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STUDY OF THE EARTH’S UPPER ATMOSPHERE USING VLF AND SHF RADIO WAVE TECHNIQUES

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Dedication

To my parents
Late Mr. Chandra Kishore and Mrs. Harshila Kishore
and to my brother Anish Pavan Kishore
Declaration

I, Amol Nitin Kishore, hereby declare that this is the result of my own work and to the best of my knowledge and belief contains no material previously published or substantially overlaps with materials submitted for the award of any other diploma or degree at any institution of higher learning, except where due acknowledgment is made.

Amol Nitin Kishore
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Abstract

The study of the Earth’s upper atmosphere using VLF and SHF radio waves is a well-developed technique. In the VLF radio wave technique, we measure lightning generated electromagnetic signals to study the nighttime lower (E-region) ionosphere. The SHF technique utilizes amplitude measurement of a geostationary satellite beacon signal at 3.925 GHz to study ionospheric scintillations at Suva.

Lightning discharges radiate a wide band of electromagnetic spectrum with the peak spectral density at around 10 kHz. These impulse-like signals known as radio atmospherics or sferics can propagate to large distances by multiple reflections between the ground and the lower ionosphere. Sferics radiated from lightning and received at long distances from the source stroke contain a great deal of information about the state of the ionosphere along the propagation path. At nights sferics that travel large distances in the Earth-ionosphere waveguide mode are appreciably dispersed near the cutoff frequency and appear as tweeks in a dynamic spectrogram. In this work tweek sferics have been used to estimate the nighttime height of the ionosphere and the total propagation distance to the source discharge. These sferics and other associated lightning generated signals such as whistlers have been received using a system installed in collaboration with the World Wide Lightning Location Network (WWLLN) The WWLLN uses the difference in the Time of Group Arrival (TOGA) of sferics from a network of similar receiver systems to locate lightning discharges to ground. A statistical analysis of WWLLN lightning detections over Viti Levu and Vanua Levu during 2003 and 2004 is also presented. WWLLN detections are compared with that obtained by the LIS detections. A whistler received at the station has been analyzed to obtain magnetospheric parameters and possible propagation mechanisms to our low latitude station are discussed. The
amplitude variations of navigational transmitter signals from NWC at 19.8 kHz and NPM at 21.4 kHz are presented. The sunset and sunrise effects on these signals are also presented.

We measured a beacon signal at 3.925 GHz from geostationary satellite Intelsat (701) situated at 180° E to study ionospheric scintillations at Suva. Scintillation is the rapid change in the phase and amplitude of trans-ionospheric radio waves as they traverse through ionization irregularities in the ionosphere. The study of ionospheric scintillation derives its interests mainly from its impacts on satellite communication and navigation communication systems. Our measurements were conducted during a low solar activity period and showed very low scintillation activity. The scintillations observed were mostly weak with $S_4$ index < 0.3. Interestingly daytime scintillations were found to be more prominent than the nighttime scintillations indicating the prominence of Sporadic-E irregularities in the ionosphere during low solar activity.
Chapter 1

Introduction

Very Low Frequency (VLF 3 – 30 kHz) technique is one of the cost-effective ground based methods that is successfully used to explore the properties of the Earth-ionosphere waveguide as well as the upper atmosphere. The upper atmosphere includes the ionosphere and magnetosphere. Lightning discharges are the natural powerful transmitters of a wide band of electromagnetic spectrum with peak spectral density at around 10 kHz (Pathak et al., 1982). Lightning energy impulse-like signals called sferics (radio atmospherics) can propagate large distances by multiple reflections through a waveguide formed by the Earth surface and the lower ionosphere. Radio atmospherics can be easily detected over distances of several thousand km. The received sferics contain information about the lower ionosphere along its propagation path and therefore is a useful tool to study the lower ionosphere. A small part of the radiated energy from a lightning discharge may propagate through the ionosphere and in the magnetosphere from one hemisphere to the other along geomagnetic field lines. These are referred as whistlers at the receiver. Whistlers have been extensively used as diagnostic tool for studying structure and dynamics of the ionosphere and magnetosphere.

Measurement of Super High Frequency (SHF 3-30 GHz) scintillation is another cost effective ground based radio wave technique used to study the ionospheric irregularities. Scintillation is a rapid change in the phase and amplitude of trans-ionospheric radio waves as they traverse through ionization irregularities in the ionosphere. Scintillations adversely affect satellite communication links in the frequency range from VHF to the S-band (Hajkowicz and Deardan, 1988). Trans-ionospheric effects of scintillation have been observed on frequencies from 20 MHz
to ~ 10 GHz (Ippolito, 1986). Beacon signal from the geostationary satellite Intelsat (701) at 3.925 GHz is used to study the ionospheric scintillation at Suva.

1.1 Ionosphere

The ionosphere is that part of the Earth’s upper atmosphere which begins with the base near the stratopause and extends upwards overlapping the exosphere. Its extension is taken from 50 km to about 400 km. Based on electron density profiles, the ionosphere is divided into D, E and F-regions which occupy the heights varying approximately from 50 – 90 km, 90 – 150 km and 150 – 400 km respectively. F-layer can be further subdivided into F1 and F2 layers during the daytime. Ionosphere can be characterized as a body of ionization in which ions and electrons coexist. Ionization depends primarily on the solar radiation and the solar activity. The electrons present in sufficient quantity can affect propagation of radio waves. Peak densities in the ionosphere vary greatly with time (diurnal, seasonal and sunspot cycle), geographical location (polar, auroral zones, mid latitudes, and equatorial regions), and certain solar related ionospheric disturbances. VLF and SHF radio wave techniques are widely used to study the ionosphere.

The presence of ions and electrons make the ionospheric medium a conductor, which reflects Extremely Low Frequency (ELF 3 – 3000 Hz) and VLF waves propagating in the Earth-ionosphere waveguide beyond its cutoff frequency. Different techniques are used to probe the different regions of the ionosphere. The swept-frequency pulse sounding known as the ionosonde was the first technique used which is still used today. The ionosonde sweeps across a series of frequency pulses that are reflected by the different electron density layers in the ionosphere. The maximum frequency (also known as the critical frequency) reflected by the E-layer is known as $f_{oE}$. The maximum frequency reflected by F1 and F2 layers are $f_{oF1}$ and $f_{oF2}$ respectively. The $f_{oE}$ and $f_{oF}$s are related to the maximum electron density of the E and F-layers, respectively. These are known as $N_{mE}$, $N_{mF1}$ and $N_{mF2}$ respectively. The reflection height is then calculated in terms of an equivalent height known as virtual height, $h'$, which is the height from which the pulse traveling at free space velocity would have reflected. With time, $t$ and $h'$ measured as a function of frequency, a
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A graph of $h'$ against $f$ called an ionogram is produced. The critical frequencies in the ionogram appear as discontinuities normally at two or more frequencies. Ionosondes can probe the E and F-regions but not the D-region, because the lowest operating frequency of the ionosonde passes through this region due to the low electron density in the D-region. The Incoherent Scatter Radar (ISR) ionospheric measurement is another technique used to probe the ionosphere, even above the $F_2$ region electron density maximum. However, the ISR returns are weak, and require high-power transmitter, large antenna, and sophisticated signal processing making such a facility large and expensive.

The use of ELF and VLF radio waves to probe the lower ionosphere is an inexpensive technique to study the lower region of the ionosphere. The ELF and VLF waves are almost completely reflected by the D and E-regions of the ionosphere making it a useful tool in probing these regions. The E-region is essentially the upper boundary of the nighttime Earth-ionosphere waveguide, and the sferic characteristics contain information about this boundary.

1.2 Radio Atmospherics

ELF and VLF parts of radio atmospherics have been studied for many years. Burton and Broadman (1933) in the early days of sferic investigation discovered two distinct VLF emissions. They were classified according to their sound when played on a speaker, as *swishes* and *tweeks*. The dynamic spectral analysis of the sferics showed the difference in the dispersion characteristics between the two signals. Burton and Broadman reported that the long tail near $\sim 1.7$ kHz in tweeks were caused by the height of the reflecting ionospheric layer indicating the cutoff frequency for the first waveguide mode. The swishes were what are now referred to as *whistlers*, which were reported earlier by Barkhausen (1930) and Eckersley (1925). Tweeks are produced when sferics propagate to large distances by multiple reflections in the Earth-ionosphere waveguide. Figure 1.1 shows a schematic propagation mechanism of tweeks. Storey (1953) proposed that whistlers are the VLF signals radiated by lightning that have propagated over extremely large distances through the Earth’s ionosphere and magnetosphere (see figure 1.1). Whistler emissions propagate in the field-aligned ducts of enhanced ionization along the Earth’s magnetic field lines.
An ideal Earth-ionosphere waveguide is generally assumed to behave like a perfectly conducting parallel plate waveguide. The radiation from the lightning discharges can be approximated as that due to a vertical dipole close to the surface of the Earth emitting transverse electric (TE), transverse magnetic (TM) and transverse electromagnetic (TEM) modes. These modes are composed of a sequence of independent field structures, which propagate with different velocities. The TE and TM modes are completely defined by their cutoff frequency, \( f_{cn} = \frac{nc}{2h} \), where \( n \) is the mode number, \( c \) speed of light in free space and \( h \) is the reflection height of the ionosphere. For these two modes if \( f > f_{cn} \) then the modes centered at \( f \) propagates with a group velocity \( v_{gn} = c \sqrt{1 - \frac{f_{cn}^2}{f^2}} \) which approaches zero as \( f \) approaches \( f_{cn} \). If \( f < f_{cn} \) then the modes are called evanescent and strongly attenuate with distance from the source in the waveguide. The TEM is the single mode with no cutoff frequency. In reality, the electrical properties of the Earth and the ionosphere, and the Earth’s magnetic field cause significant deviations from this ideal model. The major difference between ideal waveguide and the Earth-ionosphere waveguide is that the Earth and the ionospheric boundaries are not perfect conductors. The Earth has a finite conductivity which varies over land and sea. Typical conductivities, \( \sigma \) of land ranges from \( 10^{-4} - 10^{-2} \) mhos/m and for sea is around 5 mhos/m. Similarly, the
ionosphere boundary is far from a perfect conductor. The ionosphere consists of free electrons and ions in sufficient quantities to approximate as a good conductor at ELF and VLF frequencies. In the presence of the Earth’s magnetic field the ionosphere further deviates from a perfect conductor making it an anisotropic medium. As a result, an incident wave linearly polarized with either parallel or perpendicular polarization produces an elliptically polarized reflected wave with components having both parallel and perpendicular polarization. In general, since the wave polarizations are coupled at the ionosphere, pure TM and TE modes cannot exist in the Earth-ionosphere waveguide as they do in an ideal parallel plate waveguide. The propagating energy is assumed to be constituted of superposition of quasi-TM or QTM and quasi-TE or QTE modes. The QTM mode is similar to the TM mode except that it has a small axial magnetic field component resulting in a small additional transverse electric and magnetic component that are not present in the ideal TM mode (Budden, 1961). However, for frequencies less than 15 kHz the lower order QTM modes approximate pure TM modes than the higher order modes (Wood, 2004). We therefore treat the modes in the Earth-ionosphere waveguide as pure TM and TE modes in the analysis of tweeks.

The propagation mode of low latitude whistlers at present remains a controversial subject of discussion. Various workers have supported the ducted propagation of whistlers on the basis of ground based data and direction-finding measurements (Somayajulu and Tantry, 1968; Hayakawa and Ohtsu, 1973; Ondoh et al., 1979; Hayakawa et al., 1985; Ohta et al., 1989), while others indicated non-ducted propagation using satellite measurements and ray tracing computations (Smith and Angerami, 1968; Cerisier, 1973; Singh, 1976; Singh and Hayakawa, 2001). There is however a growing consensus in favour of the non-ducted pro-longitudinal (PL) mode of propagation for nighttime whistlers, and ducted mode of propagation in the presence of equatorial anomaly for daytime whistlers (Singh and Hayakawa, 2001). The PL mode was discovered by Scarabucci (1969) using satellite data with ray tracing analysis of high latitude whistlers. The PL mode is characterized by propagation with the wave normal of the waves inside a characteristic cone relative to the geomagnetic field producing travel times and down-going wave normals that are of purely longitudinal propagation along the field lines (Singh, 1976). The PL mode
propagation was first examined by Singh (1976) for the case of low latitude whistlers. He developed a model that included a maximum of electron density at the equator, and computed ray paths for whistler waves that satisfied all the conditions for the PL-mode. Andrews (1978, 1979) computed ray paths similar to that of Singh (1976) from fixed-frequency transmitter experiments without including horizontal gradients of electron density and found that the dispersion along such paths are constant with frequency even where the paths are not field-aligned. These results indicated that ducted propagation is not necessary for low latitude whistlers since all the characteristics are well produced by non-ducted propagation in the presence of horizontal or vertical gradients in the ionization density (Singh and Hayakawa, 2001).

At low latitudes, whistler data have not been used for determining ionospheric parameters mainly because the propagation paths of low latitude whistlers could not be determined from their dynamic spectrum because the nose frequencies of such whistlers are higher than 100 kHz, well above the pass band of the receiver. Dowden and Allcock (1971) described a method for determining nose frequency ($f_n$) from measurements of group delay at many frequencies along the whistler trace. This method is used for whistlers that do not reach nose frequency and so is suitable for analyzing low latitude whistlers.

In this work we used the technique employed by Prasad (1981) and Kumar et al. (1994) to determine the ionospheric reflection height and the total propagation distance of the tweek sferics. The precision of the measurements in this work is higher because the dynamic spectrograms produced are more resolved. We also analyzed a whistler received at our low latitude station and used the method given by Dowden and Allcock (1971) to determine the nose frequency and hence some of the magnetospheric parameters.

### 1.3 Lightning Measurements

Lightning measurements are extensively used by meteorological offices and the scientific community. They are useful in monitoring damages to power utilities, forest fire management, aeronautical aviation and ionospheric studies. Cloud-to-ground (CG) discharges are a major safety hazard. Return stroke peak currents could range from a few tens of kA to hundreds of kA in intense strokes. Popolansky (1972) using
624 return stroke peak current measurements reported a lower cutoff of 2 kA, including negative and positive ground discharges. The measurements indicated that 95 % of lightning had currents greater than 6 kA, a median of greater than 28 kA, while 1 % had return stroke peak currents greater than 200 kA. The first return stroke current measured at the ground rises to an initial peak of about 30 kA in some microseconds and decays to half-peak value in some tens of microseconds while exhibiting a number of subsidiary peaks associated with the branches (Rakov and Uman, 2003).

Lightning detection systems can be classified into two categories. Space-born lightning detection systems and ground-based radio location systems. Space-born systems like the Optical Transient Detector (OTD) and the Lightning Imaging Sensor (LIS) use optical detection of the luminous radiation produced by lightning flashes from the top of clouds (Christian, 1999; Christian et al., 1999).

