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MEASUREMENT OF THE NATURAL HYDROMAGNETIC WAVE SPECTRUM
AND INVESTIGATION OF POSSIBLE IONOSPHERIC
HEATING BY HYDROMAGNETIC WAVES

by

WILLIAM RICHARD SORENSON

A THESIS SUBMITTED
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INTRODUCTION

The object of this thesis is to describe and evaluate the results of a ground based experimental program designed to determine if hydromagnetic (hm) waves contribute significantly to heating of the earth's upper atmosphere observed during a magnetic storm.

A. The Purpose of the Experiment

The purpose of the experimental program described in this thesis was to determine if the energy dissipated in the ionosphere by hm waves during magnetic storms could account for the heating that is observed. In order to compute the power dissipated by hm waves, it is necessary to know the amplitude and frequency of the hm waves present in the ionosphere during a magnetic storm. This experiment provides the data necessary to compute the wave spectrum incident on the ionosphere.

The theoretical basis for this experiment was provided by Francis and Karplus (1960) and Karplus et al. (1962). They showed that vertically incident hm waves would produce magnetic variations at the earth's surface. Hydromagnetic waves, of course, do not propagate directly to the ground; the neutral atmosphere prevents this. The hm wave transforms into an electromagnetic wave as it leaves the ionized regions.
of the atmosphere at approximately 80 km altitude. Since the wave length is much larger than 80 km, the magnetic component of the wave has essentially the same amplitude at the ground as it does at 80 km. The purpose of this experiment was to measure the magnetic component of the wave at the earth's surface during a magnetic storm. Knowing the hm wave amplitude at the ground, and hence at 80 km, it is possible to compute the amplitude the wave must have had above the dissipating region of the ionosphere. When the incident spectrum is known, the power dissipation can be determined and compared with other sources of atmospheric heating. A definite statement can then be made on ionospheric heating by hm waves.

Francis and Karplus (1960) and Karplus et al. (1962) considered only vertically incident waves. Non-vertical incidence must also occur. There has been little theoretical work done, however, on non-vertical incidence. The problem is much more difficult than normal incidence. A preliminary theoretical study by Prince et al. (1964) indicates that the attenuation properties of the ionosphere are similar for vertical and non-vertical incidence.

There is another uncertainty other than non-vertical incidence inherent in this method of determining the hm wave spectrum in the ionosphere. Incident waves with amplitudes less than a certain value can not be detected with a magnetometer at the earth's surface. It will be
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demonstrated in later sections of this thesis that the sensitivity of the magnetometer is such that these undetected waves are not important with regard to ionospheric heating.

B. Density and Temperature Variations During Magnetic Storms

Atmospheric drag analysis on artificial earth satellites has revealed that the density and temperature of the upper atmosphere increase during a magnetic storm (Jacchia, 1959). A brief outline of the methods used in drag analysis to determine the density and temperature of the upper atmosphere is given, followed by a review of the pertinent experimental results.

The method outlined here is used by Jacchia and Slowey (1963). More detailed treatments are given by King-Hele (1961 a,b), Brouwer and Hori (1961), and Sehnal and Mills (1966). Atmospheric drag on an artificial satellite causes a secular decrease in the orbital period. This change in period can be determined for satellites with perigee heights up to 1000 km. From the change in period, the atmospheric density near perigee can be computed.

The optimum criteria for selecting a satellite to be used in drag analysis are spherical shape, long lifetime, an orbit that gives well-distributed observations, and a fairly high orbital eccentricity: very few satellites meet
these criteria.

Perturbing forces on an artificial satellite are generated not only by atmospheric drag, but also by solar radiation pressure, the attraction of nearby celestial objects, and by the irregular shape of the earth. The gravitational perturbations do not produce appreciable changes in the semimajor axis and, therefore, in the anomalous period (Brouwer, 1959; Kozai, 1959a). (The anomalous period is the period of revolution of the satellite about the earth with respect to its line of apsides.) For time intervals smaller than one orbital period, only atmospheric drag and solar radiation pressure have an effect. Kozai (1961) derived a method to compute the anomalous acceleration by solar radiation pressure.

Accurate observations of a satellite by the satellite tracking network are used to obtain orbital elements for the satellite. The effects of gravitational perturbations and solar radiation pressure are removed. Then \( \frac{dp}{d\tau} \) is obtained, where \( p \) is the anomalous period and \( \tau \) is time.

Sterne (1959) derived an expression for the rate of change \( \frac{dp}{d\tau} \) of the anomalous period of an artificial satellite due to drag in a rotating atmosphere. This expression can be numerically integrated to yield the density at perigee.

Nicolet (1960) demonstrated that an atmospheric model
using densities derived from satellite drag analysis gives practically a constant temperature for altitudes greater than 300 km, as predicted by Johnson (1956). Jacchia (1961 a) showed that with a suitable atmospheric model, it is possible to convert the observed densities into temperatures. Nicolet's (1961) atmospheric model was used until Jacchia's (1964) more complete model was published. Jacchia (1959) pointed out that the upper atmospheric density increases during a magnetic storm. Groves (1961) found a correlation between upper atmospheric density and geomagnetic activity indicated by Kp. Jacchia (1961 a) computed the temperature of the upper atmosphere using densities from satellite drag. He found that the temperature of the upper atmosphere increases during a magnetic storm. Jacchia (1961 b,c) and Zadunaisky et al. (1961) reported that during the geomagnetic activity of October, 1960 and November, 1960, the geomagnetic index ap followed the atmospheric drag curves rather well. It was also noted that the maximum of atmospheric perturbation occurs at nearly the same time all over the globe (Jacchia, 1961 b). Figure 1, taken from a paper by Jacchia (1964 a), illustrates the changes in density and temperature accompanying changes in ap. In Figure 2, the temperature at 730 km is plotted versus time for an individual magnetic storm. The corresponding values of ap are also shown. An important point to notice is that
the maximum of temperature lags the maximum ap by several hours. This lag is usually observed in storm induced ionospheric temperature increases. King-Hele (1963) computed densities from atmospheric drag on artificial satellites and obtained results essentially the same as Jacchia's.

Using satellite drag data from Injun III, Jacchia and Slowey (1963) concluded that on magnetically quiet days there is essentially no difference in the temperature of the upper atmosphere between the auroral zone and the equator. During a magnetic storm, however, the temperature increase appears to be much higher near the auroral zone. Jacchia (1964 a) reported that for low and moderate latitudes, the temperature increase corresponding to a unit increase in the 3-hourly ap index is 1.2 °K. For high latitudes, the temperature increase could be a factor of 3 to 5 times greater. Recent observations indicate that the high latitude enhancement of heating may only be a factor of 2 instead of 5 (Jacchia, 1967). Jacchia also reported that the temperature variations, on the average, lag five hours behind the geomagnetic variations. The atmospheric density during a large magnetic storm may change as much as an order of magnitude (Jacchia, 1964 b), and every geomagnetic variation has a counterpart in an atmospheric density variation (Jacchia and Slowey, 1964 a).
As the solar cycle progressed toward the minimum of activity, it became apparent that the value of temperature increase of 1.2 °K per unit increase in \( ap \) was valid only for periods of large magnetic activity. Jacchia and Slowey, (1964 b) determined that for quiet times ( \( Kp \leq 2 \) ) the temperature increase is 7 °K per unit increase in \( ap \). This is 6 times the value found for magnetic storm conditions. This was confirmed by Newton et al. (1965).

Newton et al. (1964, 1965), using density gages on Explorer 17, found that the measured density disagreed by a factor of 2 with densities computed from satellite drag. The densities measured by Explorer 17 are more likely to be in error than are the densities computed from satellite drag. The densities computed from atmospheric drag on a variety of satellites of different sizes and shapes are in agreement. Cook (1965) reviewed satellite drag techniques and estimated that the maximum possible error in the density estimates is 30 percent, but that a 10 percent error is more likely. Friedman (1966) reviewed the techniques used to determine atmospheric density by means of ionization gages. He concluded that since it was not possible to predict the error in the calculated gage sensitivity of atomic oxygen, the uncertainties assigned to the Explorer 17 density measurements are questionable.
It is not of great importance, however, that the two techniques yield results that are not in absolute agreement; the important point is that both techniques show relative density changes during magnetic storms that agree qualitatively.

