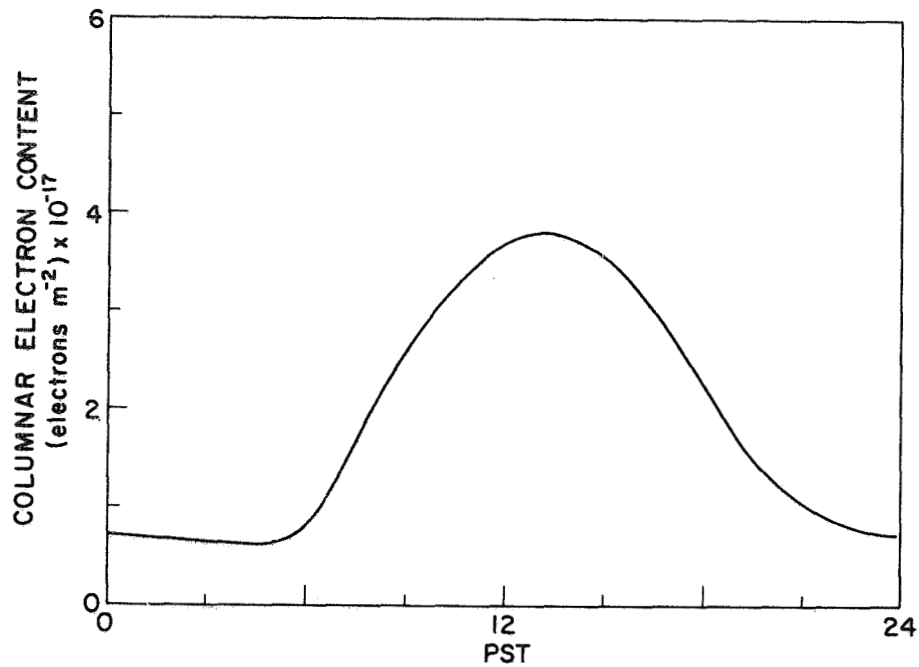


Fig. 5-1(a). AVERAGE ELECTRON CONTENT AT STANFORD FOR 1967-1969.

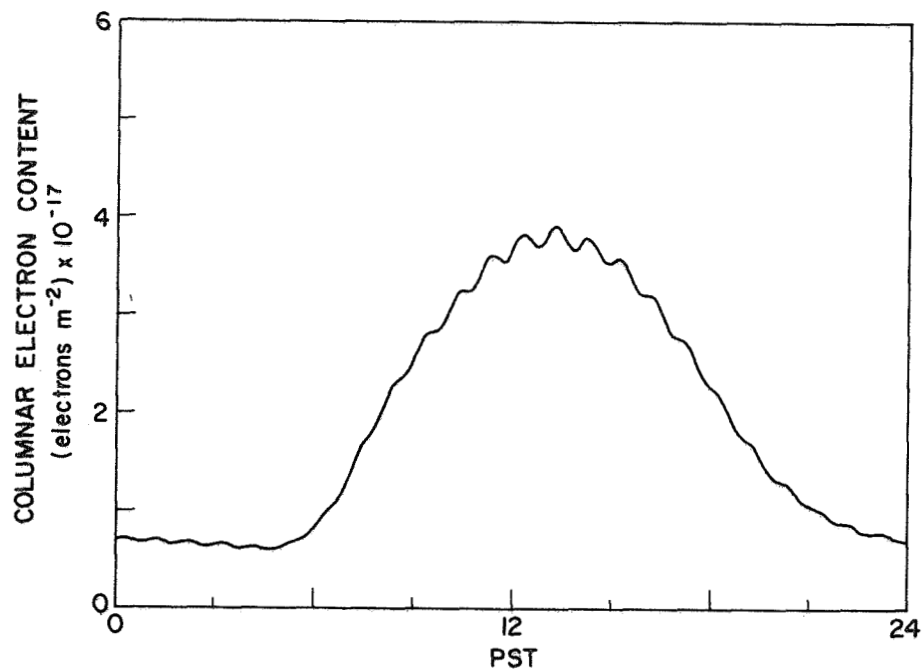


Fig. 5-1(b). RESULTS OF ADDING A SINE WAVE WITH A RELATIVE RMS AMPLITUDE OF TWO PER CENT TO THE CURVE IN (a).

shape of the content curve does not vary from day to day. Figure 5-1b shows a curve obtained using the following expression:

$$I = I_0(t) (1 + A \sin \omega t) \quad (5-2)$$

where  $I_0(t)$  is the average content for 1967-1968

$$A = 0.02828$$

and  $\omega$  corresponds to a period of one hour.

This expression produces a curve which is the average content curve with a sine wave of rms amplitude equal to two percent of the background added. By varying the size of  $A$  in Eq. (5-2) and producing various curves, the minimum detectable perturbation in  $I$  can be determined by inspection as a function of local time. Such an approach is subjective, but then so is the problem of detecting disturbances in the actual data. Using values of  $I'_{\min}$  determined in this fashion along with average values for  $I_0$ , average values of  $R_{\min}$  can be determined using Eq. (5-1). Table 5-1 shows  $I'_{\min}$ ,  $\langle R_{\min} \rangle$ , the average value of  $R_{\min}$ , and  $\sigma_{R_{\min}}$ , the standard deviation in  $R_{\min}$ , for the period 1967-1968.

The relative amplitudes of the TIDs have a probability density function (pdf) which will be called  $g(R)$ , Figure 5-2a.  $R_{\min}$  also has a pdf which will be called  $f(R_{\min})$ , Figure 5-2b. The probability of observing a disturbance of amplitude,  $R_0$ , if one is present, is given by

$$P_{\text{obs}} (R|R = R_0) = P(R_{\min} < R_0) = \int_0^{R_0} f(R_{\min}) dR_{\min} \quad (5-3)$$

$P_{\text{obs}}$  is shown in Figure 5-2c. In Figures 5-2b, c, d, the dotted lines are for the case of a well-defined terminus; i.e.,  $R_{\min}$  a constant and not a random variable.  $\delta(R_0)$  in Figure 5-2b is a delta function. Let  $h(R)$  be the pdf for the observed amplitudes. This is the distribution from which we are sampling.  $h(R)$  is given by the following

$$h(R) = \beta P_{\text{obs}}(R) g(R) \quad (5-4)$$

TABLE 5-1

MINIMUM OBSERVABLE RELATIVE AND ABSOLUTE AMPLITUDES  
1967-1968

| Local<br>Time | $I'_{\min} \times 10^{-16}$<br>electrons $m^{-2}$<br>(vertical) | $\langle R_{\min} \rangle$<br>per cent | $\sigma_{R_{\min}}$ , per cent |
|---------------|-----------------------------------------------------------------|----------------------------------------|--------------------------------|
| 2             | 0.12                                                            | 2.1                                    | 1.1                            |
| 4             | 0.12                                                            | 2.2                                    | 1.1                            |
| 6             | 0.18                                                            | 2.9                                    | 2.0                            |
| 8             | 0.23                                                            | 1.2                                    | 0.37                           |
| 10            | 0.27                                                            | 0.95                                   | 0.25                           |
| 12            | 0.27                                                            | 0.79                                   | 0.24                           |
| 14            | 0.27                                                            | 0.79                                   | 0.24                           |
| 16            | 0.29                                                            | 1.0                                    | 0.58                           |
| 18            | 0.25                                                            | 1.3                                    | 0.57                           |
| 20            | 0.17                                                            | 1.9                                    | 1.9                            |
| 22            | 0.12                                                            | 1.9                                    | 1.2                            |
| 24            | 0.13                                                            | 2.2                                    | 1.1                            |

where  $\beta$  normalizes the distribution and  $h(R)$  is shown in Figure 5-2d. As can be seen from Figure 5-2, a sharp truncation point can be obtained if only observations with a magnitude well above  $\langle R_{\min} \rangle$  are considered. "Well above" can be approximated by two standard deviations. In such a case there would be a terminus,  $\langle R_{\min} \rangle + 2\sigma_{R_{\min}}$ , for which Cohen's [1961] results could be applied. Figure 5-3 shows a graph of sample means and standard deviations along with values of  $\langle R_{\min} \rangle$  and two standard deviations (dashed lines). As the figure shows, with the present  $R_{\min}$ 's, the situation appears hopeless, except perhaps for the 0800 to 1400 PST interval, since  $\langle R_{\min} \rangle + 2\sigma_{R_{\min}}$  is usually well above the sample means. This situation results in  $\min$  effective sample sizes which are greatly reduced. Figure 5-3 illustrates the fact that detection of disturbances throughout much of the day is rather difficult and

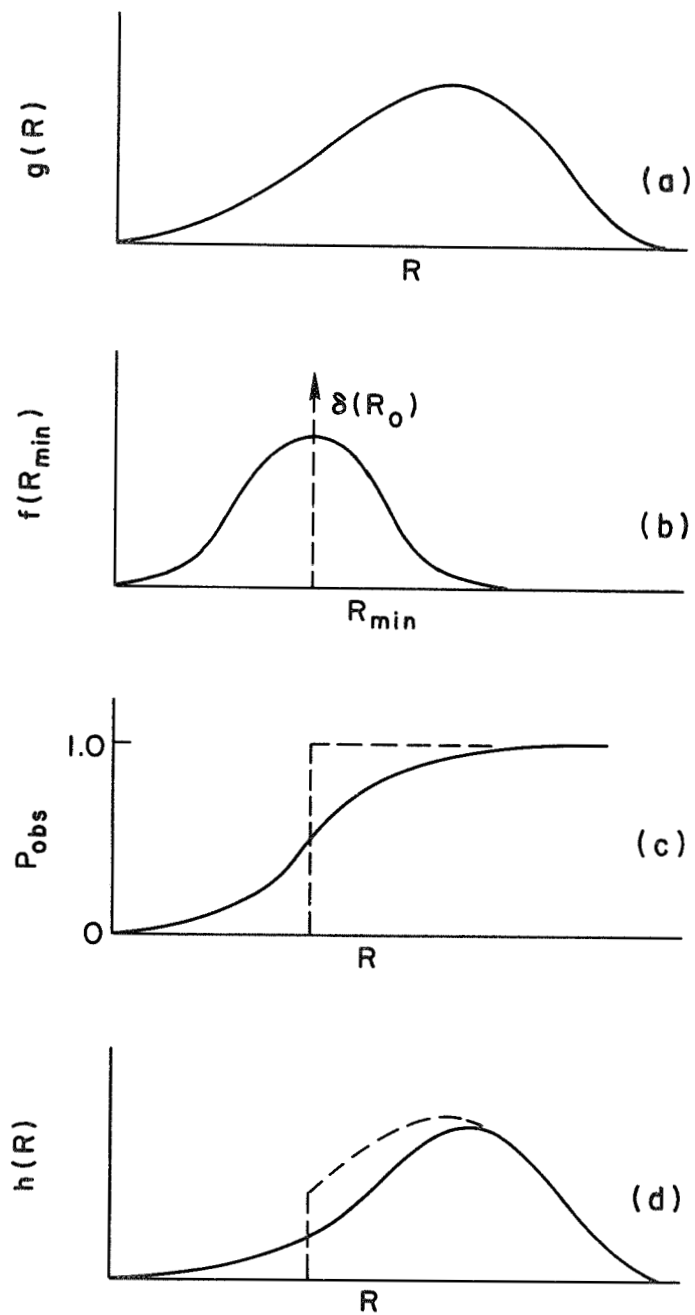


Fig. 5-2. ILLUSTRATION OF THE SAMPLING PROBLEM  
IN TERMS OF PROBABILITY DENSITY FUNCTIONS.  
The curves are explained in the text.

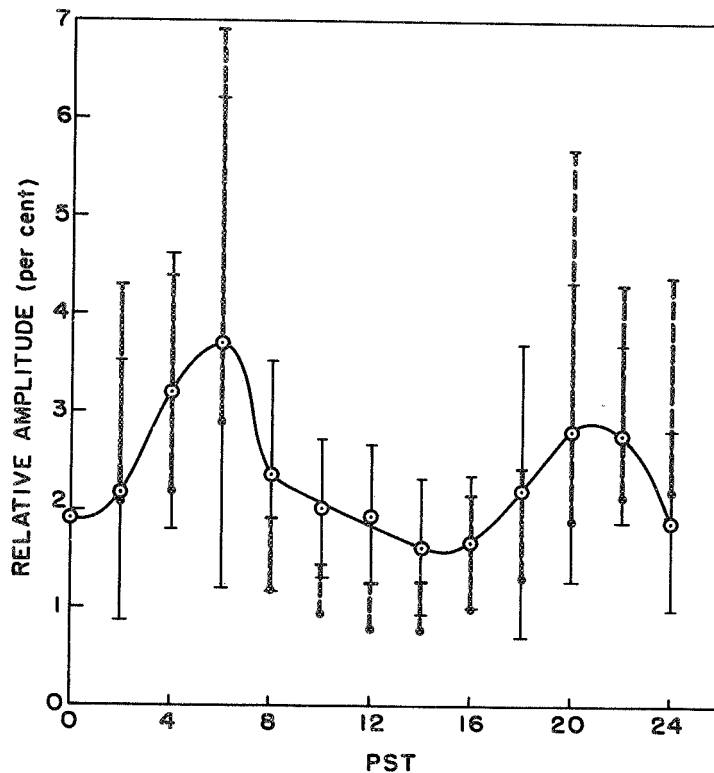


Fig. 5-3. COMPARISON OF THE AVERAGE RELATIVE AMPLITUDES OF LS TIDs (solid curve) AND THE MINIMUM DETECTABLE RELATIVE PERTURBATIONS IN ELECTRON CONTENT. Dashed lines indicate two standard deviations in  $f(R_{\min})$ .

often is only possible when the content is larger than average.

The method used to estimate values of  $R_{\min}$  will now be discussed. It has been shown in Chapter II that relative amplitudes of LS TIDs are roughly linearly related to  $K_p$ . From Figure 5-3, it can be seen that  $\langle R_{\min} \rangle + 2\sigma_{R_{\min}}$  is less than 2.3% from 0800 to 1600 PST. It is known that LS TIDs of this amplitude generally occur for average  $K_p \leq 4$ , a moderately disturbed condition. What this means is that relatively low values of  $K_p$  correspond to conditions sufficiently disturbed for TIDs to be produced with amplitudes greater than  $R_{\min}$  and which can thus be easily detected. But for these low values of  $K_p$  the content curve is not usually very distorted. For highly disturbed days (high  $K_p$ ) the electron content curves may, however, be greatly distorted; but the TID

amplitudes are then almost always so large that there is no problem detecting them. Hence, it seems most reasonable to use an average content curve based on data for all days, perhaps even excluding highly disturbed ones, for finding  $\langle R_{\min} \rangle$  values during the day. At night the situation becomes more complicated. As can be seen from Table 5-1,  $\langle R_{\min} \rangle$  is about 2% from 2000 and 0400 PST, rising to 2.9% at 0600.  $\sigma_{R_{\min}}$  is 1% or above. Disturbances with amplitudes near or above 4% generally occur only during quite disturbed conditions. Therefore, at night it might be more reasonable to determine  $\langle R_{\min} \rangle$  using an average content curve based only on data from days which are at least somewhat disturbed. It would be difficult to estimate a proper minimum level of  $K_p$ , but perhaps an average  $K_p$  equal 2, or no  $K_p$  values below 2, might be a reasonable choice. As was the case for daytime hours, highly disturbed days should also be disregarded, but at night "highly disturbed" would mean  $K_p$  greater than 7 or 8, say, while during the day it might mean  $K_p$  greater than 4 or 5.

One difficulty in using such an approach for night data is that curves of total content for disturbed days are notoriously "irregular" in shape. Hence the assumption that the shape of the curve does not vary from day to day which was made in finding  $\langle R_{\min} \rangle$  would probably be considerably less valid than during the daytime. Based on the preceding discussion it seems fair to say that the values of  $\langle R_{\min} \rangle$  determined for daytime hours are reasonable, although some small improvement might be had by not including data from very disturbed days in the calculations. For night the values of  $\langle R_{\min} \rangle$  obtained are likely quite useful in rough calculations; for any precise work they should be recalculated using only data from "disturbed" days as was discussed above.

The situation could conceivably be improved by increasing the sample sizes. This might be achieved by collecting more data or by revising the methods of finding  $R_{\min}$ . However, the fact that the truncation is so severe; i.e., the truncation point is too near the sample mean, will result in problems when using Cohen's [1961] approach. In addition, there is reason to believe that the samples are not from a normal distribution; Cohen has given results only for a normal distribution. Theoretically, a similar approach could be applied to any distribution. However,

the actual distribution is not known. It must be concluded that at this time the average relative amplitudes of the disturbances cannot be determined throughout the day. While the results of this section are essentially negative, they are of value for two reasons. First, they illustrate the difficulty of the problem. Second, they show rather vividly the most important observational difficulty encountered in using the electron content approach, namely, the large diurnal variation in sensitivity to disturbances.

It should be pointed out that if the average relative amplitudes were determined, they would still have to be converted into amplitudes of disturbances in the neutral atmosphere. This could be accomplished using Eq. (8-6). However, in order to use such an approach, an important assumption would have to be made. The precise relationship between the relative amplitude of a TID and the amplitude of the atmospheric disturbance causing it depends upon the distribution of ionization and neutrals and upon the character of the atmospheric disturbance. The relationship depends upon local time and it would have to be assumed that it depends only upon local time. In a given time interval, the day-to-day variation of ionization would be included only by determining R using the observed content. Unless measured electron concentration profiles were available on an event-to-event basis, in which case a normalization could be attempted, an average profile would have to be used for all data, admittedly a very crude method.