The OTD onboard an Earth-orbit satellite provided lightning occurrence data on a global scale. It consisted of a compact combination of lens, detector array and circuitry to convert the electronic output into useful data. The detector resembled a TV camera that detected intensity changes over cloud cover, which indicated the occurrence of lightning. The instrument had a spatial resolution of 10 km and a temporal resolution of 2 ms (Christian et al., 1996). Flashes were determined by comparing the luminance of adjoining frames of OTD data. If the difference is more than a specific threshold value, an event is recorded. One or more adjacent events in the same 2 ms time frame were recorded as a group. One or more groups in a sufficiently small time period was classified as a flash. Lightning flashes were detected during both daytime and nighttime with a detection efficiency ranging from 40 – 65 % depending upon external conditions such as glint and radiation (Christian, 1999). The LIS instrument was launched aboard the Tropical Rainfall Measuring Mission (TRMM) satellite in November 1997. The TRMM orbit has an inclination of 35° to the equator and the LIS instrument can observe lightning activity between latitudes 35° S and 35° N. The lightning sensor consists of a staring imager that uses a wide field-of-view expanded optics lens with a narrow-band pass filter in conjunction with a high speed charge-coupled device detection array (Christian et al., 1999). A Real Time Event Processor (RTEP), inside the electronics unit is used to determine
when a lightning flash occurs. With its ability to detect daytime lightning signals, which are otherwise hard to detect due to background illumination, the LIS system has a detection efficiency of up to 90%. Lightning measurements from the OTD and LIS are not suitable for studying localized lightning incidence because a given location is sensed only for a few minutes (Christian et al., 1999). The LIS and OTD cannot differentiate Intra-cloud (IC) lightnings against CG lightnings. However results from a long-term study give the global distribution of lightning activity and can be used to infer the lightning activity in specific areas.

Ground based lightning location systems using a network of radio receivers are of two main classes. The first class determines the direction of lightning at each receiving station by monitoring the magnetic field associated with the lightning radiation. These directions are then triangulated to locate the lightning. The second class uses the difference in the Times-of-Arrival (TOA) of the lightning sferics at independent pair of receiving stations. The US National Lightning Detection Network (NLDN) uses a combination of magnetic direction finding and measurements of TOA of sferics using a number of stations covering the entire United States. The Lightning Position and Tracking System (LPATS) uses the TOA of the leading edge of the lightning pulse at each station. Only the first few microseconds of the lightning wave train is used to avoid the reflected wave from the ionosphere. The data obtained contains frequency components in the MF band (0.3 – 3 MHz) (Dowden et al., 2002). Such an impulse is not dispersed so its time of arrival is determined to within one microsecond. Since the technique uses the ground wave propagation, which has high attenuation at the high frequencies, it requires a close network of radio receivers with separations of a few hundred km. The NLDN is a dense network of ground stations (~ 100) covering the contiguous US.

A subclass of ground based lightning location system uses the VLF band of the lightning impulse. The sferic radiation at VLF propagates with little attenuation in the Earth-Ionosphere Waveguide (EIWG) to large distances (Mm) by multiple reflections. The EIWG disperses the initial sharp pulse of the lightning stroke into a wave train lasting a few milliseconds. The amplitude of the received wave train rises slowly (a few hundred microsecond) from the noise floor and so there is no sharply defined TOA. The whole wave train is therefore used to get the TOA. There are two
ways of using the whole VLF train. In the first the Arrival Time Difference (ATD) is measured by cross-correlation of the full VLF wave received at a pair of stations. The second measures the rate of change of the sferic phase with respect to frequency at a trigger time to find the time of group arrival (TOGA) at each receiver site. In this work we utilize the second class of the ground based lightning location system. The TOGA theory is given in chapter 5. The lightning location is identified from the difference in TOGA at pairs of stations. A lightning occurrence is confirmed when a minimum of 4 stations record it. At present WWLLN uses widely spaced network of receivers (~24) all over the world to measure the TOGA from the VLF sferics to determine lightning locations on real time.

1.4 Ionospheric Irregularities

Equatorial ionospheric irregularities are the irregular structures in the electron density formed due to various plasma instability mechanisms. The ionospheric irregularities can be divided into E and F-region irregularities.

E-region irregularities are called the sporadic E layer or Es and are formed between 100 km to 200 km. These irregularities are associated with the east-west electric field which drives the electrojet. These are present both during day and night even when the electron density is greatly reduced and are highly elongated along the Earth’s magnetic field. A number of mechanisms have been proposed which describe the formation of Es. The ionospheric wind shear theory is one in which the east-west winds in the E-region result in vertical movement compressing the ions into thin layers of high density ionization forming the Es (Whitehead, 1961). The wind shear theory confirms well with experiments in the mid-latitudes but not at other latitudes (Whitehead, 1989). In 1933 Appleton and Naismith found some correlation with thunderstorms and Es but were later concluded to be inaccurate. The recent studies on the interaction of thunderstorm phenomena like red sprites and blue jets with the E-region have renewed focus on correlation between thunderstorm and Es. However, no strong correlations have yet been identified. The occurrence of Es in the equatorial region is generally accepted to be due to the electrojet currents flowing in the E region (Bhattacharyya, 1991).
The plasma bubbles that occur in the post-sunset equatorial F-region are collectively known as the Equatorial Spread F (ESF). It is generally accepted that these plasma bubbles are generated at the bottom side of the F-region via the Gravitational Rayleigh Taylor (GRT) instability. The plasma density in the F-region increases with altitude up to about 300 km and then decreases. This situation of high-density plasma on top of low-density plasma is unstable (GRT instability). Under certain conditions if the equilibrium is disturbed a lump of low-density layer will rise and cause density irregularities. The growth rate of the instability depends on the density gradient. This gradient becomes steeper after sunset thus the irregular structures usually appear in nights. The equatorial plasma bubbles are generated at the magnetic equator due to the GRT instability and rises to the higher low-latitudes due to the non-linear evolution of the $\mathbf{E} \times \mathbf{B}$ drift, where $\mathbf{E}$ is the ionospheric electric field and $\mathbf{B}$ geomagnetic field in the ionosphere. Both are perpendicular to each other.

Ionization in the upper atmosphere is far from static. It is mainly dependent on solar radiation, geomagnetic activity, and solar cycle. Ionospheric irregularities are the major cause of scintillations on satellite communication signal in the equatorial low latitude and auroral regions and have been an important subject of study for many years. However, significant gaps still exist in understanding the complex occurrence pattern of F-region irregularities especially at GHz frequencies.

1.5 SHF Ionospheric Scintillations

Scintillations are caused by the electron density irregularities in the ionosphere. Radio waves undergo random amplitude and phase fluctuations due to the random fluctuations of the medium refractive index. The amplitude fluctuations are caused by the interference between different components of the wave front while propagating through the ionosphere. The phase fluctuations are due to the changes in the wave path length. Figure 1.2 shows the ionospheric irregularities effects on the satellite signal. Scintillation occurrence in the VHF is very common and severe in the equatorial anomaly region during solar maximum period. However, the magnitude of scintillation decreases with increasing frequency (Basu et al., 2002).
With the increasing demand for Global Positioning Services, scintillation studies in the L-band have increased significantly. Studies have showed that the severity of scintillation depends on the radio signal frequency, location, elevation angle of the line-of-sight to the satellite, time of day, the season of year, the degree of magnetic disturbance and the sunspot cycle phase (Babayev, 2001). Fujita et al. (1978) measured 1.7 GHz scintillation in Japan using a geostationary satellite during May to August 1977 and found enhanced scintillations at night in June with maximum peak-to-peak variation of scintillation of about 2.3 dB. Total Electron Content (TEC) measurements made simultaneously showed that irregular electron density structures play an important role in 1.7 GHz scintillation. During the solar maximum 20 dB fades at 1.5 GHz are often encountered in the anomaly region (19 - 23 ° of geomagnetic equator) after sunset (Babayev, 2001). Scintillation studies in the C-band (used in the present work) are very limited. Dabas et al. (1992) studied equatorial plasma bubble dynamics using scintillation observations from 4 GHz telemetry signals from two geostationary satellites, Indian National Satellites (INSAT) 1B (74° E) and 1C (94° E), in the Indian region. The signals were characterized by weak scintillations during low solar activity periods in the summer. Babayev (2001) reported that peak daily scintillation activity occurs about one hour after sunset and
scintillations on GHz signals of any significant amplitude occur only within approximately $\pm 30^\circ$ of the geomagnetic equator. Our station is located just outside the boundary of the equatorial anomaly region (geomag. Lat. $21.2^\circ$ S) and ionospheric scintillations at 3.925 GHz during a solar minimum period is expected to be weak and rare. There are several proposed scintillation indices, and a widely used index is the $S_4$ index, which is defined as the standard deviation of the received power divided by the mean value of the received signal power. This index determines the severity of amplitude scintillation and measures the time-averaged change in the power of the detected wave (Briggs and Parkin, 1963). The scintillation levels are classified according to the Global Ionospheric Scintillation Model (GISM) scintillation classifications weak ($S_4 \leq 0.25$), moderate ($0.25 < S_4 \leq 0.5$) and strong scintillation ($S_4 > 0.5$) [http://www.projectstation.co.uk/DIFS/glossaryesa.html#_ftn1]

1.6 Objectives

The main objectives of this work are:

i. Study the propagation characteristics of lightning generated ELF-VLF signals in the Earth-ionosphere waveguide to determine ionospheric parameters.

ii. Applications of lightning generated ELF-VLF signals to study lightning discharge occurrence.

iii. Study of ELF-VLF lightning emissions from experimental and theoretical viewpoints.

iv. Record GHz signals from geostationary satellite INTELSAT 701 and analyze scintillation occurrence and spectral properties.

1.7 Thesis Organization

Chapter 2 presents the theoretical understanding of ELF-VLF wave propagation in the Earth/Sea-ionosphere waveguide. The theory of propagation in a parallel plate waveguide is presented first. The waveguide is assumed a parallel plate with perfectly conducting boundaries. For a more realistic waveguide finite conducting boundaries are then considered. The Earth and sea propagation path are treated separately.
Mathematical formulations of ELF-VLF attenuation are made using the Wait’s theory for Earth and sea propagation paths.

Chapter 3 describes the experimental setup and the techniques used for data recording. The lightning detection system and the satellite receiving system are described.

Chapter 4 presents the results and discussion of analysis of the ELF-VLF signals received at our station. The ionospheric reflection heights and propagation distances are calculated from lightning generated tweek sferics. A low latitude whistler observation is shown and the calculated magnetospheric parameters are presented. Preliminary observations of navigational transmitter signal amplitudes from NWC at 19.8 kHz and NPM at 21.4 kHz in daytime are presented. The sunset and sunrise effects on these signals are also discussed.

Chapter 5 presents an analysis of lightning detections made over Fiji during the years 2003 and 2004. The lightning data have been obtained from the World Wide Lightning Location Network (WWLLN).

Chapter 6 presents the scintillation observations of the beacon signal from the geostationary satellite Intelsat (701) at 3.925 GHz. $S_i$ index calculated from the amplitude measurements are presented and discussed. The spectral characteristics could not be determined because scintillation durations were of insufficient duration.

Chapter 7 finally summarizes the important findings of this work.

**1.8 Research Papers Published and Presented**


Chapter 2

Theoretical Considerations

Lightning emits a wide spectrum of electromagnetic radiation during its breakdown. Lightning could be cloud-to-ground (CG), intra-cloud (IC), cloud-to-cloud (CC) or cloud-to-air. In CG lightning stroke, the stepped leader descending from the base of a thundercloud triggers the first return stroke. This stroke is responsible for most of the charge transfer within few tens of microseconds causing a current of tens of kA. The radio impulse emitted by the CG strokes, commonly referred as sferics, has the highest spectral density in the Very Low Frequency band (VLF band 3 – 30 kHz) with the peak at ~10 kHz (Pathak et al., 1982). The emission from the lightning discharges can be approximated as that due to a vertical dipole close to the surface of the Earth emitting transverse electromagnetic or TEM, transverse magnetic or TM and transverse electric or TE modes. The sferic radiation propagates with little attenuation in the Earth-ionosphere waveguide (EIWG) to large distances (Mm) by multiple reflections. The EIWG disperses the sharp sferic impulse to a wave train lasting a millisecond or more.

There are three primary mathematical formulations for VLF propagation in the Earth-ionosphere waveguide that are most widely utilized. These theories were developed by J. Galejs (Galejs, 1972), J.R. Wait (Wait, 1970), and K.G. Budden (Budden, 1961). All these theories treated the time-harmonic concept of propagation at a single frequency. The differences were in the treatment of the ionospheric and ground boundaries. Using impedance boundary conditions Galejs formulated fields of vertical and horizontal dipole sources in free space region between two spherical shells represented as the ionospheric and ground boundaries. The time-harmonic form of the Maxwell’s equations was used to find the solutions in the waveguide. A solution for propagation to large distances from the source was obtained by various
asymptotic expansions of the Legendre polynomials (Galejs, 1972). The ionosphere was treated sharply stratified with free space between the Earth and the ionosphere. Within each layer the ionosphere was assumed to be homogenous. In the real ionosphere, the electron density smoothly varies with height. This makes the Galejs theory of little use for calculations involving different ionospheric heights. The theory developed by Budden specifies the upper and lower waveguide boundaries completely in terms of general reflection coefficients (Budden, 1961). By this approach the waveguide can be assumed to be comprised of any medium, sharply bounded or stratified, or even anisotropic. This theory details a more realistic waveguide between the Earth and the ionosphere but is far complicated for the purpose of this work. The Wait’s theory assumes free space region between two spherical shells and specifies the conductivity, permittivity, and permeability of the upper and lower boundaries of the waveguide. By similar asymptotic expansions and approximations to that of Galejs, Wait obtained the solutions for fields in free space region as a sum of traveling wave modes (Wait, 1970). The theory also extends to a stratified ionosphere with smooth transitions from one layer to another. For simplicity the ionosphere was assumed to be isotropic. However, the real ionosphere is anisotropic, as it is comprised of cold magnetized plasma. To identify and understand the nature of propagation inside the waveguide bounded by spherical boundaries of variable conductivity, this chapter first deals with propagation of electromagnetic waves in a parallel plate waveguide. The theory is then extended to spherical waveguide (Wait’s theory). When the sferics are detected after traveling large distances (Mm) to our station Suva, Fiji they will be traveling partially or wholly in waveguides formed by the Earth or sea surface as the lower boundary of the waveguide. Hence the Wait’s theory is used to investigate the effects of the different boundaries. The propagation modes are assumed to be similar to that of a parallel plate waveguide. In section 2.1 which follows, the treatment outlined in Ramo et al. (1984) and Rao (2004) is used to describe propagation of plane electromagnetic waves in perfectly conducting Earth-ionosphere waveguide.
2.1 Propagation of Plane Electromagnetic Waves in a Parallel Plate Waveguide

2.1.1 Transmission and reflection of electromagnetic waves on a dielectric-dielectric interface

![Diagram of wave incident at an angle, \( \theta_i \) on a dielectric-dielectric interface.](image)

**Figure 2.1:** Wave incident at an angle, \( \theta_i \) on a dielectric-dielectric interface.

- a. Polarization with the electric field parallel to plane of incidence (TM mode).
- b. Polarization with electric field perpendicular to plane of incidence (TE mode) [reproduced from Ramo et al., 1984].

