3.1 Introduction

The world-wide characteristics of upper atmospheric storms are obtained from a synoptic study of local observations at different locations on the globe. Such observations may include variations in geomagnetic field and electron density distributions, occurrence of spread-F and sporadic E, visual aurora and red arcs, airglow, and radio absorption. With the advent of methods for topside probing, variations in ionic composition, electron and ion density, and temperature in the higher regions have also been observed.

The disturbed value of a geophysical parameter $X$ is defined as

$$\Delta X = X_{\text{obs}} - X_\text{q}$$

where

- $X_{\text{obs}}$ is the observed value of $X$ during a disturbance
- $X_\text{q}$ is the quiet day value of $X$ which includes the regular $S_\text{q}$ and $L$ variations.

Following Chapman's (1936) method for the study of the geomagnetic field components, a widely-used method for studying a disturbed parameter $\Delta X$ is to classify it into two parts -
1) The $D_{st}$ storm-time variation which is the average value of $\Delta X$ around a circle of constant latitude.

2) The DS disturbance local-time inequality or the SD solar daily disturbance.

The deviation $\Delta X$ is thus a function of geomagnetic latitude $\lambda$, the geographic longitude $\theta$, and storm-time $T$, and can be written as $\Delta X (\lambda, \theta, T)$. It is a component of the disturbance field $D$ which can be written as

$$D = D_{st} + DS$$

in which $\bar{D}$ is the average of $D$ over $24$ hours. $D_{st}$ being a function of $\lambda$, and $T$, and independent of longitude, can be expressed as

$$D_{st} (\lambda, T) = \frac{1}{2\pi} \int_{0}^{2\pi} D(\lambda, \theta, T) d\theta$$

DS being a function of local time is expressed as

$$DS (\lambda, \theta, T) = D (\lambda, \theta, T) - D_{st} (\lambda, T)$$

When the parameter considered is the critical frequency of the $F_2$ layer, these components become $D_{st}(f_0 F_2)$, $DS (f_0 F_2)$, and $SD (f_0 F_2)$ respectively.

(a) $D_{st} (f_0 F_2)$ studies

Storm-time variations are computed by taking the hour
nearest to the time of sudden commencement of a magnetic storm (as deduced from magnetic records) as the zero-hour, irrespective of its local hour value. The value of $f_{o F_2}$ at this hour divided by the monthly median value of $f_{o F_2}$ for the same hour, gives the ratio $f_{o F_2 (\text{storm})}/f_{o F_2 (\text{median})}$. This ratio is an estimate of the percentage change in $f_{o F_2}$ for that hour, because

$$\frac{f_{o F_2 (\text{storm})} - f_{o F_2 (\text{median})}}{f_{o F_2 (\text{median})}} = \frac{f_{o F_2 (\text{storm})}}{f_{o F_2 (\text{median})}} - 1$$

Some workers have simply used the quantity $f_{o F_2 (\text{storm})}$ minus $f_{o F_2 (\text{median})}$ as an estimate of the degree of disturbance, but it is better to divide it by $f_{o F_2 (\text{median})}$, as then the dependence of disturbance on local time is also taken into account, and eliminated. The ratio $f_{o F_2 (\text{storm})}/f_{o F_2 (\text{median})}$ is written down for each of the hours following the zero-hour of the storm (these may be 48, 72 or 96 hours as required), and if so desired, for a few hours before the zero-hour of the storm, to obtain control day behaviour. When plotted, these values give the storm-time variation of $f_{o F_2}$ for the particular storm.

If the statistical behaviour of a large number of storms is required, the ratios for say the 72 hours following the zero-hour for each storm are written out on consecutive lines, with the zero-hour ratios lying one
below the other. The average of each column is then taken, and the 72 values thus obtained, represent the storm-time variation of $f_o F_2$ for the particular group of storms. This variation is independent of local time, as it is assumed that there is a uniform distribution of the hours of storm-commencement over all hours of the day. In reality however, some amount of diurnal component remains in the storm-time variation.

(b) $DS (f_o F_2)$ - disturbance local time inequality

The $DS (f_o F_2)$ for the same group of storms is computed by re-arranging the above data in a different way. The final row of 72 average ratios is written below the 72 values for each individual storm, and the difference for each column is taken. If the storm considered starts at 06 hr. local time, then the first 24 differences obtained correspond to the local times 06 hr. through 00 hr. to 05 hr., the next 24 differences correspond to the next 06 hr. through 00 hr. to 05 hr., and so on.