Density variations at low altitudes (170 km) during magnetic storms are in phase with density variations at higher levels (Zirm, 1964). Jacchia (1966) interpreted this as showing that the heating occurs lower than 170 km, possibly in the E layer.

Jacchia noted that both the solar plasma velocity and the exospheric temperature are related to the geomagnetic indices. He then determined that the exospheric temperature varies with the solar wind velocity in a nearly linear fashion (Jacchia, 1965).

In order to compute the temperatures at a certain altitude knowing only the density, an atmospheric model is used. The models used most frequently are those of Nicolet (1961) and Jacchia (1964 c). The models have the same fixed boundary conditions at the reference altitude of 120 km. Both models predict densities close to the densities determined by drag analysis. Stein and Walker (1965) used a number of different boundary conditions at 120 km and found that many models generated densities close to that observed. However, when temperatures are computed
from these different models, the values obtained differ as much as 25 percent. This is due primarily to the uncertainty in the concentration of atomic oxygen. They conclude that upper atmospheric temperatures deduced from satellite drag are uncertain by as much as 25 percent.

As illustrated in Figure 2, there is a time delay between the maximum temperature of the ionosphere and the maximum geomagnetic activity. Roemer (1966), using a large number of storms, found the average time delay to be 5.4 ± 0.4 hours. He rules out any dependence of the time delay on local time or on the intensity of the geomagnetic disturbance, at least in the range Kp = 4 to 9. Jacchia et al. (1967) stated that the time delay is 6 hours at high latitudes and 7 hours at low latitudes. Jacchia (1967) found a 5.0 hour time delay using low altitude satellites.

The difference in heating for a unit increase in ap for quiet and stormy conditions was discussed briefly before. More data has made a definite statement possible. In Figure 3, taken from a paper by Jacchia (1966), the differences are clearly seen. The most accurate relation between the temperature increase $\Delta T$ and the ap index at the present time is given by Jacchia (1967), and is

$$\Delta T = 1.0^\circ ap + 100^\circ [1 - \exp(-0.08ap)]$$
C. Hydromagnetic Wave Attenuation in the Ionosphere

In order to explain atmospheric heating during magnetic storms, Dessler (1959 a, b; 1965) proposed that hydromagnetic (hm) waves generated during the storm, presumably at the boundary of the magnetosphere, may dissipate enough energy in the ionosphere during a magnetic storm to produce the observed heating. Dessler used an order of magnitude calculation to show that a hm wave incident on the ionosphere with an amplitude of 30 gammas (3 x 10^{-8} weber/m^2) at a frequency of one cycle per second could produce significant heating. Since the hm wave spectrum above the ionosphere had not been measured at that time, no definite conclusion regarding the heating could be reached. Fejer (1960) derived more exact expressions for hm wave propagation and dissipation. His work confirmed Dessler's order of magnitude calculations.

Francis and Karplus (1960) and Karplus et al. (1962) numerically integrated the hm wave equations in the ionosphere for four different ionospheric conditions to determine hm wave dissipation and attenuation. These papers will now be discussed in some detail because of their importance to this thesis. No attempt will be made to distinguish between them since the 1962 paper is essentially a continuation of the 1960 paper.
Plane hm waves were assumed incident vertically on a horizontally stratified atmosphere with a magnetic field \( \mathbf{B} = B_0 \hat{z} \) at a dip angle of 60° above a perfectly conducting earth. They considered simple harmonic time dependence of magnetic and electric fields \( \mathbf{B} \) and \( \mathbf{E} \) and the current density \( \mathbf{j} \). Maxwell's equations were then written as

\[
\nabla \times \mathbf{E} = i \omega \mathbf{B} \\
\n\nabla \times \mathbf{B} = \mu_0 \mathbf{j}.
\]

For the frequencies under consideration, the displacement current is negligible. The current density is

\[
\mathbf{j} = \sigma_0 \hat{z} (\mathbf{E} \cdot \hat{z}) + \sigma_1 [\mathbf{E} - \sigma_0 (\mathbf{E} \cdot \hat{z})] \\
+ \sigma_2 (\mathbf{E} \times \hat{z}),
\]

where \( \sigma_0 \) is the zero-field conductivity; \( \sigma_1 \) is the Pederson conductivity; and \( \sigma_2 \) is the Hall conductivity.

Using the auxiliary expressions

\[
\sigma_3 = \sigma_1 + \frac{\sigma_2}{\sigma_1}
\]
\[ \sigma_{r} = \cos^2 \psi + \left( \frac{\sigma_{i}}{\sigma_{o}} \right) \sin^2 \psi \]

\[ \sigma_{z} = \cos^2 \psi + \left( \frac{\sigma_{i}}{\sigma_{o}} \right) \sin^2 \psi, \]

where \( \psi \) is the dip angle; the equations for the electric field, magnetic field, and the current density were written as follows:

\[ B_{y} = \frac{i}{\omega} \frac{dE_{z}}{dx} \]
\[ B_{z} = -\frac{i}{\omega} \frac{dE_{y}}{dx} \]

\[ J_{y} = \frac{\sigma_{x}}{\sigma_{i}} E_{z} \cos \psi + \frac{\sigma_{z}}{\sigma_{i}} \sigma_{i} E_{y} \]
\[ J_{z} = \frac{\sigma_{x}}{\sigma_{i}} E_{z} - \frac{\sigma_{z}}{\sigma_{i}} \sigma_{i} E_{y} \cos \psi \]

\[ \left[ \frac{1}{M_{o}} \frac{d^2}{dx^2} + i \omega \sigma_{i} \frac{\sigma_{z}}{\sigma_{i}} \right] E_{y}(x) + i \omega \sigma_{z} \frac{\cos \psi}{\sigma_{i}} E_{z}(x) = 0 \]

\[ \left[ \frac{1}{M_{o}} \frac{d^2}{dx^2} + i \omega \sigma_{i} \frac{1}{\sigma_{i}} \right] E_{z}(x) - i \omega \sigma_{z} \frac{\cos \psi}{\sigma_{i}} E_{y}(x) = 0. \]
The coordinate system used is shown in Figure 4.

The atmosphere was divided into two regions, the neutral atmosphere and the ionosphere.

\[ x = 0 \quad : \quad \text{surface of the earth} \]
\[ 0 < x < d = 80 \text{ km} \quad : \quad \text{neutral atmosphere} \]
\[ d < x < D = 550 \text{ km} \quad : \quad \text{ionosphere} \]
\[ x = D = 550 \text{ km} \quad : \quad \text{upper limit of integration} \]

They numerically integrated the wave equations using the following boundary conditions: the electric field vanishes at \( x = 0 \), the tangential component of the electric field and its derivative are continuous at \( x = 80 \text{ km} \). They obtained two solutions. The first for ordinary wave incidence at 550 km, and the second for extraordinary wave incidence at 550 km. Ordinary waves are plane polarized in the \( Y \) direction (\( E_x = 0 \)), and extraordinary waves are plane polarized in the \( Z \) direction (\( E_y = 0 \)).

The local time average power dissipation was computed from

\[ P(x) = \frac{1}{2} \Re\{ E(x)^* \cdot J(x) \} . \]

It is particularly important to point out at this time that the incident hm wave spectrum was not known. To illustrate the effects of the ionosphere on hm waves, Francis and Karplus and Karplus et al. computed the power dissipation
and attenuation for ordinary and extraordinary incident waves at frequencies of 1, 5, 10, and 30 radians/second with an incident flux of $10^{-3}$ watts/m$^2$.

The numerical integration of the wave equation was carried out for ionospheric conditions existing at day sunspot-minimum, night sunspot-minimum, day sunspot-maximum, and night sunspot-maximum. Figures 5 and 6 show for day sunspot-maximum $|E|^2$ and $|B|^2$ respectively, as a function of altitude. Figure 7 gives the power dissipation as a function of altitude for the four frequencies considered, for an incident energy flux of $1.0 \times 10^{-3}$ watts/m$^2$ at each frequency. In Figure 5, the electric field goes to zero at zero altitude because the earth was assumed to be a perfect conductor. Figure 6 is especially important. From it we see that at zero altitude, the magnetic field is finite. The wave amplitude at the earth's surface is essentially the same as the wave amplitude at 80 km. It is possible to relate the amplitude of the magnetic field of the wave at 80 km to the wave amplitude above the ionosphere. It is then possible to compute the power dissipated by the wave. Figure 8 gives the amplitude transmission through the ionosphere as a function of frequency for extreme phases of the sunspot cycle. It is seen that waves with frequencies less than 1 radian/second (0.16 c/s) suffer little attenuation in passing through the ionosphere and, therefore, do little
atmospheric heating. The highest frequency considered was 30 radians/second (4.8 c/s). They concluded that incident hm waves with amplitudes of 50 gammas would produce significant changes in the atmospheric density. A 50 gamma hm wave at 550 km has an energy flux approximately equal to that of extreme ultraviolet solar radiation. Ershkovish (1963) using similar techniques obtained almost identical results.