First consider the transmission and reflection of plane electromagnetic waves of arbitrary polarization on a dielectric – dielectric interface. The incident ray makes an arbitrary angle \( \theta_i \) with the normal to the interface. The incident \( \mathbf{E} \) vector is resolved into two components, one parallel and the other perpendicular to the plane of incidence as shown in figure 2.1. The plane \( x-z \) is taken to be the plane of incidence.
The component with $E$ parallel to the plane of incidence is known as the transverse magnetic (TM) mode while $E$ normal to the plane of incidence is called transverse electric (TE) mode. Figure 2.1a and b shows the TM and TE mode waves respectively. The $\theta_r$ is the angle of reflection and $\theta_t$ is the angle of transmission. In figure 2.1a, $H_i$, $H_r$ and $H_t$ are in the $+y$ direction (out of page). In figure 2.1b $E_i$ and $E_t$ are in the $+y$ direction and $E_r$ is in the $-y$ direction. The ratio of the magnitudes of $E$ and $H$ fields for the waves gives the intrinsic impedance, $\eta$ of the respective media.

$$\frac{E_i}{H_i} = \frac{E_r}{H_r} = \eta_1 = \sqrt{\frac{\mu_1}{\varepsilon_1}}$$  \hspace{1cm} (2.1)$$

$$\frac{E_i}{H_i} = \eta_2 = \sqrt{\frac{\mu_2}{\varepsilon_2}}$$  \hspace{1cm} (2.2)

By the laws of reflection and refraction it can be shown that

$$\theta_i = \theta_r$$  \hspace{1cm} (2.3)$$

and $\theta_i = \sin^{-1}\left(\sqrt{\frac{\mu_1\varepsilon_1}{\mu_2\varepsilon_2}} \sin \theta_i\right)$  \hspace{1cm} (2.4)

Consider the reflection and transmission of waves at the interface, the reflection coefficient, $\mathcal{R}$ and the transmission coefficient, $\mathcal{T}$ for both parallel and perpendicular polarizations are defined as

$$\mathcal{R} = \frac{E_r}{E_i}$$  \hspace{1cm} (2.5)$$

$$\mathcal{T} = \frac{E_i}{E_i}$$  \hspace{1cm} (2.6)

The reflection and transmission coefficients for parallel and perpendicular polarizations respectively, are given by

$$\mathcal{R}_\parallel = \frac{\eta_1 \cos \theta_i - \eta_2 \cos \theta_i}{\eta_1 \cos \theta_i + \eta_2 \cos \theta_i}$$  \hspace{1cm} (2.7)$$

$$\mathcal{R}_\perp = \frac{2\eta_2 \cos \theta_i}{\eta_1 \cos \theta_i + \eta_2 \cos \theta_i}$$  \hspace{1cm} (2.8)$$

$$\mathcal{R}_\parallel = \frac{\eta_1 \cos \theta_i - \eta_2 \cos \theta_i}{\eta_1 \cos \theta_i + \eta_2 \cos \theta_i}$$  \hspace{1cm} (2.9)
$S_\perp = \frac{2\eta_2 \cos \theta_i}{\eta_1 \cos \theta_i + \eta_2 \cos \theta_i}$  \hspace{1cm} 2.10

These are defined as the Fresnel’s coefficients.

### 2.1.2 Reflection of electromagnetic waves on a dielectric-conductor interface

#### Figure 2.2: Wave incident at an angle, $\theta_i$ on a dielectric-conductor interface.

- **a.** Polarization with the electric field parallel to plane of incidence.
- **b.** Polarization with electric field perpendicular to plane of incidence [reproduced from Ramo et al., 1984].

Consider reflection of plane electromagnetic waves on a perfect conductor. The polarization of the waves and space orientation considered here is the same as in figure 2.1. The medium above the conductor is assumed to be free space. There is no transmitted wave for a perfect conductor interface as shown in figure 2.2. The incident wave is propagating in the $+\delta$ direction and the reflected wave in the $-\delta'$
direction. The phase factors for the incident and reflected waves can be expressed as $e^{-j k \delta}$ and $e^{j k \delta'}$ respectively, where $k$ is the wave number. The total $E$ field in medium 1 ($z < 0$) is given by the sum of the incident and reflected waves and can be written as

$$E(x, z) = E_i e^{-j k \delta} + E_r e^{j k \delta'}$$  

2.11

where $E_i$ and $E_r$ are the incident and reflected electric field vectors respectively. *Note: only the spatial variation of the field vectors is considered here.* From figure 2.2a, $\delta$ and $\delta'$ can be expressed in the rectangular coordinate system as

$$\delta = x \sin \phi + z \cos \phi$$  

2.12

$$\delta' = -x \sin \theta + z \cos \theta$$  

2.13

The field components for parallel and perpendicular polarization are shown separately in sections 2.1.3 and 2.1.4.

### 2.1.3 Polarization with electric field parallel to the plane of incidence (TM mode)

For the TM mode, the electric field vector have components in the $x$ and $z$ directions. Substituting Eq. 2.12 and Eq. 2.13 in Eq. 2.11 and separating into their $x$ and $z$ components gives

$$E_x(x, z) = E_i \cos \phi e^{-j k (x \sin \phi + z \cos \phi)} - E_r \cos \phi e^{j k (-x \sin \phi + z \cos \phi)}$$  

2.14

$$E_z(x, z) = -E_i \sin \phi e^{-j k (x \sin \phi + z \cos \phi)} - E_r \sin \phi e^{j k (-x \sin \phi + z \cos \phi)}$$  

2.15

The magnetic field component $H_y(x, z)$ is

$$H_y(x, z) = H_i e^{-j k (x \sin \phi + z \cos \phi)} + H_r e^{j k (-x \sin \phi + z \cos \phi)}$$  

2.16

The intrinsic impedance of a perfect conducting medium, $\eta_2$ approaches zero as the conductivity, $\sigma \to \infty$. There is no transmitted wave as the Fresnel’s equations for parallel polarization reduces to

$$R_\parallel = 1$$  

2.17

$$S_\parallel = 0$$  

2.18

It follows from Eq. 2.5, that

$$E_i = E_r$$  

2.19
Substituting these results in Eq. 2.14 – Eq. 2.16, the final expressions for the field components at any point \( z < 0 \) are

\[
E_y(x, z) = -2j E_i \cos \theta_i \sin(k z \cos \theta_i) e^{-j k x \sin \theta_i} \quad \text{Eq. 2.20}
\]

\[
E_x(x, z) = -2 E_i \sin \theta_i \cos(k z \cos \theta_i) e^{-j k x \sin \theta_i} \quad \text{Eq. 2.21}
\]

\[
\eta_i H_y(x, z) = 2 E_i \cos(k z \cos \theta_i) e^{-j k x \sin \theta_i} \quad \text{Eq. 2.22}
\]

The boundary condition for the electric field is that on the surface \( (z = 0) \) of the perfect conductor \( E_x \) must be zero for all values of \( x \). From Eq. 2.20 it is clear that the boundary condition is satisfied. The factor \( e^{-j k x \sin \theta_i} \) shows the \( x \) dependence, while \( \sin(k z \cos \theta_i) \) and \( \cos(k z \cos \theta_i) \) shows the \( z \) dependence of the field components.

Thus the field has the character of a traveling wave in the \( x \)-direction, but that of a standing wave in the \( z \)-direction.

### 2.1.4 Polarization with electric field perpendicular to the plane of incidence (TE mode)

In the TE mode the incident \( \mathbf{E} \) field is directed in the \(+y\)-direction (perpendicular to the plane of incidence) and the reflected \( \mathbf{E} \) field in the \(-y\)-direction. The \( \mathbf{E} \) and \( \mathbf{H} \) fields are as shown in figure 2.2b. The field components of the two waves in the Cartesian coordinates can be derived in a similar way. The field components are

\[
E_y(x, z) = E_i e^{-j k (x \sin \theta + z \cos \theta)} - E_r e^{j k (x \sin \theta + z \cos \theta)} \quad \text{Eq. 2.23}
\]

\[
\eta_i H_x(x, z) = -E_i \cos \theta_i e^{-j k (x \sin \theta + z \cos \theta)} - E_r \cos \theta_i e^{j k (x \sin \theta + z \cos \theta)} \quad \text{Eq. 2.24}
\]

\[
\eta_i H_z(x, z) = E_i \sin \theta_i e^{-j k (x \sin \theta + z \cos \theta)} - E_r \sin \theta_i e^{j k (x \sin \theta + z \cos \theta)} \quad \text{Eq. 2.25}
\]

By a similar reasoning as in the TM mode as \( \sigma \to \infty \) for a perfect conductor, \( \eta_2 \to 0 \), the Fresnel’s equations (Eq. 2.9 and Eq. 2.10) gives

\[
\Re = 1 \quad \text{Eq. 2.26}
\]

\[
\Im = 0 \quad \text{Eq. 2.27}
\]

From Eq. 2.5 and Eq. 2.26,

\[
E_i = E_r \quad \text{Eq. 2.28}
\]

Hence the field components Eq. 2.23 – Eq. 2.25 reduces to

\[
E_y(x, z) = -2j E_i \sin(k z \cos \theta_i) e^{-j k x \sin \theta_i} \quad \text{Eq. 2.29}
\]
\[ \eta_i H_x(x, z) = -2 E_i \cos \theta_i \cos(k z \cos \theta_i) e^{-jk x \sin \theta_i} \quad 2.30 \]
\[ \eta_i H_z(x, z) = -2 j E_i \sin \theta_i \sin(k z \cos \theta_i) e^{-jk x \sin \theta_i} \quad 2.31 \]

The boundary conditions are satisfied since \( E_y \) and \( H_z \) are zero and \( H_x \) is a maximum at \( z = 0 \). These fields also exhibit traveling wave pattern in the \( x \)-direction and standing wave pattern in the \( z \)-direction.

### 2.1.5 Parallel plate waveguide

![Diagram of a parallel plate waveguide](image)

**Figure 2.3:** Uniform Plane wave propagation in a parallel plate waveguide.

By placing a perfectly conducting plate at some \( z < 0 \), we have a parallel plate waveguide. Note: the lower boundary is at \( z = 0 \) and is positive in the downward direction. From Eq. 2.20 and Eq. 2.29, the tangential components of \( E \) namely \( E_x \) and \( E_y \) will be zero for all values of \( x \) when \( k z \cos \theta_i = \pm n \pi \), where \( k = 2\pi/\lambda \) and \( n \) is a positive integer. Thus

\[ z = -\frac{n \lambda}{2 \cos \theta_i} \quad 2.32 \]

where the \( - \) sign is used because the plates are placed at \( z < 0 \) to form the waveguide. If a perfectly conducting plate is kept at \( z \) satisfied by Eq. 2.32, the boundary conditions are not violated and the resultant wave will propagate in the \( x \)-direction. As shown in figure 2.3, a plate is placed at \( z = -\lambda/2 \cos \theta_i \), the boundary conditions at the surface on this plate are satisfied and the waves can propagate. In the EIWG only the TM and TEM modes have a non-zero electric field in the plane of incidence. The ELF-VLF recording instrument used in this work measures the vertical component of
the electric field of lightning generated sferics near the ground so only the TM and TEM modes are considered.

### 2.1.6 Cutoff frequency

![Parallel Plate Waveguide Diagram](image)

**Figure 2.4:** Cutoff condition in a parallel plate waveguide.

Consider that the conducting plates are situated in planes $z = 0$ and $z = -h$ as shown in figure 2.4a. From Eq. 2.32

$$h = \frac{n \lambda}{2 \cos \theta_i}$$  \hspace{1cm} 2.33

$$\cos \theta_i = \frac{n \lambda}{2h} = \frac{nc}{2hf}$$  \hspace{1cm} 2.34

where $c = \frac{1}{\sqrt{\mu_0 \varepsilon_0}}$ is the speed of light in free space ($\mu_i = \mu_o$ and $\varepsilon_i = \varepsilon_o$). From Eq. 2.34 it is clear that for different values of angle $\theta_i$, waves of different frequencies are reflected between the plates. As frequency, $f$ decreases, to satisfy the boundary conditions $\theta_i$ should decreases and the wave reflects more and more closely as shown in figures 2.4b and c. When $\theta_i = 0$, the wave reflects normal to the plates, as shown in figure 2.4d, with no propagation in the $x$-direction. This leads to the cutoff condition and is given by
Chapter 2 Theoretical Considerations

\[ f_{cn} = \frac{n c}{2 h} \tag{2.35} \]

where \( f_{cn} \) is the cutoff frequency of the \( n^{th} \) mode. Propagation of a particular mode is only possible if \( f \) is greater than \( f_{cn} \) for that mode. For \( f < n c/2h \), \( \cos \theta_i > 1 \), and \( \theta_i \) has no real solution, indicating that propagation does not occur for these frequencies in the waveguide mode.

### 2.1.7 Phase velocity

Using Eq 2.20, the phase constant of the wave in the \( x \)-direction is defined as

\[ \beta_x = k \sin \theta_i \tag{2.36} \]

where \( \sin \theta_i \) obtained using Eq. 2.34 and Eq. 2.35 is given by

\[ \sin \theta_i = \sqrt{1 - \left( \frac{f_{cn}}{f} \right)^2} \tag{2.37} \]

This gives the expression for the phase constants as

\[ \beta_x = \frac{2 \pi}{\lambda} \sqrt{1 - \left( \frac{f_{cn}}{f} \right)^2} \tag{2.38} \]

Eq. 2.38 shows that the phase constant is a function of frequency. The variation of \( \beta_x \) with frequency is known as the dispersion curve (\( \omega \) vs. \( \beta_x \)) and is plotted in figure 2.5. The guide wavelength \( \lambda_g \) is the wavelength along the guide in the \( x \)-direction and is given by

\[ \lambda_g = \frac{2 \pi}{\beta_x} = \frac{2 \pi}{k \sin \theta_i} = \frac{\lambda}{\sqrt{1 - \left( \frac{f_{cn}}{f} \right)^2}} \tag{2.39} \]

where \( \lambda \) is the free space wavelength. The phase velocity of the wave in the \( x \)-direction is given by

\[ v_p = \frac{\omega}{\beta_x} = \frac{\omega}{k \sin \theta_i} = \frac{c}{\sin \theta_i} = c \left[ 1 - \left( \frac{f_{cn}}{f} \right)^2 \right]^{-1/2} \tag{2.40} \]
From Eq. 2.40 the phase velocity for a particular mode along the waveguide is a function of the frequency. The components of a signal with a band of frequency will travel with different phase velocities as they propagate in the waveguide. This is known as dispersion. As frequency approaches infinity, $v_p$ approaches $c$ as indicated by the gradient of line A in figure 2.5. The gradient of line B is the phase velocity at radian frequency $\omega_1$. It will be shown in section 2.1.8 the gradient of the tangent at $\omega_1$ (line C) is the velocity of a band of waves whose frequency is centered at $\omega_1$.