Each set of 72 differences is now in terms of local time, beginning at the hour of commencement of the individual storm considered. They are written in consecutive rows starting from the column containing the appropriate local time. The average of each column is obtained. The first 24 values represent $DS (1)$ which is
the disturbance local time inequality for the first day of
the group of storms, the next $2^h$ values represent $DS' = (2')$, 
the same parameter for the second day of storms, and so on.

This DS component depends on local time, and can be 
regarded as constant in form and position (but not in 
intensity) as viewed from the sun.

(c) $SD\left(f_0 F_2\right)$ - solar daily disturbance

$SD\left(f_0 F_2\right)$ does not differ much from $DS\left(f_0 F_2\right)$ save in 
the method of computation. Based on the $K_p$ magnetic index 
values of the different days, the IAGA publishes the dates 
of the 5 quietest days of the month (the International Quiet Days denoted by $Q$) and of the 5 most disturbed days 
of the month (the International Disturbed Days denoted by $D$). The hourly values of $f_0 F_2$ averaged for the 5$Q$ days are 
denoted by $S_q\left(f_0 F_2\right)$ and the hourly values of $f_0 F_2$ 
averaged for the 5$D$ days are denoted by $S_d\left(f_0 F_2\right)$.

The solar daily disturbance variation $SD\left(f_0 F_2\right)$ 
(also called disturbance daily variation) is then given as

$$SD\left(f_0 F_2\right) = S_d\left(f_0 F_2\right) - S_q\left(f_0 F_2\right)$$

It consists of $2^h$ hourly values ranging from 00 hr. 
to $2^h$ hr. and is clearly a function of local time.
(d) N-h studies - Electron density distributions at true heights.

The above methods of computing $D_{st}$ ($f_{o}F_2$) and $DS$ ($f_{o}F_2$) are applied to groups of storms. Instead of such statistical studies, some workers study the detailed characteristics of individual storms (Somayajulu 1963, Matsushita 1963). This approach leads to essentially similar conclusions in most cases, and the separation into $D_{st}$ and SD components can be made just as clearly. Each storm, however, has its own individuality and characteristics, and it is seldom that any two storms resemble each other exactly.

$f_{o}F_2$ studies refer only to changes in $N_{max}F_2$ the peak electron density. During a storm, there are significant changes in the altitude-distribution of $F$-region ionisation, and in the earlier days, these were studied through the variation of the virtual height of reflection of ionosonde echoes. The method was rather unsatisfactory, as the virtual height is strongly influenced by group retardation of the sounding pulses, in the lower ionospheric regions lying below the $F_2$ peak.

The computation of electron densities at true heights from ionograms is done by either Model or Integral methods, (Ratcliffe 1951; Schmerling 1958; Budden 1954;
Thomas and Vickers (1958) and is a long and complicated process. Recently, the development of electronic computer techniques for true-height analysis of ionograms has quickened and simplified such computations, and has helped in the investigation of changes in the altitude-distribution of electron densities during storms. Such analyses confirm that during storms, large changes in electron density do occur, especially above heights of 180 km - 200 km.

3.2 Review of earlier work on the bottomside F-region

The earliest reports of geomagnetic storm effects on the F-region were made by Appleton and Ingram (1935) when they found that at high-latitude stations Slough and Tromsø, $f_o F_2$ was reduced. Berkner et al., (1939), and Berkner and Seaton (1940), found that for an equatorial station Huancayo, the opposite was observed, the effect of storms being to increase $f_o F_2$.

Martyn (1953 b) from a study of mid-latitude stations Washington, Canberra, and Watheroo, found that storms generally had the effect of depressing $f_o F_2$ in summer and the equinoxes, but the effect was much smaller in winter; in fact $f_o F_2$ seemed to be slightly increased in winter. Other workers have shown this seasonal variation of F-region storm-effects (Rastogi 1962, Skinner and Wright 1955) by correlating the variation of $(f_o F_2)^2$ i.e. $N_{max} F_2$ with $K_p$ index for summer and winter separately. $F_2$ layer disturbances seem to lag behind magnetic disturbances by the order of a day (J.W. King, 1961).
Systematic morphological studies of F-region disturbances were made by Appleton and Piggott (1952), Martyn (1955), Obayashi (1952, 1954, 1959) and others. A detailed study of ionospheric disturbances for Canadian stations was done by Meek (1952). Matsushita (1959), studied $D_{st}$ ($f_o F_2$) for 109 storms at 38 observatories situated all over the world, by grouping them into eight geomagnetic latitude zones, and classifying the storms into strong and weak ones. These results are shown in Fig. 3.1 (Pg. 68 (a)) and they confirm the general characteristics obtained by earlier workers.