Akasofu (1960) computed the attenuation coefficients for 0.01, 0.1, and 1.0 c/s incident hm waves. He determined that the attenuation coefficient for a 1.0 c/s wave was $8 \times 10^6$. He then concluded that since 1 c/s micropulsations are observed on the ground, they must be generated below the ionosphere. In order to have traversed the ionosphere, the incident amplitude of the hm wave would have been at least $10^4$ gammas, which he considered totally unreasonable. Using dispersion equations developed by Hines (1953), Akasofu computed the relative amplitudes at 60 and 500 km for plane hm waves at different local times (Akasofu, 1965). Akasofu's results are not correct. His attenuation factor of $8 \times 10^6$ is at least a factor of $10^3$ greater than that computed in other papers (which are consistent with each other). His approximations were crude; he only considered a small region of the ionosphere and then generalized to the
entire ionosphere. It is now generally accepted that the majority of Pc 1 micropulsations are generated above the ionosphere. This would mean, as Akasofu pointed out, that the hm wave above the ionosphere would have an amplitude of at least $10^4$ gammas. This corresponds to an energy flux orders of magnitude greater than solar ultraviolet radiation. Atmospheric heating has not been observed during Pc 1 micropulsation events, so Akasofu's result is incorrect.

Greifinger and Greifinger (1965) and Field and Greifinger (1965, 1966) derived analytical solutions for hm wave attenuation in the ionosphere using a model similar to Francis and Karplus (1960). Their results were similar to Karplus et al. (1962). In addition they made the interesting discovery that the transmission coefficient had a pronounced resonance at a frequency of 0.15 c/s. They suggested this might account for micropulsations observed near this frequency.

D. Previous Measurements of the Hydromagnetic Wave Spectrum

As was previously discussed, Dessler (1959 a, b), Francis and Karplus (1960), and Karplus et al. (1962) assumed an incident spectrum of hm waves because the incident spectrum had not been experimentally determined at that time.
Hydromagnetic waves propagating in a variety of modes have been observed in the magnetosphere (see e.g. Coleman et al. 1960, Judge and Coleman 1962, Sonett et al. 1962, Ness et al. 1962, and Patel and Cahill 1964). The upper frequency limit in these papers was 2 c/s; most observations were of much lower frequencies. Invariably, the observations were made at distances greater than an earth radius from the surface of the earth. For use in computing hm wave heating in the ionosphere, the observation should ideally be made just above the altitude where wave attenuation becomes important. This corresponds to altitudes from 600 - 1000 km. The magnetometer should be able to respond to at least a 7 c/s hm wave. The observations would have to be made during a magnetic storm. Such satellite measurements have not yet been made.

Another area that needs more discussion is the frequency range assumed in the theoretical papers on hm wave attenuation. There is general agreement that frequencies below 0.1 c/s pass through the ionosphere with little attenuation. The highest frequencies considered varied from 5 to 10 c/s. It is reasonable to ask why nothing above 10 c/s was considered. The reason is that small magnetic variations, call micropulsations, have been observed for years at the earth's surface. Micropulsations
occur at frequencies less than 5 c/s (e.g. Jacobs et al. 1964, and Jacobs and Westphal 1963). This does not, of course, rule out higher frequencies. It is not necessary to consider frequencies from 7 to 26 c/s because observations have already been made in this frequency range. The Schuman resonance phenomena, Schuman (1952), is a propagation of electromagnetic waves at certain frequencies in a wave guide formed by the earth and ionosphere. The principal modes occur at frequencies of approximately 8, 14, 20, and 26 c/s (Balser, 1964). The magnetic field intensity is generally less than a milligauss. Observations in this frequency range show nothing that correlates with geomagnetic activity (e.g. Keefe and Polk, 1964, and Gendrin and Stefant, 1964). There are, however, seasonal changes that are due to variations in global thunderstorm activity. Also, nuclear explosions in the ionosphere produce marked changes in the spectrum. Frequencies above 26 c/s need not be considered since there are few places in the magnetosphere where hm waves could propagate with such a high frequency.
EXPERIMENTAL APPARATUS

A. Design Requirements

It has been theoretically predicted that if plane h m waves propagate vertically into the ionosphere, attenuation will occur. The amplitude below the ionosphere will be less than the amplitude above the ionosphere. The purpose of this experiment is to measure the wave amplitude at the earth's surface so that the amplitude above the ionosphere and the power dissipated in the ionosphere can be computed. The amplitude of the electric field or the magnetic field could be measured. However, Figure 5 shows that the electric field below the ionosphere goes to zero if we assume a perfectly conducting earth. It is much easier to measure the magnetic field. Figure 6 illustrates the relative magnetic amplitudes above and below the ionosphere at day sunspot-maximum.

In order to design a system to measure the wave spectrum, it is necessary to know in what amplitude and frequency range the waves that might heat the ionosphere should occur. The lower limit on the frequency should be approximately 0.1 c/s. Waves with frequencies less than this pass through the ionosphere with little attenuation and, therefore, do little ionospheric heating. In section D of the Introduction, it was pointed out that observations of the Schuman resonance
have revealed nothing in the frequency range 7 to 26 c/s that correlates with geomagnetic activity. The sensitivity of the measurement was such that waves with amplitudes small enough to be undetected at earth's surface would produce negligible heating. Therefore, the highest frequency that needs to be considered is 7 c/s.

Karplus et al. (1963) stated that an incident hm wave with an amplitude of 50 gammas will produce significant ionospheric heating. In order to produce significant heating, the incident energy flux would have to be comparable to the flux of extreme ultraviolet (EUV) solar radiation, thought to be the principle heat source of the upper atmosphere. The flux of EUV is about $1 \times 10^{-3}$ watts/m$^2$. (e.g. see Harris and Priester, 1963; Hunt and Van Zant, 1961). A hm wave with an amplitude of 50 gammas at 550 km altitude has an energy flux of approximately $1 \times 10^{-3}$ watts/m$^2$. From Figure 8, it is seen that for day conditions between solar minimum and maximum the percentage amplitude transmission is from one to twenty percent, depending on the frequency of the incident wave. For a 50 gamma amplitude incident wave, the amplitude on the ground should be from 0.5 to 10 gammas.

B. Rubidium Vapor Magnetometer System

1. Description of Equipment
There are two basic types of magnetometers used to measure magnetic fluctuations in the amplitude and frequency range of interest. They are the rubidium vapor magnetometer and the induction coil magnetometer. The rubidium magnetometer measures only the total magnetic field intensity, in increments as small as 0.01 gammas, and gives no information about the direction of the magnetic fluctuations. The frequency response is flat from zero frequency up to kilocycle per second variations.

An induction coil magnetometer will resolve magnetic fluctuations as small as 0.005 gammas in the frequency range of interest. The coil responds only to the component of the magnetic field directed along the axis of the coil. A system of three mutually orthogonal coils can be used to determine the magnitude and direction of the fluctuating magnetic field. It is necessary to calibrate the coils using a magnetic field of a known frequency and amplitude.

The rubidium magnetometer was chosen to be the principal instrument of this investigation. For details of the theory of operation, the reader is referred to Bloom (1962) and Skillman and Bender (1958). Two induction coil magnetometers were constructed for use as a back-up system for the rubidium magnetometer, and also to record Pc 1 micropulsations. They will be described in detail in the next section.

A Model V-4938 Rubidium Magnetometer was purchased from Varian Associates. The magnetometer consists of three units.
The rubidium oscillator produces a signal whose frequency, the Larmor frequency, is directly proportional to the total magnetic field intensity. The proportionality constant is 4.667 cycles per second per gamma for rubidium 85. This signal is conducted along a 200 foot coaxial cable to the readout unit where the Larmor frequency goes to a mixer circuit and the difference frequency is converted by a discriminator to a DC voltage. As the frequency changes, the voltage changes such that a one gamma change will produce a 7.779 millivolt change in the discriminator output. The discriminator has a flat frequency response from zero frequency to 20 c/s. The discriminator voltage is recorded by an analog recorder and is also available for recording by other methods.