2.1.8 Group velocity

Consider the field patterns of two waves with frequencies $f_A$ and $f_B$ propagating in the $x$-direction inside the parallel plate waveguide. The guide wavelengths are $\lambda_{gA}$ and $\lambda_{gB}$ and phase velocities $v_{pA}$ and $v_{pB}$ respectively. $v_{pB} > v_{pA}$. Figure 2.6 shows the two field patterns in the waveguide.
Figure 2.6: Illustration of the concept of group velocity of two waves propagating in a parallel plate waveguide [reproduced from Rao, 2004].
The waves travel with their respective phase velocities along the waveguide. Assume that the positive peak of the two fields numbered 0 are aligned at \( t = 0 \) (see figure 2.4a). As the waves travel with their respective phase velocities these two peaks get out of alignment. At some later time, \( \Delta t \) the two peaks numbered \(-1\) on the two waves will align at some distance \( \Delta z \) from the location of the initial alignment of the 0 peaks. At \( t = \Delta t \) the \(-1\) peak of wave A has traveled a distance \( \lambda_{gA} + \Delta z \) with a phase velocity of \( v_{pA} \) but the \(-1\) peak of wave B, has traveled a distance of \( \lambda_{gB} + \Delta z \) with a phase velocity of \( v_{pB} \). Then

\[
\Delta z + \lambda_{gA} = v_{pA} \Delta t \quad 2.41
\]

\[
\Delta z + \lambda_{gB} = v_{pB} \Delta t \quad 2.42
\]

Solving Eq. 2.41 and Eq. 2.42 for \( \Delta t \) and \( \Delta z \) gives

\[
\Delta t = \frac{\lambda_{gA} - \lambda_{gB}}{v_{pA} - v_{pB}} \quad 2.43
\]

\[
\Delta z = \frac{\lambda_{gA} v_{pB} - \lambda_{gB} v_{pA}}{v_{pA} - v_{pB}} \quad 2.44
\]

The group velocity, \( v_g \) of the two waves is then defined as the velocity of a group of these two waves with same phase traveling as a whole, and is given as \( \Delta z / \Delta t \).

Dividing Eq. 2.44 by Eq. 2.43 gives

\[
v_g = \frac{\omega_B - \omega_A}{\beta_{gB} - \beta_{gA}} = \frac{\Delta \omega}{\Delta \beta} \quad 2.45
\]

For a narrow band of signal, the group velocity of the entire group is best represented as \( d\omega / d\beta \), which is the slope of dispersion curve at the center frequency. Hence

\[
v_g = \frac{d\omega}{d\beta} \quad 2.46
\]

As shown in figure 2.5, the gradient of line C is the group velocity at \( \omega_1 \). The expression for the group velocity is obtained by differentiating Eq. 2.38 with respect to frequency and is given as

\[
v_g = c \sqrt{1 - \frac{f_{on}^2}{f^2}} \quad 2.47
\]
2.1.9 Propagation distance of tweeks in conducting waveguide

Lightning generated sferic signals propagate to large distances in the EIWG by multiple reflections. From Eq. 2.47 as \( f \) approaches \( f_{cn} \), the group velocity is reduced considerably. In the TM mode, components of frequencies, which are slightly higher than \( f_{cn} \) and travel large distances to the receiver, arrive with noticeable time delay i.e. these frequency components arrive slightly later than the higher frequency components. They appear as vertical traces on a spectrogram with a curve near the cutoff frequency. Such signals are known as tweeks. Figure 2.7 shows a pictorial representation of a spectrogram of a typical tweek sferic. When the received sferic signals are displayed on a spectrogram, the components which have not suffered dispersion appear as vertical traces only.

Assuming a parallel plate EIWG, tweek sferics can be used to estimate the total propagation distance from the source. Tweeks received using the WWLLN experimental setup is used to estimate their propagation distance. From the dispersion in the tweek traces, two close frequency components with their absolute times of arrival are accurately estimated by graphical examination. In figure 2.7, \( f_{c1}, f_{c2}, f_{c3}, \) and \( f_{c4} \) are the cutoff frequencies of their respective modes. The frequency
components $f_1$ and $f_2$ in the dispersed portion of the tweek trace are received at absolute times $t_1$ and $t_2$. The group velocities, $v_{gf1}$ and $v_{gf2}$ at these two frequencies can be calculated using Eq. 2.47. The total propagation distance, $D$ of the tweek in the EIWG is given by

$$D = v_{gf1} (t_1 - t_s)$$  \hspace{1cm} 2.48
$$D = v_{gf2} (t_2 - t_s)$$  \hspace{1cm} 2.49

where $t_s$ is the absolute time of the lightning stroke causing the tweek. The group velocities given by Eq. 2.47 of these two frequencies and their time difference are used to calculate the total propagation distance. From Eq. 2.48 and Eq. 2.49

$$\Delta t = D \left( \frac{v_{gf1} - v_{gf2}}{v_{gf1}v_{gf2}} \right)$$  \hspace{1cm} 2.50

where $\Delta t = t_2 - t_1$. Rearranging Eq. 2.50 and using Eq. 2.47 gives the total propagation distance, $D$ written as

$$D = \Delta t \ c \ \left[ \left(1 - \frac{f_2^2}{f_1^2} \right) - \left(1 - \frac{f_2^2}{f_1^2} \right) \right]^{\frac{\gamma}{2}}$$  \hspace{1cm} 2.51

Eq. 2.51 is the same as that reported by Prasad (1981).

### 2.2 Attenuation of Tweek Sferics in Finitely Conducting Earth-Ionosphere Waveguide

The actual EIWG is spherical with finite conducting surfaces. Consider that the Earth and the lower ionosphere form waveguide with concentric spherical walls of finite conductivities with free space between the two boundaries. The lightning can be considered as vertical electric dipole. For a vertical electric dipole source, the propagation characteristics of the modes can be described by a factor $e^{-jks,D}$ (Wait, 1957), where $k$ is the wave number, $S_n$ is the sine of a complex angle between the wave normal and the vertical, and $D$ is the distance traveled by tweeks in the waveguide. Since $S_n$ is complex, writing $S_n = X_n + jY_n$ in the exponential form gives

$$e^{-jks,D} = e^{-jX_n,D} e^{jY_n,D}$$  \hspace{1cm} 2.52
where $kX_n$ and $kY_n$ represents the wave number and attenuation, $\alpha_n$ of the $n^{th}$ mode respectively. The mode equation for VLF wave propagation (Wait, 1970) is given by

$$khC_n^2 - \pi n C_n - j\Delta = 0$$

where $h$ is the reflection height, $C_n$ is the cosine of the complex angle and $\Delta$ is a parameter given by the refractive indices of the conducting boundaries and the space between them for which the ground-ionosphere and sea-ionosphere propagation paths is written as

$$\Delta_g = N\left(\frac{1}{N_i} - \frac{1}{N_g}\right)$$

and

$$\Delta_is = N\left(\frac{1}{N_i} - \frac{1}{N_s}\right)$$

where $N$ with the subscripts $g$, $s$, and $i$ represent the refractive indices of the ground, sea, and ionospheric surfaces respectively and $N$ is the refractive index of the medium inside the waveguide. These refractive indices are complex and are given by

$$N_g^2 = \varepsilon_g - j\frac{\sigma_g}{\omega\varepsilon_o}$$

$$N_s^2 = \varepsilon_s - j\frac{\sigma_s}{\omega\varepsilon_o}$$

$$N_i^2 = \varepsilon_i - j\frac{\sigma_i}{\omega\varepsilon_o}$$

where $\varepsilon_g$, $\varepsilon_s$, and $\varepsilon_i$ are the relative permittivites and $\sigma_g$, $\sigma_s$ and $\sigma_i$ are the conductivities of the earth, sea, and ionosphere respectively. The mode equation given in Eq. 2.53 is quadratic in $C_n$. Assuming $|\Delta| kh << 1$, a solution to this equation is written as

$$C_n = \frac{\pi n}{kh} + j\frac{\Delta}{\pi n}$$

The sine of the complex angle is given as $S_n = (1-C_n^2)^{1/2}$. Substituting $C_n$ from Eq. 2.59, $S_n$ can be written as
Expanding Eq. 2.60, gives the real \( X_n \) and the imaginary \( Y_n \) parts of \( S_n \)

\[
X_n = \left\{ \frac{1}{2} \left[ 1 - \left( \frac{n \pi}{kh} \right)^2 + \left( \frac{2 \Delta}{kh} \right)^2 \right] + \frac{1}{2} \sqrt{1 - \left( \frac{n \pi}{kh} \right)^2 + \left( \frac{2 \Delta}{kh} \right)^2} \right\} \left( \frac{2|\Delta|}{kh} \right)^\frac{1}{2} 
\]

\[
Y_n = -\left( \frac{|\Delta|}{khX_n} \right) 
\]

The conductivity and dielectric constants of the Earth and seawater can be specified with some confidence over most paths. However, the distribution of charged particles in the ionosphere depends on solar radiation and thus changes with the hour of the day, season, geomagnetic coordinates and sunspot cycle. One of the simplest ionospheric profiles is an exponential variation of conductivity with height. Wait and Spies (1964) formulated such a profile from a parameter \( \omega_r(h) \). The parameter, \( \omega_r(h) \) is given as the electron plasma frequency, \( \omega_p^2(h) \) divided by the particle-neutral collision frequency, \( v(h) \) (Wait and Spies, 1964)

\[
\omega_r(h) = \frac{\omega_p^2(h)}{v(h)} = 2.5 \times 10^5 \exp[\beta(h - H')] 
\]

where \( \omega_p(h) \) is the electron plasma frequency given as \( \omega_p(h) = \sqrt{N_e e^2 / e_o m} \) and \( v(h) \) is the effective electron neutral collision frequency. Both \( \omega_r(h) \) and \( v(h) \) are the functions of the height, \( h \), in km. \( \omega_r(h) \) varies exponentially with height at a rate determined by a sharpness factor, \( \beta \) (in km\(^{-1}\)). \( H' \) is a reference height (in km) at which \( \omega_r(h) = 2.5 \times 10^5 \text{ rad/s} \) or \( H' = h \). The conductivity, \( \sigma_i \) of the ionosphere can be obtained from the ohms law,

\[
J = N_e e \dot{x} = \sigma_i E 
\]

where \( N_e \) is the number of electrons per cubic meter, \( e \) is the electronic charge and \( \dot{x} \) is the electron drift velocity. The \( \dot{x} \) is obtained from the equation of motion for an
electron in presence of an electric field, \( \dot{x} = \frac{eE}{m(\nu(h) + j\omega)} \). Substituting \( \dot{x} \) and \( \omega_n^2(h) \) in Eq. 2.64, the conductivity of the ionosphere can be written as

\[
\sigma_i = \varepsilon_o \frac{\omega_n^2(h)}{[\nu(h) + j\omega]}
\]

In the D region, \( \nu(h) >> \omega \), in the order \( 10^5 - 10^6 \text{ s}^{-1} \) (Budden, 1961), the ionospheric conductivity reduces to

\[
\sigma_i = \varepsilon_o \frac{\omega_n^2(h)}{\nu(h)}
\]

Substituting Eq. 2.63 into Eq. 2.66 gives

\[
\sigma_i = 2.5 \times 10^3 \varepsilon_o \exp[\beta(h - H')]
\]

The attenuation factor, \( \alpha_n \) defined as the number of decibels (dB) per 1000 km of path length is given by Wait (1957a) and Hayakawa et al. (1995)

\[
\alpha_n = -\frac{8.86 \times 10^3 \times \omega \times Y_n}{c}
\]

where \( \omega \) is the angular frequency of the tweek signal and \( c \) is the velocity of light in free space. Substituting Eq. 2.62 into Eq. 2.68 the \( \alpha_n \) in terms of \( X_n \) can be written as

\[
\alpha_n = \frac{8.86 \times \omega \times (|A|)}{cX_n} \times 10^3
\]
Chapter 3
Experimental Setup and Data Recording

This chapter gives an overview of the lightning location detection system that uses the Time of Group Arrival (TOGA) of lightning generated sferics to determine lightning location. The system was setup in collaboration with the World Wide Lightning Location Network (WWLLN) in 2003. The system was extended to record the lightning generated ELF-VLF signals (tweeks and whistlers). A ground-based receiving system used for ionospheric scintillation measurements in the C-band at 3.925 GHz from geostationary satellite Intelsat (701) was set-up. The general functions and usage of the different components in the satellite receiving system are described.

3.1 ELF-VLF Experimental Setup
The Block diagram of the ELF-VLF experimental setup used for TOGA determination is shown in figure 3.1. The system consists of a VLF antenna, pre-amplifier, GPS antenna, Service unit, and a microcomputer. The components that were used are briefly described.

*VLF Antenna* is a 5 m long stranded wire with 1 mm diameter, encased within a 2 m length of 40 mm PVC water pipe. The wire is folded to fit the inside of the pipe laterally. A small pill bottle is threaded with the wire forming a whip type antenna. The whip antenna arrangement is approximately 1.5 m in length and is attached to the pipe with the bottle end at the top connecting to the cover. A short (~ 100 mm) coaxial wire is connected to the end of the antenna to the BNC to make a secure connection. The coaxial shield is earthed through a thin wire. This design minimizes the electric field induced at the antenna surface by extreme thunderstorm fields to avoid attracting lightning strikes and damage to setup.
The VLF antenna is connected to the pre-amplifier fixed at the bottom of the PVC pipe. The antenna with the preamplifier unit is mounted on the roof of a two-storey building.

**VLF pre-amplifier** is attached to the support pipe immediately below the antenna. It is vertically mounted with the antenna input at the top. The pre-amplifier is encased in a rainproof metal box with 2 mm drain holes near the output socket, the lowest point on the pipe. The pre-amplifier is grounded to a lightning protection ground. The preamplifier consists of a RC network of filters having attenuation in dB proportional to the square root of frequency. For example, the attenuation is about 6 dB at 40 kHz and 24 dB at 640 kHz. The VLF input signal is clamped between ±10 V by fast 0.5 A diodes for protection against lightning impulse fields of up to 100 kV/m. The op-amp 744 is utilized for a high linearity having a gain of 11. Pre-amplifier gives flat-response in the VLF range. The output of the pre-amplifier is connected to the Service Unit (SU) with Amphenol plug on one end (pre-amplifier) and DIN on the other (SU).
Global Positioning System (GPS) - The lightning detection system uses the National Marine Electronic Association (NMEA) time code from the GPS to determine the precise timing measurements. Given below is a brief overview of the GPS system. The Global Positioning System (GPS) is a satellite based radio navigation system. It consists of three segments: satellite constellation, ground control network, and user equipment. The satellite constellation comprises of 24 satellites in six earth centered orbital planes at an altitude of 20,200 km. The orbital period of a GPS satellite is 12 hours with a nearly circular orbit equally spaced about the equator at a 60° separation angle and an inclination of 55° relative to the equator. Any user with a clear view of the sky will have a minimum of four satellites in view at anytime. The ground control network operates the system and provides command and control functions for the satellite constellation. The user equipment referred as “GPS receivers” receives L-band signals from the satellites and processes them for user position, velocity and time. The GPS provide two levels of service: Standard Positioning Service (SPS) and a Precise Positioning Service (PPS). SPS usage is available for general civil use with accuracies of 100 m horizontally and 156 m vertically, and time dissemination accuracy within 340 ns. It provides a highly accurate positioning, velocity and timing service. The position accuracy is 22 m horizontally and 27.7 m vertically and time transfer accuracy to UTC within 200 nanoseconds. The PPS is the data transmitted in the L1 and L2 frequencies. GPS applications generally fall into 5 major categories. These are location finding, navigation, tracking, mapping and timing. In this work the GPS NMEA – 0183 timing code provides the precise time for sferic detection to within a few hundred nanoseconds. The NMEA time from the GPS receiver is connected through the serial port RS-232 that automatically sets the PC clock. The NMEA time keeps the computer referenced to an atomic time standard.
**Note:** The pre-amplifier power is isolated from SU and PC ground by 1 kV. The middle pin of 5-pin DIN plug is connected only to the pre-amplifier PCB ground and metal case. Thus the voltages shown above as “-13 0 +13 V” are with respect to the pre-amplifier PCB ground.