It does not seem possible to determine the source radiation pattern by studying the diurnal variation in relative amplitudes. However, it may be possible to do this using somewhat different approaches. Two methods will be mentioned here; they will be pursued in a separate study. The first approach, which is the less attractive but simpler of the two, makes use of the information shown in Figure 5-3 and the damping rates estimated in Chapter IV. Assume that the source radiates isotropically. Using the damping rates estimated in the preceding chapter, and including any diurnal variation in the coupling between the atmospheric disturbance and content, hypothetical amplitudes, relative to the amplitude at some time, such as local noon, could be estimated, since the average distance traveled by the disturbances is known. From Figure 5-3, it can

be seen that an estimate of  $\langle R \rangle$  may be obtained at 1000 or 1200 PST. If the calculated amplitude greatly exceeded the sample average at some time, this would be grounds for assuming that disturbances arriving at that time originated from the source in a direction in which the "gain" was small. Note that this approach is valid since the observational limitations are such that the observed sample averages are larger than the true mean. Such an approach should allow determination of the location of nulls and peaks in the radiation pattern, but is very crude. A second, more difficult, method is available. The directional character of the source would be determined if both the input and the output amplitudes were known. It seems reasonable to estimate the input by using the amplitude of the source substorms and using the measured amplitudes of the TIDs as the output. If the source substorm could not be clearly identified, AE indices near the time of interest might be used. Such an approach would require consideration of the variation in damping from disturbance to disturbance as the distance traveled varied, as well as any variation in coupling between atmospheric disturbance and electron content throughout the day. The day could be divided into a number of time intervals and the results in each interval averaged, resulting in an average radiation pattern.

#### D. Discussion

It seems reasonable to seek and to use experimental information in an effort to learn more about the mechanisms involved in producing large scale TIDs. Unfortunately, one of the principal characteristics of the source, its radiation pattern, remains unknown. However, it is still possible to use the known information as the basis for some speculation. As has been pointed out, the evidence points toward some process acting in the evening sector of the auroral oval during polar substorms. What is notable about the evening sector? In auroral activity the most noteworthy feature during substorms appears to be the westward traveling surge [Akasofu, Kimball and Meng, 1965]. This surge consists of a large scale fold which travels westward along pre-existing auroral arcs during the expansive phase of the auroral substorm. The speed of travel is often highly supersonic ( $> 1000 \text{ m s}^{-1}$ ). These high speeds persist

throughout the expansive phase of the substorm (about 30 min), so that the surge will travel 1000 to 2000 km along the oval [Akasofu, Meng and Kimball, 1966]. The surge leads the negative bay in the horizontal component of the geomagnetic field as it travels. After the peak of the expansive phase of the substorm, the surge no longer leads the region of the negative bay. Such surges are common, but very complicated phenomena. The expansion evident in the surge is not due to motions in the atmosphere but is rather due to the motion of the source of high energy particles which produces the aurora, that is, the point at which energy is deposited moves, often supersonically, along the oval. This energy would appear in the form of heat in the region of the surge. In addition, due to the expanding electrojet, Lorentz forces could be expected to affect the neutral atmosphere.

Recently, Chimonas and Hines [1970b] have suggested that the moving shadow produced by a solar eclipse could result in the generation of atmospheric gravity waves. It is possible that a westward traveling surge may be quite analogous as a source. Both the shadow and the surge are supersonic, both have horizontal scales in the direction of motion of more than a thousand kilometers and widths of the order of hundreds of kilometers. Their times of duration are also somewhat similar. The possible detection of an eclipse generated TID has been reported [Davis and da Rosa, 1970]. The possibility of a similar source in the westward traveling surge is worthy of consideration. French [1968] suggested a source for the TID he observed in a supersonic auroral motion. That auroral motion was in the evening sector. Production of disturbances by such a mechanism would probably result in a highly directional radiation pattern; if a bow wave associated with the surge were the cause of the disturbances, little effect might be expected at points behind the surge. Unfortunately at that time, approximately local midnight, an observer using electron content measurements is in a very poor position for studying TIDs. However, Georges' [1967, 1968a] work has shown that large scale TIDs are observed at night. In fact, because of observational biases in his method of observation, he seems to see these disturbances only at that time, rather than primarily during the day, as observed during the present study. In both cases the observational limitations

are due to use of ionization, which varies diurnally, as a tracer for motions in the neutral atmosphere. Perhaps use of electron content measurements and CW Doppler sounders together would provide good coverage throughout the day. At any rate Georges' observation of many disturbances at night would seem to rule out a highly anisotropic radiation pattern such as might be expected from a bow wave. It is possible, however, that the disturbances are produced by more than a single mechanism, and that a bow wave associated with surges in the evening sector may produce disturbances which are observed at midlatitudes during certain times of the day. At other observing times (equivalent to other directions away from the source) other processes might be more efficient in producing disturbances. Clearly, the subject of the source merits considerable additional attention.



## PART II

PRODUCTION OF LARGE SCALE TRAVELING  
IONOSPHERIC DISTURBANCES BY  
DISTURBANCES PROPAGATING IN THE NEUTRAL ATMOSPHERE



## CHAPTER VI

### THEORY

Interpretation of TIDs as manifestations of disturbances propagating in the neutral atmosphere is now generally accepted. It is of interest to study how such disturbances are related to the large scale TIDs considered in Part I. Such a study involves two important problems. The first is the specification of the characteristics of the disturbance in the neutral atmosphere. The second is the determination of how the disturbance produces changes in the ambient ionization. The present chapter is devoted to the problem of how an atmospheric disturbance, once its properties are specified, affects the local ionization. The results obtained in this chapter are quite general in the sense that they are not limited by the models specified in the following chapter. They could be used equally well with other, perhaps better, models for the disturbances in the neutral atmosphere. The principal result of this chapter is an equation describing the changes in electron concentration produced by atmospheric disturbances in the height range 200 to 600 km. Such perturbations are easily integrated to determine the accompanying perturbation in electron content. The numerical analysis required to solve the equation describing the changes in the electron concentration is also presented.

#### A. Background

In recent years several quantitative discussions of the ionospheric effects of atmospheric waves have become available (Georges [1967, 1968b], Hooke [1968a, 1969], Thome [1968], Thome and Rao [1969]). The interested reader is referred to Hooke [1968a] for a discussion of earlier work.

Hooke [1968a] has presented the most complete discussion available of the ionospheric irregularities produced by internal atmospheric gravity waves. He gives considerable attention to perturbations caused by these waves in the motion of the ionization and in photoionization rates. However, in obtaining his final results, he neglects perturbations in the diffusion of the ionization and in the production and loss processes. He

obtains a sinusoidal steady state solution considering only the dominant effect of the waves on the ionization, i.e., the movement of the ionization along the geomagnetic field lines with a velocity equal to the component of the neutral gas velocity along those field lines. Georges [1968b] and Thome [1968] have obtained similar results.

In a subsequent paper Hooke [1969] extends his analysis considerably for the case of E-region ionospheric irregularities. Because of the heights of interest, perturbations in the diffusion of the ionization are negligible. Electron loss is assumed to be proportional to the square of the electron concentration and a Chapman production function is used to describe the unperturbed photoionization rate. Perturbations in both production and loss are included and the recombination coefficient is assumed to be temperature dependent. A solution is obtained using a sinusoidal steady state approach. Hooke [1968a, 1969] uses a single slab, nonviscous, isothermal model for the neutral atmosphere.

Thome and Rao [1969] also neglect perturbations in plasma diffusion as well as neglecting perturbations in production. Perturbations in the loss rate are considered, although any temperature dependence of the reaction rates is neglected. The loss rate is assumed to be given by the parallel loss law of Ratcliffe [1956], although the applicability of that expression, which is based on an assumption of photochemical equilibrium, during the passage of an atmospheric wave, is not justified. Again a sinusoidal steady state solution is obtained. Thome and Rao use the equations of Pitteway and Hines [1963] which describe an atmospheric wave propagating in a viscous, isothermal atmosphere. A multi-slab model for the neutral atmosphere with constant temperature in each slab is used.

No detailed analysis has been carried out specifically for the case of large scale TIDs. Of particular importance for disturbances of this type is inclusion of the perturbation in the plasma diffusion throughout the analysis. Although it is of questionable importance for the disturbances of interest here, the perturbation in the photoionization rate has not yet been studied using a realistic model. A study of the effect of atmospheric waves on electron content based on a detailed model for the coupling between the atmospheric disturbance and the

ionization is not available in the literature. Very recently, however, Georges and Hooke [1970] have used a coupling model equivalent to that of Hooke [1968a] to study perturbations in electron content caused by atmospheric waves.

## B. Outline of Analysis

The approach used by previous workers (e.g., Hooke [1968a, 1969] and Thome and Rao [1969]) to obtain the changes in electron concentration caused by a disturbance in the neutral atmosphere has been based on the linearized continuity equation for electrons. The same approach will be used here.

It is assumed that changes in all quantities of interest are small enough so that a perturbation approach can be used and second order terms can be ignored. In the analysis to follow unsubscripted variables will refer to electrons; those with subscripts "i" and "n" refer to ions and neutrals respectively.

For perturbed conditions, in the presence of an atmospheric disturbance, variables will be written in the following form:  $N = N_0 + N'$ ,  $U = U_0 + U'$ , etc. The zero subscript indicates the unperturbed state and the primed term is the perturbation in the quantity of interest.

Eq. (6-1) is the continuity equation for electrons.

$$\frac{\partial N}{\partial t} + \nabla \cdot (N \vec{U}) = q - \ell \quad (6-1)$$

where  $N$  is the electron concentration and  $\vec{U}$ ,  $q$  and  $\ell$  are the electron velocity, production and loss, respectively. The first order perturbation equation is

$$\frac{\partial N'}{\partial t} + \nabla \cdot (N_0 \vec{U}' + N' \vec{U}_0) = q' - \ell' \quad (6-2)$$

The changes which an atmospheric disturbance produces in the electron concentration are the result of perturbations caused by the disturbance in the motion of the ionization, in the electron photoionization rate, and in the electron loss rate. The remainder of this chapter is devoted to dealing with these various effects. Figure 6-1 shows

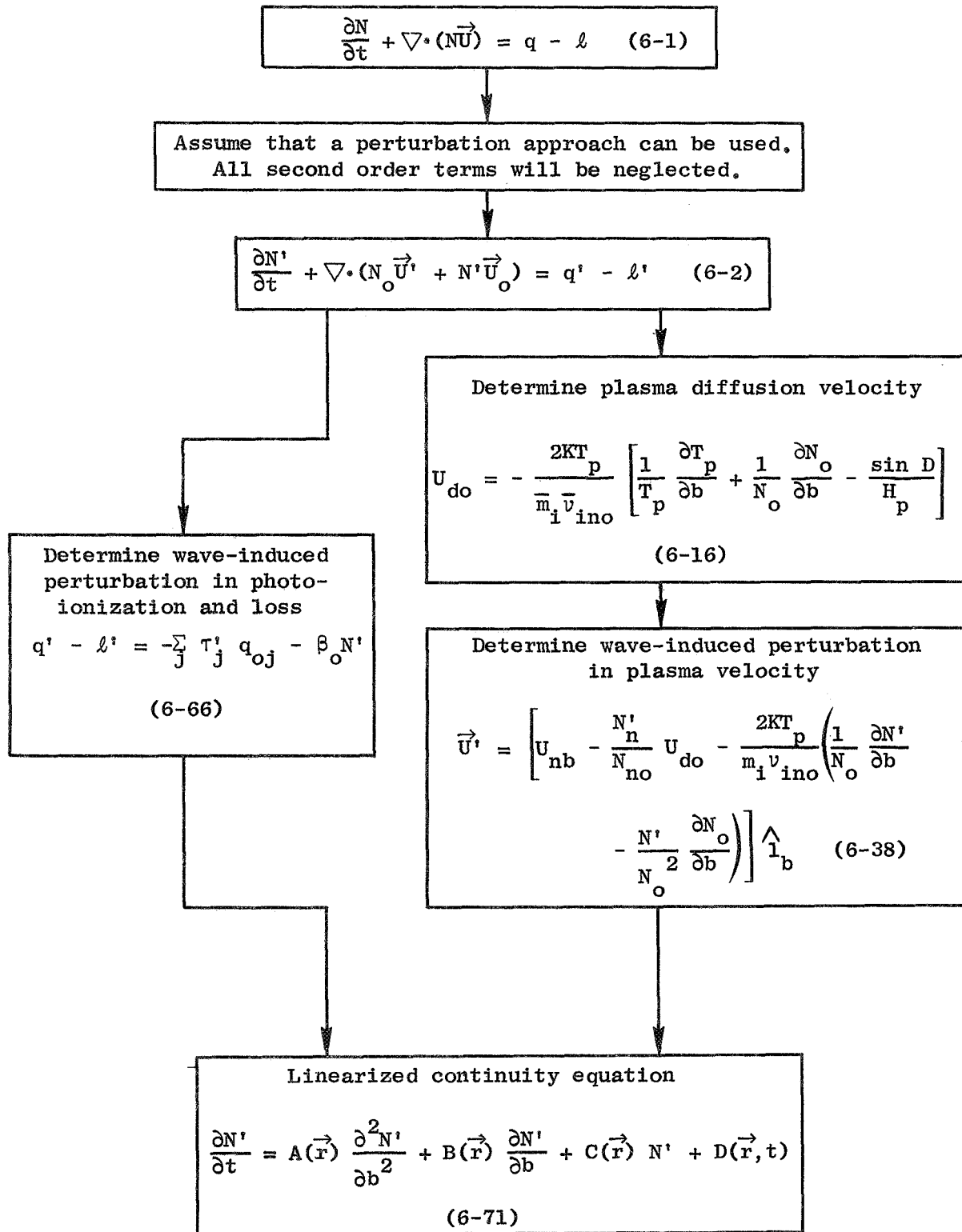


Fig. 6-1. OUTLINE OF ANALYSIS USED

a block diagram of the scheme used. The numbers in parentheses are the numbers of equations in the text; the meanings of the quantities in the equations are given in the text. In the sections which follow, the perturbed and unperturbed divergence in the electron flux will be found first, followed by studies of the changes caused by an atmospheric disturbance in electron production and loss. Finally these results will be used along with Eq. (6-2) to determine the accompanying changes in the electron concentration.

### C. Motion of the Ionization

The divergence term in Eq. (6-2) may be expanded as follows:

$$\nabla \cdot (N_0 \vec{U}' + N' \vec{U}_0) = \nabla \cdot (N_0 \vec{U}') + N' \nabla \cdot \vec{U}_0 + \vec{U}_0 \cdot \nabla N' \quad (6-3)$$

We are interested in each of the terms on the r.h.s. of Eq. (6-3). In order to study them further, expressions are needed for both the perturbed and unperturbed motion of the ionization.

#### 1. Unperturbed motion of the ionization

The purpose of this section is to obtain the unperturbed, steady state diffusion velocity of the ions and electrons, the plasma diffusion velocity. The approach used, a standard one, is to write the equations of motion for both the ions and electrons and then to combine them. Due to the fact that the electron mass is much less than the ion mass, whereas the ion and electron concentrations are the same, a simplified expression for the plasma diffusion velocity may be obtained. Figure 6-2 outlines the procedure followed.

It will be assumed that an equilibrium situation exists in which all ions have the same mean velocity,  $U_0$ . It will also be assumed that this velocity is the same as that of the electrons, due to the electrical forces between them. In addition, it will be assumed that the neutral atmosphere is at rest in the unperturbed state. Winds certainly do exist; however, it is felt that they cannot be specified in a manner which would add meaningfully to the discussion.