The wave spectrum is determined by performing a power spectrum analysis on the data that has been obtained with the magnetometer. In order that the analysis can be performed on an electronic computer, the magnetometer output needs to be in digital form. The equipment that was used to digitize the magnetometer output is shown schematically in Figure 9. The rubidium oscillator signal goes to the readout unit where it is converted by a discriminator circuit to a voltage that is directly proportional to the frequency of the signal from the oscillator and, therefore, directly proportional to the magnetic field intensity. As the magnetic
field intensity changes, so does the output voltage of the discriminator. This voltage is recorded by a strip chart recorder. The output of the discriminator also goes to a low-pass filter and a high-pass filter. From the low-pass filter the DC level and lower frequency components are sampled by the second channel of a two channel analog-to-digital (A-D) converter. The higher frequency components from the high-pass filter are amplified by a DC amplifier and then pass through a band-pass filter. The output of the filter goes to the first channel of the A-D converter. The numbers from the A-D converter are recorded on an incremental tape recorder. A detailed discussion of this system will now be given.

The manufacturer of the rubidium magnetometer, Varian Associates, states that the minimum amplitude signal that is resolvable is 0.01 gammas. A change in the magnetic field intensity of 0.01 gammas produces a change in the discriminator voltage of 0.077 millivolts. In order to measure this with an A-D converter the voltage must be amplified. Due to the low frequencies involved a DC amplifier must be used. The DC voltage level from the discriminator is about 3 volts. The amplifier gain needs to be about thirty. It is obvious that the DC level of the discriminator must not be amplified. The DC level of the discriminator was eliminated by using a simple R-C hi-pass filter. Low leakage
capacitors were used. The amplifier used was an Astrodata Model 885 wideband DC amplifier.

In order to prevent aliasing, another filter must be used to remove all components of the signal with frequencies greater than the Nyquist frequency. The Nyquist frequency is defined as the reciprocal of twice the time interval between samples, \( N_y = \frac{1}{2\Delta t} \). A Krohn-Hite Model 330-A(R)-4 band-pass filter follows the amplifier. The lower frequency cut-off is set at 0.01 c/s to prevent unwanted low frequency components from entering the A-D converter. The upper frequency cut-off is set at the Nyquist frequency. The output of the filter is connected to channel 1 of the A-D converter. The frequency response curve for this part of the system is shown in Figure 10. The DC level and low frequency components blocked by the hi-pass filter are of interest and should be recorded. Therefore, a two channel A-D converter was needed. That is, a converter that samples one input for a specific number of times then switches to a different input for a sample, then switches back to the original input and the cycle starts over again. The A-D converter needs to have a sampling rate such that the Nyquist frequency is above the frequencies of interest. Since in this investigation the frequency range extends up to 7 c/s the sampling rate should be at least 14 samples per second (SPS).
In order to prevent aliasing, the signal from the discriminator to channel 2 goes through an RC low-pass filter. Low leakage capacitors were used. From the low-pass filter, the signal goes to channel 2 of the A-D converter. The frequency response curve for this part of the system is shown in Figure 11.

The output of the A-D converter is a binary number. This number could have been recorded on punched paper tape, on a regular (non-incremental) magnetic tape recorder, or on an incremental recorder. The paper tape recorder was rejected because the Rice IBM 7040 computer does not have a paper tape input system. The non-incremental recorder was rejected because it would be necessary to repack the data on the tape before it would be compatible with the IBM 7040 magnetic tape input system.

The incremental tape recorder used was a Precision Instrument Model PI-1107, Type A. It used regular IBM magnetic computer tape and writes one tape character at a time. Each tape character consists of seven channels across the tape as illustrated in Figure 12. Only six of the channels are used for data, the seventh is used for a parity bit. The parity bit is used by the computer to determine if a writing or reading error has occurred in the six bits of data in the tape character.
There are three tape formats used for computer input. The first has 200 tape characters per inch, the second has 556, and the third has 800 characters per inch. The Precision Instruments recorder writes 556 characters per inch, and is capable of writing from 0 to 300 tape characters per second. The tape that is written is IBM compatible with a 3.4 inch load gap, a 0.75 inch inter-record gap (IRG), and a 3.4 inch end-of-file gap (EOF).

After a reel of tape is properly loaded on the recorder, the recorder operates in the following way. A write pulse into the recorder causes either 0 or 1 bit, depending on the number being written, to be recorded in each of the six writing locations in a tape character. A parity bit is written on the seventh. The transport capstan rotates and moves the tape 1/556 of an inch to the next tape character. The recorder waits until the next write pulse occurs. The time between write pulses can vary from 1/300 of a second to weeks or months.

The A-D converter used was built by Test Equipment Corporation in Houston, Texas specifically for this project. It is designed to operate in conjunction with the incremental recorder and generates all pulses to operate the recorder.

Each converter sample consists of twelve bits, eleven
for the magnitude of the voltage and one for a plus or minus sign. Twelve bits per sample means that the tape recorder must record two six bit characters for each sample. The way this is written on the tape is illustrated in Figure 12.

The A-D converter produces pulses to write the inter-record gaps. Inter-record gaps are used to divide the data on the tape into convenient lengths for computation by the computer. There cannot be more numbers between IRG's than there is storage location in the computer memory. Generally there are considerably fewer numbers since room must be left for the computer program. The A-D converter generates a write IRG pulse at 100, 200, 400, or 800 second intervals. A switch on the A-D converter determines this. Since the time between IRG's is known, they are used as timing marks. A crystal oscillator in the A-D converter acts as a clock and is accurate to within a minute over a period of 6 days. This is accurate enough for the purposes of this investigation.

An IRG takes 0.7 ± 0.1 seconds to write. In order to keep the number of samples per record a constant, the converter stops sampling for exactly one second while the IRG is written.

When the stop button on the A-D converter is pushed, an end-of-file mark is written. An end-of-file mark is
used to identify the end of a series of records.

Each A-D converter measurement is accurate to ± 0.05% ± 1/2 the least significant bit. For small voltages the ± 1/2 bit is the most important and it corresponds to ± 2.5 millivolts. A gain setting of 30 in the DC amplifier gives an uncertainty in the measurement of a 0.05 gamma variation of about 20 percent. For a 1 gamma variation, the uncertainty is approximately 2 percent and is acceptable. These uncertainties were confirmed by using the calibration coil to generate a signal at the sensor and then comparing the numbers recorded on the magnetic tape with amplitudes of the variation on the strip-chart recorder.

The output voltage of the amplifier is limited to ± 10 volts peak AC. It would take an 86 gamma (peak-to-peak) fluctuation to saturate the amplifier and this is unlikely to occur.

For this investigation, the sampling rate of channel 1 was 15 SPS yielding a Nyquist frequency of 7.5 c/s. For channel 2 the sampling rate was 0.1 SPS for a Nyquist frequency of 0.05 c/s. The record length was 400 seconds.

The A-D converter-incremental tape recorder combination works in the following manner. A reel of magnetic tape is loaded on the recorder as described in the tape recorder manual. The start button is pushed and the converter makes the first sample. The first six bits are recorded and the
recorder increments. Then the next six are written and the recorder waits for the next sample. One fifteenth of a second after the start of the first sample the second sample is made. This continues until exactly eight seconds after the first sample. Then the converter switches to channel 2 and takes one sample and then returns to channel 1. Then ten seconds later another sample from channel 2 is taken. This continues until it is time to write an IRG. The setting of 400 seconds per record means that at the end of 399 seconds (actually at the end of 15 X 399 = 5985 samples) it is time to write an IRG. The A-D converter stops sampling for one second while the mark is written. The reason the second channel is sampled 8 seconds after the start of sampling and each 10 seconds thereafter is to prevent the occurrence of the second channel sample during the one second the IRG is being written.

There are 15 channel 1 samples missing each 400 seconds because of the IRG and 40 missing because of the 40 channel 2 samples. There are, however, no gaps in channel 2 samples in an entire data tape.