Figure 3.2: Block diagram of the Service Unit.
**Service Unit and the GPS Engine** - The Service unit encases all the electronics of the ELF-VLF receiver. This unit was provided by the WWLLN leaders. The SU supplies an isolated power (± 15 V) to the pre-amplifier, and the isolation transformer (1:1). The SU contains the GPS “engine” which generates the PPS and the NMEA time code. The GPS antenna is mounted on the roof of the building housing the system and is “open” to the sky. A 10 m coaxial cable connects the GPS antenna to the SU. The PPS is a 10 μs pulse. Its leading edge is well within 1 μs of true UTC, while the NMEA time code identifies the second. The pre-amplifier connects to the SU through a 5-pin DIN socket. The NMEA output connects to the PC at COM 1 via serial port RS-232. Two sound cards were installed in the PC. The SU has three parallel VLF outputs. Two of them are the stereo outputs for connection to the sound cards while the other one connects through a BNC socket. The network uses one of the stereo outputs for lightning location identification while the second stereo output is used for recording the ELF-VLF signal analyzed in this work. Note: The second sound card installed is an additional modification for the Suva station only. Figure 3.2 shows the schematic diagram of the SU. The third BNC output has been used to measure amplitude of two VLF transmitters.

### 3.2 ELF-VLF Data Recording and Analysis

The lightning sensor software is loaded in the computer and is operated remotely by the network administrator in Dunedin, New Zealand. It contains the Red Hat Linux operating system and is connected to the Internet. The lightning software continuously monitors the ELF/VLF signals. The sound card continuously samples at 48 kHz. The samples are kept in a buffer for a short time. A sferic is detected when the amplitude of the signal rises above a set threshold. At this instant a trigger time is recorded. The trigger time $t_o$ is the time when a lightning stroke is detected. A total of 64 samples are taken from the buffer; 16 sample before $t_o$ and 48 samples after $t_o$. TOGA is determined from the phase slope of the received sferic at $t_o$ and transmitted in packets to the central processing stations in Dunedin and University of Washington, Seattle, USA for stroke location. Lightning is located only if at least four sites have detected the same stroke.
The data are sent through port 5555 using UDP (User Datagram Protocol). The data from each station is provided in the form of spectrograms for public access in the web site http://webflash.ess.washington.edu/. The spectrograms are updated every 10 min. The network administrators send lightning location data on CDs for every month to all host stations. The lightning data for the years 2003 and 2004 over Viti Levu and Vanua Levu have been analyzed and presented in chapter 5.

The sferics data using the second sound card was initially recorded for 20 min at predefined times (00 hrs, 01 hrs, 02 hrs, 03 hrs, 20 hrs, 21 hrs, 22 hrs, 23 hrs LT) during September 2003 – January 2004. Daytime recordings were also conducted but showed no tweeks. Due to the large data size (~11 MB/min) the recording was reduced to every hour for 2 min from February 2004 – July 2004. The data files are used for tweek and whistler analysis presented in chapter 4. The data are analyzed using MATLAB codes that produce the dynamic spectrograms (one every 1 s) of the received ELF-VLF signals. Tweeks are then identified from the spectrograms. The spectrograms form the basis of the tweek analysis technique used in this work. Figure 3.3 shows an example of a tweek spectrogram recorded at 2200 hrs 33 s LT on 31 March 2004. The horizontal lines in the spectrogram are those produced by VLF MSK (minimum shift keying) transmitters. A whistler observed at the station was analyzed similarly.

The VLF output (BNC) from the SU is connected to a spectrum analyzer that produced the frequency spectrum of the VLF transmitter signals received at the station. The signals from NPM Hawaii at 19.8 kHz and NWC Australia at 21.4 kHz have been
analyzed and presented in chapter 4. Figure 3.4 shows a typical example of the frequency spectrum of the transmitter signals as seen on the spectrum analyzer.

**Figure 3.4:** VLF transmitter signals when displayed on a spectrum analyzer.

### 3.3 SHF Ionospheric scintillation Setup

The essential components of a ground based satellite scintillation recording system: the satellite dish antenna with *Low Noise Block Amplifier* (LNB), receiver, and recording equipment and software are shown in figure 3.5.

**Figure 3.5:** Setup of a ground based satellite scintillation receiving system.
**Dish Antenna** - Conventional ground based radio wave receiving systems utilize dipole or array antennas which are suitable for reception of frequencies less than 1 GHz. Low loss cables are used for interconnections between the antenna and receiver. For frequencies higher than 1 GHz, transmissions through cables suffer considerable attenuation. However for frequencies less than 1.3 GHz, low loss transmission cables can be used provided they are kept short (< 10 m). Metallic horn antennas and waveguides are used to transmit the wave energy to the receiver. Curved reflectors concentrate the electromagnetic radiation into the feed horn. Parabolic dish-shaped reflectors are used because it provides high gain in proportion to its aperture. Reflector surfaces with mesh less than 1/8th the received wavelength appear like a smooth surface. Hence for frequencies in the C-band (3.7 – 6.4 GHz), a reflector surface made of metallic mesh (6 – 10 mm) act as good reflector. The dish antenna used in this work is a 5 m centre-fed dish with 3.5 mm Aluminium reflector mesh suitable for the frequency measured here. Figure 3.6 shows the dish antenna used in this work.

![Dish antenna used for scintillation measurement.](image)
Figure 3.7: Block diagram of LNB.

**Low Noise Block Downconverter** - The reflector focuses the radiowaves into the receiving device known as the Low Noise Block Downconverter (LNB). The LNB construction consists of a feedhorn at the receiving end with a waveguide at the center. Figure 3.7 shows the block diagram of LNB. The feedhorn is set at the focal point of the dish parabola. The dish collects the signal from the satellite and focuses it into the small opening in front at the front of the feedhorn. An internal probe inside the guide is the actual microwave antenna, which converts the signal into electrical current. These are weak signals that are usually amplified by a Low Noise Amplifier (LNA) by about 50 dB. The LNA is one of the critical elements in determining the earth station performance as a system element. The LNA is designed to amplify the desired signal while injecting very little self induced noise. LNAs were originally made using ordinary transistors and the arrangement was immersed in liquid nitrogen or helium to reduce noise. GaAsFETs are now used in LNAs that provide more efficient amplification with little noise. Figure 3.8 shows the electronic block diagram of a LNB. The GaAsFET transistors are cascaded in series to provide the necessary amplification. The LNB then downconverts the incoming signal. This component of the system is known as the Low Noise Converter (LNC).
Heterodyning or mixing accomplishes the down-conversion of the frequency. In this work 3.925 GHz signal is mixed with the local oscillator frequency in the LNC at 5.194 GHz, which give intermediate frequency of 1.269 GHz. The frequency of the local oscillator is determined by the dielectric resonant oscillator in the LNB. The frequency is dependent on the power supplied. Hence by varying the voltage at the power supply, different down conversions could be obtained. The receiver used does not provide DC power to the LNB. A stabilised 15 V-dc supply is used to power the LNB. The circuit diagram used is shown in Fig.3.9.
Figure 3.9: Circuit diagram to power LNB.

A selective filter removes the higher frequency, and let pass the lower frequency signal, which in this case is 1.269 GHz. This 1.269 GHz signal contains all the information of the 3.925 GHz input signal and is tuned at 1.269 GHz on the receiver. A coaxial cable RG-75 is used as the feedline to the receiver.

**Receiver** - The receiver used in this work is an ICOM IC-R8500 communications receiver specially modified to measure ionospheric scintillations. The modification was done at Total Electronic Concepts, Lincoln, MA 01773, USA by Prof. J.A Klobuchar.

### 3.4 Scintillation Data Recording and Analysis

1. The geostationary satellite IntelSat (701) is located at 180°E. It transmits a beacon signal at 3.925 GHz. The calculated azimuth and elevation angle are 4.86° and 68.45° respectively. The ionospheric scintillation were measured at University of the South Pacific (USP), Suva (18.13° S, 178.47° E), Fiji.

2. The LNB downconverted the received beacon signal to 1.269 GHz. The communications receiver (model IC-R8500) was used to detect the signal. Amplitude measurements were made from a special output in the receiver. The receiver was modified to produce an amplified diode load output for such measurements. The output is on an RCA-type plug at the rear of the receiver.
3. The output from the receiver is connected to the LabVIEW PCI data acquisition board (NI-LAB PCI-6023E) for recording in the computer. A LabVIEW program is developed to monitor the amplitude variations. The temporal resolution of the measurements is 1 second.

4. The data are recorded on excel spreadsheet files which includes the date, time and amplitude of the beacon signal in volts.

5. The measurements were done for 24 hours during the period December 2003 to June 2004.

6. The data recorded were analyzed for scintillation occurrence and $S_4$ index.
Chapter 4  Propagation Features of ELF-VLF Radio Waves: Applications

In this chapter the tweeks recorded using WWLLN system at Suva have been used to calculate the ionospheric reflection height and the total propagation distance of tweeks from the source point of occurrence. The calculated attenuation factor for different modes of tweeks have been calculated for the Earth-ionosphere and sea-ionosphere paths during day and nighttime conditions to explain the occurrence of tweeks. An analysis of a whistler observed on 22 November 2003 is presented. Sub-ionospheric propagation of transmitter signals from amplitude measurements of NPM Hawaii (21.4 kHz) and NWC Australia (19.8 kHz) for daytime period is also presented.

4.1 Introduction

The major part of the lightning wave energy travels in the Earth-ionosphere waveguide by multiple reflections and is received as radio atmospherics or sferics. Tweeks result when sferics are ducted in the Earth-ionosphere waveguide to large distances, up to several thousand kilometers particularly at night (Reeve and Rycroft, 1972). The received signals are slightly dispersed and have a sharp low frequency cutoff at around 1.5-2.0 kHz. Tweek sferics contains a lot of information about the state of the ionosphere along its propagation path and therefore forms a useful tool for probing the lower ionosphere. From the oscillograms of tweeks, Burton and Boardman (1933) deduced the nighttime variation of the lower limiting frequency. Outsu (1960) calculated attenuation coefficients near cutoff frequencies of first and second mode tweek sferics using TM mode waveguide theory. Reeve and Rycroft (1971) observed VLF signals during a solar eclipse of 7 March 1970 and using tweeks they determined the ionospheric reflection
height using the TM mode waveguide theory. Yamashita (1978) investigated the propagation mechanism of tweeks near cutoff frequency using the waveguide mode theory. He examined the propagation characteristics in the frequency range of 1.5 kHz – 10 kHz considering a homogenous and anisotropic ionosphere. Kumar et al. (1994) studied the propagation characteristics of tweeks observed at Bhopal a low latitude ground station in India. Hayakawa et al. (1995) estimated the propagation distance and reflection height using tweek sferics near the cutoff frequencies. Singh and Singh (1996) reported the observations of higher harmonic tweeks at low latitude station Varanasi (geomag. Lat. 14º 55’ N), India and modeled propagation features using Wait’s theory. At our station tweek occurrence in the nighttime are common phenomena. Tweek sferics are only prominent at nighttime and thus does not provide any information on the daytime ionosphere. In this chapter an analysis of up to 503 selected tweeks observed during a period of September 2003 to July 2004 is presented.

In addition to propagation in the Earth-ionosphere waveguide, a small part of the wave energy from lightning discharges can penetrate into the ionosphere and propagate in ducted modes along geomagnetic field lines to the opposite hemisphere through the magnetosphere (Smith and Helliwell, 1960; Sazhin et al., 1992). Lightning signals that make large angles with respect to the geomagnetic field propagate in the non-ducted mode (Edgar, 1976) and are received by satellites as whistlers. The signals propagating with wave normal approximately along the geomagnetic field are believed to propagate in the ducted mode (Helliwell, 1965). These waves are mostly observed at ground facilities at conjugate points, but are rarely observed on satellites due to relatively small volumes occupied by the ducts. The main interest in whistler research lies in interpreting data to obtain the ionization in the plasmasphere; it’s coupling with the ionosphere, and to understand the propagation mechanism at different latitudes. Data are mainly in the form of spectrograms, which show variation of frequency with time. These spectrograms form the foundation on which the analysis of whistler data is based. Barkhausen was the first to identify whistlers. Many studies followed thereafter (eg. Eckersley 1925; Burton and Boardman, 1933; and Potter, 1951), but the most convincing interpretation on whistlers
based on observations and theory was presented by Storey (1953). The Theory of Barkhausen and Eckersley predicted that the frequency, $f$ in the whistler is related to the dispersion, $D$ and travel time $\tau$, after the original flash by the expression $\tau = D \times f^{-1/2}$ (Storey, 1953). Whistlers recorded on spectrograms fall into one of four different classes (Sazhin et al., 1992). These are: 1. the initiating sferic and the nose can be identified, 2. the initiating sferic can be identified but not the nose, 3. the nose can be identified but not the initiating sferic, 4. neither the sferic nor the nose can be identified. Whistlers have been used quite effectively for diagnostics of magnetospheric parameters (Sazhin et al., 1992) such as equatorial gyrofrequency, the magnetic induction at the whistler $L$ value, electric field and plasmasphere electron content. The propagation mechanisms of low latitude whistlers have also been discussed. The whistler reported here falls in the second class and is given in figure 4.8.

The VLF signals radiated by navigational transmitters also propagate in the Earth-ionosphere waveguide. Field measurements from VLF transmitters have been used very successfully to determine daytime ionospheric parameters (e.g. Thomson, 1993; McRae and Thomson, 2000). Variations in the D-region of the ionosphere leads to changes in the propagation conditions for VLF waves inside the waveguide, which in turn causes changes in the observed amplitude of VLF transmissions. Variations in time and space in the lower boundary (e.g. reflections from mountain ranges (Bar and Armstrong, 1996)) of the waveguide can also lead to changes in the VLF propagation. Lightning discharges lead to changes in characteristics of the waveguide and thus perturbations in the received phase and amplitude of VLF transmissions (Rodger, 2003). The D-region of the ionosphere is quite stable in the daytime with reflection heights ranging from 70 – 75 km (McRae and Thomson, 2000). Variations are mostly due to changes in the Lyman-$\alpha$ flux with solar zenith angle (McRae and Thomson, 2000). Other factors affecting the D-region are solar flares (Thomson and Clilverd, 2001), total solar eclipses (Clilverd et al., 2001) and long-term changes due to solar cycle changes (Thomson and Clilverd, 2000). Thomson (1993) used VLF field measurements to determine $H'$ and $\beta$ parameters. Thomson and Clilverd (2000) used field strength measurements from a number of VLF
transmitters to examine changes in the daytime attenuation rates during a solar cycle (1986 – 1996). The lightning detection system used in this work provided an opportunity to measure amplitude variations of two strong transmitter signals during the period January 2004 – June 2004.