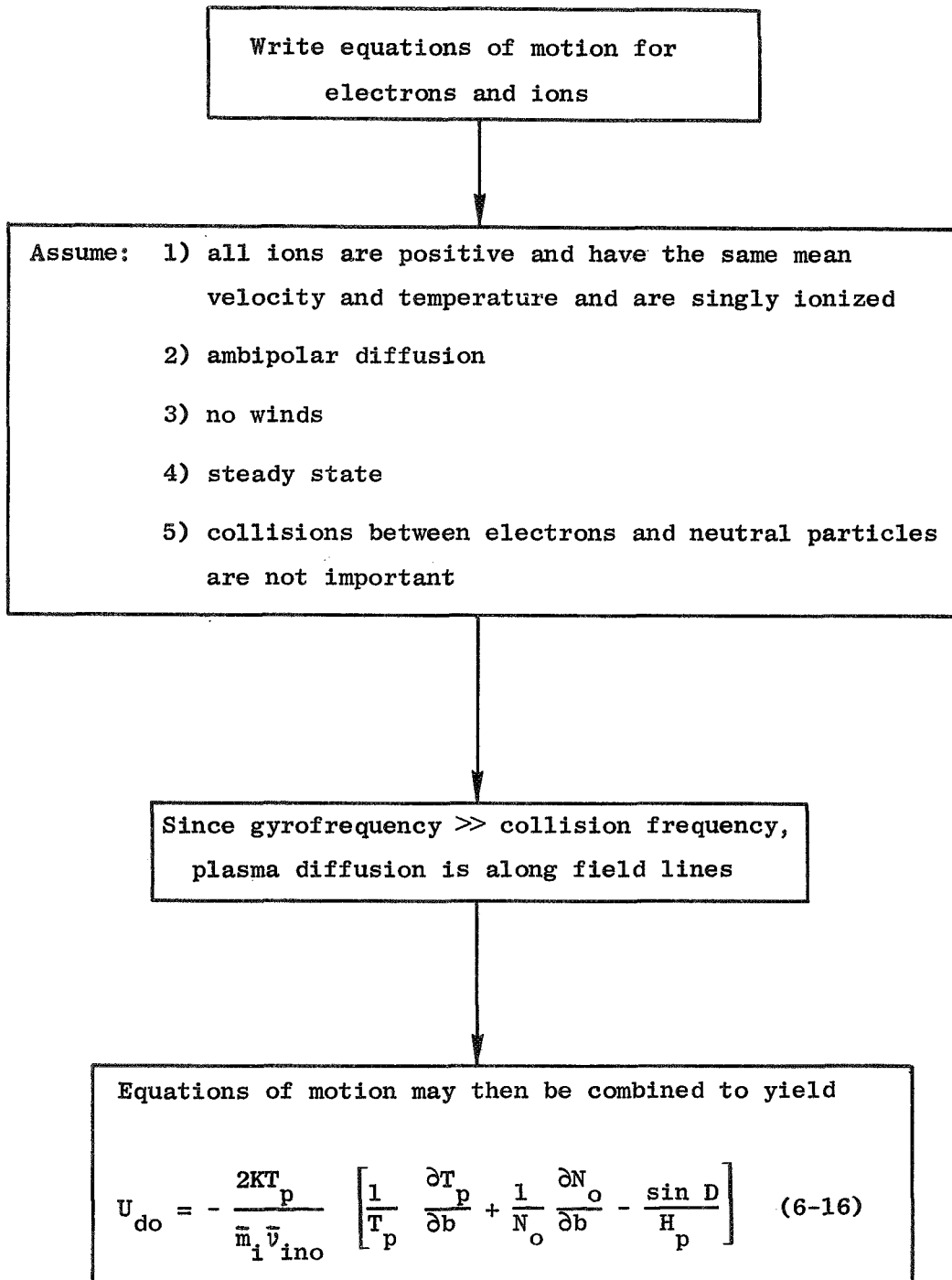


Fig. 6-2. DETERMINATION OF PLASMA DIFFUSION VELOCITY

In what follows  $m_j$  will refer to the mass of a particle, of species  $j$ ,  $N_j$  will refer to its concentration,  $\vec{U}_j$  to its velocity, and  $T_j$  to its temperature.  $\nu_{jk}$  is the collision frequency of species  $j$  with species  $k$ .  $\vec{g}$  is the acceleration due to gravity and  $\vec{E}$  and  $\vec{B}$  are electric and magnetic fields.  $K$  is Boltzman's constant.

The unperturbed equation of motion for ion species  $j$  is as follows:

$$N_{jo} m_j \frac{D\vec{U}_{jo}}{Dt} = -\nabla(N_{jo} K T_{jo}) + N_{jo} m_j \vec{g} + N_{jo} m_j \sum_k \nu_{jko} (\vec{U}_{ko} - \vec{U}_{jo}) \\ + N_{jo} m_j \nu_{ejo} (\vec{U}_o - \vec{U}_{jo}) + N_{jo} e_j [\vec{E}_o + \vec{U}_{jo} \times \vec{B}] = 0 \quad (6-4)$$

The acceleration term is negligible for long spatial scales. Species  $k$  are neutrals. Assume that all ions are positive and singly ionized, good assumptions for the height of interest. Using the assumptions made above concerning the ion and electron velocities, Eq. (6-4) may be simplified.

$$\nabla(N_{jo} K T_{jo}) = N_{jo} m_j \vec{g} - N_{jo} m_j \vec{U}_o \sum_k \nu_{jko} + N_{jo} e [\vec{E}_o + \vec{U}_o \times \vec{B}] \quad (6-5)$$

Adding the equations for all ion species gives

$$\sum_j \nabla(N_{jo} K T_{jo}) = \vec{g} \sum_j N_{jo} m_j - \sum_j [N_{jo} m_j \vec{U}_o \sum_k \nu_{jko}] + e [\vec{E}_o + \vec{U}_o \times \vec{B}] \sum_j N_{jo} \quad (6-6)$$

Assume all ions have the same temperature,  $T_{io}$ ; note that  $\sum_j N_{jo} = N_{io}$  and make the following definitions

$$\bar{m}_i \equiv \frac{\sum_j N_{jo} m_j}{\sum_j N_{jo}} \quad (6-7)$$

$$\bar{m}_i \bar{\nu}_{ino} \equiv \frac{\sum_j [N_{jo} m_j \sum_k \nu_{jko}]}{\sum_j N_{jo}} \quad (6-8)$$

Eq. (6-6) may now be rewritten

$$\nabla(N_{io} K T_{io}) = N_{io} \bar{m}_i \vec{g} - N_{io} \bar{m}_i \bar{v}_{ino} \vec{U}_o + N_{io} e [\vec{E}_o + \vec{U}_o \times \vec{B}] \quad (6-9)$$

A similar expression may also be written for electrons

$$\nabla(N_o K T_o) = N_o m \vec{g} - N_o m \bar{v}_{eno} \vec{U}_o - N_o e [\vec{E}_o + \vec{U}_o \times \vec{B}] \quad (6-10)$$

Assume the following:  $m_i \gg m$  (ion mass much greater than electron mass),  $N_{io} = N_o$  (all ions were assumed negative),  $\bar{m}_i \bar{v}_{ino} \gg m \bar{v}_{eno}$  (collisions with neutral particles not important for electrons). Furthermore, at the heights of interest the plasma diffusion velocity is essentially parallel to  $\vec{B}$ , i.e.,  $\vec{U}_o = U_{do} \hat{1}_b$  where  $\hat{1}_b$  is a unit vector along  $\vec{B}$ . Using these assumptions, Eqs. (6-9) and (6-10) become, when added:

$$\frac{\partial}{\partial b} (N_o K (T_o + T_{io})) = N_o \bar{m}_i \vec{g} \cdot \hat{1}_b - N_o \bar{m}_i \bar{v}_{ino} U_{do} \quad (6-11)$$

$$\therefore U_{do} = - \frac{1}{\bar{m}_i \bar{v}_{ino}} \left[ \frac{1}{N_o} \frac{\partial}{\partial b} (N_o K (T_o + T_{io})) - \bar{m}_i g \sin D \right] \quad (6-12)$$

where  $D$  is the magnetic dip angle (positive in the northern hemisphere) and  $b$  and  $U_{do}$  are positive northwards.  $b$  is a coordinate along the field lines.

Define a plasma temperature

$$T_p \equiv (T_o + T_{io})/2 \quad (6-13)$$

$$\text{Then } U_{do} = \frac{-2KT_p}{\bar{m}_i \bar{v}_{ino}} \left[ \frac{1}{T_p} \frac{\partial T_p}{\partial b} + \frac{1}{N_o} \frac{\partial N_o}{\partial b} - \frac{\sin D}{2(KT_p/\bar{m}_i g)} \right] \quad (6-14)$$

Define a plasma scale height

$$H_p \equiv \frac{2KT_p}{\bar{m}_i g} \quad (6-15)$$

Then

$$U_{do} = \frac{-2KT_p}{\bar{m}_i \bar{v}_{ino}} \left[ \frac{1}{T_p} \frac{\partial T_p}{\partial b} + \frac{1}{N_o} \frac{\partial N_o}{\partial b} - \frac{\sin D}{H_p} \right] \quad (6-16)$$

In what follows the bars indicating that the ion mass and the ion-neutral collision frequency are averages will be dropped. However, it should be recalled whenever these quantities occur that they are average values.

Some of the terms in Eq. (6-3) may be simplified somewhat. Since  $\vec{U}_o = U_{do} \hat{1}_b$ ,

$$\nabla \cdot \vec{U}_o = \frac{\partial U_{do}}{\partial b} \quad (6-17)$$

and

$$\vec{U}_o \cdot \nabla N' = U_{do} \frac{\partial N'}{\partial b} \quad (6-18)$$

## 2. Perturbed motion of the ionization

An atmospheric disturbance modifies the motion of the ionization in two ways. First, because of collisions between the neutrals and ions, the plasma tends to follow the motions of the neutral gas. Second, due to changes in the distribution of the ionization, its temperature and the ion-neutral collision frequency, the plasma diffusion velocity is altered. The approach which is used here to obtain the changes in the velocity of the ionization is essentially that used by Hooke [1968a]. The perturbed equation of motion for ions is written. Utilizing the fact that the disturbance alters the motions of the ionization in the two ways mentioned above, this equation can be separated into two parts. These two equations are then combined with similar equations for electrons; the change in the plasma velocity caused directly by collisions and that due to the perturbation in the diffusion velocity may then be found. The method used is outlined in Figure 6-3.

Assuming that  $N'_i = N'$ , the perturbed equation of motion for ions is as follows

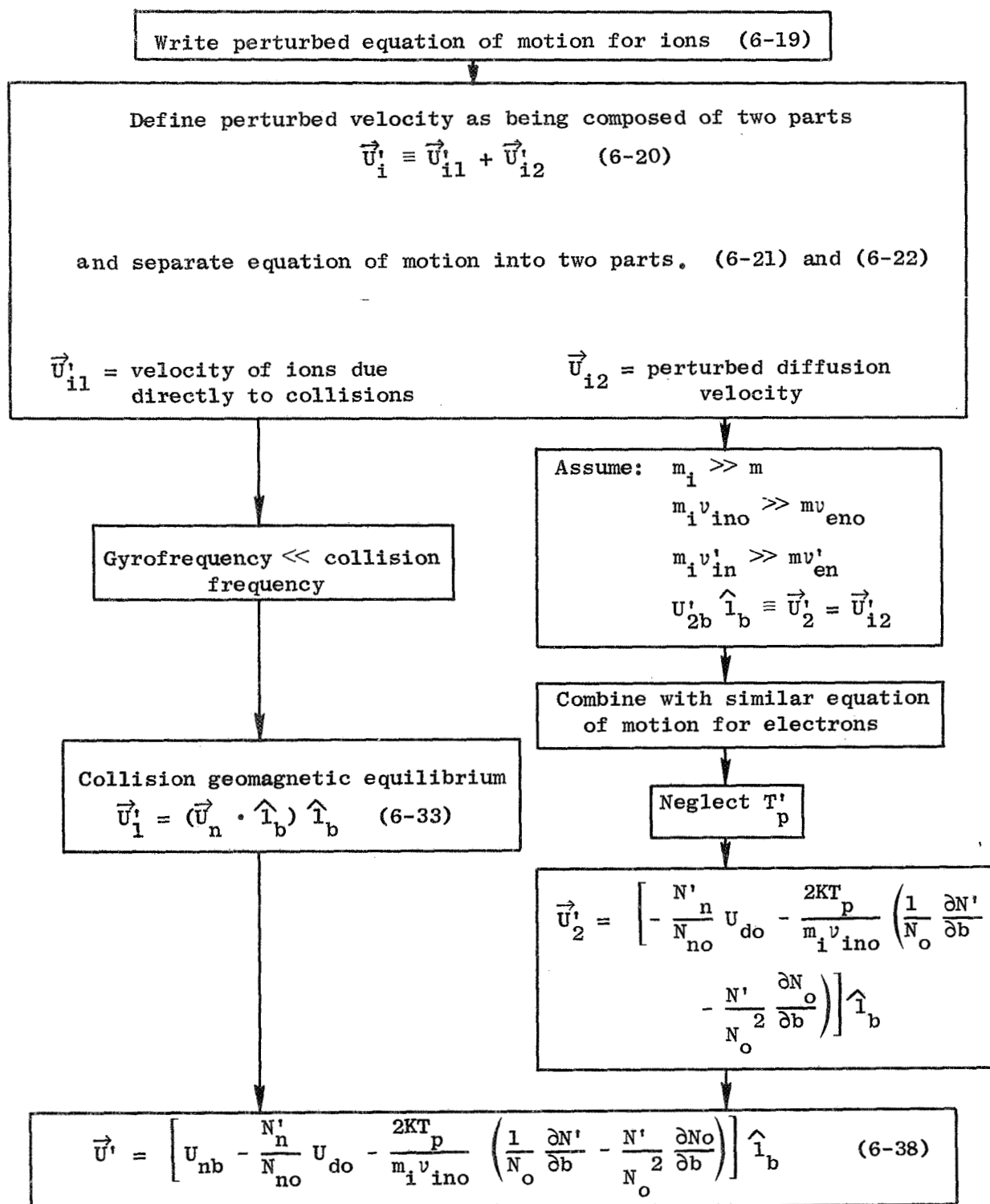


Fig. 6-3. DETERMINATION OF THE PERTURBED MOTION OF THE IONIZATION

$$\begin{aligned}
m_i \left( N' \frac{D\vec{U}_o}{Dt} + N_o \frac{D\vec{U}'_i}{Dt} \right) = & -\nabla(N'KT_{io} + N_oKT'_i) + N'm_i\vec{g} \\
& - N'm_i v_{ino} \vec{U}_o - N_o m_i v'_{in} \vec{U}_o + N_o m_i v_{ino} (\vec{U}'_n - \vec{U}'_i) \\
& + N'e [\vec{E}_o + \vec{U}_o \times \vec{B}] + N_o e [\vec{U}'_i \times \vec{B}]
\end{aligned} \tag{6-19}$$

The perturbation in  $\vec{E}$  is neglected. This is apparently justifiable [Hooke, 1968a]. The inertia terms will be neglected subsequently because of the long spatial and time scales of interest.

Using Hooke's [1968a] notation, define

$$\vec{U}'_i \equiv \vec{U}'_{i1} + \vec{U}'_{i2} \tag{6-20}$$

where  $\vec{U}'_{i2}$  satisfies

$$\begin{aligned}
0 \equiv & -\nabla(N'KT_{io} + N_oKT'_i) + N'm_i\vec{g} - N'm_i v_{ino} \vec{U}_o \\
& - N_o m_i v'_{in} \vec{U}_o - N_o m_i v_{ino} \vec{U}'_{i2} + N'e (\vec{E}_o + \vec{U}_o \times \vec{B}) \\
& + N_o e (\vec{U}'_{i2} \times \vec{B})
\end{aligned} \tag{6-21}$$

and  $\vec{U}'_{i1}$  satisfies

$$0 \equiv N_o m_i v_{ino} (\vec{U}'_n - \vec{U}'_{i1}) + N_o e (\vec{U}'_{i1} \times \vec{B}) \tag{6-22}$$

By comparing Eqs. (6-21) and (6-9), it is apparent that  $\vec{U}'_{i2}$  is the perturbation in  $\vec{U}_i$  caused by perturbations in  $N_i$ ,  $T_i$  and  $v_{in}$ .  $\vec{U}'_{i2}$  will be essentially parallel to  $\vec{B}$ .

Writing an equation similar to Eq. (6-21) for electrons gives

$$\begin{aligned}
0 = & -\nabla(N'KT_o + N_oKT') + N'm\vec{g} - N'mv_{eno} \vec{U}_o \\
& - N_o m v'_{en} \vec{U}_o - N_o m v_{eno} \vec{U}'_2 - N'e (\vec{E}_o + \vec{U}_o \times \vec{B}) \\
& - N_o e (\vec{U}'_2 \times \vec{B})
\end{aligned} \tag{6-23}$$

Assume that  $m_i \gg m$  and that  $m_i v'_i \gg m v'_{en}$ ; the latter is reasonable since  $m_i \gg m$  and since there is no reason to expect a much larger percentage increase in  $v_{en}$  than in  $v_{in}$ . It seems reasonable to assume that the perturbed motion is also along  $\vec{B}$  (collision frequency  $\gg$  gyrofrequency) and that the perturbation in diffusion is ambipolar, i.e.,  $U'_{2b} \hat{1}_b \equiv \vec{U}'_2 = \vec{U}'_{i2}$  is assumed. Using these assumptions, Eqs. (6-21) and (6-23) may be added to yield:

$$0 = - \frac{\partial}{\partial b} (N' K T_{i0} + N_o K T'_i + N' K T_o + N_o K T') + N' m_i g \sin D - N' m_i v_{ino} U_{do} - N_o m_i v'_{in} U_{do} - N_o m_i v_{ino} U'_{2b} \quad (6-24)$$

Solving for  $U'_{2b}$ ,

$$U'_{2b} = \frac{1}{N_o m_i v_{ino}} \left\{ - \frac{\partial}{\partial b} (N_o K (T'_i + T') + N' K (T_{i0} + T_o)) - m_i U_{do} (N' v_{ino} + N_o v'_{in}) + N' m_i g \sin D \right\} \quad (6-25)$$

Define a perturbation in the plasma temperature

$$T'_p \equiv (T'_i + T')/2 \quad (6-26)$$

Then

$$U'_{2b} = - \frac{2K}{m_i v_{ino}} \left\{ \frac{\partial T'_p}{\partial b} + \frac{T'_p}{N_o} \frac{\partial N_o}{\partial b} + \frac{N'}{N_o} \frac{\partial T_p}{\partial b} + \frac{T_p}{N_o} \frac{\partial N'}{\partial b} - \frac{N'}{2N_o K} m_i g \sin D \right\} - U_{do} \left[ \frac{N'}{N_o} + \frac{v'_{in}}{v_{ino}} \right] \quad (6-27)$$

Rearranging this equation, substituting Eq. (6-16) for  $U_{do}$  and simplifying gives

$$U'_{2b} = - \frac{2K T_p}{m_i v_{ino}} \left[ \frac{1}{N_o} \frac{\partial N'}{\partial b} + \frac{1}{T_p} \frac{\partial T'_p}{\partial b} + \left( \frac{T'_p}{T_p} - \frac{N'}{N_o} \right) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right] - \frac{v'_{in}}{v_{ino}} U_{do} \quad (6-28)$$

Eq. (6-28) gives the perturbation in the plasma diffusion velocity. It can also be obtained directly from Eq. (6-16) by writing the perturbation equation.