At a sampling rate of 15 SPS, a 2400 foot reel of magnetic tape lasts six days.

The analog recorder was used as a monitor. The chart speed was 1 inch/hour with a full scale deflection of 50 gammas. The recorder has a 50 position bias stepper. If
an interesting event occurs, the time of occurrence can be determined to within 3 minutes from an examination of the chart. Then knowing that each record on the magnetic tape is 400 seconds long, the number of the record in which the event was recorded can be determined. A computer program then can find the desired record on the tape.

The frequency of the signal from the rubidium oscillator near Houston is approximately 243 kilocycles/second. The discriminator circuit in the readout unit operates for input frequencies between 50 and 3000 c/s. The readout unit has a number of crystals with different frequencies that work in mixer circuits to drop the frequency of the rubidium oscillator signal into the range of the discriminator. If on a magnetically quiet day, the frequency into the discriminator is 2000 c/s, it is unlikely that during a magnetic storm the rubidium oscillator output frequency will change enough to drive the frequency of the signal into the discriminator out of the 50 to 3000 c/s range. Only a very unusual increase of 200 gammas or a decrease of 400 gammas would accomplish this.

A sensor-telephone line coupler was built by Varian Associates. It consists of a mixer circuit, to reduce the rubidium oscillator output frequency to 2000 c/s, and an amplifier. When using the coupler, only the rubidium sensor and a power supply need be at the site of the observatory.
An audio grade telephone line was used 24 hours a day to transmit the signal to the Rice University campus. The signal from the telephone line came into the discriminator and the system operated as previously described.

The telephone line coupler was used for about a year. When operating properly there was no noticeable degradation of the magnetic measurements. On occasion, however, there were problems. Lightning struck the telephone line twice and destroyed the output transistors in the coupler amplifier. There was occasional noise generated somewhere on the telephone line circuit that ruined the data. Sending data over the telephone line was discontinued in September, 1966 for these reasons and reasons to be discussed in the next section.

A calibration coil was used to provide a varying magnetic field at the rubidium sensor for testing purposes. The coil was placed 6.7 m due south (magnetic) of the sensor. It was octagonal in shape with a radius of 1.04 m and had 52 turns of insulated aluminum wire. Since the coil was used for low frequency signals, the wave lengths were orders of magnitude greater than the distance between the calibration coil and the sensor. Therefore, the magnetostatic component of the dipole is dominant and the field at a point on the axis of the coil is given by
B (in gammas) = \( \frac{M_0 M_c R_0^2}{2 R^3 R_s} V_o \)

where \( M_0 \) is the magnetic permeability of free space, \( M_c \) is the number of turns of wire on the coil, \( R_0 \) is the radius of the coil, \( V_o \) is the voltage across a 879 \( \Omega \) resistor in series with the coil, \( R \) is the distance of the sensor from the coil, and \( R_s \) is the value of the series resistor. Putting in the appropriate values yields

\[
B \text{ (in gammas)} = 6.0 \times 10^{-2} V_o
\]

This coil is driven by a low frequency oscillator.

2. Equipment Problems

A number of unexpected problems arose soon after the equipment was assembled. The first problem was with the Krohn-Hite band-pass filter. Variations in the AC power line voltage produced a variation in the output level of the filter. These variations were as large as some of the smaller signals. The amplifier was placed before the filter so that the variations would not be amplified. An AC line voltage stabilizer was used to reduce the variations even further.

After the system was operating, spurious EOF marks were found on the tape. Violent AC line voltage changes triggered these. The AC line voltage stabilizer reduced
the number of spurious marks considerable, but a switch grounding the wire from the A-D converter carrying EOF pulses solved the problem completely. Just before pushing the stop button the wire was disconnected from ground.

Spurious inter-record gaps were found on each data tape. The problem was never solved completely, but was reduced to the extent that only one or two were found on each data tape containing six days of recording.

In September, 1966 all the equipment was moved to the site of the observatory. The AC line voltage was much more stable. The generally quiet electrical surroundings reduced the number of spurious tape marks considerably.

3. Computer Programs

The computer programs used to separate the samples from the first channel of A-D converter from the samples of a second channel will now be briefly discussed.

In order to perform an analysis on the data from channel 1 of the A-D converter, it was necessary to remove the channel 2 samples. This was done in the following way. The program goes to the desired record and stores it in the computer memory. It checks to see if 5985 samples are present. If they are not, the record is defective. Generally, it is not worth the effort to try and get as much data as possible from a defective record since for the purpose of this experiment, they were rare. The computer
prints a message telling that the record is defective and, on a sense switch option, will print out the defective record.

If the record is of the proper length, the program proceeds to eliminate the channel 2 samples. They will occur in the same position in each record so the program finds these positions in the memory and replaces these numbers by the average of the numbers on each side. This introduces little error since there are 149 channel 1 samples for each channel 2 sample. Then the program proceeds to whatever computation is called for.

The samples from the second channel of the A-D converter contain information on very low frequency magnetic variations. Many computations on very low frequency magnetic variations would need more than 40 numbers, the number of second channel samples per record. Therefore, a data transfer program was written to search down the data tape, find the second channel samples and write them on another tape. Each record on the output tape is 1080 samples (3 hours) long.

If the data tape is written with no defects, the second channel samples will have no interruption from the start to the end. In order that this be the case on the output tape, the second channel samples must be obtained from each record. If a parity error occurs in
a record, the input system on the Rice University IBM 7040 computer will reject it and all the second channel samples will be lost. The data transfer program has a subroutine that turns off this function of the 7040 input system. A record with a parity error will be taken into the memory and a message printed telling it which input and output records the parity error occurred.

The data transfer program can also find the second channel samples if the records are not of the proper length, or if up to five spurious inter-record gaps occur. If the defect is too serious for the program to handle, a message is printed saying so and 40 zeros are written on the output tape.

C. Induction Coil Magnetometers

An induction coil magnetometer consists of a coil with many turns of wire connected to an amplifier, an integrating circuit, and a band-pass filter, and a strip chart recorder as shown in Figure 13. Two coil magnetometers were constructed. One was aligned in the magnetic north-south direction and the other in the magnetic east-west direction.

The design for the magnetometers was taken from Tepley (1961). The interested reader is referred to Tepley's paper for details of the design. Each sensor
consisted of two identical pick-up coils placed around a high-permeability core of Allegheny Ludlum No. 5 relay steel. The core is 6 feet long and 1.5 inches in diameter. The output of the coils is amplified by a Geotechnical Corporation Model 4300 Phototube Amplifier. The amplifier has a gain of 250,000. The signal from the amplifier passes through a Krohn-Hite Model 330-A (R)-4 variable electronic band-pass filter. The signal then goes through an integration circuit, Figure 14, and is recorded on a three channel Beckman Instrument Type SC Dynagraph recorder.

A calibration coil was constructed and placed so that it produced a field of equal magnitude at both sensors. The calibration coil consisted of 23 turns of wire wound around an octagonal form 1.5 meters in diameter.

Using equations similar to that used for the coil in the rubidium system, we get

\[ B \text{ (in gammas)} = 1.1 \times 10^{-3} V_0 \]

where \( V_0 \) is the voltage across the series resistor. The coil system magnetometer frequency response curve is shown in Figure 15.

The calibration coil was turned on once each hour for one minute by a timer. The voltage across the series resistor was recorded on the third channel of the Beckman recorder.
D. The Magnetic Observatory Site

The observatory was built on property owned by Rice University located 48 km (30 miles) north-east of Houston, Texas. The geographic coordinates were 95.2° W and 30.1° N. The magnetic latitude was 41° N. This property had an area of 3.6 km² (900 acres). The observatory was near the center of the property, and the nearest buildings were 2 to 3 km away. The Gulf of Mexico was approximately 140 km to the south.

A small building was used to house the equipment. The physical layout is shown in Figure 16. The sidewalk and instrument pads were made of concrete that contained, of course, no reinforcing steel. A concrete pier 3 feet by 3 feet by 6 feet was used to isolate the Geotechnical amplifiers from vibration.

For the maximum signal-to-noise ratio, the axis of the rubidium oscillator must be in the magnetic north-south direction and be at a 45° angle to the magnetic field vector. A plexiglass holder, as shown in Figure 17, kept the sensor in the proper position. A plexiglass box protected the sensor from the weather. During the winter it was necessary to add insulation to keep the sensor at the proper operating temperature.