4.2 Experimental Data and Analysis

The experimental setup used for ELF-VLF recording has been described in chapter 3. Using the setup, the sferics data are recorded everyday for two minutes every hour during February 2004 – July 2004. The recordings were initially (September 2003 – January 2004) conducted for 20 min at predefined times (00 hrs, 01 hrs, 02 hrs, 03 hrs, 20 hrs, 21 hrs, 22 hrs, 23 hrs LT). The lightning software records data in one file of 11 MB per minute. The files are analyzed using a MATLAB code via the Matlab software, which produces a dynamic spectrogram of one-second duration. Tweeks atmospherics are identified from spectrograms and then analyzed in two steps. In the first step, the cutoff frequency \( f_{cn} \) and two other close frequencies \( f_1 \) and \( f_2 \) near the cutoff and corresponding times \( t_1 \) and \( t_2 \) from the trace are determined graphically using Microsoft Paint software. The spectrograms produced are in bmp format and uploaded into Microsoft Paint. The Paint software displays spectrograms on an identical coordinate system with arbitrary values. Figure 4.1 shows the Paint display screen of a spectrogram. The frequency and time components from the tweeks are determined in arbitrary units. The second step involves calibration of these coordinates in Microsoft Excel spreadsheet to produce the actual frequencies and times. A program is developed in excel to convert these coordinates to the actual frequency and time. The resolution from the graphical analysis in the frequency and time components are 70 Hz and 2 ms respectively.
Ionospheric parameters such as reflection height and total propagation distance are then calculated using equations 2.35 and 2.54. The whistler analysis is done in a similar way. The VLF output from the SU is connected to a spectrum analyzer that provided the frequency spectrum of the received signals at our station. Figure 4.2 shows the signal viewed from the spectrum analyzer. The two VLF transmitter signals monitored were centered one at a time and recorded for signal strength in dBμV every hour during the daytime manually. This value represents the signal strength at the spectrum analyzer, after the preamplifier. The spectrum analyzer was set at the following configuration: SPAN 10.0 kHz, REF (reference level) 77.0 dBμV, SWP (sweep time) 670 ms, RBW (resolution bandwidth) 300 Hz, VBW (video bandwidth) 100 Hz. Four quick readings were taken for every measurement. The hourly readings were averaged and plotted to show (see figure 4.11) the daytime variation of both the signals.
Sunset and sunrise measurements were made at 15 min intervals for few selected days. The measurements from the navigational transmitter signals are only preliminary and limited to the study of diurnal variation of signal strength of 19.8 kHz and 21.4 kHz from NWC and NPM VLF transmitters respectively.

4.3 Results and Discussion

4.3.1 Occurrence of tweeks and determination of $h$ and $D$

The occurrence pattern of tweeks was determined from the measurements made during September to December 2003. No tweeks were from the daytime data so data were analyzed for nighttime only. A total of 2428 tweeks were observed from the recordings during this period. The tweeks were classified into two categories; pre-midnight and post-midnight. Pre-midnight tweeks are those that occurred between 18 – 00 hrs LT. Post-midnight tweeks occurred between 00 – 06 hrs LT. Table 4.1 shows the statistical analysis of tweek occurrence.
Table 4.1: Statistical analysis of tweek occurrence from measurements during the period September – December 2003.

<table>
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<th>Harmonics</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
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<td>Occurrence</td>
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<td>194</td>
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<td>0</td>
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<td></td>
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<td>27.83</td>
<td>2.45</td>
<td>0.14</td>
<td>0</td>
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<tr>
<td>Post-midnight</td>
<td>Occurrence</td>
<td>1104</td>
<td>526</td>
<td>87</td>
<td>18</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>% Count</td>
<td>63.41</td>
<td>30.21</td>
<td>5.00</td>
<td>1.03</td>
<td>0.23</td>
</tr>
</tbody>
</table>

Tweek sferics occur throughout the night, but more in the post-midnight period. Higher harmonic tweeks \((n \geq 2)\) occur more often in the post-midnight period than in the pre-midnight period. Because of the large number of tweeks observed at our station, the analysis for the reflection height and propagation distance was done using selected tweeks (503) observed during September 2003 – July 2004. The selections of tweeks were based on the clarity of dispersion observed on the spectrogram. This selection includes a total of 320 tweek events analyzed in Table 4.1. Figure 4.3a – j show some examples of high harmonic tweeks recorded at Suva. The reflection height \(h\) and total propagation distance \(D\) calculated for tweeks as shown in figure 4.3a – j are depicted in table 4.2. The \(h\) varies in the range 80 – 93 km and \(D\) in the range 400 km – 12000 km. From the overall analysis the maximum propagation distance calculated from a tweek was 11837 km (see figure 4.2d). The average propagation distance was 2251 km. Figure 4.4 shows the occurrence rate for propagation distance from the selected tweeks in this analysis. The maximum occurrence rate was observed in distance range 2000 – 3000 km at 33.6 % (169 tweeks). The larger propagation ranges were measured in the smaller fractions but contributed a significant amount from the total tweeks analyzed. In total 47.1 % (237) tweeks measured distances greater than 3000 km. Using similar method, Kumar et al. (1994) found \(D\) to vary from 1500 km to around 2600 km for the tweeks observed in India. The ionospheric reflection heights in this work are comparable with that of Kumar et al. (1994) and Singh and Singh (1996).
Figure 4.3 (a – e): Selected multimode tweek sferics observed in the nighttime.
WB200312091302 stands for year 2003, month 12, day 09, time 1302 UT.
Figure 4.3 (f – j): Selected multimode twepoch sferics observed in the nighttime.
Table 4.2: Ionospheric reflection height \( h \) and propagation distance \( D \) estimated from tweek sferics observed in spectrograms a – j in figure 4.3.

<table>
<thead>
<tr>
<th>Spectrogram</th>
<th>Time  (LST)</th>
<th>Mode number ( n )</th>
<th>Earth-ionosphere Waveguide height ( h^* ) (km)</th>
<th>Total Propagation Distance ( D^{**} ) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>0102 am</td>
<td>1</td>
<td>86</td>
<td>4512</td>
</tr>
<tr>
<td></td>
<td></td>
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<td>88</td>
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<td>b</td>
<td>0102 am</td>
<td>1</td>
<td>83</td>
<td>3185</td>
</tr>
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<td></td>
<td>2</td>
<td>88</td>
<td>3100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3</td>
<td>92</td>
<td>3432</td>
</tr>
<tr>
<td>c</td>
<td>0108 am</td>
<td>1</td>
<td>84</td>
<td>2252</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2</td>
<td>87</td>
<td>2008</td>
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<td></td>
<td>3</td>
<td>88</td>
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<td></td>
<td>4</td>
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</tr>
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<td>j</td>
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<tr>
<td></td>
<td></td>
<td>4</td>
<td>86</td>
<td>481</td>
</tr>
</tbody>
</table>

* Height, \( h \) calculated using equation 2.35.

** Propagation distance, \( D \) calculated using equation 2.51.
However, the distance of tweeks reported here are higher than that reported by Kumar et al. (1994) and Singh and Singh (1996). It is because of different path conditions in the atmospheric waveguide where the lower boundary of the waveguide is Earth for the Indian observations but largely sea for observations made here. The average cutoff frequency and average reflection heights estimated from the analysis of 503 selected tweek sferics are presented in figure 4.5 and 4.6 respectively. The analysis shows that the average cutoff frequency for the first harmonic is 1.8 kHz, which varies as the traveling path conditions and the time of the day changes. Figure 4.5 shows the average cutoff frequency ($f_c$) of tweeks upto 6th harmonics and the variation of the estimated middle base cutoff frequency ($f_c/n$) with mode number ($n$). The $f_c/n$ varies between 1.7 kHz – 1.8 kHz. It is evident that higher harmonic cutoff frequencies ($n > 1$) are not exact multiples of the mode number. They are slightly lower as shown by the trend line of $f_c/n$. The cutoff frequency depends on the reflection height, which in turn depends on the electron density. The negative gradient of $f_c/n$ trend line suggests the presence of an electron
Figure 4.5: Variation of middle base cutoff frequency with mode number, $n$. The error bars indicate standard error in the estimated frequency. The numbers in the parenthesis “( )” indicates the total number of tweeks analyzed.

Figure 4.6: Middle base ionospheric reflection height with mode number, $n$. The error bars indicate standard error in the estimated height. The numbers in the parenthesis “( )” indicates the total number of tweeks analyzed.
density gradient at the reflecting height of the ionosphere. It also indicates that the higher order waves penetrate slightly deeper into the lower ionosphere while traveling to the receiver by multiple reflections. The $h$ calculated from the cutoff frequency for different modes using Eq. 2.35 for a perfectly conducting Earth-ionosphere waveguide is found to vary in the range 78 – 95 km during the nighttime. The calculated $h$ from $f_c$ of modes $n = 1$ to $n = 6$ is shown in figure 4.6. The ionospheric reflection heights increase with increase in $n$. This result confirms that the ionosphere is not sharply bounded and is not homogeneous indicating the presence of an electron density gradient. Shvets and Hayakawa (1998) estimated the steepness of the electron density profile from the fundamental cutoff frequencies of multimode tweek sferics. There results also showed a negative gradient for $f_c/n$ indicating an increase in the electron density from 28 – 224 cm$^{-3}$ in the altitude range of 2 km at the reflection height of 88 km. The temporal variations of $f_c$ and $h$ are shown in Figure 4.7. The plots have been made only for $n = 1 – 3$ harmonic tweek sferics due to the small number of tweeks observed for the higher harmonics. The $f_c$ for $n = 1 – 3$ decrease during 18 – 23 hrs LT. For the first harmonic, $f_c$ decreases up to 22 hrs and then recovers with a maximum at around 00 hrs and then gradually decreases up to 0500 hrs. The second and third harmonic $f_c$ shows a periodic type variation with maximum at around 18 hrs. The ionospheric reflection heights vary in the same way as the cutoff frequencies, since they are related by the expression $h = nc/2f_c$, i.e. when $f_c$ increases $h$ decreases as shown in figure 4.7.
Figure 4.7: Temporal variation of cutoff frequency \( (f_c) \) and reflection height \( (h) \).
4.3.2 Attenuation of ELF-VLF sferics propagating in partially conducting Earth/sea-ionosphere waveguide

![Graph showing attenuation of sferics](image)

**Figure 4.8:** Attenuation of tweek sferics propagating in a waveguide formed by (a) the Earth (ground) and the lower ionosphere (b) sea surface and the lower ionosphere.

To explain the occurrence of tweeks at this station, attenuation factor, $\alpha_n$, is calculated for modes $n = 1 – 3$ using equation 2.69, $\alpha_n = \frac{8.86 \times 10^3 \times \omega \times |A|}{c k h X_n}$. In this Equation $X_n$ is
a function of mode number and frequency, and the refractive index of the reflecting boundaries of the waveguide. Equations 2.54 and 2.55 clearly show that for a perfectly conducting Earth/sea-ionosphere waveguide, $\Delta$ becomes zero. The attenuation factor, $\alpha_n$ of each mode vanishes and higher harmonic tweeks should be observed at any time. In practice, tweeks are not observed always and harmonics higher than third occur less often (Kumar et al., 1994; Singh and Singh, 1996). Our results show that the Earth, sea and ionospheric boundaries are not perfectly conducting and the propagation path of the received tweek sferics provide sufficient conductivity to reduce attenuation allowing higher harmonic tweeks to occur more often. In this analysis the parameters used are; ionospheric dielectric constant $\varepsilon_i/\varepsilon_o = 0.5$, ground dielectric constant $\varepsilon_g/\varepsilon_o = 20$, sea dielectric constant $\varepsilon_s/\varepsilon_o = 81$, ground conductivity $\sigma_g = 10^{-4}$ mhos/m, sea conductivity $\sigma_s = 5$ mhos/m. The ionospheric reflection heights $h$ for daytime were taken as 71 and 74 km and for nighttime as 86 km and 90 km. The lower ionosphere is characterized as a “Wait ionosphere” defined by a reference height, $H'$ in km and the exponential sharpness factor, $\beta$ in km$^{-1}$ as considered by McRae and Thomson (2000). The ionospheric conductivity $\sigma_i$, has been calculated using the height dependent conductivity parameter $\omega_i(h) = \omega_o^i(h)/\nu(h) = 2.5 \times 10^5 \exp[\beta(h-H')]$ where $\omega_o^i(h)$ and $\nu(h)$ are the electron plasma and the effective electron collision frequencies at the reflection height $h$ (km). For daytime and nighttime conditions $H'$ and $\beta$ are 70 km, 0.3 km$^{-1}$ and 80 km, 0.5 km$^{-1}$ respectively. The magnitude of $\omega_i(h)$ used in this work range from $3.4 \times 10^5$ in the daytime to $3.7 \times 10^7$ in the nighttime. It can be seen from figure 4.8a that the attenuation increases sharply as the frequencies approach the cutoff frequency. Attenuation also increases when the reflecting layer height is lowered for all the modes. Therefore the attenuation is higher in the daytime than that in the nighttime. Figure 4.8b shows a plot of $\alpha_n$ versus $f$ for the waveguide formed by sea surface and the lower ionosphere. The frequency dependent ocean conductivity has been assumed to be 5.0 mho m$^{-1}$ (Westerlund, 1974). The attenuation is larger in the daytime ($h = 71$ and
74 km) than that during nighttime \((h = 86 \text{ and } 90 \text{ km})\) and increases sharply as frequency approaches the cut-off frequency. Wait (1958) studied the attenuation rate for perfectly conducting ground and partially conducting ionosphere \(\omega_r(h) = 0.47 \times 10^5 - 3.1 \times 10^9\) for different values of ionospheric reflection height. Wait (1958) calculated attenuation factors to vary between 0.5 - 16 dB/1000 km for frequencies 4 to 16 kHz at \(h = 90 \text{ km}\). Hayakawa et al. (1995) studied the attenuation rate at the frequencies below 3.0 kHz for \(0^{\text{th}}, 1^{\text{st}}\) and \(2^{\text{nd}}\) order TM modes. They defined the cut-off frequency as that at which the attenuation rate exceeds 30 dB/Mm. In this work the attenuation has been calculated for \(n = 1, 2, 3\) for both day and nighttime conditions and for Earth and sea-ionosphere waveguides. The results of attenuation are consistent with that reported by Hayakawa et al. (1995) for Earth-ionosphere waveguide. Sukhorukov (1996) studied the attenuation rate of the QTE1 mode in the cut-off frequency region (1.5-1.9 kHz) under the night-time ionospheric conditions and found the attenuation rate to be markedly high at the frequency around 1.6 kHz. The average cutoff frequency for the first harmonic tweek through measurements in this work is 1.8 kHz and attenuations calculated below this frequency are very high. The geographical location of our station is a small island in the South Pacific region. On a global scale, the total area of the island is very small and is surrounded by ocean waters to large distances. Tweek sferics propagate through a significant portion in the waveguide formed by the sea and the lower ionosphere. The comparison of attenuation shown in figure 4.8a and b indicates that the lower attenuation is offered by the waveguide between the sea and the ionosphere than that between the earth and the ionosphere for all modes both during day and nighttime. This could be one of the factors for the more probable occurrence of tweeks and their higher harmonics compared to those reported where the Earth surface dominates as lower wall of the waveguide in the total propagation path.
4.3.3 Whistler analysis

The ELF-VLF data were also analyzed for whistler occurrence. The analysis show that the occurrence of whistler at low latitude is very low. There had been only one whistler of class 2 (initiating sferic but no nose) in the entire set of data analyzed, which is presented in figure 4.9.