Consider  $\vec{U}'_{i1}$  again. Eq. (6-22) may be rewritten as

$$\frac{m_i v_{ino}}{eB} (\vec{U}'_n - \vec{U}'_{i1}) + \vec{U}'_{i1} \times \hat{1}_b = 0 \quad (6-29)$$

Define  $\rho$ , the ratio of collision frequency to gyrofrequency

$$\rho \equiv \frac{m_i v_{ino}}{eB} \quad (6-30)$$

Then

$$\rho (\vec{U}'_n - \vec{U}'_{i1}) + \vec{U}'_{i1} \times \hat{1}_b = 0 \quad (6-31)$$

The solution of Eq. (6-31) is

$$\vec{U}'_{i1} = \frac{1}{1 + \rho^2} \left[ \rho^2 \vec{U}'_n + \rho \vec{U}'_n \times \hat{1}_b + (\vec{U}'_n \cdot \hat{1}_b) \hat{1}_b \right] \quad (6-32)$$

Eq. (6-32) gives the velocity of the ions caused by collisions with the neutrals when the only other force present is due to the geomagnetic field. The electrons are assumed to be dragged along with the ions ( $\vec{U}'_1 = \vec{U}'_{i1}$ ). This result has been presented by MacLeod [1965] who labeled the situation "collision-geomagnetic equilibrium".

For the heights of interest,  $\rho \ll 1$  and Eq. (6-32) becomes

$$\vec{U}'_1 = (\vec{U}'_n \cdot \hat{1}_b) \hat{1}_b \quad (6-33)$$

Combining (6-28) and (6-33), the complete expression for the perturbed plasma velocity is as follows

$$U' = (\vec{U}'_n \cdot \hat{1}_b) - \frac{v'_{in}}{v_{ino}} U_{do} - \frac{2KT_p}{m_i v_{ino}} \left[ \frac{1}{N_o} \frac{\partial N'}{\partial b} + \frac{1}{T_p} \frac{\partial T'_p}{\partial b} + \left( \frac{T'_p}{T_p} - \frac{N'}{N_o} \right) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right] \quad (6-34)$$

$U'$  is along the field lines.

Consider next the ion-neutral collision frequency. The collision frequency for momentum transfer has been given by Chapman [1956].

$$\nu_{in} = 2.6 \times 10^{-15} N_n M^{-1/2} \quad [\text{m.k.s.}] \quad (6-35)$$

$M$  is the average value of ion and neutral particle mass given in a.m.u. It is assumed that the temperatures of all species are the same. Since it was previously assumed that all ion temperatures are the same and since the ion and neutral temperatures are approximately the same, this is an acceptable approximation. Eq. (6-35) gives an average collision frequency as discussed earlier in this chapter and hence is an appropriate expression to use in this discussion. Assuming that the mean molecular weight is constant:

$$\nu'_{in} = 2.6 \times 10^{-15} N'_n M^{-1/2} \quad (6-36)$$

Therefore

$$\frac{\nu'_{in}}{\nu_{ino}} = \frac{N'_n}{N_{no}} \quad (6-37)$$

It will be assumed that perturbations in the plasma temperature can be neglected, i.e.,  $T'_p = 0$  will be assumed.

Eq. (6-34) may now be rewritten as follows

$$\vec{U}' = \left[ U_{nb} - \frac{N'_n}{N_{no}} U_{do} - \frac{2KT_p}{m_i \nu_{ino}} \left( \frac{1}{N_o} \frac{\partial N'}{\partial b} - \frac{N'}{N_o^2} \frac{\partial N_o}{\partial b} \right) \right] \hat{1}_b \quad (6-38)$$

where  $U_{nb}$  is defined by

$$U_{nb} \equiv \vec{U}_n \cdot \hat{1}_b \quad (6-39)$$

Eq. (6-38) is the final, simplified expression for the perturbation in plasma velocity.

The divergence term given in Eq. (6-3) may now be developed further. Using Eqs. (6-17), (6-18), and (6-38), Eq. (6-3) becomes

$$\begin{aligned} \nabla \cdot (N_o \vec{U}' + N' \vec{U}_o) = N' \frac{\partial U_{do}}{\partial b} + U_{do} \frac{\partial N'}{\partial b} + \nabla \cdot \left[ N_o \left( U_{nb} - \frac{N'_n}{N_{no}} U_{do} \right. \right. \\ \left. \left. - \frac{2KT_p}{m_i v_{ino}} \left( \frac{1}{N_o} \frac{\partial N'}{\partial b} - \frac{N'}{N_o^2} \frac{\partial N_o}{\partial b} \right) \right) \hat{1}_b \right] \end{aligned} \quad (6-40)$$

Expanding

$$\begin{aligned} \nabla \cdot (N_o \vec{U}' + N' \vec{U}_o) = N' \frac{\partial U_{do}}{\partial b} + U_{do} \frac{\partial N'}{\partial b} + \frac{\partial}{\partial b} (N_o U_{nb}) - \frac{\partial}{\partial b} \left( \frac{N'_n}{N_{no}} N_o U_{do} \right) \\ - \frac{\partial}{\partial b} \left[ \frac{2KT_p}{m_i v_{ino}} \left( \frac{\partial N'}{\partial b} - \frac{N'}{N_o} \frac{\partial N_o}{\partial b} \right) \right] \end{aligned} \quad (6-41)$$

Consider the last term on the r.h.s. of Eq. (6-41)

$$\begin{aligned} \frac{\partial}{\partial b} \left[ \frac{2KT_p}{m_i v_{ino}} \left( \frac{\partial N'}{\partial b} - \frac{N'}{N_o} \frac{\partial N_o}{\partial b} \right) \right] = \frac{\partial}{\partial b} \left( \frac{2KT_p}{m_i v_{ino}} \right) \frac{\partial N'}{\partial b} + \frac{2KT_p}{m_i v_{ino}} \frac{\partial^2 N'}{\partial b^2} \\ - \frac{\partial}{\partial b} \left( \frac{2KT_p}{m_i v_{ino}} \cdot \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right) \cdot N' - \frac{2KT_p}{m_i v_{ino}} \frac{1}{N_o} \frac{\partial N_o}{\partial b} \frac{\partial N'}{\partial b} \end{aligned} \quad (6-42)$$

Make the following definitions, where  $\vec{r}$  is a position vector.

$$A(\vec{r}) \equiv \frac{2KT_p}{m_i v_{ino}} \quad (\text{diffusion constant}) \quad (6-43)$$

$$B(\vec{r}) \equiv - U_{do} + \frac{\partial A(\vec{r})}{\partial b} - A(\vec{r}) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \quad (6-44)$$

which upon substituting for  $U_{do}$  and simplifying becomes

$$B(\vec{r}) \equiv A(\vec{r}) \left[ \frac{1}{T_p} \frac{\partial T_p}{\partial b} - \frac{\sin D}{H_p} \right] + \frac{\partial A(\vec{r})}{\partial b} \quad (6-45)$$

Hence Eq. (6-41) may be written as

$$\begin{aligned} \nabla \cdot (N_o \vec{U}' + N' \vec{U}_o) = & - A(\vec{r}) \frac{\partial^2 N'}{\partial b^2} - B(\vec{r}) \frac{\partial N'}{\partial b} + \left[ \frac{\partial U_{do}}{\partial b} + \frac{\partial}{\partial b} \left( A(\vec{r}) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right) \right] N' \\ & + \frac{\partial}{\partial b} (N_o U_{nb}) - \frac{\partial}{\partial b} \left( \frac{N'}{N_{no}} N_o U_{do} \right) \end{aligned} \quad (6-46)$$

Eq. (6-46) is the final expression for the divergence in the plasma flux caused by a disturbance in the neutral atmosphere.

#### D. Effect of Atmospheric Disturbances on Photoionization

The change in photoionization caused by an atmospheric disturbance can be determined from the expression for the perturbed photoionization rate. By straightforward algebraic manipulation, it is shown that the perturbation in ionization can be written as two terms, one due to the perturbation in the optical depth caused by the disturbance and the other due to the changes in the local concentration of the neutrals which the disturbance produces. The approach used is outlined in Figure 6-4.

The rate of production of electrons per unit volume by photoionization,  $q$ , is given by

$$q = \sum_j (\Phi_j e^{-\tau_j} \sum_k \eta_{jk} \sigma_{jk} N_k) \quad (6-47)$$

where summations are over the neutral species  $k$  and wavelength bands  $j$ , and the quantities have the following meanings

$\Phi_j$  = flux in  $j$ th band outside atmosphere,

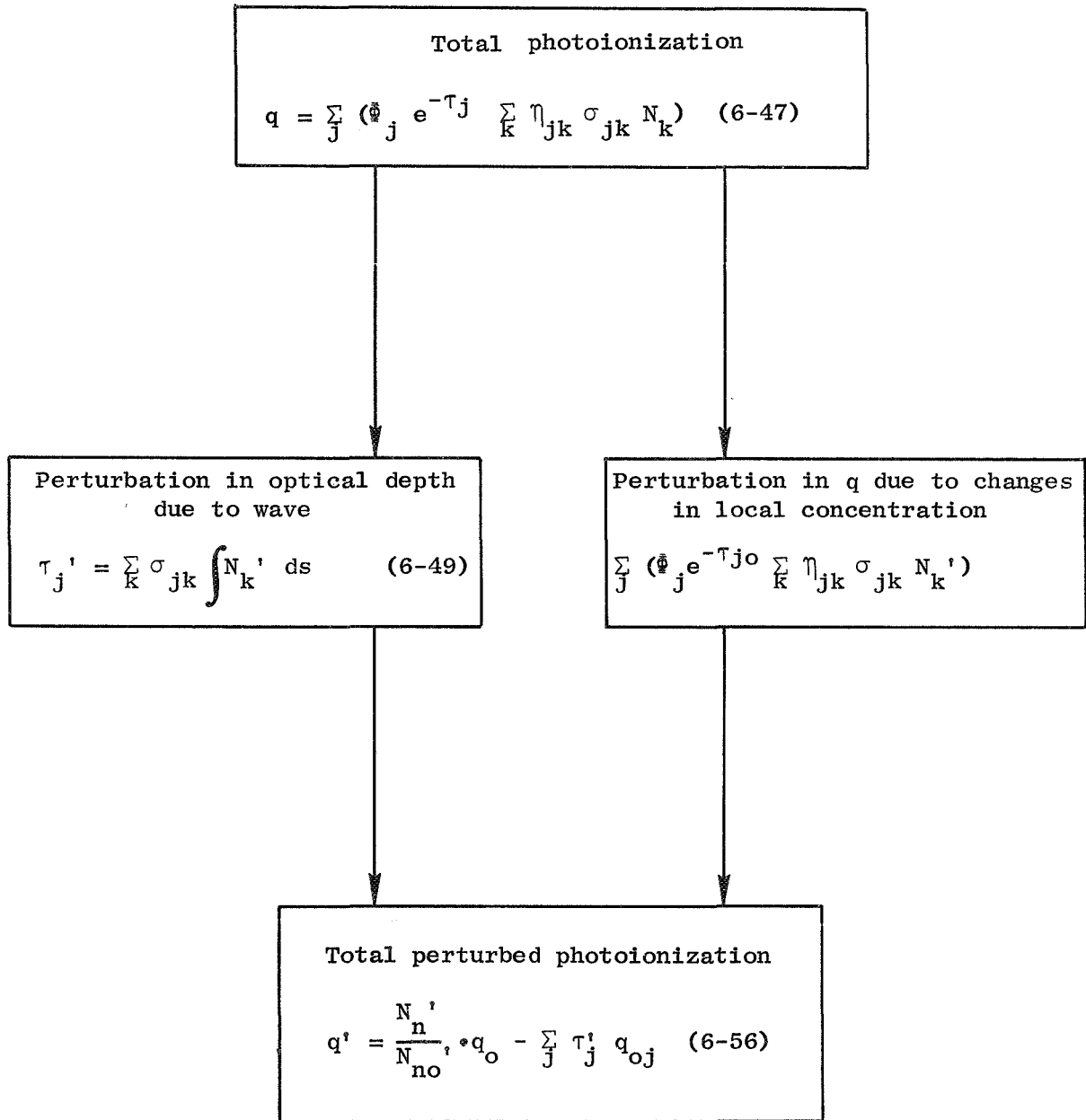


Fig. 6-4. DETERMINATION OF THE PERTURBED PHOTOIONIZATION

$\tau_j$  = optical depth in jth wavelength band,

$\eta_{jk}$  = ionizing efficiency of kth species for jth wavelength band,

$\sigma_{jk}$  = absorption cross-section of kth species for jth wavelength band,

and  $N_k$  = concentration of kth neutral constituent.

Optical depth is given by the following

$$\tau_j = \sum_k \sigma_{jk} \int N_k ds \quad (6-48)$$

therefore the perturbation in optical depth caused by an atmospheric disturbance may be given as

$$\tau'_j = \sum_k \sigma_{jk} \int N'_k ds \quad (6-49)$$

Consider the total production as being composed of a perturbed and an unperturbed part

$$q_o + q' = \sum_j \left[ \frac{1}{j} e^{-(\tau_{jo} + \tau'_j)} \sum_k \eta_{jk} \sigma_{jk} (N_{ko} + N'_k) \right] \quad (6-50)$$

Examine the factor  $e^{-(\tau_{jo} + \tau'_j)}$

$$e^{-(\tau_{jo} + \tau'_j)} = \left[ e^{\tau'_j/\tau_{jo}} \cdot e \right]^{-\tau_{jo}} \quad (6-51)$$

Assume  $\tau'_j/\tau_{jo} \ll 1$ , then

$$e^{-(\tau_{jo} + \tau'_j)} \approx (1 + \tau'_j/\tau_{jo})^{-\tau_{jo}} \cdot e^{-\tau_{jo}} \quad (6-52)$$

Taking the binomial expansion of the term in parenthesis and simplifying gives

$$e^{-(\tau_{jo} + \tau'_j)} \approx e^{-\tau_{jo}} - \tau'_j e^{-\tau_{jo}} \quad (6-53)$$

therefore

$$q_o + q' = \sum_j [\Phi_j e^{-\tau_{jo}} (1 - \tau'_j) \sum_k \eta_{jk} \sigma_{jk} (N_{ko} + N'_k)] \quad (6-54)$$

whence

$$q' = \sum_j [\Phi_j e^{-\tau_{jo}} \sum_k \eta_{jk} \sigma_{jk} N'_k] - \sum_j [\Phi_j \tau'_j e^{-\tau_{jo}} \sum_k \eta_{jk} \sigma_{jk} N_{ko}] \quad (6-55)$$

Assume that  $N'_k/N_{ko}$  is independent of species; then

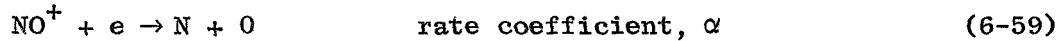
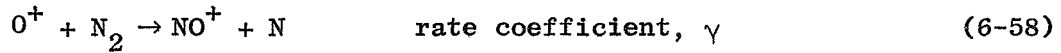
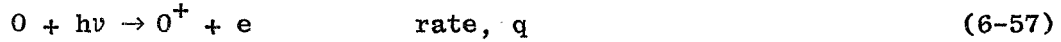
$$q' = \frac{N'_n}{N_{no}} q_o - \sum_j \tau'_j q_{oj} \quad (6-56)$$

where  $q_{oj}$  is the unperturbed production due to the  $j$ th band.

Eq. (6-56) is an expression for the perturbation in production caused by any disturbance which changes the distribution of the neutral atmosphere. Each of the two terms on the r.h.s. of Eq. (6-56) has a very definite physical meaning. The first gives the change in photoionization due to changes in the neutral concentration at the point of interest. The second gives the variation due to changes in the solar flux reaching the point of interest caused by the redistribution of neutral particles along the path traveled by the radiation. This point was discussed by Hooke [1968a]; his approach was less general than the one used here, since he assumed a monochromatic flux of radiation and a neutral atmosphere containing a single constituent in order to obtain an expression for  $q'$  which could be dealt with analytically.

### E. Effect of Atmospheric Disturbances on Loss Processes

At the heights of interest the photoionization of molecular ions can be neglected. The reactions which will be considered are the following:



In the above set of equations N stands for nitrogen and not for electron concentration as in the rest of the text. At the altitudes of interest, the reaction described in Eq. (6-58) proceeds more slowly than that in Eq. (6-59) due to the lack of molecules of  $N_2$  and is therefore the controlling reaction at those heights. A linear formula may be used to describe the electron loss

$$l = \beta N \quad (6-60)$$

where

$$\beta = \gamma [N_2]$$

The perturbation in the loss caused by the presence of an atmospheric disturbance is given by

$$l' = \beta' N_O + \beta_O N' \quad (6-61)$$

where

$$\beta' = \gamma' [N_2]_O + \gamma [N_2]' \quad (6-62)$$

$\gamma$  will be assumed independent of temperature. Neglecting any perturbation in  $\gamma$ , Eq. (6-61) may be written

$$l' = \gamma [N_2]' N_O + \beta_O N' \quad (6-63)$$

or

$$\ell' = \left( \frac{N'_n}{N_{no}} N_o + N' \right) \beta_o \quad (6-64)$$

#### F. Continuity Equation: Conclusion

Now that all the terms in Eq. (6-2) have been examined in detail, the results of the preceding sections will be combined here with that equation to give a complete expression for the linearized electron continuity equation.