1) In high latitudes, a marked depression of $f_o F_2$ is seen during storms. The largest effect is centred near the auroral zone, and it has a pronounced diurnal variation.

2) In equatorial and low latitudes there is generally an increase of $f_o F_2$ accompanying a storm. A depression of $f_o F_2$ is a rare occurrence, which happens only during a very severe storm.

3) At mid-latitudes, a decrease of $f_o F_2$ is definitely observed during summer and equinoxes, but in winter there is occasionally an increase of $f_o F_2$.

The local time of commencement of a storm is found to have an effect on the behaviour exhibited by $f_o F_2$ (Appleton and Piggott 1952). For mid-latitudes,
Fig. 3.1 $D_{st}$ ($\Delta H$) variations for strong and weak storms for the geomagnetic latitude zones
(1) $60^\circ - 55^\circ$ (2) $55^\circ - 50^\circ$ (3) $50^\circ - 45^\circ$ (4) $45^\circ - 40^\circ$
(5) $40^\circ - 30^\circ$ (6) $30^\circ - 20^\circ$ (7) $20^\circ - 10^\circ$ (8) $10^\circ$ to $(-10^\circ)$
(Matsushita 1959).
Thomas and Venables (1966) find that the major changes in $f_0F_2$, depend on the local time of the main phase of the magnetic storm; if this be in the night hours, $f_0F_2$ shows an immediate depression, but for a daytime main phase, this sudden drop in $f_0F_2$ is not seen.

One of the storm-effects seen in ionograms is the visibility of a clear $F_1$ 'cusp', often at times when it is not normally seen. This is believed to be an increase of the parameter.

$$G = \frac{\beta^2}{\alpha q}$$

where \(\beta\) - attachment coefficient, \(\alpha\) - recombination coefficient, and \(q\) - production rate

This is probably due to an increase in the loss coefficient \(\beta\), and G.A.M. King (1962 a; 1967) finds that $G$ generally increases when the magnetic disturbance figure $K$ increases.

During storms the $hF$ curves of the $F_2$ layer are greatly changed, and the great increase in virtual height, often up to 600 km was previously considered to be real. With the increasing use of techniques for true height reduction of ionograms, it is becoming increasingly clear
(Thomas and Robbins 1958; Becker 1964) that these fantastic increases of virtual height are due to group retardation of the probing radio waves in the lower ionospheric layers, and that the changes in storm-time $h_{\text{max}}^{F_2}$ depart by only ± 30 km from quiet days.

Matsushita (1963) and Somayajulu (1963) have analysed the storm-effects at several American stations by using hourly N-h data. Their main conclusions are:

1) Most changes in N-h distribution occur in the region above 180 km in summer and the equinoxes, seldom below that. In winter the changes are predominant above 300 km.

2) The sub-peak total electron content $N_T$ behaves in the same way as $f_{\sigma F_2}$.

3) The peak height of the $F_2$ layer ($h_{\text{max}}^{F_2}$) and its semi-thickness ($y_{\text{m}}^{F_2}$) generally increase during a storm both in summer and in winter, though an occasional decrease occurs in winter.

It is believed that the electromagnetic radiations during a solar flare do not affect the F-region as much as the corpuscular radiation or storm effects, but a few isolated cases of photon effects have been noted on the F-region. Sato (1957 b) and Brown and Wynne (1967) found a reduction
of about 10% in $N_{F1}$ following a flare. Dieminger et al. (1950) reported such a case for Lindau on November 19, 1949; Shapley and Knecht for Okinawa, and Bhargava and Subrahmanyan (1962) for Kodaikanal on February 23, 1956, and Knecht and McDufffe on November 12th, 15th, 1960. The height of the $F_2$ peak in all these cases was below a critical height i.e. below 300 km. It is interesting to note that all these cases occurred in the local winter months. Kotadia (1966) pointed out that daytime solar flares caused a marked increase in $f_{o}F_2$ in winter.