![Figure 4.9: Whistler observed on 22 November 2003 at 00:11:22 hrs LT.](image)

\[ t_o = 0.34 \]

The whistler shown in figure 4.9 was recorded on 22 November 2003 at 00:11:22 hrs LT. The twoek atmospheric at time \( t_o = 0.34 \) s may have initiated this whistler. The whistler nose frequency does not appear in the spectrogram. Observations made by Storey (1953) confirmed that the quantity \( f^{-1/2} \) increased linearly with time. Whistlers which followed atmospherics were found to intercept the \( f^{-1/2} \) against \( t \) graph at the time \( (t_o) \) of the causative atmospheric (Storey, 1953). The well known Eckersley’s approximation for dispersion of whistlers is the relation between travel time, \( t - t_o \) (\( t_o \) is the time of causative sferics) and the wave frequency \( f \) (Helliwell, 1965) as follows:

\[
\tau = t - t_o = \frac{D}{\sqrt{f}} \quad 4.1
\]

where \( D \) is the constant called dispersion. Dowden and Allcock (1971) introduced the function
which was more convenient for approximating the whistler dispersion. A plot between $Q(f)$ and $f$ produces a straight line and was also approximated to be (Dowden and Allcock, 1971):

$$Q(f) = \frac{1}{D(f)}$$

4.2

where $\alpha = 3.09 \pm 0.04$, and $Q_o = 1/D_o$ at zero frequency. An intercept on the $f$ axis gives $f_o = \alpha f_n$. From the intercept the nose frequency, $f_n$ is determined. This technique to determine the nose frequency is particularly useful for Eckersley whistlers. To remove the slight nonlinearity in the function $Q(f)$, Dowden and Allcock (1971) estimated an error in the determination of $f_n$ given by the equation:

$$\Delta f_n = 0.12 \left( \tilde{f} - 0.75 f_n \right)$$

4.4

where $\tilde{f}$ is the whistler mean frequency. The electron cyclotron frequency, $f_{He}$ is obtained using $f_n$ (Sazhin et al., 1992):

$$f_{He} = \frac{f_n}{0.4}$$

4.5

The regions inside the magnetopause are generally described in terms of the $L$-shell parameter or McIlwain parameter determined by $L = N_{Re} / \cos^2(\Phi)$ where $N_{Re}$ is the number of Earth radii and $\Phi$ is the geomagnetic latitude of the station. The $L$ gives the radial distance from the center of the Earth (in units of Earth radii) to the minimum $B$ along the line intersecting the magnetic equator. Hence, $L = 1$ means the point is at the equator on the surface of the Earth. The $L$ value of the field-aligned duct can be calculated using $f_{He}$ (Sazhin et al., 1992) and is given by

$$L = \frac{9.56}{f_{He}^{1/3}}$$

4.6

where $f_{He}$ is in kHz from the measured value of $f_n$. The magnitude of the magnetic field, $B_o$ at the Earth’s equator is calculated using the equation (Ratcliffe, 1972) and is given as
The dispersion graph \( t \) vs \( f^{-1/2} \) of the whistler is shown in figure 4.10. The \( f \) and \( t \) were estimated from the whistler spectrogram with an error of 70 Hz and 2 ms respectively. From the graph it is evident that the quantity \( f^{-1/2} \) increases linearly with \( t \). The intercept on the time axis occurs at \( t_0 = 0.331 \) s. However, there is a very strong tweek sferic at 0.340 s and is believed to be the causative sferic. The intercept on the time axis at 0.331 s suggests that the time of the lightning stroke is about 9 ms earlier than the sferic arrival time. The dispersion \( D \) computed using Eq. 4.1 is found to be 12.5 \( s^{1/2} \). Dispersion of low latitude whistlers has been reported in the range 10 \( s^{1/2} \) to 70 \( s^{1/2} \) (Singh and Hayakawa, 2001). Hayakawa and Tanaka (1978) obtained an empirical relationship

\[
f_{He} = \frac{eB_o}{m_e}
\]

where \( e \) is the electron charge and \( m_e \) the electron mass. For a dipole field, the magnitude of magnetic induction \( B_{eq} \) at the value of \( L \) is given by (Sazhin et al., 1992)

\[
B_{eq} = \frac{B_o}{L^3}
\]
between geomagnetic latitude and dispersion based on nighttime whistlers data recorded at the Japanese station. This relation is \( D = 1.22 (\Phi - 0.72) \) where \( \Phi \) is the geomagnetic latitude in degrees. From this relation the maximum dispersion of whistlers observed at Suva should be about 25 \( \text{s}^{1/2} \). This suggests that whistlers with dispersion higher than this are high latitude whistlers. The whistler observed at our station is therefore a low latitude whistler. Dispersion analysis is often used to determine electron density and large-scale electric field distribution in the inner plasmasphere (Singh, 1993). However, with the single whistler observed at this station, such an analysis is not possible. Whistlers are known to be used as novel diagnostics tools to magnetosphere (Hayakawa et al., 1995). The magnetospheric parameters calculated using equations 4.4 – 4.8 are summarized in table 4.3

### Table 4.3: Magnetospheric parameters estimated from the whistler spectrogram.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Calculated values</th>
</tr>
</thead>
<tbody>
<tr>
<td>( f_o )</td>
<td>547.489 kHz</td>
</tr>
<tr>
<td>( f_n )</td>
<td>177.181 kHz</td>
</tr>
<tr>
<td>( \Delta f_n )</td>
<td>- 15.088 kHz</td>
</tr>
<tr>
<td>( f_n + \Delta f_n )</td>
<td>162.093 kHz</td>
</tr>
<tr>
<td>( f_{He} )</td>
<td>405.234 kHz</td>
</tr>
<tr>
<td>( L )</td>
<td>1.3</td>
</tr>
<tr>
<td>( B_o )</td>
<td>2.305 ( \mu ) T</td>
</tr>
<tr>
<td>( B_{eq} )</td>
<td>1.069 ( \mu ) T</td>
</tr>
</tbody>
</table>

The corrected nose frequency is 162.093 kHz. This frequency is well above the pass band of the receiver, which is only 24 kHz and therefore not observed. At low latitudes, whistler data have not been used for determining ionospheric parameters because the propagation paths of low-latitude whistlers cannot be determined from their dynamic spectra on frequency-time spectrograms, since the nose frequencies are higher than 100 kHz (Singh and Hayakawa, 2001).
To explain the propagation mechanism of the observed whistler, the lightning activity detected by the WWLLN system in the conjugate area and the $L$ parameter calculated from the nose frequency of the whistler is considered. The lightning occurrence data provided by the WWLLN was analyzed to find the causative stroke. A lightning stroke was detected on 22 November 2003 at 00:11:22.263 LT, located at the geographic lat. 32.3009° N (geomagnetic lat. 28.82° N), long. 181.3029° E provided the best match for the whistler detected at the station. The $L$ value of the estimated lightning stroke is 1.3. The estimated time of the source stroke from the whistler analysis is 00:11:22.331 LT. Our station is located at geomagnetic lat. 21.2° S which is not too far from the conjugate point of the stroke. Furthermore, the $L$ value of our station is 1.2 which is at the boundary of low latitude. We therefore assume the stroke to be the causative stroke for the whistler observed. The difference in geomagnetic latitude of 28.82° N – 21.2° S indicates the mixed mode propagation. The mixed-mode propagation occurs when the signal propagating in the field aligned duct escapes at the conjugate point and then propagates in the Earth-ionosphere waveguide before being received or first travels in the geomagnetic field duct and then in the Earth-ionosphere waveguide.

Although there have been controversy over the propagation mechanism of low-latitude whistlers in the ionosphere and magnetosphere over the years, the method adopted here is valid for propagation in the ducted mode. The $L$ value of 1.3 estimated from the whistler data and the $L$ value of the matched lightning stroke location of 1.3 suggests ducted propagation. However various studies have been proposed against ducted propagation of low latitude and equatorial whistlers (James, 1972; Cerisier, 1973; Singh and Tantry, 1973; Hayakawa and Iwai, 1975; Tanaka and Cairo, 1980). Singh and Tantry (1973) provided the first evidence against the ducted mode of propagation for low-latitude whistlers. Using ray-tracing computations in a realistic model of a field-aligned duct, they showed that the trapping of whistler waves at an altitude of 300 km and 25° latitude required a 400 % increase in the ionization density along the central field line above the background density. They concluded that since such a high enhancement of ionization density is unlikely, ducted propagation for low-latitude whistlers is
improbable. Hayakawa and Iwai (1975) obtained similar results from their ray tracing computations, and suggested that the necessary enhancement could be provided by the equatorial anomaly. Singh and Hayakawa (2001) proposed a new type of propagation mechanism, known as the pro-longitudinal (PL) mode for low-latitude whistlers. The PL-mode is defined as involving propagation in which the wave normal angle along the ray path is always less than a characteristic wave normal angle for which the component of the refractive-index vector along the magnetic field line is minimum. Their results indicated that for low and very low-latitude whistlers, ducted propagation is not necessary because all the major characteristics of such whistlers are well reproduced by the non-ducted propagation in the presence of horizontal or vertical gradients in the ionization density. In this analysis the $L$ value of lightning occurrence in the conjugate area and the $L$ value of propagation calculated from the whistler spectrogram are the same. Assuming the whistler to be associated to this lightning the analysis suggests that the whistler propagated along the geomagnetic field duct from the point of discharge to its conjugate point and then traveled along the Earth-ionosphere waveguide to our station.
4.3.4 Analysis of 19.8 kHz and 21.4 kHz transmitter signals

The 19.8 kHz and 21.4 kHz transmitters are situated in North West Cape (NWC), Australia (21.82° S, 114.16° E) and Lualualei, Oahu, Hawaii (NPM) (21.43°N, 158.16° W), and radiate 1000 kW and 500 kW power respectively. The signal from the NPM transmitter has an east-west component in its propagation path compared to the signal...
from NWC, which is mostly west-east propagation. The great circle paths (GCP) for NPM and NWC are shown in Figure 4.11. The propagation distance to our station is 5.4 Mm for NPM and the 7.4 Mm for NWC. The amplitudes of NWC and NPM transmitters above \( 1 \mu V \) in dB were measured using a spectrum analyzer between 08-18 hrs LT during the months of January to June 2004. The amplitudes were taken every hour, and three values were taken and then averaged to avoid the amplitude variations due to lightning. Short-term (10 – 100 ms) fluctuations could not be recorded since the measurements were done manually. On average the received signal strength for NWC is 60 dB\( \mu V \) and the signal from NPM is 56 dB\( \mu V \). This corresponds to the received signal ratio of 1.071. The ratio of the transmitter power (NWC/NPM) of the transmitter signals to our station is 2. These ratios would have been the same had the loss only been due to the propagation distance. But the propagation path of NWC is over Earth and sea whereas for NPM it is mostly sea. As shown in section 4.3.2 the path over Earth offers more attenuation than over sea. The signal from NWC suffers more attenuation as it propagates in the mixed Earth/sea-ionosphere waveguide than the NPM signal, which propagates mostly in the sea-ionosphere waveguide. Also the distance of propagation of NWC is higher than that of the NPM signal. Hence the received signal strength of NWC is lower resulting in the reduced signal ratio. Figure 4.12 shows daytime variation in the observed amplitudes. The error bars indicate standard deviations. NPM to Suva shows a smoothly varying signal with a maximum at 12 hrs LT and a decrease in the afternoon (15 – 18 hrs LT). The signal from NWC to Suva shows a fairly constant trend between 11 – 17 hrs LT. There is a minimum at 09 hrs for NWC. Both show maximum signal strength between 12 – 14 hrs LT. The sunrise effect is more prominent at NWC whereas the NPM shows a prominent sunset effect. The daytime amplitude variations for NWC peak later than 12 hrs LT compared to the NPM signal, which peak at 12 hrs LT. The NPM signal has north-south component in its propagation path.
McRae and Thomson (2000) made similar observations. They observed signals from Omega Japan (10.2 and 13.6 kHz) and Omega Hawaii (10.2 and 13.6 kHz) to Dunedin (170.5° E) and showed that the west-east propagation path (Omega Japan) peaked later than 12 LT while the signals from Hawaii arriving from the east peaked before 12 LT. Figure 4.13 shows sunset effects on measurements taken at 15-minute intervals on 16 May 2004.
Figure 4.13: Sunset effects on NPM (21.4 kHz) and NWC (19.8 kHz) signals to Suva.

The signal strengths are reduced during the sunset transition between the transmitter and receiver. It can be noted that the minimum occurred at around 1745 hrs for NPM where as at around 1845 hrs for NWC. The minimum in NPM is ahead of NWC, which can be attributed to the change in height of the waveguide at the day/night terminator. Figure 4.14 shows the sunrise effects in 19.8 kHz signal from NWC Australia.
There is a sharp decrease in the signal during the day/night transition. Similar trends are shown on measurements done by McRae and Thomson (2000) for the signals from Omega Japan (10.2 kHz, 13.6 kHz) and Omega Hawaii (10.2 kHz, 13.6 kHz) to Dunedin, New Zealand. Lynn (1967) studied VLF transmissions from NLK to Smithfield, South Australia. He observed the signal strength decreased during sunrise and sunset transitions. Crombie (1964) outlined a waveguide theory to explain observed signal fading and phase variations due to sunrise – sunset effects. He explained these variations by considering mode conversion at the discontinuity in the earth-ionosphere waveguide between the day and night sections of the path. VLF radio signals are subjected to vector interference that can produce deep minima in the signal strength as seen at 6.25 LT in Figure 4.14. These signal minima are called modal interference nulls because they occur as a result of destructive interference between modes propagating in the Earth-ionosphere waveguide. The variation shown in Figures 4.13 and 4.14 may be due to the variations in the structure of the lower ionosphere during path transition from night to day and vice versa. The decrease in signal strengths as reported by Crombie (1964) and Lynn (1967) due modal conversions or interference could be responsible for sunrise sunset effects observed here.
Chapter 5
Lightning Occurrence over Fiji

The equipment provided by WWLLN is primarily intended to identify the location of the lightning occurrence globally. The network’s main data processing center in Otago, New Zealand provides participating hosts monthly records of lightning, the network has identified. To complement the analysis of the sferics, the location of the points of origin of the sferics (i.e. lightning location) in the two larger islands of the Fiji islands are analyzed. The WWLLN has good location accuracy, however the detection efficiency is low (Jacobson et al. 2006). Hence the absolute values presented in this work should be treated with caution. This chapter presents a statistical analysis of lightning occurrence over Viti Levu (Latitude 17.25 - 18.25° S, Longitude 177.25 - 178.65° E) and Vanua Levu (Latitude 16.20 - 17.00° S, Longitude 178.45 - 180.05° E) for the years 2003 – 2004. There were a total of 8849 flashes recorded by WWLLN during this period. Lightning occurrences have been analyzed to study the seasonal and diurnal variations. There is a higher lightning activity in summer than in the winter months (local). Diurnal variation shows maximum flashes in the afternoon. The spatial distribution of lightning flashes over both the islands shows high lightning activity in the western parts. To analyze the dependence of lightning on rainfall, it is necessary to consider the rainfall in the area where the lightning is studied. Suva is one of the areas in Viti Levu which records high rainfall. The rainfall data for Suva is compared with lightning occurrence over Viti Levu. With no other lightning location detection system available for the Fiji group, the lightning data from LIS onboard the TRMM satellite recorded during 2003 – 2004 are used to compare with WWLLN lightning detections. The LIS detections were analyzed for the same geographical limits for Viti Levu and Vanua Levu given above. Although
the LIS does not provide continuous data for a given location, the observations provide a good comparison for large-scale pattern of total flashes.