In Eq. (6-2) there is a term of the form  $q' - \ell'$ ; using Eqs. (6-56) and (6-64) this becomes

$$q' - \ell' = \frac{N'_n}{N_{no}} q_o - \sum_j \tau'_j q_{oj} - \left( \frac{N'_n}{N_{no}} N_o + N' \right) \beta_o \quad (6-65)$$

Note that below the F-layer peak photochemical equilibrium is a good approximation during unperturbed conditions and that  $q_o = \beta_o N_o$ . Therefore, Eq. (6-65) shows that the change in loss rate due to a change in  $\beta$  effectively cancels the change in production due to local concentration changes at the point of interest. At higher levels where photochemical equilibrium is no longer a good approximation, the importance of the production and loss terms is greatly decreased. Therefore, the cancellation will be assumed to exist at all heights (although it is strictly true only when photochemical equilibrium exists) and Eq. (6-65) may be simplified to give

$$q' - \ell' = -\beta_o N' - \sum_j \tau'_j q_{oj} \quad (6-66)$$

Substituting Eqs. (6-46) and (6-66) into Eq. (6-2) results in the following:

$$\begin{aligned} \frac{\partial N'}{\partial t} - A(\vec{r}) \frac{\partial^2 N'}{\partial b^2} - B(\vec{r}) \frac{\partial N'}{\partial b} + \left[ \frac{\partial U_{do}}{\partial b} + \frac{\partial}{\partial b} \left( A(\vec{r}) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right) \right] N' + \frac{\partial}{\partial b} (N_o U_{nb}) \\ - \frac{\partial}{\partial b} \left( \frac{N'_n}{N_{no}} N_o U_{do} \right) = -\beta_o N' - \sum_j \tau'_j q_{oj} \end{aligned} \quad (6-67)$$

The following definition of C will be made.

$$C(\vec{r}) \equiv - \frac{\partial U_{do}}{\partial b} - \beta_o - \frac{\partial}{\partial b} \left( A(\vec{r}) \frac{1}{N_o} \frac{\partial N_o}{\partial b} \right) \quad (6-68)$$

Upon simplification this becomes

$$C(r) \equiv - \beta_o + \frac{\partial}{\partial b} \left[ A(\vec{r}) \left( \frac{1}{T_p} \frac{\partial T_p}{\partial b} - \frac{\sin D}{H_p} \right) \right] \quad (6-69)$$

Another new quantity D will also be defined.

$$D(\vec{r}, t) \equiv - \sum_j \tau'_j q_{oj} - \frac{\partial}{\partial b} (N_o U_{nb}) + \frac{\partial}{\partial b} \left( \frac{N'_n}{N_{no}} N_o U_{do} \right) \quad (6-70)$$

Eq. (6-67) may now be rewritten as follows:

$$\frac{\partial N'}{\partial t} = A(\vec{r}) \frac{\partial^2 N'}{\partial b^2} + B(\vec{r}) \frac{\partial N'}{\partial b} + C(\vec{r}) N' + D(\vec{r}, t) \quad (6-71)$$

Eq. (6-71) is the partial differential equation which must be solved in order to determine changes in the electron concentration produced by the passage of an atmospheric disturbance. This equation cannot be solved analytically; numerical solutions will be considered in the following section. Once the perturbations in the electron concentration have been found, it is very simple to determine the fluctuation in the electron content. Columnar electron content is given by

$$I = \int (N_o + N') ds \quad (6-72)$$

where ds is an element of path length along which the content is being determined. The perturbation in the electron content is then simply

$$I' = \int N' ds \quad (6-73)$$

The details involved in finding I' will be discussed in the following section.

## G. Numerical Analysis

### 1. Boundary and initial conditions

Before the arrival of the disturbance in the neutral atmosphere, the electron concentration was unperturbed. Hence  $N' = 0$  would appear to be an appropriate initial condition for solving Eq. (6-71).

Atmospheric disturbances induce changes in the ionosphere primarily because the ionization is coupled by collisions to motions in the neutral atmosphere. At low altitudes the electron concentration becomes small enough so that fluctuations in the ionization caused by the disturbances cannot be detected. Therefore, a reasonable lower boundary condition would be  $N' = 0$  at an appropriate height. At high altitudes  $N'$  also approaches zero due not only to small electron concentration but also due to the fact that the disturbance in the neutral atmosphere is damped by viscous forces. The upper boundary condition which will be used is also  $N' = 0$ . Based on Thome's [1968] observations of LS TIDs,  $N' = 0$  would be roughly true at 150 km and again at 600 or 700 km. Precise limits are difficult to assign because of the inaccuracy of the data at high and low altitudes. Tesud and Vasseur [1969] present excellent data for LS TIDs for the very important 200 to 400 km height range but, unfortunately, do not extend their observations to heights where we need to establish our boundary conditions. However, the precise heights at which the boundary conditions are applied is probably not too important when studying changes in the electron concentration and even less when studying perturbations in electron content just as long as they are near or outside the 200 to 400 km range.

### 2. Numerical solution

A finite difference approach will be used to obtain solutions to Eq. (6-71). Let  $N'_{i,j}$  denote the perturbation in the electron concentration at the  $i$ th height step and the  $j$ th time step. Let the time step be  $k$  and the increment in displacement along the field lines be  $h$ .

Using the Crank-Nicholson implicit method [Crank and Nicholson, 1947], the following difference equation may be written for Eq. (6-71):

$$\frac{N'_{i,j+1} - N'_{i,j}}{k} = \frac{A_i}{2h^2} [N'_{i+1,j+1} - 2N'_{i,j+1} + N'_{i-1,j+1} + N'_{i+1,j} - 2N'_{i,j} + N'_{i-1,j}] + \frac{B_i}{h} [N'_{i+1,j} - N'_{i,j}] + C_i N'_{i,j} + D_{i,j} \quad (6-74)$$

For brevity let  $x_i = N'_{i,j+1}$

$$N'_i = N'_{i,j} \quad (6-75)$$

$$D_i = D_{i,j}$$

Then

$$\begin{aligned} \frac{x_i}{k} - \frac{A_i}{2h^2} x_{i+1} + \frac{A_i x_i}{h^2} - \frac{A_i x_{i-1}}{2h^2} &= \frac{N'_i}{k} + \frac{A_i N'_{i+1}}{2h^2} - \frac{A_i N'_i}{h^2} \\ &+ \frac{A_i N'_{i-1}}{2h^2} + \frac{B_i}{h} N'_{i+1} - \frac{B_i}{h} N'_i + C_i N'_i + D_i \end{aligned} \quad (6-76)$$

Let  $\alpha_i = -\frac{A_i}{2h^2}$

$$\beta_i = \frac{1}{k} + \frac{A_i}{h^2} \quad (6-77)$$

$$\gamma_i = N'_i \left( \frac{1}{k} - \frac{A_i}{h^2} - \frac{B_i}{h} + C_i \right) + N'_{i+1} \left( \frac{A_i}{2h^2} + \frac{B_i}{h} \right) + N'_{i-1} \left( \frac{A_i}{2h^2} \right) + D_i$$

Then

$$\alpha_i x_{i-1} + \beta_i x_i + \alpha_{i+1} x_{i+1} = \gamma_i \quad i = 1, 2, \dots, N-1 \quad (6-78)$$

The boundary conditions are  $X_0 = 0$  and  $X_N = 0$ .

Also  $N'_{i,1} = 0$  implies  $\gamma_i = D_i$  for  $j = 1$ .

One now has the following system of equations

$$\begin{aligned}
 0 + \beta_1 X_1 + \alpha_1 X_2 &= \gamma_1 \\
 \alpha_2 X_1 + \beta_2 X_2 + \alpha_2 X_3 &= \gamma_2 \\
 \alpha_3 X_2 + \beta_3 X_3 + \alpha_3 X_4 &= \gamma_3 \\
 &\vdots \\
 &\vdots \\
 &\vdots \\
 \alpha_{N-1} X_{N-2} + \beta_{N-1} X_{N-1} + 0 &= \gamma_{N-1}
 \end{aligned} \tag{6-79}$$

These equations can be solved using Gauss' elimination method.

$X_i$  is determined in terms of  $X_{i+1}$ .

$$X_i = E_i X_{i+1} + F_i \tag{6-80}$$

where

$$E_i = \frac{-\alpha_i}{\beta_i + \alpha_i E_{i-1}} \tag{6-81}$$

$$F_i = \frac{\gamma_i - \alpha_i F_{i-1}}{\beta_i + \alpha_i E_{i-1}} \tag{6-82}$$

$$E_0 = 0, F_0 = X_0 = 0$$

so

$$E_1 = -\frac{\alpha_1}{\beta_1}, F_1 = \frac{\gamma_1}{\beta_1}$$

Since  $X_N = 0$ ,  $X_{N-1}$  may be found and so on by back-substitution until  $X_1$  is found. Since the initial perturbation profile is known (equivalent to knowing initial  $\gamma$ 's) the profile at  $j = 2$  can be found and so on for later times.

### 3. Integrated effects

Referring to Eq. (6-73), the perturbation in electron content may be written as a summation,  $\sum_j N'_j \Delta s_j$ , where the  $\Delta s_j$  are increments along the path of interest and the  $N'_j$  are electron concentration perturbations at the center of each increment.

Eq. (6-71) must be solved once for each  $N'_j$  desired. The solution is carried out along a field line and the point at which that field line intersects the path along which the perturbed content is being calculated provides one  $N'_j$ . As the solution of the equation is obtained, a series of values for  $N'$  at some height along the path of interest is determined. Such a process is repeated along many field lines until enough  $N'_j$ 's are determined to use the summation. Assuming that  $n$  height steps are used and that  $m$  iterations of  $I'$  are desired, then  $nm$  iterations of Eq. (6-71) must be carried out. The process involved can be quite time consuming if small step sizes are used.

## Chapter VII

### MODELS

In order to carry out the numerical solution of Eq. (6-71) various models must be adopted. Models are needed for the atmospheric disturbance, the ionosphere, the neutral atmosphere, and the solar spectrum. Specifying a model for the atmospheric disturbance is a particularly difficult task. Most of the theoretical work published so far on the ionospheric effects of atmospheric disturbances has been based on the assumption that the observed ionospheric disturbances have a quasi-periodic character and that, therefore, use of a sinusoidal steady state analysis is a valid approach. Such a method implicitly assumes that the attenuation is sufficiently small so that use of a purely real horizontal wave number is a good approximation. As a result of the work presented in Part I, the validity of such assumptions becomes questionable for the case of large scale TIDs. It was shown in Chapter IV that for the case of such TIDs there is considerable attenuation of the atmospheric disturbance, a fact not previously demonstrated. The attenuation appears to be sufficiently high to prevent any substantial oscillation of the atmosphere. Therefore, the disturbances should tend to propagate as single pulses, as was indeed shown to be the case at midlatitudes (Chapter III). It was also shown that because of the regularly recurring source of the disturbances, there is often an appearance of periodicity in a train of TIDs. This may have been one of the reasons why previous workers have considered the sinusoidal steady state analysis approach as adequate when considering these disturbances. The question of the validity of such a technique should be carefully examined in a future investigation. Nevertheless, it should be pointed out that Hines [1968], using a sinusoidal atmospheric gravity wave model, has been able to obtain profiles of the perturbations in the neutral concentration which qualitatively agree with observed perturbations in electron concentration (determined by incoherent scatter means) which were due to a large scale TID. For the case of medium scale TIDs, Thome and Rao [1969] have obtained encouraging results with the coupling between a sinusoidal wave

and the ionization included. The results which will be presented here are primarily an extension of the work of Hines [1968], who investigated the influence of molecular transport dissipative processes on the shape of phase surfaces of atmospheric gravity waves. However, he did not consider the coupling between the atmospheric wave and the ionization which is included here. As previous workers [e.g. Hines, 1968; Hooke, 1968a; Thome and Rao, 1969] have done, we will use a monochromatic atmospheric wave having a purely real horizontal wave number, well aware that this arbitrary assumption and the arbitrary choice of a wave number are not really defensible for large scale TIDs. This is done in order to obtain a general feeling for a problem whose solution is beyond the scope of this work.

The models which are used for the atmospheric disturbance are based on the results of Hines [1960] and those of Pitteway and Hines [1963], as extended by Thome and Rao [1969], for internal atmospheric gravity waves. The first of these models [Hines, 1960] does not include the effects of viscosity or heat conduction; hence the wave amplitude grows with height in a manner which is unacceptable for disturbances extending to very high altitudes. Therefore, for this study results based on the second model are more reasonable, although calculations are done using both for the sake of comparison. The results of Pitteway and Hines [1963] which are used here are for the case of an isothermal atmosphere with constant kinematic viscosity. In order to include the vertical variation of temperature and, in particular, the exponential increase of kinematic viscosity with height, a number of isothermal layers are used; within each the kinematic viscosity is constant. The multiple layer approach used is similar to that of Thome and Rao [1969], and seems to be more reasonable than a single isothermal layer. However, a separate dispersion relation must be solved for each layer and the results obtained for each layer are therefore independent of the results for other layers, which is not a completely satisfactory situation.

The horizontal wavelength which will be used (3000 km) is roughly twice the "length" of a typical TID pulse. In other words, the pulse is treated as the equivalent of a half cycle. The period of the monochromatic wave used (90 minutes) is about twice the "width" of a typical pulse. Because of the rate at which substorms recur during disturbed

periods, very similar values would be obtained, by coincidence, if a train of large scale TIDs were (falsely) regarded as periodic.

In addition to the limitations already mentioned concerning the atmospheric waves used, several others also apply. For the very long wavelengths considered, the assumption of a plane earth, used in obtaining the equations describing the wave, approaches the limits of its validity. The effect of ion drag on the waves is also neglected. One serious difficulty relates to the use of the dispersion relation of Pitteway and Hines [1963]. Both viscous and thermal conduction effects operate together and are of the same order of magnitude. In the analysis used here the thermal conduction effects are included by an ad hoc increase in the coefficient of viscosity, although it was assumed that the process is adiabatic in obtaining the dispersion relation.

It is believed that the models chosen for the ionosphere, neutral atmosphere, and solar spectrum are the most reasonable available for a study of the sort done here. However, for an investigation of a particular event, use of measured parameters for the time of interest would result in a definite improvement.

The models which will be used do not correspond to geomagnetically disturbed conditions. As was shown in Chapter II, these disturbances generally occur during disturbed conditions. However, as was also shown (cf. Figure 2-10) many occur for relatively undisturbed conditions (average  $K_p$  less than 4). The disturbances occurring during highly disturbed periods have very large amplitudes and it is possible that the perturbation approach used in the preceding chapter is not applicable at such times. However, for relatively undisturbed conditions the perturbation analysis seems very reasonable and at such times the models presented in this section become useful. The numerical results which will be obtained may become less meaningful under highly disturbed conditions. The following sections describe in detail the models used.

#### A. Neutral Atmosphere

During the period of interest, 1967-1969, the average value of 10 cm solar flux was 147.8, in units of  $10^{-22} \text{ W m}^{-2} \text{ Hz}^{-1}$ . For 1967 the average value was 143.2, while for 1968 and 1969 it was 149.1 and 151.2

respectively. The model used for the neutral atmosphere is the COSPAR International Reference Atmosphere 1965 [CIRA, 1965] model number 5. This model is for moderate solar activity and corresponds to a 10 cm flux level of 150. Since for all three years of interest the 10 cm flux level was near 150 and varied little from year to year, this seems a reasonable choice.

The density and temperature profiles for CIRA model 5 are shown in Figure 7-1 for day (1200 LT) conditions. Figures 7-2a and 7-2b show profiles of several parameters which were calculated using these data,

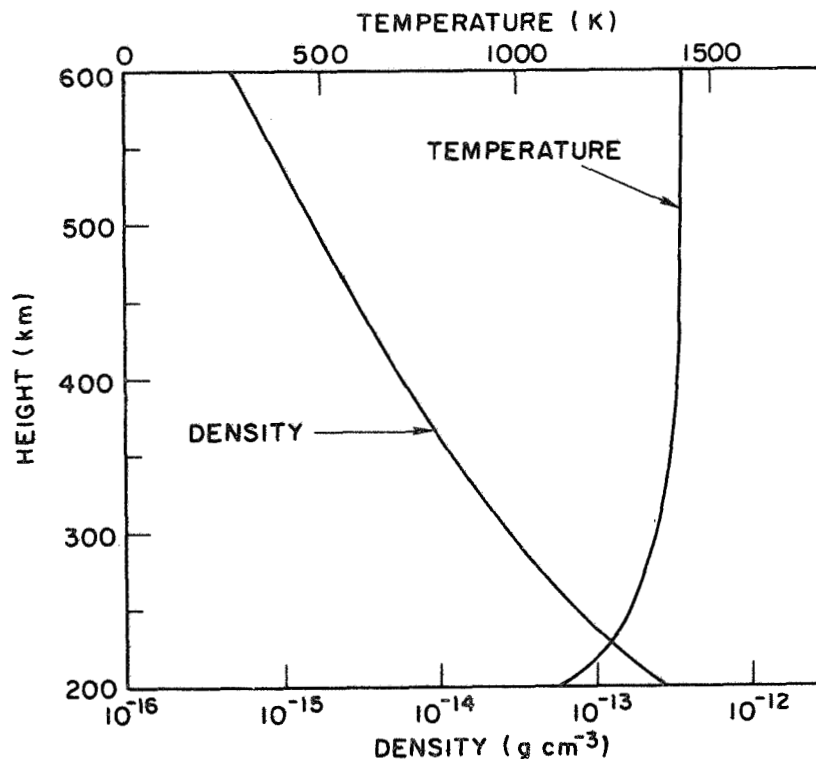


Fig. 7-1. TEMPERATURE AND DENSITY PROFILES FOR THE MODEL ATMOSPHERE USED.

namely; the neutral atmosphere scale height, the kinematic viscosity, the speed of sound and the Brunt period. A ratio of specific heats of 1.67 was used (monatomic gas). The value of the coefficient of viscosity ( $\mu$ ) used was  $5.4 \times 10^{-5} \text{ kg m}^{-1} \text{ s}^{-1}$ . The coefficient of viscosity will be assumed independent of height; the value given was calculated

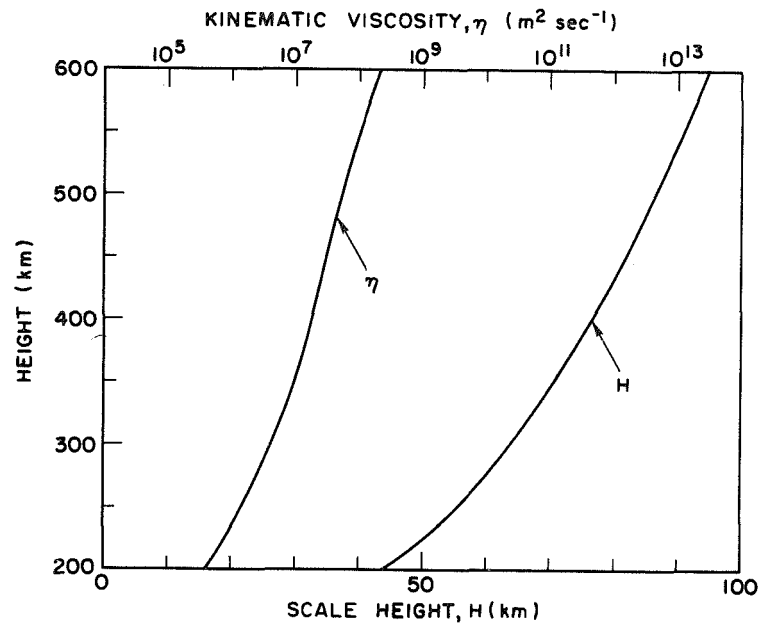


Fig. 7-2(a). PROFILES OF KINEMATIC VISCOSITY AND SCALE HEIGHT FOR THE MODEL ATMOSPHERE USED.