The 200 foot coaxial cable from the rubidium oscillator
to the building was enclosed in 1 inch diameter plastic pipe to protect it from the weather and small gnawing rodents. The pad on which the rubidium oscillator was placed was surrounded by a 5 foot high fence made entirely of wood to repel cows and woodland creatures.

The induction coils were in wooden boxes placed on the concrete, and were covered by larger wooden boxes to prevent wind vibrations.

Figures 18, 19, and 20 are photographs taken at the observatory site. Figure 18 shows the rubidium oscillator and its plexiglass support. Figure 19 shows the building used to house the equipment. Figure 20 was taken in the building and shows the magnetic tape recorder, the A-D converter, the rubidium magnetometer readout unit, and the strip chart recorder.
EVALUATION OF THE RESULTS OF THE EXPERIMENT

The equipment began recording on a 24 hours per day basis in January, 1965. Equipment failures and a paucity of magnetic activity resulted in no useful data during 1965. In 1966 four magnetic storms occurred while the equipment was running properly. The storms began on March 19, March 23, September 8, and October 4. The methods used to evaluate the experimental results will now be discussed.

A. Power Spectrum Analysis

Power spectra were computed using Blackman and Tukey's (1959) method. The following expressions were used to obtain numerical estimates of the power spectrum:

\[ C_r = \frac{1}{m-r} \sum_{g=0}^{m-r} X_g \cdot X_{g+r} \]

\[ r = 0, 1, 2, \ldots, m \]

where \( C_r \) is the mean lagged products, \( m \) is the number of samples, \( X_g \) is the value of sample number \( g \), and \( m \) is the number of lags to be used.
\[ v_r = 2 \Delta t \left[ c_0 + 2 \sum_{q=1}^{m-1} c_q \cdot \cos \frac{qn\pi}{m} + c_m \cdot \cos r\pi \right] \]

where \( v_r \) is the raw spectral density estimate, and \( \Delta t \) is the time interval between samples. Refined spectral density estimates were obtained by smoothing the raw estimates using a process called hanning. Hanned estimates were obtained as follows:

\[ u_0 = 0.5 v_0 + 0.5 v_1 \]

\[ u_r = 0.25 v_{r-1} + 0.5 v_r + 0.25 v_{r+1} \]

\[ u_m = 0.5 v_{m-1} + 0.5 v_m \]

\[ 1 \leq r \leq m-1 \]

The frequency corresponding to \( r \) is \( \frac{r}{2am\Delta t} \). Thus, the highest frequency estimate is \( \frac{1}{2\Delta t} \), and the lowest non-zero frequency estimate is \( \frac{1}{2am\Delta t} \). The computed
values of power density will be distributed as $\chi^2$ with $2m/m$ degrees of freedom. The mean and linear trends were removed from the data before analysis.

Power spectra were computed for the first 253 seconds of every other 400 second tape record. A maximum lag of 150 was used with 3788 data points to give 50 degrees of freedom per estimate. The time between samples was 1/15 of a second, corresponding to a Nyquist frequency of 7.5 c/s. A maximum lag of 150 was used to obtain estimates separated by 0.05 c/s, resulting in a frequency resolution of 0.1 c/s. Fifty degrees of freedom per estimate was used in order that the 90 percent confidence limits would be acceptably small. The 90 percent confidence limits for 50 degrees of freedom as determined by Blackman and Tukey's test of significance were obtained by multiplying each refined spectral estimate by 1.45 to get the maximum value, and by 0.74 to get the minimum value. The true value of power in the band will be between these extremes 90 percent of the time.

Table 1 gives the three-hour Kp indices for the four magnetic storms considered in this thesis. In each storm the time interval used for the analysis began one hour before the onset of the storm and ended seven hours after the conclusion of the storm. The analysis extended seven hours past the conclusion of the storm to include any lag between
<table>
<thead>
<tr>
<th></th>
<th>THREE-HOUR RANGE INDICES $K_p$</th>
</tr>
</thead>
<tbody>
<tr>
<td>MARCH 19, 1966</td>
<td>2+ 2+ 4- 3+ 3- 5- 5-</td>
</tr>
<tr>
<td>MARCH 20, 1966</td>
<td>4+ 2- 2+ 2+ 2- 2+ 2+</td>
</tr>
<tr>
<td>MARCH 23, 1966</td>
<td>3+ 3- 5 7- 7- 6 7 4-</td>
</tr>
<tr>
<td>SEPTEMBER 8, 1966</td>
<td>4+ 5+ 5- 6- 5 5- 4- 5-</td>
</tr>
<tr>
<td>OCTOBER 4, 1966</td>
<td>2+ 3+ 2+ 3+ 4- 5+ 6</td>
</tr>
<tr>
<td>OCTOBER 5, 1966</td>
<td>5+ 5+ 4+ 5+ 3+ 5- 5- 3-</td>
</tr>
</tbody>
</table>

UNIVERSAL TIME
heating of hm waves and Kp. Figures 21 and 22 are typical examples of the spectra obtained. The spectrum plotted in Figure 21 was computed from data recorded at approximately 1200 U.T., 23 March 1966, and Figure 22 was computed from data recorded at approximately 0100 U.T., 8 September 1966. The power spectra shown in Figures 21 and 22 are representative of all the power spectra computed. No spectrum was significantly different from those illustrated.

B. Computation of Power Dissipation

A copy of the computer program used by Francis and Karplus (1960) and Karplus et al. (1962) to numerically integrate the hm wave equations in the ionosphere was obtained from Mr. W. E. Francis. This program computes the power dissipation in the ionosphere for a vertically incident hm wave of a given frequency and amplitude. It was modified to compute the wave amplitude at 550 km given the amplitude and frequency at zero altitude. The program then computes the power dissipation for this incident wave.

In order to use this program it was necessary to obtain the wave amplitude at the earth's surface from the power spectra that had been computed. It is possible to convert the power spectrum into an amplitude spectrum. The computational procedure of Blackman and Tukey predicts the average energy per unit bandwidth measured in (gammas)$^2$/c/s. The conversion to an amplitude spectrum is made by determining
the bandwidth of distinct spectral peaks. There were, however, no large distinct peaks in the frequency range of interest, as seen in Figures 21 and 22. The rather flat sections of the spectra are due primarily to instrument noise. The power density level from 0.2 to 7.5 c/s is almost exactly that predicted for instrument noise. Spectra from magnetically quiet periods were compared with spectra from the four magnetic storms. This comparison revealed there were no essential differences in the frequency range 0.2 to 7.5 c/s between the spectra from active and quiet times. That is, the spectra were generally within the 90 percent confidence limits of each other. Figure 23 is a typical power spectrum for magnetically quiet times (Kp ≤ 2). In the range 0.2 to 7.5 c/s, it is very similar to Figures 21 and 22. The power density estimates for frequencies less than 0.2 c/s were larger for the magnetic storm spectra.

These results demonstrate that the amplitudes of the magnetic fluctuations in the range 0.2 to 7.5 c/s were quite small during the magnetic storms. The amplitude of the 0.1 c/s fluctuations, although larger during magnetic storms, is not nearly large enough to account for atmospheric heating. Hydromagnetic waves, therefore, produce little heating in the ionosphere. An upper limit on energy deposition will now be determined.

Since we are interested in placing an upper limit on
heating by hm waves, it is not necessary to compute power dissipation for all ionospheric conditions. It is only necessary to consider one day-ionosphere and one night-ionosphere. If the upper limit on hm waves heating is not important for these extreme cases, it will not be important for the intermediate conditions.

If average day and night ionospheres for the sunspot conditions existing when the measurements were made had been used, the attenuation coefficients computed would have been smaller than those actually existing in the ionosphere. Atmospheric heating during the magnetic storms increases the temperature ($\Delta T = 250^\circ$ for $Kp = 7$) and density in the ionosphere in the region of interest. The ionospheric conditions existing during the storm are much closer to sunspot maximum conditions than is the average ionosphere. The ionospheric models for day and night sunspot maximum used by Karplus et al. (1962) were used in the calculations. Use of these models assures that an upper limit on power dissipation will be obtained.

It is not possible to say what portion (if any) of the power density estimates was due to hm waves. In order to obtain an upper limit on heating, it was assumed that the entire spectrum was due to hm waves.