Artificial nuclear explosions have offered another means of studying disturbances in the F-region (Kotadia 1967). The tests in the Pacific, and at Novaya Zemlya showed the effect of travelling pressure waves on the F-region (Obayashi 1962; Oksman and Kataja 1962). The pressure wave caused an anomalously large increase in electron density followed by fluctuations of the same, as it moved from near to distant stations with an average speed of 400 m/sec.

3.3 Review of topside studies of the F-region

The topside is the name given to that part of the ionosphere which lies above the $h_{max}F_2$ level. Till 1958, ground-based ionosondes had supplied most of the world-wide electron density data, while rocket measurements had yielded most of the information about ionic composition, plasma temperatures, and ionising radiations, all upto the
level of $h_{max} F_2$. The advent of satellites and other methods of directly probing the higher regions of the ionosphere since the IGY, has led to much progress in our knowledge of the topside on a world-wide basis.

(a) **Methods of probing the topside ionosphere**

A large variety of in situ experiments can be put aboard a satellite, and these relay direct measurements of the parameters concerned, to ground.

The incoherent back-scatter radar technique is another way of studying the topside as well as the bottom-side. It is capable of estimating ionisation densities and temperatures throughout the ionosphere to heights of over 1000 km. Its working principle is that the total power scattered back by electrons in a pulse volume at a particular height is proportional to the number of electrons in that volume. The characteristic width of the spectrum observed depends on the ratio $T_e/T_i$ where $T_e$ is electron temperature, and $T_i$ is ion temperature.

Three stations exist in the 70°W longitude region - Jicamarca at equatorial latitudes, Arecibo at 30°N, and Millstone Hill at 55°N. Three others are Prince Albert in Canada, Nancay in France, and Malvern in England. These six stations have yielded valuable information on the
time-variations of electron and ion densities, and temperatures.

Faraday rotation and Doppler shift of signals reflected from satellites or the moon (Klobuchar et al., 1964) and Whistler studies (Carpenter 1966) are two other methods of inferring electron densities in the topside.

(b) The topside during quiet conditions

The main ionic constituents in the topside are $O^+$, $He^+$, $H^+$ and a small amount of $N^+$. The gases separate out under diffusive equilibrium, according to their atomic weights, with the heavier nitrogen and oxygen below, then helium to form the heliosphere, and hydrogen above to form the protonosphere. Their relative concentrations depend on altitude, latitude, local time, season, and level of solar activity. Under conditions of sunspot minimum, $He^+$ is found to be a minor constituent (less than 30%) at all times (Carlson and Gordon 1966) with its maximum occurrence around 500 km by night, and around 1000 km by day. The transition from $O^+$ to predominantly $H^+$ occurs around 700 km at night, and 900 km by day (Gordon 1966). As in the lower ionosphere, sudden and rapid changes occur in the topside at sunrise and sunset. Satellite studies show that above 800 km at night, a deep equatorial trough in $He^+$ is formed with a corresponding rise in $O^+$, possibly arising from the charge-exchange
In the height range of about 500 km, the geomagnetic anomaly shows up in the daytime latitudinal variation of $O^+$, with maxima occurring at ± 20° geomagnetic latitude (Chandra and Goldberg 1964; Hanson and Moffett 1966). This is again an instance of geomagnetic and gravitational effects on ionospheric plasma.

Satellite and back-scatter studies have yielded useful information on ion temperature ($T_i$), electron temperature ($T_e$), and neutral temperature ($T_n$) in the higher regions. In the daytime $T_i$ increases from around 1000°K near the $F_2$ peak (about 400 km.) to a few thousand degrees Kelvin at about 800 km and remains isothermal to beyond 1000 km. Night-time values are almost half the daytime ones. $T_e \approx 3T_i$ near the $F_2$ peak, but approaches $T_i$ at greater heights. The daytime pattern seems fairly stable with time, but the night-time pattern seems to vary. Seasonal changes are not marked (Gordon 1966). Both $T_e$ and $T_i$ increase polewards from the equator, with maxima between 40° and 50° geomagnetic latitude (Brace et al., 1967).

(c) The topside during disturbed conditions

Studies of the topside show that magnetic disturbances can cause changes in the following parameters:
a) $T_e$ and $T_i$, and neutral temperature and density
b) concentration of different ionic species
c) spatial distribution of electron density

All these vary with altitude, latitude, local time, and storm-time, and often the large changes in the topside do not manifest themselves at the level of $h_{max} F_2$. Very often too, results of the topside obtained by different methods do not tally. A good review of the topside during geomagnetic storms has been given by Warren (1969).