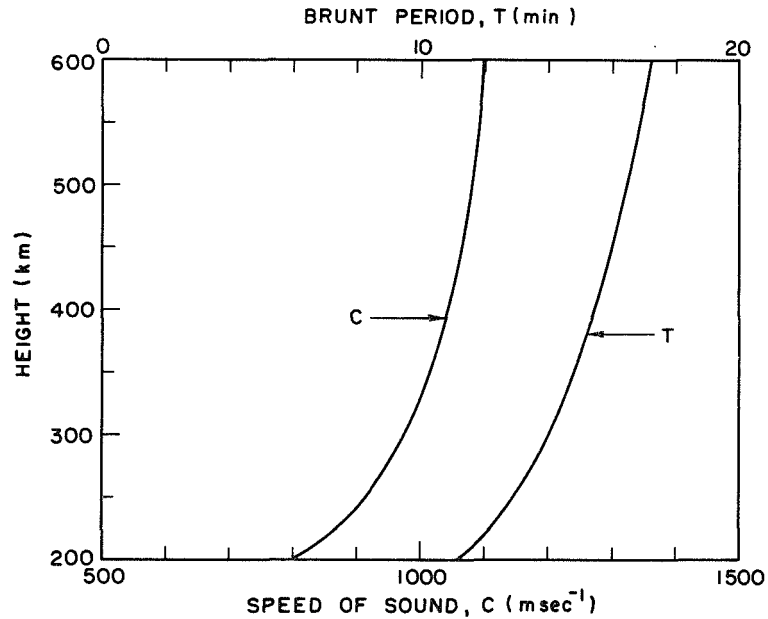


Fig. 7-2(b). PROFILES OF THE ISOTHERMAL BRUNT PERIOD AND THE SPEED OF SOUND FOR THE MODEL ATMOSPHERE USED.

using an expression given by Dalgarno and Smith [1962] for the viscosity of atomic oxygen as a function of temperature. The value was chosen to give the best fit to  $\mu$  over the height range of interest. The values of  $\mu$  determined from Dalgarno and Smith's expression differ by ten per cent or less over the 200-600 km range from the value given above.

Thermal conductivity damps atmospheric gravity waves in much the same manner as viscosity [Pitteway and Hines, 1963] but by a slightly larger factor (1.26 times as much). The coefficient of viscosity will be enhanced by this amount in order to include the effects of thermal conductivity, albeit in only an approximate fashion [cf. Hines, 1968]. The value of " $\mu$ " used to obtain the "kinematic viscosity" curve shown in Figure 7-2a was  $1.22 \times 10^{-4} \text{ kg m}^{-1} \text{ s}^{-1}$ .

#### B. Solar Spectrum and Production and Loss

The solar spectrum used is essentially that of Hinteregger, Hall and Schmidtke [1965] extrapolated to the period of interest by assuming it to be proportional to 10 cm flux measurements. Table 7-1 shows the wavelength bands used and the energy flux assumed for each band. Wavelengths below  $31 \text{ \AA}$  were neglected since they only affect the atmosphere below the level of interest. The cross-sections given in Table 7-1 are those of Hinteregger et al. [1965].

Only the photoionization of atomic oxygen is considered, although the absorption by molecular oxygen and nitrogen is included. The production of  $\text{O}^+$  ions is obtained using Eq. (6-47). At 200 km a rate of  $1300 \text{ cm}^{-3} \text{ s}^{-1}$  is obtained, while at 300 km the rate falls to 360. These values were determined for Stanford at local noon with the solar declination equal to zero.

The rate  $\gamma$  for the reaction in Eq. (6-58) is determined as follows: photochemical equilibrium is assumed to apply at 200 to 220 km, i.e.,  $q = \ell = \beta N$  is assumed. Using values of  $q$  obtained as discussed in the preceding paragraph and values of  $N$  from the model ionosphere described in the following section,  $\beta$  is determined at 200, 210, and 220 km. Using values of the concentration of molecular nitrogen from the model neutral atmosphere,  $\gamma$  is found at each of those heights and then averaged. This value of  $\gamma$  is then used to find values of  $\beta$  at all heights using the

TABLE 7-1

SOLAR SPECTRUM AND CROSS-SECTIONS OF CONSTITUENTS

| BAND<br>NO. | WAVELENGTH<br>RANGE (Å) | EFFECTIVE<br>WAVELENGTH | ENERGY FLUX<br>(erg cm <sup>-2</sup> s <sup>-1</sup> ) | ABSORPTION<br>CROSS-SECTION |                |      | IONIZATION<br>CROSS-SECTION |
|-------------|-------------------------|-------------------------|--------------------------------------------------------|-----------------------------|----------------|------|-----------------------------|
|             |                         |                         |                                                        | N <sub>2</sub>              | O <sub>2</sub> | O    |                             |
| 1           | 911-1027                | 980                     | 0.52                                                   | 11.8                        | 5.6            | 0.0  | 0.0                         |
| 2           | 796-911                 | 880                     | 0.60                                                   | 4.9                         | 14.2           | 3.0  | 3.0                         |
| 3           | 700-796                 | 780                     | 0.23                                                   | 24.6                        | 27.2           | 4.2  | 4.2                         |
| 4           | 600-700                 | 620                     | 0.25                                                   | 23.8                        | 34.6           | 11.5 | 11.5                        |
| 5           | 500-600                 | 550                     | 0.34                                                   | 24.0                        | 29.0           | 13.0 | 13.0                        |
| 6           | 400-500                 | 470                     | 0.27                                                   | 22.4                        | 25.4           | 12.4 | 12.4                        |
| 7           | 280-400                 | 320                     | 1.18                                                   | 6.7                         | 19.5           | 9.6  | 9.6                         |
| 8           | 205-280                 | 250                     | 0.54                                                   | 3.4                         | 12.6           | 6.3  | 6.3                         |
| 9           | 165-205                 | 180                     | 1.41                                                   | 2.1                         | 6.6            | 3.3  | 3.3                         |
| 10          | 138-165                 | 160                     | 0.17                                                   | 1.4                         | 4.2            | 2.1  | 2.1                         |
| 11          | 103-138                 | 120                     | 0.18                                                   | 0.9                         | 2.2            | 1.1  | 1.1                         |
| 12          | 62-103                  | 80                      | 0.51                                                   | 0.5                         | 1.1            | 0.6  | 0.6                         |
| 13          | 41-62                   | 51                      | 0.24                                                   | 0.2                         | 0.4            | 0.2  | 0.2                         |
| 14          | 31-41                   | 36                      | 0.15                                                   | 0.07                        | 0.2            | 0.1  | 0.1                         |

(Cross-sections in cm<sup>2</sup> x 10<sup>18</sup>)

assumed concentration of  $N_2$ . A value for  $\gamma$  of  $1.8 \times 10^{-12} \text{ cm}^3 \text{ s}^{-1}$  is obtained for Stanford at local noon. A value of  $\beta$  of  $6.8 \times 10^{-3} \text{ s}^{-1}$  is obtained at 200 km while a value of  $4.8 \times 10^{-4} \text{ s}^{-1}$  is found at 300 km.

### C. Ionosphere

The model ionosphere was obtained using a simulation program based on the coupled continuity and heat flow equations [Waldman, 1970]. The program makes use of observed values of electron content as one of the necessary boundary conditions. The electron concentration profile for 1200 LT is shown in Figure 7-3a. The average columnar electron content observed at Stanford for the period 1967-1969 is shown in Figure 5-1a. As can be seen, the value of content at 1200 LT is  $3.7 \times 10^{17}$  electrons  $\text{m}^{-2}$ . This value was used with the simulation program.

Figure 7-3b shows profiles of electron and ion temperature which were obtained using the simulation program and which are used to determine the plasma temperature.

### D. Atmospheric Disturbances

It has been suggested that large scale TIDs are caused by free internal atmospheric gravity waves whose properties are determined by the characteristics of the thermosphere. This explanation has been advanced by Georges [1968a] and is based primarily on the observed wave-front tilts of the TIDs. It has also been suggested that the TIDs are the result of surface waves [Tolstoy, 1967; Thome, 1968; Tolstoy and Pan, 1970].

In the numerical calculations two models for internal waves are used. The first model is that of Hines [1960], the second that of Pitteway and Hines [1963] which includes the effects of viscosity. In both cases a multislabs approach is used with all parameters assumed constant within each layer.

#### 1. Internal waves in an isothermal atmosphere

It will be assumed that the atmosphere is isothermal, of uniform composition, nonviscous and that the wave motions of interest have only

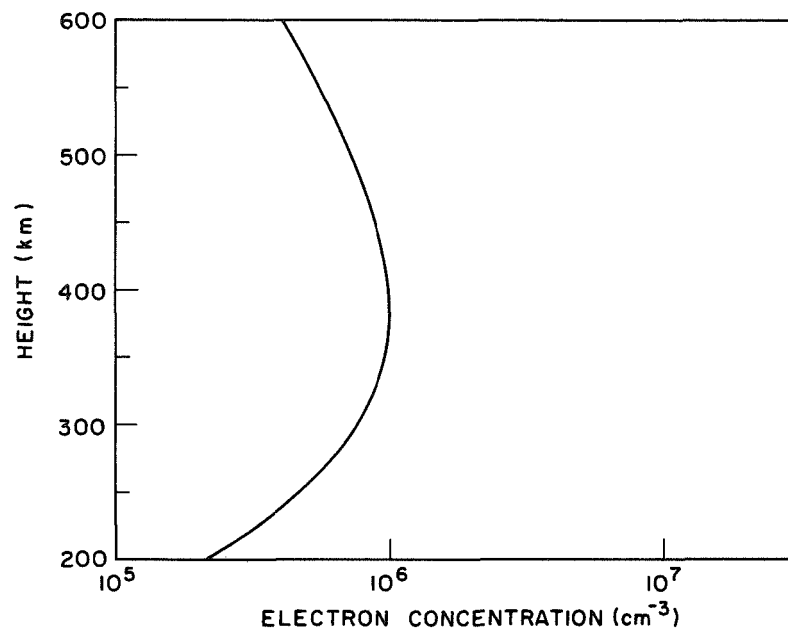


Fig. 7-3(a). PROFILE OF ELECTRON CONCENTRATION USED.

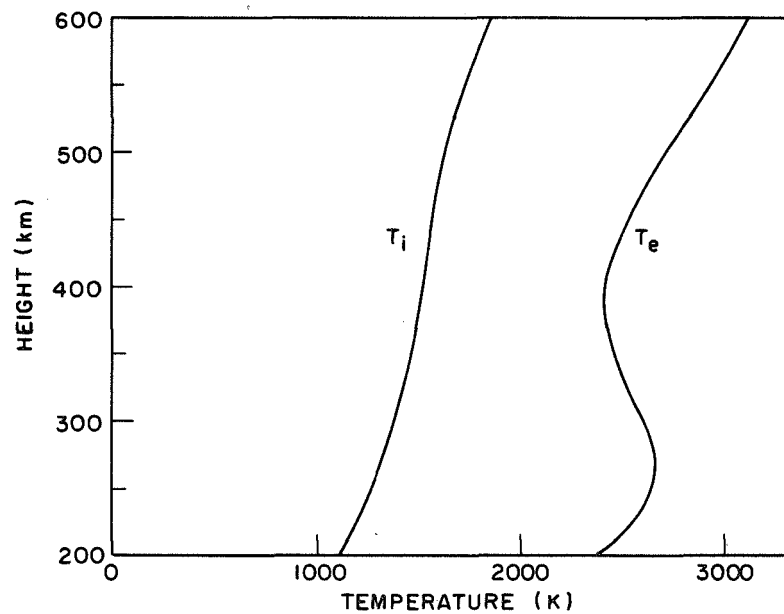


Fig. 7-3(b). PROFILES OF ELECTRON AND ION TEMPERATURE USED IN THE CALCULATIONS.

perturbation magnitude. The following equations then govern the motions of interest.

$$\frac{\partial(\rho_{no} \vec{U}_n)}{\partial t} = \rho_n \vec{g} - \nabla p_n \quad (7-1)$$

$$\frac{\partial p_n}{\partial t} + \vec{U}_n \cdot \nabla p_o = c^2 \left( \frac{\partial \rho_n}{\partial t} + \vec{U}_n \cdot \nabla \rho_{no} \right) \quad (7-2)$$

$$\frac{\partial \rho_n}{\partial t} + \nabla \cdot (\rho_{no} \vec{U}_n) = 0 \quad (7-3)$$

These are the linearized equations of motion, adiabatic state and continuity.  $c$  is the speed of sound, given by

$$c = [\gamma p_{no} / \rho_{no}]^{\frac{1}{2}} \quad (7-4)$$

$p$  is the pressure,  $\rho$  is the density and  $\gamma$  is the ratio of specific heats.

A solution of Equations (7-1) to (7-3) will be sought such that

$$p'_n / p_{no} P = \rho'_n / \rho_{no} R = U_{nx} / X = U_{nz} / Z = A \exp i(\omega t - K_x x - K_z z) \quad (7-5)$$

Using these solutions and the above equations, the dispersion relation below is found

$$\omega^4 - \omega^2 c^2 (K_x^2 + K_z^2) + (\gamma - 1) g^2 K_x^2 + i \gamma g \omega^2 K_z = 0 \quad (7-6)$$

Assuming  $K_x = k_x$  (purely real),  $K_z$  is purely imaginary or

$$K_z = k_z + i/2H \quad (7-7)$$

where  $H = c^2 / \gamma g$

The latter choice is appropriate for internal waves. It can then be shown that the following "polarization relations" relate the parameters of interest

$$P = \gamma \omega^2 [k_z - i(1 - \gamma/2) g/c^2] \quad (7-8)$$

$$R = \omega^2 k_z + i(\gamma - 1) g k_x^2 - i\gamma g \omega^2 / 2c^2 \quad (7-9)$$

$$X = \omega k_x c^2 [k_z - i(1 - \gamma/2) g/c^2] \quad (7-10)$$

$$Z = \omega [\omega^2 - k_x^2 c^2] \quad (7-11)$$

The preceding discussion has been based on the results of Hines [1960]. For details the interested reader is referred there or to Georges [1967].

By assuming values for  $\omega$  and  $k_x$ ,  $k_z$  can be found using the dispersion relation. Using the polarization relations and a model atmosphere, the quantities of interest,  $U_{nx}$ ,  $U_{nz}$ , and  $\rho'_n$  may be found in terms of one another.

## 2. Internal waves in an isothermal atmosphere with viscosity

In order to obtain solutions for the form of Eq. (7-5), the kinematic viscosity will be assumed constant. In that case the equation of motion becomes [Pitteway and Hines, 1963]

$$\frac{\partial(\rho_{no} \vec{U}_n)}{\partial t} = \rho_n \vec{g} - \nabla p_n + (\nabla \cdot \mu \nabla) \vec{U}_n + \nabla(\mu \nabla \cdot \vec{U}_n)/3 + (\nabla \mu \times \nabla) \times \vec{U}_n \quad (7-12)$$

where  $\mu$  is the coefficient of viscosity. The dispersion relation now becomes [Pitteway and Hines, 1963]

$$\omega(\omega - i\eta\varphi)\{\omega^2 - 4i\omega\eta/3 + c^2\}\varphi + k_x^2 (g + i\omega\eta/H)\{g(\gamma - 1) - 2i\omega\eta/3H\} = 0 \quad (7-13)$$

where  $\eta = \mu/\rho_{no}$  is the kinematic viscosity

and  $\varphi = k_x^2 + k_z^2 + 1/4H^2$

The polarization relations for this case have been obtained by Thome and Rao [1969]. Those relations are

$$P = \gamma \omega^2 K_z - i\omega^2/H + i\eta K_x^2 K_z \omega\gamma/3 + \eta K_x^2 \omega\gamma/H - \frac{4}{3} \eta K_x^2 \omega/H - \eta K_z^2 \omega/H - i\frac{4}{3} \eta K_x^2 K_z \omega\gamma - i\eta K_z^3 \omega\gamma \quad (7-14)$$

$$R = \omega^2 K_z + i(\gamma - 1) g K_x^2 - i\omega^2/H - i\eta K_x^2 K_z \omega + i\eta K_z \omega/H^2 - i\eta K_z^3 \omega + \frac{1}{3} \eta K_x^2 \omega/H - 2\eta K_z^2 \omega/H \quad (7-15)$$

$$X = \omega K_x K_z c^2 - i g \omega K_x + i\eta K_x K_z \omega^2/3 + \eta K_x \omega^2/H \quad (7-16)$$

$$Z = \omega^3 - \omega K_x^2 c^2 - \eta K_z \omega^2/H - i\frac{4}{3} \eta K_x^2 \omega^2 - i\eta K_z^2 \omega^2 \quad (7-17)$$

As for the nonviscous case, the dispersion relation and the polarization relations are used to determine the relationships between  $U_{nx}$ ,  $U_{nz}$ , and  $\rho_n'$ .