It was not possible, or necessary, to compute the wave amplitude above the ionosphere for all 150 power density
estimates. Only five frequencies were considered: 1, 10, 20, 30 and 40 radians/second. If power dissipation in the ionosphere is not significant for these frequencies, it will not be for the intermediate frequencies. Amplitudes at these frequencies were obtained from the power spectra by assuming a peak of unit bandwidth. Then the amplitude is just the square root of the power density estimate.

The power spectra were divided into day and night groups. Spectra computed from data recorded near local noon formed the day group, and spectra computed from data recorded near local midnight formed the night group. The ionospheric attenuation is largest near local noon and smallest near local midnight. The maximum amplitude at each of the five frequencies was determined for the day and night groups. Corrections were made for the frequency response curve of the system.

Since we assumed the magnetic field of the wave was in the horizontal plane, it was necessary to multiply the amplitudes by two. The rubidium magnetometer only responds to changes in the total field, and since the dip-angle is approximately 60°, a horizontal magnetic field variation will change the total field magnitude by one-half the magnitude of the variation.

The computer program supplied by Mr. Francis uses as a boundary condition a perfectly conducting earth. The
real earth is not, of course, perfectly conducting. It was, therefore, necessary to multiply the amplitudes by a correction factor. The true value of the correction factor lies between 1 and 2, but is not known exactly. However, since an upper limit on heating was desired, the factor of 2 was used. The amplitudes used in the computation are given in Table 2.

Figures 24 through 35 and Table 2 show the results of this computation. It was necessary to compute the power dissipation for both ordinary and extraordinary wave incidence since either or both could have produced the spectrum observed at the ground. The plots shown are for extraordinary waves. As seen in Table 2, the extraordinary waves are attenuated more than the ordinary (except for 1 radian/second, night). The power dissipated at a given altitude by ordinary waves was generally a factor of 2 to 4 less than that of extraordinary waves. The shape of the curves were very similar.

In Figures 30 through 34, the power dissipation versus altitude is plotted for the five frequencies considered for night ionospheric conditions. The maximum power dissipation occurs for a wave with a frequency of 20 radian/second at an altitude of 280 km and is $10^{-13}$ watts/m$^3$.

The power dissipation for day ionospheric conditions is shown in Figures 24 through 28. The peak power dissipation
TABLE 2: The amplitudes of vertically incident hydromagnetic waves at zero and 550 km altitude.
The frequency of the wave (radians/second) is given. Extraordinary wave incidence is indicated by EX, and ordinary wave incidence is indicated by 0. The amplitudes are in gammas.

**DAY**

<table>
<thead>
<tr>
<th>Frequency</th>
<th>Mode</th>
<th>0 km</th>
<th>550 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>EX</td>
<td>0.11</td>
<td>0.28</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>0.11</td>
<td>0.16</td>
</tr>
<tr>
<td>10</td>
<td>EX</td>
<td>0.053</td>
<td>5.5</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>0.053</td>
<td>1.9</td>
</tr>
<tr>
<td>20</td>
<td>EX</td>
<td>0.051</td>
<td>7.9</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>0.051</td>
<td>4.7</td>
</tr>
<tr>
<td>30</td>
<td>EX</td>
<td>0.049</td>
<td>14.3</td>
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<td>0</td>
<td>0.049</td>
<td>8.8</td>
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<td>40</td>
<td>EX</td>
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<tr>
<td></td>
<td>0</td>
<td>0.048</td>
<td>8.2</td>
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**NIGHT**

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<th>550 km</th>
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</thead>
<tbody>
<tr>
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<td>EX</td>
<td>0.10</td>
<td>0.03</td>
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<td>0.10</td>
<td>0.04</td>
</tr>
<tr>
<td>10</td>
<td>EX</td>
<td>0.054</td>
<td>0.31</td>
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<td>0</td>
<td>0.054</td>
<td>0.11</td>
</tr>
<tr>
<td>20</td>
<td>EX</td>
<td>0.051</td>
<td>0.33</td>
</tr>
<tr>
<td></td>
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<td>0.051</td>
<td>0.23</td>
</tr>
<tr>
<td>30</td>
<td>EX</td>
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<td>0.049</td>
<td>0.21</td>
</tr>
<tr>
<td>40</td>
<td>EX</td>
<td>0.047</td>
<td>0.27</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>0.047</td>
<td>0.26</td>
</tr>
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of $9 \times 10^{-11}$ watts/m$^3$ occurs at 300 km altitude for an incident wave of 30 radians/second.

In Figures 29 and 35 there is plotted the total power absorbed above a certain altitude versus altitude for extraordinary waves for the day and night conditions. This is power dissipated by both the incident and reflected wave. For day conditions, the $2.4 \times 10^{-5}$ watts/m$^2$-column absorbed above 80 km is approximately the incident energy flux on the ionosphere since the reflected wave is very much smaller than the incident wave. To produce the observed changes in density and temperature, the incident energy flux would have to be comparable with the incident flux of extreme ultraviolet (EUV) solar radiation. This is thought to be about $1 \times 10^{-3}$ watts/m$^2$ (e.g. see Harris and Priester, 1963; Hunt and Van Zandt, 1961). For day conditions the upper limit on the energy flux is $2.4 \times 10^{-2}$ that of EUV. For night conditions, the upper limit on the energy flux is $10^{-4}$ that of EUV.

C. Conclusions

It has been demonstrated that the upper limits of both the power dissipation per unit volume and incident energy flux are much too small to account for the observed ionospheric heating. It is, therefore, concluded that hm waves do not produce the changes in density and temperature
observed during magnetic storms. This result is only valid for mid-latitudes.

This conclusion is strengthened by the recent observation that the heating rate $\Delta T/\Delta Kp$ is essentially the same for the day and night sides of the earth. Because the attenuation factor of the night is considerably smaller than that of the day ionosphere, the amplitudes of the incident $h m$ waves necessary to produce equivalent heating would have to be much larger on the night side than those on the day side. Since we had only one observatory, it was not possible to measure simultaneously the wave spectrum for both the day and night sides of the earth. A comparison of power spectra of day and night data from the single observatory, however, revealed no important differences. This lack of asymmetry is more evidence that $h m$ waves do not heat the ionosphere to any appreciable extent.

Geomagnetic micropulsations have very nearly the same amplitudes and frequencies as those used in the computation of power dissipation. It can be concluded that micropulsations do not contribute significantly to the heat budget of the ionosphere.

D. Additional Theories of Ionospheric Heating

There have been several other mechanisms proposed to
account for the ionospheric heating observed during magnetic storms. A brief review of the most often mentioned theories is given below.

Cole (1961) calculated the increase in ionospheric temperature due to joule heating by ionospheric currents. He estimated the current flowing in the ionosphere needed to produce the geomagnetic disturbances observed on the ground, and the ionospheric conductivities. He then computed temperature increase due to the joule heating. The increase in temperature produced by joule heating for a moderate magnetic disturbance is within a factor of 2 of those actually observed. Cole predicted that since there is an enhancement of geomagnetic activity in the auroral zone the joule heating should be greater by a factor of approximately 2. An enhancement of heating in the auroral zone by a factor of 2 to 3 has been observed. The maximum joule heating, according to Cole, should occur at 150 km altitude. Satellite observations indicate the heating occurs below 200 km. Cole computed joule heating only for a day-ionosphere. Since the conductivities in the night ionosphere are considerably different, it is difficult to see how this theory, as it now stands, could explain the observed equivalent heating rates in the day and night ionospheres.

Cummings (1966) computed the power input into the ionosphere caused by joule heating from the ionospheric cur-
rents of a partial ring current system. The power input into the ionosphere for the recovery phase of a typical magnetic storm is approximately $10^{11}$ watts. The estimated power input of solar EUV is $10^{12}$ watts. It is possible, however, that the power input into the ionosphere by joule heating could exceed the power input of EUV in the regions where the currents are flowing.

Hines (1965) suggested that internal atmospheric gravity waves could deposit energy in the ionosphere. The waves would transport energy from the auroral zones to lower latitudes. The energy in the auroral zones comes, presumably, from joule heating by ionospheric currents and particle precipitation.

It is apparent that there is no generally accepted explanation of ionospheric heating during magnetic storms.
FIGURE CAPTIONS

Figure 1: Densities and temperatures derived from the drag of the Explorer 9 satellite compared with the geomagnetic index \( a_p \) and the 10.7-cm solar flux (from Jacchia, 1964 a).