Studies of air-drag force on satellites show that the neutral air density increases with magnetic disturbance, and is strongly latitude-dependent, being especially high at high latitudes (Jacchia 1961; Newton et al., 1965; Schilling and Whitney 1959). Groves (1961) expressed the variation of neutral density with $K_p$ as

$$\delta = \delta_0 \left( 1 + \alpha K_p \right)$$

with $\alpha = 0.2$ for low latitudes

and $\alpha = 0.6$ for high latitudes

Thomson scatter studies at Millstone Hill, Arecibo, and Prince Albert, show in general that not taking diurnal variations into account:
$T_e$ (disturbed) $\geq$ $T_e$ (quiet) by 300°K to 600°K
but $T_1$ (disturbed) $\geq$ $T_1$ (quiet) by less (Evans 1965; Rao 1968)

Topside sounder studies agree with backscatter radar results on temperatures, and an increase of 4% in temperature per unit increase of $K_p$ was observed (Watt 1966). Jacchia (1964) showed that the effect of magnetic storms on temperature can be expressed as

$$\Delta T = 1.2 a_p \, ^0K$$

where $a_p$ - 3-hour geomagnetic planetary index
$\Delta T$ - change in temperature

Direct studies by satellite-borne probes however give different results. Willmore (1965) using the Langmuir probe on Aeriel I, invariably found a decrease in $T_e$ during disturbances both by day and night. Reddy et al., (1967) too found similar results for the mid-latitudes from their electrostatic probe on Tiros VII. Both of them found an increase in $N_e$ occurring simultaneously with a drop in $T_e$.

Mariner II results (Coleman et al., 1962) showed that both electron density and magnetic field intensity in the topside increase during the initial phase of a storm. Explorer 26 (Cahill et al., 1966) found that the particles responsible for the main phase appeared over the equator
between $L = 2$ and $L = 6$ in the late noon and evening latitudes. Gledhill et al., (1967) found a strong correlation between particle precipitation near the magnetic shell $L = L_0$ and the occurrence of ionospheric disturbances. Whistler studies by Carpenter (1963) showed that the "knee" in electron density moved to lower $L$ values with increasing $K_p$, indicating that $N_e$ at very high altitudes decreases.

From Faraday rotation studies of radio signals reflected from the moon, Klobuchar et al., (1964) found that the total electron content $N_T$ on a disturbed day was only half that on a quiet day. Millman (1964) reflected radio signals from the satellite Echo I, to find that while $N_T$ increased with $K_p$ by day, it did not do so at night. Hibberd and Ross (1967) used the differential Doppler technique on radio transmissions from Transit IV-A to find that at mid-latitudes, the slab-thickness on storm days was larger than on a quiet day. Titheridge and Andrews (1967) found that the scale height gradient $dH/dh$ increased greatly during the positive phase of a storm, then sharply decreased. This decrease in $dH/dh$ was interpreted as an increase in the transition height at which $O^+$ is replaced by lighter ions. Watt (1966) could not obtain any correlation between scale height and $K_p$ by day, but at night the $O^+ - H^+$ transition level rose at the rate of 30 km/unit increase of $K_p$. 
To summarise topside results:

At the equator

King et al., (1967) found that levels near $N_{\text{max}} F_2$ are more sensitive to storm changes than higher levels. In the topside too, the equatorial anomaly vanishes during disturbances.

At mid-latitudes

For very severe storms - both $N_T$, $N_{\text{max}} F_2$ decrease
For moderate storms - $N_{\text{max}} F_2$ decreases but $N_T$ increases
For mild disturbances - $N_T$ increases but $T_e$ decreases.

At polar latitudes

Nishida (1967) finds amidst irregularities, certain features of permanence in topside electron density such as

1. The polar peak
2. the auroral peak
3. the mid-latitude trough and evening increase of ionisation. Storms are found to enhance the peaks and deepen the troughs. Above 500 km, both $N_e$ and $T_e$ increase (Ondoh 1967, Reddy et al., 1967).

Topside studies have decisively shown that the depression phase of F-region storms at middle and high latitudes represents an actual decrease in the total number of electrons in the ionosphere, and is not due to a mere spatial re-distribution of the ionisation present.