### 3. Multislab approach

Over the height range of interest the parameters in the dispersion equation and polarization relations vary a great deal. Therefore, a number of slabs will be used; within each slab the temperature is taken as constant. Each slab is 10 km thick. The dispersion equation and polarization relations are applied in each layer to determine  $U_{nx}$ ,  $U_{nz}$ , and  $\rho_n'$ . Such an approach is particularly important in dealing with the viscous case, since  $\eta$  was assumed independent of height, whereas it actually increases exponentially.

## CHAPTER VIII

### RESULTS

This chapter is devoted to a study of results obtained using the theory developed in Chapter VI along with the models of Chapter VII. Except for the limited height range used, the results of Chapter VI are quite general. The limitations of the models of Chapter VII have already been discussed and should be recognized when considering the results presented here. The wave parameters which will be used in this chapter, unless otherwise stated, are as follows: horizontal wavelength, 3000 km, period, 90 minutes, wave amplitude (horizontal velocity of the neutral gas),  $20 \text{ m s}^{-1}$  at 200 km. The direction of travel of the wave is assumed to be from dipole north to south. These parameters will be referred to as the "standard" parameters. They imply a horizontal speed of travel of  $556 \text{ m s}^{-1}$ . In addition, all calculations are for local noon at Stanford (geographic coordinates:  $37.4^\circ \text{ N}$ ,  $122.2^\circ \text{ W}$ ; dipole coordinates:  $43.8^\circ \text{ N}$ ,  $299.6^\circ \text{ E}$ ), unless otherwise stated. Figure 8-1 shows pictorially the relationship between the wave vector and the dipole

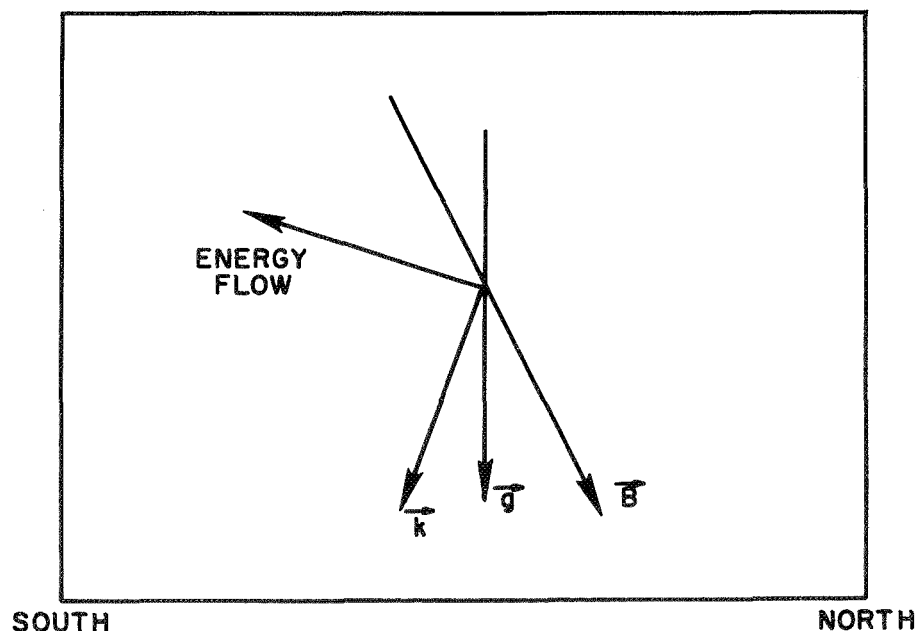


Fig. 8-1. PICTORIAL VIEW OF QUANTITIES ASSOCIATED WITH THE WAVE.

field for the waves of interest. The wave vector,  $\vec{k}$ , is nearly vertical, while the energy flow (group velocity) is roughly horizontal.

The present chapter will present several examples which show how the theoretical effect of an internal atmospheric gravity wave on the local ionization depends upon the complexity of the coupling model used. This will be followed by a discussion of some additional approximations which were used in obtaining these examples. Next the dependence on height of the coupling between the gravity wave and the ionization will be discussed in order to explain the examples which are presented. Finally, attention will be given to the effect of a gravity wave on columnar electron content.

#### A. Examples

Figure 8-2 shows results obtained using the theory developed in Chapter VI and the models given in the previous chapter. Contours of constant per cent fluctuations in electron concentration are plotted for three cases, all using the standard wave parameters. In obtaining the plots in Figure 8-2a and 8-2b, Eq. (6-71) was used in the following form

$$\frac{\partial N'}{\partial t} = - \frac{\partial}{\partial b} (N_o U_{nb}) \quad (8-1)$$

which will be called the collision geomagnetic equilibrium (C.G.E.) case.  $\mu = 0$  was also used in Figure 8-2a so the results correspond to a wave in an atmosphere with no viscosity. Two cycles of the disturbance are shown, with the negative portions of the cycles omitted in order to improve clarity. Because the horizontal wavelength of the wave used is 3000 km, the contours are nearly horizontal. The tilt of the wavefront in the ionization is about  $8^\circ$  from the horizontal. When no viscosity is considered in the solution, the amplitude of the neutral gas velocity grows exponentially with height and "perturbations" of 100 percent are reached at 450 km.

Because of the fact that in the simple case, when  $\mu = 0$ , the wave amplitude grows exponentially with height, such a model is unsuitable for a study of disturbances of large vertical extent, although it may

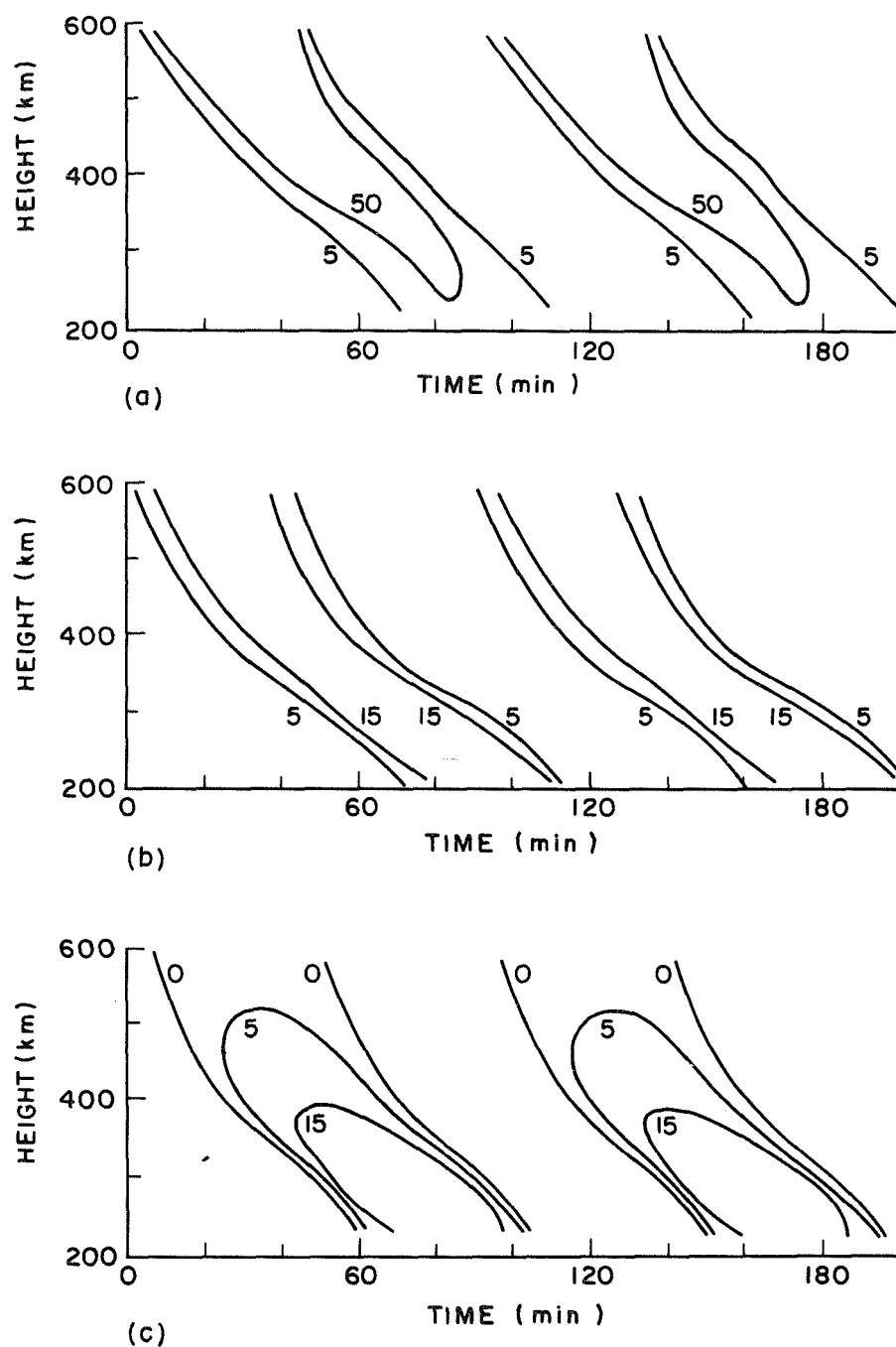


Fig. 8-2. CONTOURS OF CONSTANT PER CENT FLUCTUATIONS IN THE ELECTRON CONCENTRATION PRODUCED BY AN INTERNAL ATMOSPHERIC GRAVITY WAVE. The curves are discussed in the text.

provide useful insights when considering disturbances restricted to the region below the F-layer peak. Figure 8-2b show contours obtained for the C.G.E. case when viscosity is added. The addition of viscosity is the only difference between Figures 8-2a and 8-2b. The addition of viscosity results in an upward turning of the contours at high levels. This has been discussed by Hines [1968] and has also been shown by Thome and Rao [1969].

Note that there is a small inflection of the contours in Figure 8-2b near 300 km. The slight bending downward of the curves below that level is due to the variability of the coupling mechanism with height. Close inspection of the contours in Figure 8-2a reveals a slight upward bending for the case when  $\mu = 0$ . This bending is due to the fact that there is some decrease of the vertical wave number with height in the multi-slab model used even when viscosity is not considered.

Figure 8-2c presents the results for the case in which the complete Eq. (6-71) is used. Only the perturbation in production due to the perturbed optical depth term was omitted. This omission will be justified in the next section. Note that due to the cancellation of the perturbed production and loss effects mentioned following Eq. (6-65), some "perturbation" in loss and production is always included even in the simplest cases. The contours obtained using the complete model (i.e., all terms in Eq. (6-71) included) approach vertical somewhat more than the others. The changes at low levels result from including the effect of losses, while the inclusion of the perturbation in diffusion depresses the concentration perturbations at higher levels and thus causes the closing of the contours.

#### B. Additional Approximations

The effect of omitting the perturbation in the production caused by the variation of the ionizing flux reaching the point of interest is negligible for the disturbances studied here at levels above 200 km. Hooke [1968a] has shown that the effect on the flux is a maximum when the real part of the wave vector is perpendicular to the sun's rays. Since the wave vector for the disturbances of interest is nearly vertical, this would require the sun to be near the horizon. At such times the

production effects should be negligible due to the large zenithal angle and the resulting very low solar flux. At local noon, the time for which the calculations in this section were carried out, it has been verified numerically that the effect is negligible. The model used in this verification included viscous effects and production and loss (no diffusion). A wave of 3000 km horizontal wavelength, 90 minute period,  $30 \text{ m s}^{-1}$  amplitude (at 200 km) and a direction of travel from north to south was used. At the altitude near 200 km where production effects should be a maximum (within the range of interest), perturbations due to diffusion are negligible and inclusion of these effects does not affect the radiation reaching lower levels. Using a height range of 200-370 km, the perturbation in electron concentration was found to change by 16% at 210 km when production was removed, but by only 4% at 230 km. The integrated effect was a less than 1% change when production was removed.

For the model used, the electron content below 200 km is very small, about  $1.5 \times 10^{16}$  electrons  $\text{m}^{-2}$ , whereas above that level it is about  $3.5 \times 10^{17}$  electrons  $\text{m}^{-2}$ . It is expected that the amplitude of electron concentration perturbations below 200 km is severely depressed by losses, but even when the effect of losses is neglected, the integrated perturbation below 200 km is less than 10% of the total content perturbation below 600 km under very extreme conditions, i.e., when perturbation amplitude does not decrease below 200 km. Therefore, it is felt that levels below 200 km may be reasonably neglected. There are important practical considerations involved in making this assumption. The time step size used in solving Eq. (6-71) must be smaller than the time constants of the loss process, otherwise the solution will become unstable. Since for the model used  $1/\beta$  falls below 60 seconds near 180 km, very small step sizes are required to include levels below 200 km. Typically, at these levels lowering the lower boundary by 20 km requires doubling the computation time. In obtaining the results of this chapter, step sizes of 7 km and 12 minutes were used in the C.G.E. cases and steps of 14 km and 6 minutes were used in the others.

#### C. Height Variation in the Coupling between the Atmospheric Wave and the Ionization

The relationship between the contours of constant percent fluctuation

in the electron concentration and the phase surfaces of the atmospheric gravity wave depends upon the height and the wave parameters. The discussion which follows will be restricted to large scale disturbances, although it could be readily extended to other types of TIDs.

When considering the coupling between the wave and the ionization, it is useful to consider Eq. (6-71) with various terms neglected, depending upon the height of interest. At low levels, well below the layer peak, diffusion is negligible and Eq. (6-71) may be rewritten as

$$\frac{\partial N'}{\partial t} = -\beta_o N' - \frac{\partial}{\partial b} (N_o U_{nb}) \quad (8-2)$$

In the sinusoidal steady state case (time varying quantities having the form  $e^{i\omega t}$ ), Eq. (8-2) may be solved for  $N'$ .

$$N' = -\frac{1}{\beta_o + i\omega} \frac{\partial}{\partial b} (N_o U_{nb}) \quad (8-3)$$

The partial derivative of  $N_o U_{nb}$  with respect to  $b$  is the wave-induced flux divergence and is the principal driving term in Eq. (6-71). From Eq. (8-3), it can be seen that the phase relationship between  $N'$  and the divergence depends upon the relative sizes of  $\beta_o$  and  $\omega$ . At low levels  $\beta \gg \omega$ , and  $N'$  and the divergence are in anti-phase. At such levels a quasi-static equilibrium exists in which the loss and divergence terms nearly balance each other. The phase relationship between  $U_{nb}$  and the divergence term depends upon the relative importance of the rate of growth of  $N_o$  and  $U_{nb}$  with height and the rate of change of phase of  $U_{nb}$  with height. When the rate of growth of  $N_o$  or  $U_{nb}$  is much larger than the rate of change of phase, the velocity and the divergence will be in phase. The phase relationships determined using the full coupling model are shown in Figure 8-3a for 228 km.  $N'$  determined using the C.G.E. model leads the divergence term (labeled DIV in Figure 8-3) by 90 degrees and is larger than the  $N'$  shown since inclusion of the loss damps the electron concentration perturbation.

Figure 8-3b shows the phase relationships at 354 km. At this level the importance of the perturbation in loss is relatively small.

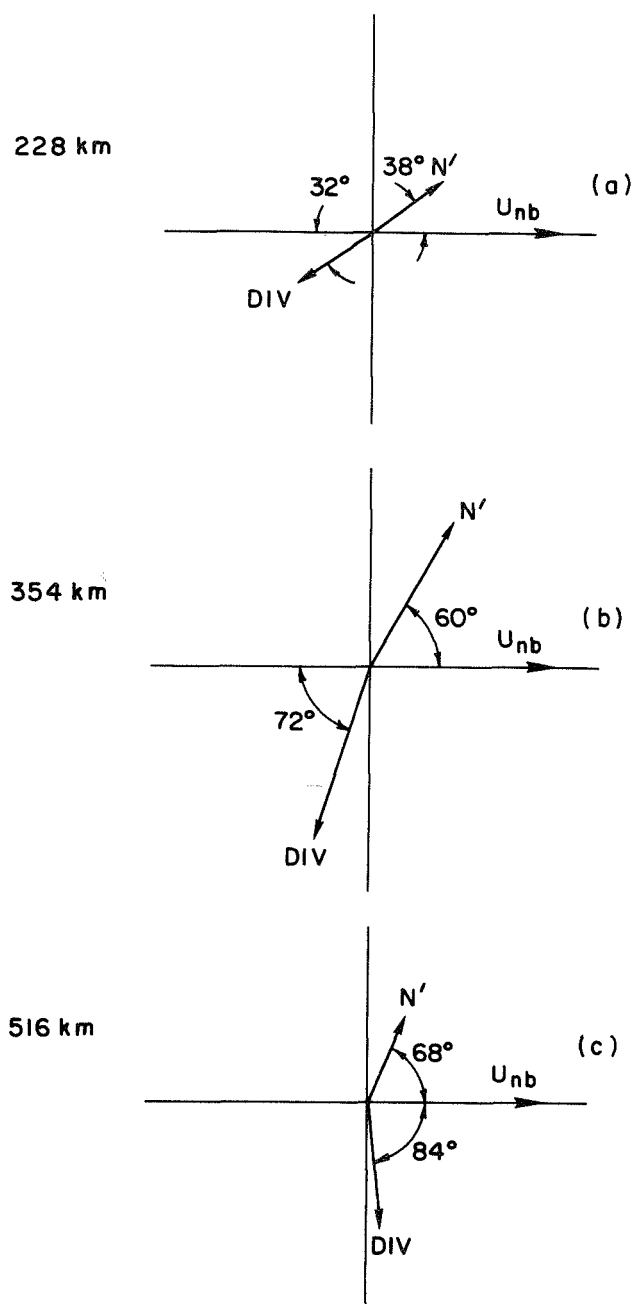


Fig. 8-3. PHASE RELATIONSHIPS BETWEEN THE NEUTRAL GAS VELOCITY ALONG THE FIELD LINE,  $U_{nb}$ , THE ELECTRON CONCENTRATION PERTURBATION,  $N'$ , AND THE WAVE INDUCED DIVERGENCE,  $DIV$ . Results are given for three heights.