Figure 2: Atmospheric temperatures derived from the drag of Explorer 9 satellite during a magnetic storm (from Jacchia, 1964 b).

Figure 3: Atmospheric heating \( \Delta T \) as a function of the three-hour geomagnetic indices \( K_p \) and \( a_p \) (from Jacchia, 1966).

Figure 4: The coordinate system used by Francis and Karplus (1960) and Karplus et al. (1962).

Figure 5: The square of the magnitude of the electric field versus altitude above the earth's surface for ordinary incident waves of 1, 5, 10, and 30 radians/second frequency each with an incident
energy flux of $1.0 \times 10^{-3}$ watts/m$^2$ for day sunspot maximum conditions (from Karplus et al., 1962).

Figure 6: The square of the magnitude of the magnetic field versus altitude above the earth's surface for ordinary incident waves of 1, 5, 10, and 30 radians/second frequency each with an incident energy flux of $1.0 \times 10^{-3}$ watts/m$^2$ for day sunspot maximum conditions (from Karplus et al., 1962).

Figure 7: Power dissipation versus altitude above the earth's surface for ordinary incident waves of 1, 5, 10, and 30 radians/second frequency each with an incident energy flux of $1.0 \times 10^{-3}$ watts/m$^2$ for day sunspot maximum conditions (from Karplus et al., 1962).

Figure 8: The amplitude transmission of propagating and evanescent modes versus angular frequency for daytime conditions (from Karplus et al., 1962).
Figure 9: The rubidium vapor magnetometer system.

Figure 10: The frequency response curve for the first channel of rubidium magnetometer system.

Figure 11: The frequency response curve for the second channel of the rubidium magnetometer system.

Figure 12: This diagram illustrates the manner in which the number from the A-D converter is written on magnetic tape by the incremental tape recorder.

Figure 13: The induction coil magnetometer system.

Figure 14: The circuit used to integrate the output of the induction coil.

Figure 15: The frequency response curve of the induction coil magnetometers.

Figure 16: The layout of the magnetic observatory.
Figure 17: The support for the rubidium oscillator.

Figure 18: A photograph of the rubidium oscillator and its plexiglass support.

Figure 19: A photograph of the building used to house the equipment. The concrete amplifier pier is behind the building.

Figure 20: A photograph taken inside the building showing the magnetic tape recorder, the A-D converter, the rubidium magnetometer readout unit, and the strip chart recorder.

Figures 21 and 22: Typical examples of the power spectra obtained from data recorded during magnetic storms.

Figure 23: A typical power spectrum computed from data recorded during quiet magnetic conditions.

Figures 24 - 28: Power dissipation versus altitude above the earth's surface for extraordinary incident waves, day ionosphere conditions.
Figure 29: The total power absorbed by the ionosphere above a certain altitude due to extraordinary incident waves, day ionospheric conditions.

Figures 30 - 34: Power dissipation versus altitude above the earth's surface for extraordinary incident waves, night ionospheric conditions.

Figure 35: The total power absorbed by the ionosphere above a certain altitude due to extraordinary incident waves, night ionosphere conditions.
FIGURE 1

EXPLORER IX

LOG DENSITY
REDUCED TO A STANDARD
HEIGHT OF 730 km.

EXOSPHERIC TEMPERATURE

GEOMAGNETIC INDEX

SOLAR FLUX AT 10.7 cm

1961
FIGURE 2
FIGURE 9
<table>
<thead>
<tr>
<th>Parity</th>
<th>Parity</th>
<th>Channel 7</th>
</tr>
</thead>
<tbody>
<tr>
<td>± Sign</td>
<td>2^5</td>
<td>Channel 6</td>
</tr>
<tr>
<td>2^10</td>
<td>2^4</td>
<td>Channel 5</td>
</tr>
<tr>
<td>2^9</td>
<td>2^3</td>
<td>Channel 4</td>
</tr>
<tr>
<td>2^8</td>
<td>2^2</td>
<td>Channel 3</td>
</tr>
<tr>
<td>2^7</td>
<td>2^1</td>
<td>Channel 2</td>
</tr>
<tr>
<td>2^6</td>
<td>2^0</td>
<td>Channel 1</td>
</tr>
</tbody>
</table>

NOT TO SCALE
20/027
FIGURE 14

$R_1 = 1.0 \times 10^4 \, \Omega$

$R_2 = 1.0 \times 10^6 \, \Omega$

$C = 16 \, \mu \text{FARADS}$

A = OPERATIONAL AMPLIFIER
FIGURE 15

FIELD STRENGTH (B/B_{MAX})
NORMALIZED MAGNETIC FREQUENCY (cps)
$\omega = 1$

DAY

EXTRAORDINARY WAVE

POWER DISSIPATION (WATTS/\(M^3\))

\[10^{-15} \quad 10^{-14} \quad 10^{-13}\]

ALTITUDE (Km)

0 \quad 100 \quad 200 \quad 300 \quad 400 \quad 500 \quad 600

FIGURE 24
\( \omega = 10 \)

DAY

EXTRAODINARY WAVE

FIGURE 25
$\omega = 20$

DAY

EXTRAORDINARY WAVE

POWER DISSIPATION (WATTS/M$^3$)

ALTITUDE (Km)

FIGURE 26
FIGURE 27
$\omega = 40$

DAY

EXTRAORDINARY WAVE

POWER DISSIPATION (WATTS/M$^3$)

ALTITUDE (Km)

FIGURE 28
\[ \omega = 1 \text{ is approximately } 10^{-3} \text{ that of } \omega = 10. \]

FIGURE 29
\[ \omega = 1 \]

NIGHT

EXTRAORDINARY WAVE

POWER DISSIPATION (WATTS/M^3)

ALTITUDE (Km)

FIGURE 30
\( \omega = 10 \)

**NIGHT EXTRAORDINARY WAVE**

**FIGURE 31**

- **Y-axis:** POWER DISSIPATION (WATTS/M\(^3\))
- **X-axis:** ALTITUDE (Km)
ω = 30
NIGHT
EXTRAORDINARY WAVE

FIGURE 33
$\omega = 40$

NIGHT
EXTRAORDINARY WAVE

POWER DISSIPATION (WATTS/M$^3$)

$10^{-13}$

$10^{14}$

$10^{15}$

ALTITUDE (Km)

FIGURE 34
NIGHT EXTRAORDINARY WAVE

Power Absorbed (Watts/m²-column)

Altitude (Km)
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Jacchia, L. G., J. Slowey, and F. Verniani, Geomagnetic
perturbations and upper-atmosphere heating, J.

Jacobs, R. L., Atmospheric density derived from the drag
of eleven low-altitude satellites, J. Geophys. Res.,
72, 1571-1581, 1967.

Jacobs, J. A., Y. Kato, S. Matsushita, and V. A. Troitskaya,
Classification of geomagnetic micropulsations, J.

Jacobs, J. A., and K. O. Westphal, Geomagnetic micro-
pulsations, Physics and Chemistry of the Earth, 5,

Johnson, F. S., Temperature distribution of the ionosphere
under control of thermal conductivity, J. Geophys.

Judge, D. L., and P. J. Coleman, Jr., Observations of low-
frequency hydromagnetic waves in the distant geo-
magnetic field: Explorer 6, J. Geophys. Res., 67,
5071-5089, 1962.

Karplus, R., W. E. Francis, and A. J. Dragt, The attenuation
of hydromagnetic waves in the ionosphere, Planetary

Keefe, T. J. and C. Polk, Results of simultaneous ELF
measurements at Brannenburg (Germany) and Kingston,
R. I., National Bureau of Standards Report No. 8815,
Contribution 12, 1964.

King-Hele, D. G., Theory of Satellite Orbits in an Atmosphere,

King-Hele, D. G., The contraction of satellite orbits under
the influence of air drag, Proc Roy. Soc. A267, 541-
557, 1961b.

King-Hele, D. G., Decrease in upper atmosphere density since
the sunspot maximum of 1957-1958, Nature, 198, 832-843,
1963.

Kozai, Y., The motion of a close earth satellite, Astron. J.,


Topley, L. R., A study of hydromagnetic emissions, Lockheed Missiles and Space Division Rept. AF 19(604)-5906, 1961.