Using the C.G.E. model,  $U_{nb}$  and  $N'$  would be nearly in phase.

At higher levels the importance of the perturbation in diffusion becomes very important. Figure 8-3c shows the phase relationships at 516 km. Again the phase relationship between the divergence term and  $U_{nb}$  depends upon the relative importance of the rate of change of phase and amplitude with height. Figure 8-3c indicates that the rate of change of phase is somewhat the more important at 516 km.

At high levels, if the perturbation in plasma diffusion becomes large enough, diffusion will tend to remove any perturbations in the ionization, and a quasi-static situation will exist. This is approximately true at 516 km for the model used. The precise phase between the driving term and  $N'$  depends upon the amplitudes and phases of the terms with coefficients A, B, and C in Eq. (6-71).

The bending upwards of the contours in Figure 8-2c is principally due to the effect of molecular dissipation (viscosity and heat conduction) on the gravity wave as was suggested by Hines [1968]. The inclusion of the coupling mechanism between the ionization and the wave and, in particular, consideration of the effects of loss and diffusion also affect the shape of the profiles. The bending of the contours in the ionization toward the vertical has been reported experimentally [Thome, 1964; Georges, 1968a].

Figure 8-4 shows schematically the relationship between the contour

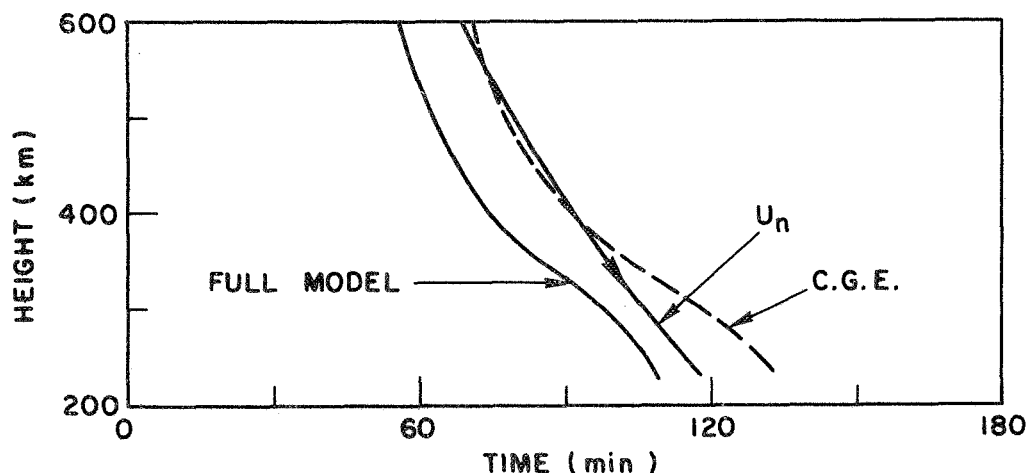


Fig. 8-4. RELATIONSHIP BETWEEN PHASE SURFACES IN THE ATMOSPHERIC WAVE AND CONTOURS IN THE PERTURBED IONIZATION.

of maximum neutral gas velocity and the contours of maximum wave-induced fluctuations in the electron concentration for the two models used. The figure shows that interpretation of contours in the ionization as phase surfaces may be somewhat misleading.

#### D. Integrated Effects

As might be expected from an inspection of Figure 8-2, the theoretical effect of gravity waves on electron content depends upon the model used for the interaction of the wave with the ionization. Table 8-1 shows some results for Stanford and for Edmonton, Alberta (geographic coordinates: 53.4° N, 113.5° W; dipole coordinates: 60.9° N, 302.7° E). The standard wave model was used for all calculations. For a path of integration vertically through the ionosphere the maximum perturbation in content changes only slightly when the full model is used instead of the C.G.E. model. The fluctuations in the electron concentration are decreased by including loss and diffusion, but the change in the shape of the contours compensates for this decrease.

Calculations are also shown for slant paths through the ionosphere. The paths used are along the raypaths from the stations to the geostationary satellite ATS-1, whose sub-satellite point is 150° west longitude. Most of the electron content data which are used in this study were obtained along these paths. The elevation angle at Edmonton when viewing ATS-1 is considerably lower than at Stanford. The same models were used at both locations.

Since the relationship between the wave amplitude,  $U_{nb}$ , and the concentration fluctuations are related by a linear equation, one may write the following expression

$$N' = N_o m U_n \quad (8-4)$$

where  $m$  is a constant of proportionality that can be determined numerically using the models which have been developed.  $N'$  and  $U_n$  are amplitudes. A similar expression may be written for the perturbation in electron content.

$$I' = I_o m_h U_{nh} \quad (8-5)$$

TABLE 8-1

THEORETICAL ELECTRON CONTENT PERTURBATIONS  
CAUSED BY INTERNAL ATMOSPHERIC GRAVITY WAVES

|                                                        |                            | <u>Stanford</u><br>(electrons m <sup>-2</sup> )<br>x 10 <sup>-17</sup> | <u>Edmonton</u><br>(electrons m <sup>-2</sup> )<br>x 10 <sup>-17</sup> |
|--------------------------------------------------------|----------------------------|------------------------------------------------------------------------|------------------------------------------------------------------------|
| $\int_{230}^{590} N_o dz$ (vertical)                   |                            | 2.61                                                                   | 2.61                                                                   |
| $\int_{230}^{590} N' dz$                               | Full<br>Model              | 0.29                                                                   | 0.17                                                                   |
|                                                        | C.G.E.<br>( $\mu \neq 0$ ) | 0.26                                                                   | 0.19                                                                   |
| $\int_{230}^{590} N_o ds$ (slant - to ATS-1 at 150° W) |                            | 3.90                                                                   | 5.50                                                                   |
| $\int_{230}^{590} N' ds$                               | Full<br>Model              | 0.37                                                                   | 0.32                                                                   |
|                                                        | C.G.E.<br>( $\mu \neq 0$ ) | 0.29                                                                   | 0.22                                                                   |

where  $U_{nh}$  is the wave amplitude at some height  $h$  and  $m_h$  is an average value of  $m$  which is weighted by the profiles of electron concentration and wave velocity and depends upon the height  $h$  chosen. Again,  $m_h$  can be determined numerically using the models developed. The amplitude of a disturbance causing a TID can be estimated using Eq. (8-5) along with models for the coupling process and wave amplitude and concentration profiles, when the electron content and the perturbation in content are known.

The relative amplitude of the perturbation in electron content will be defined as follows:

$$R = I' / I_0 \quad (8-6)$$

$$= m_h U_{nh}$$

Therefore,  $R$  is a measure of the amplitude of the wave and is more meaningful than  $I'$ , since the value of  $I'$  is proportional to  $I_0$ .

The most important information to be obtained from Table 8-1 is a comparison of  $m$  at the two stations for the slant path. For the C.G.E. case  $m_{\text{Edmonton}} / m_{\text{Stanford}} = 0.54$ . For the full model the ratio becomes 0.62. Therefore, for a wave with the given parameters, larger perturbations in electron content are expected at Stanford than at Edmonton.

Figure 8-5 shows the relative sensitivity of electron content measurements along the Stanford to ATS-1 path to the effects of atmospheric waves of the standard type as a function of the direction of the wave normal.  $180^\circ$  in the plot means propagation from dipole north to south. The symmetry with respect to the north-south direction is obvious. Waves coming from within  $\pm 40^\circ$  of dipole north are the most readily detected. The sensitivity is down by nearly one half for waves coming from the south.

The large anisotropy in Figure 8-5 is caused by the variability in coupling between the atmospheric wave and the ionization as the wave normal direction varies. The component of the neutral gas velocity along the field lines is the most important factor in determining the amplitude of the fluctuations in the electron concentration. When the wave is traveling from east to west or vice versa, the only velocity component along the field lines is due to the vertical component of the

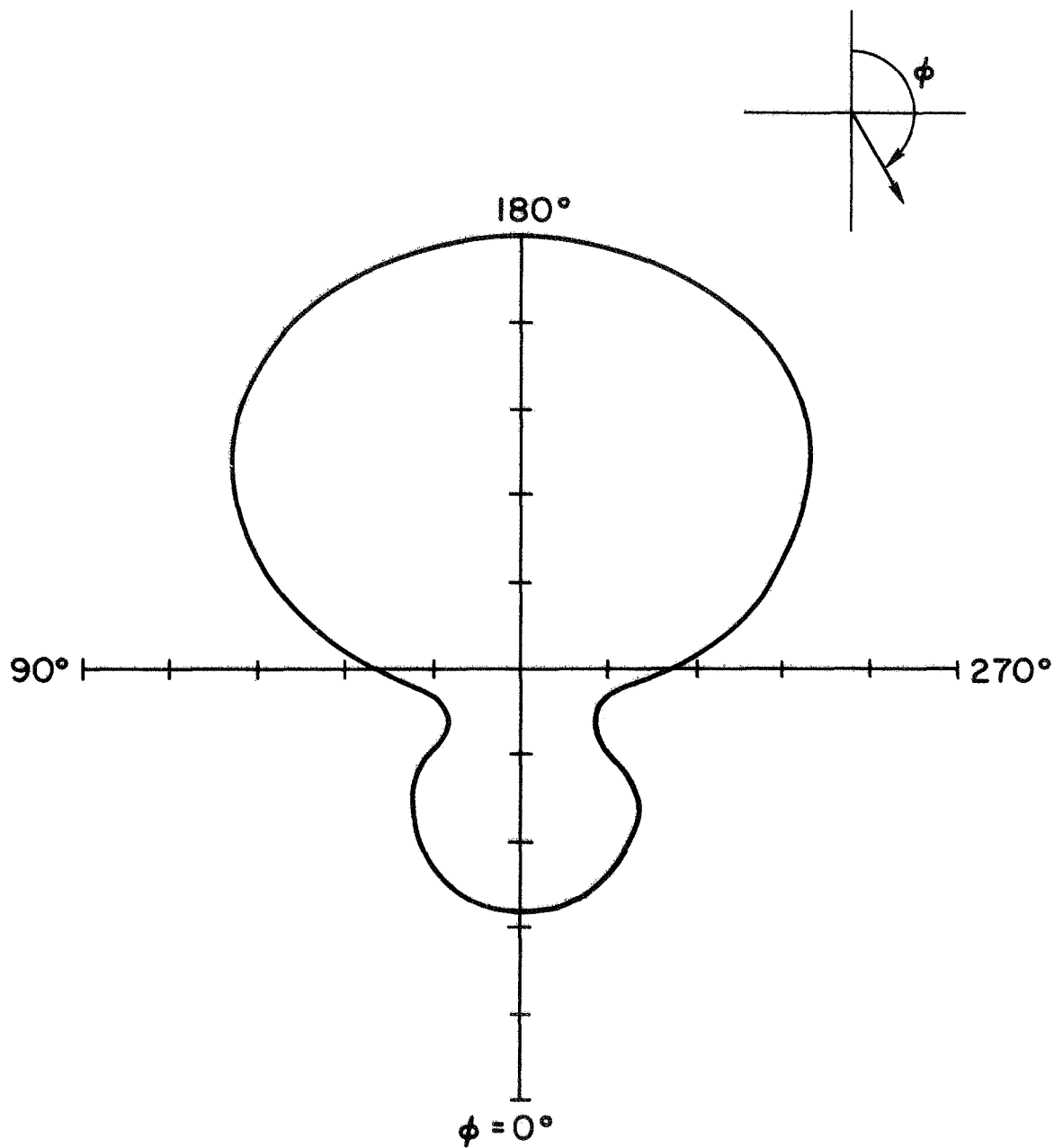


Fig. 8-5. VARIATION IN SENSITIVITY OF MEASUREMENTS TO DIRECTION OF TRAVEL OF THE ATMOSPHERIC WAVE. Calculations are for the Stanford to ATS-1 raypath.

neutral gas velocity. Since this is much smaller than the horizontal component, the concentration fluctuations are much less than for north-to-south propagation. These smaller fluctuations in electron concentration result in a smaller perturbation in the electron content. The perturbation produced by a wave traveling poleward is smaller than for one going toward the equator because the latter has its neutral gas velocity more in the direction of the field (cf. Figure 8-1). The calculations done to obtain the results in Figure 8-5 used the full model for coupling between the wave and the ionization. The integration was over the height range 230 to 590 km. The sensitivity pattern obtained depends upon the parameters of the wave, the path of integration and the observer location. More detailed studies of this sort which were based on simpler models, but which considered variation in other parameters besides direction of propagation, have been carried out by Hooke [1968b] and Georges and Hooke [1970].

#### E. Discussion

The importance of diffusion in the results presented undoubtedly depends to a fairly large extent on the models used for the ionosphere and neutral atmosphere and the results obtained may not apply well to any particular day. However, they do indicate the importance of diffusion at high levels in calculations of the sort done here.

Interpretation of contours observed in the ionization in terms of phase surfaces of the atmospheric waves may be somewhat misleading. Workers who study these disturbances using radio techniques should take care when evaluating their results directly in terms of atmospheric waves. Other problems also affect the interpretation of atmospheric wave-induced fluctuations in electron content, as has been discussed by Georges and Hooke [1970]. These problems are related to the integration through the electron concentration fluctuations and the loss of information in this fashion. However, the difficulties they discuss do not have such a large influence for the type of disturbance studied here. As was shown in Chapter II, LS TIDs travel within about  $\pm 20^\circ$  of dipole north to south at midlatitudes. Figure 8-5 shows that the sensitivity to disturbances drops to one-half for travel at  $\pm 75^\circ$  from dipole

north to south. Therefore, disturbances coming from well outside the observed range would be detectable if they were present. In general the problems besetting the use of electron content measurements in studying smaller scale disturbances are of much less importance when dealing with disturbances of the scale discussed here.

## CHAPTER IX

### PRINCIPAL CONCLUSIONS

The principal purpose of this study has been to relate the observed characteristics of a type of TID to a pattern of phenomena occurring at high latitudes. In particular it has been shown that these TIDs originate in the evening sector of the auroral oval and that their source is related to the occurrence of polar substorms. In addition, numerous details concerning the morphology of the disturbances have been presented, including information about their velocities, amplitudes and times of occurrence. The rate at which the atmospheric disturbances which cause these TIDs are damped has been studied and a possible source related to the occurrence of westward traveling surges has been suggested.

Throughout the preceding chapters the term "large scale TID" has been used to describe the disturbances of interest. The name is suggested by the large spatial scales found in the disturbances and has been used, sometimes with minor variations, by numerous workers. Since the relationship between these disturbances and polar substorms has been firmly established, it now seems reasonable to adopt a different name, one related to the source of the disturbances. The term "polar substorm related TID" has been used by Georges and Hooke [1970] and seems to be most appropriate.

In addition to considering the morphology of substorm related TIDs, this work has also investigated the problem of the effect of atmospheric disturbances on local ionization. The results have included a general treatment of the perturbation in photoionization, an inclusion of the effects of a perturbation in the plasma diffusion in the numerical results and an illustration of the importance of doing this. The models used in performing the calculations related to the coupling between the disturbance and the ionization suffer from a number of serious limitations. However, it is felt that the approach used allows some useful results to be obtained, particularly when considering the region above the F-layer peak.

There are numerous aspects of the subject investigated here which are worthy of additional study. In particular, there is a need for

additional and better data of the type already used. Other than such an effort, which would benefit all areas of the subject, the most important single area for future work is related to the subject of the source. The following work should be carried out:

1. A study of the source radiation pattern such as was discussed in Chapter V.
2. A comparison of substorm related TIDs with other phenomena, in particular auroral motions and possibly auroral infrasonic waves.
3. Any positive results obtained using the first two suggestions would allow more complete models for the physics of the source to be developed. In any such study the theoretical radiation pattern should be determined for comparison with the results from (1).
4. The theory which has been developed implies that disturbances should be produced at conjugate points during substorms. This should be verified for a number of individual events.
5. A more realistic model for atmospheric disturbances should be used in Part II. However, overcoming all the difficulties discussed there will be very difficult.
6. A more complete theory for damping should be developed which is not limited to the small damping case. Such a model plus more data would allow a more accurate determination of the rate of damping of the atmospheric disturbances.
7. Obtaining more data would permit further study of the location of the source of individual TIDs in order to investigate what parameters, if any, affect its precise location.
8. If the theory of these disturbances were extended sufficiently, it might be possible to use them as naturally occurring probes of the upper atmosphere during geomagnetically disturbed periods.

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