5.1 Introduction

Lightning is a large electric spark between regions having oppositely charged particles, which is produced by the separation of charges within a growing thunderstorm. Lightning occurs by a series of electrical processes in which charge is transferred along a discharge channel. The charge transfer could occur within a thundercloud (intra-cloud discharge, IC), between cloud and the Earth’s surface cloud-to-ground, CG or ground-to-cloud discharge, between two different clouds intercloud or cloud-to-cloud discharge, CC or between a cloud and the air (air discharge). As opposed to IC lightning, the CG lightning is a major safety hazard. It was originally believed that WWLLN would only detect CG lightning as mentioned in Dowden et al. (2002). However recent work have shown that as much as half of the lightning locations provided by WWLLN are cloud flashes (Rodger et al., 2005). The WWLLN detects CG lightning from sferic signals in the VLF band, but there is an unknown fraction of contribution from IC flashes in this analysis.

Studies on lightning occurrence in the tropical areas are limited. Some of the studies conducted show diurnal and seasonal variations. Lopez and Holle (1986) analyzed 50,664 negative CG Flashes recorded in Florida during 1983 and found strong diurnal variation with a maximum occurrence between 14 – 15 hrs LT and a minimum occurrence between 09 – 10 hrs LT. In Darwin, Australia, Williams and Heckman (1993) found the peak activity at ~ 17 hrs LT. In Southern Germany the peak occurred at 16 – 17 hrs LT (Finke and Hauf, 1996). Orville et al. (1997) published results of geographical distribution of CG lightning flashes obtained for about 1 year in the tropical area of Papua New Guinea. They found a highest flash density of 2.0 flashes/km² per year. The diurnal distribution of lightning in Papua New Guinea showed two peaks at 15 hrs LT and 02 hrs LT. Pinto et al. (1999a) and Pinto et al. (1999b) reported on geographical distribution, time variation and the flash characteristics of CG lightning in southeastern Brazil. They found the lightning occurrence to be strongly dependent on season but not so with geographical topography. The CG activity during the summer was higher that that in the winter
season. More recently Kandalgaonkar et al. (2003) studied the diurnal variation of lightning activity over the Indian region using the LIS data and found a slightly earlier peak time at 10 hrs UT and another peak at 16 hrs UT (LT = UT + 5.5 hrs for the Indian region). The probable cause of variations in peak time and distribution may be associated with variations in the local mesoscale circulation. These mesoscale circulations are set up in response to the interplay of large scale circulation impacting on the diurnal cycle of insolation and the underlying topographic features of the region (Kandalgaonkar et al., 2003). Ramachandran et al. (2003) also studied spatial and diurnal variation of CG flashes over Viti Levu, Fiji from February to April 2003. Spatial distribution revealed a higher frequency of lightning occurrence in the western part of the island.

5.2 Data Analysis
Lightning strokes detected by WWLLN are mostly CG discharges. The lightning data consists of the location in latitude and longitude, a residual time error value which is indicative of the spatial uncertainty, and the number of stations involved in locating that stroke. Matlab codes are used to analyze the lightning data for the two larger islands of the Fiji group. Lightning occurrence during 2003 and 2004 has been analyzed for diurnal and spatial variations. The analysis for spatial distribution firstly involved dividing the two main islands into grids of area ~ 20 km × 20 km (0.2° × 0.2° resolution). Total CG strokes for each of these grids are then computed, and the spatial distribution is plotted in the form of contours. Contour plots are made using Matlab code. It is essential to note here that the WWLLN data indicated in this work are only a small and unknown fraction of the total global lightning activity. The detection efficiency for the South Pacific region is unknown. Thus the statistical analyses given in the sections that follow are completely relative and should not be considered as absolute values of the total global activity. A study by Rodger et al. (2005) indicated ~26 % CG detection efficiency and ~10 % IC detection efficiency for the South East Australian region.
5.3 TOGA Theory

The TOGA theory is described by Dowden et al. (2002) in detail. Since the lightning location network uses the TOGA to determine lightning locations, the theory is summarized here. The discharge current in a typical lightning return stroke reaches its maximum value in about 2 μs and decays to half the maximum in about 40 μs. This results in a short pulse of around 100 μs covering a wide band from ULF to optical frequencies. The source of the radiation is a current element. At distance, \( r \) from the lightning stroke and at time \( t \), the wave field can be expressed by its electric field, \( E \) given by

\[
E(r,t,\omega) = \sum A(\omega) \cos(\phi(\omega)) \tag{5.1}
\]

At any one Fourier component of frequency \( \omega \), the phase \( \phi(\omega) \) is given by

\[
\phi(\omega) = \omega t - k(\omega) r + \phi_0 \tag{5.2}
\]

where \( k \) is the wave vector and \( \phi_0 \), is an initial phase constant. \( k(\omega) = \beta(\omega) = \beta_s \).

From Eq. 2.38, it is clear that \( \beta_s \) is dependant on frequency. Differentiating Eq. 5.2 with respect to frequency at any time \( t \) and distance \( r \), gives

\[
\frac{d\phi}{d\omega} = t - r \frac{dk}{d\omega} \tag{5.3}
\]

From Eq. 2.46, \( d\beta_s / d\omega = dk / d\omega = 1 / v_g (\omega) \), and so

\[
\frac{d\phi}{d\omega} = t - \frac{r}{v_g (\omega)} \tag{5.4}
\]

By the definition of the group velocity, \( r / v_g (\omega) \) is the time \( t_g (\omega) \), taken by the wave group to travel from the lightning stroke to the receiver. This group travel time is dependant on frequency, but not so strongly if measurements are kept well above the Earth-ionosphere waveguide cutoff frequency of the dominant mode. Eq. 5.4 can be rewritten as

\[
t_g (\omega) = t - \frac{d\phi}{d\omega} \tag{5.5}
\]

Consequently, \( t = t_g (\omega) \) when \( d\phi / d\omega \) is zero.
TOGA is defined, as an absolute time in UTC, at which the group of waves is received at the station. Then if the lightning stroke occurs at an absolute time, \( t_s \), in UTC the relationship is,

\[
\text{TOGA} = t_o + t_s \omega \quad (5.6)
\]

However TOGA is determined at an absolute known time, \( t_o \), in UTC. \( t_o \) is known as the trigger time i.e. the time at which a lightning stroke is detected. Although \( t_o \) is an adequate substitute for TOGA in some applications, it introduces both random and systematic errors. Random errors of up to 20 µs arise because the trigger time is digitized in approximately 20 µs steps, the reciprocal of the sampling frequency. Some sound cards used by different host stations sample at 48 kHz, and some at 50 kHz. Systematic errors arise because the trigger threshold is reached earlier for strong sferics than of a weak sferic. The sferic from a lightning stroke is strongest at the nearest receiver and weakest at furthest. Therefore, an early trigger occurs at the closest receiver and a late trigger at the furthest. The trigger time is precisely defined as a certain point in the digitized VLF sferics waveform. It can be determined to within a few hundred nanoseconds. TOGA is obtained by adding a correction (positive or negative) to \( t_o \).

The sferic can be detected when the amplitude of the signal rises above a set threshold. An effective way of detecting sferics is by using the rate of change of the amplitude. The signal is sampled at about 50 kHz, and so consecutive samples are 20 µs apart. The magnitude of the difference between consecutive samples, \( i \) and \( i + 1 \) are monitored. When this difference is above a set threshold value the time of the second sample \( (i + 1) \) is recorded as the trigger time. This time is within 100 µs of the TOGA (Dowden et al., 2002). Substituting Eq. 5.5 in Eq. 5.6 gives

\[
\text{TOGA} = t_o - \frac{d\phi}{d\omega} \quad (5.7)
\]

where \( t_o = t + t_s \). \( t_o \) is a precisely known time determined from the GPS. The only confusion remains in the determination of \( d\phi/d\omega \) which can only be measured at a finite range of frequency. In fact the sferics have a 3 dB wide bandwidth of ~14 kHz with maximum amplitude at ~10 kHz. A band of 6 – 22 kHz has the highest sferics amplitude and so is used for phase measurement. The lower limiting frequency is well
above the EIWG cutoff having low dispersion and the higher limiting frequency is less than half the sampling frequency (25 kHz) satisfying the Nyquist criterion. Hence, \( \frac{d\phi}{d\omega} \) is calculated from the slope of a least squares regression line from the phase measurements in the frequency range 6 – 22 kHz. TOGA is determined from the phase slope, \( \frac{d\phi}{d\omega} \) at time \( t_0 \) from Eq. 5.7.

Lightning is then located from the TOGAs measured by other stations in the network. The difference in TOGAs from independent pairs of stations is used to locate the stroke. A stroke is registered only when a minimum of four stations have recorded it. The use of the TOGA algorithm came into effect after 31 July 2003. The lightning was earlier detected using TOA of the received sferics. In this analysis the data used includes both TOA and TOGA measurements during the period January 2003 – December 2004.

5.4 Results and Discussion

5.4.1 Diurnal and seasonal variation of lightning occurrence

The detection efficiency of the WWLLN is not known for all locations in the globe. Comparative study of WWLLN and NLDN (USA) estimate the detection efficiency of less than 1 % (Jacobson et al. 2006). These findings suggest that the detection efficiency is low compared to the other lightning location systems. This may be due to the fact that the threshold current for WWLLN is 26 kA (high) compared to that of NLDN (11 kA). The detection efficiency will determine the absolute number of strikes. It is however expected that the detection efficiency will not show any temporal variation. Hence the diurnal and seasonal variation of lightning occurrences will be still valid, even though the absolute counts of lightning flashes may vary. A total of 8849 lightning flashes over Viti Levu (Latitude 17.25 - 18.25° S, Longitude 177.25 -178.65° E) and Vanua Levu (Latitude 16.20 - 17.00° S, Longitude 178.45 - 180.05° E) were registered in the year 2003 and 2004, out of which 2620 were in the year 2003 and 6229 occurred in the year 2004. The low number of detections in 2003 could be attributed to the low number of detecting stations (11) at that time. With the increase in the number of stations (24 active stations at present) the detection efficiency increased and more flashes were detected in 2004.
The diurnal variation for the year 2003 and 2004 is shown in figure 5.1. The occurrence pattern shows that lightning activity is more prominent between 13 – 18 hrs LT. There is a maximum occurrence between 14 – 16 hrs LT and minimum between 07 – 08 hrs LT for both Viti Levu and Vanua Levu. In comparison these results are similar to those obtained by other workers given in section 5.1. The peak lightning activity in the afternoon period suggests a close association to the
convective activity inside a cloud after the ground surface is warmed by solar radiation.

Fiji is a tropical island nation in the South Pacific region. The seasons in a year are broadly classified as summer and winter. May to October and November to April are classified as winter and summer months respectively. Figure 5.2 shows seasonal variation of lightning occurrence over Viti Levu and Vanua Levu during the year 2003 and 2004. The flash counts are an average for the two years. As shown in figure 5.2, lightning activity is more pronounced during summer and less during the winter months for both the islands.

5.4.2 Spatial distribution

As stated earlier in 5.4.1 the detection efficiency of WWLLN is low. Even though one of the VLF receiver is situated in Suva, Viti Levu, this does not dictate higher detection efficiency around Suva since the location system is based on detecting the sferic signals at least at four sites. For the period presented in this thesis the other stations close to Fiji were Brisbane, Darwin, Perth (Australia) and Dunedin (New Zealand). For the stations in Australia, locations in Viti Levu and Vanua Levu are almost at the same distances. Hence we conclude that the WWLLN data is representative of the spatial distribution of lightning though the absolute values reported are expected to be higher. The lightning occurrence over Viti Levu and Vanua Levu in grid areas of 20 × 20 km is presented in the figures 5.3 and 5.4 respectively. The graphs show lightning contours estimated from flash counts at the center of each grid. The contour levels show an average lightning occurrence for two years (2003 and 2004). In figure 5.3, CG flash incidence in the west of Viti Levu is higher than that in the east of Viti Levu. The pocket of maximum flash is greater than 300 strokes. This is equivalent to a flash density of 0.8 flash km\(^{-2}\) per year. The flash density of 2.0 flash km\(^{-2}\) per year in Papua New Guinea is much larger than here. However it should be noted that the flash density of 0.8 flash km\(^{-2}\) per year determined here is only a small fraction of the total CG activity. The absolute value is expected to be much higher but yet unknown at this stage. The topography of this location does not reveal anything unusual. The altitude at 18° S, 177.5° E corresponding to the maximum occurrence is approx. 380 m.
Figure 5.3: CG Flash distribution over Viti Levu during 2003 – 2004. Total number of flashes included in this analysis was 6424 out of which 1832 occurred in 2003 and 4592 were recorded in 2004.

Figure 5.4: CG Flash distribution over Vanua Levu during 2003 – 2004. Total number of flashes included in this analysis was 2425 out of which 788 occurred in 2003 and 1637 were recorded in 2004.
There are mountainous terrain in the same vicinity e.g. at 17.75° S, 177.75° E the altitude is approx. 900 m. Ramachandran et al. (2005) studied the spatial distribution of CG Flashes over Viti Levu for the year 2003 and found two pockets of peak activity around the same areas. The contours in this work show an average for two years (2003 and 2004) and are strokes with residuals less than 300 µs representing a more accurate stream. The total CG Flashes recorded during 2003 with this criterion (< 300 µs) are far less (Viti Levu 1832, Vanua Levu 788) than those recorded during 2004 (Viti Levu 4592, Vanua Levu 1637). This may be due to the low detection efficiency of the system at that time. It was only after 31 July 2003 when the processing algorithm was modified to determine the sferic TOGA at each station thereby improving the network efficiency. Lightning was detected using TOA prior to this modification (Rodger et al., 2005).

The lightning distribution over Vanua Levu shown in figure 5.4 reveals pronounced lightning activity in the western part of the island. A maximum of 138 strokes occurred in the grid 16.6 – 16.8° S, 178.85 – 179.05° E, corresponding to a flash density of 0.35 flash km⁻² per year. The highest average flash density in Viti Levu was found to be 0.8 flashes km⁻² per year whereas in Vanua Levu the highest average flash density was 0.35 flashes km⁻². The total land area of Viti Levu (10,544 km²) and Vanua Levu (5,535 km²) is small compared to Papua New Guinea, which has a total land area of 452,860 km². On a global scale, the total land area of Fiji is very small and is often engulfed in cool oceanic winds. Accounting for the detection efficiency of the WWLLN, the lightning incidence in the two main islands is still small. A possible reason for the low flash density on both the islands is the cooling effect by these winds, which inhibit the formation of convective updraft reducing the stroke incidence.
5.4.3 Comparison of lightning events and rainfall

Figure 5.5: Monthly variation of lightning activity over Viti Levu and rainfall over Suva during 2003 – 2004.

Figure 5.5 shows the monthly variation of total rainfall over Suva during 2003 – 2004 and the lightning occurrence over Viti Levu. The Fiji Meteorological Services in Nadi provided the rainfall data used for comparison. The rainfall records in Suva show a seasonal pattern with slightly reduced levels during the winter months except for the month of August, which has a significantly high record. The month of November, which falls in the summer season, received significantly low rainfall in the two years.

The correlation between lightning flashes and rainfall has been studied in the past (Tapia et al., 1998; Fehr and Dotzek, 2003). Tapia et al. (1998) and Fehr and Dotzek (2003) obtained the spatial distribution of rainfall estimates from weather satellite radar observations and compared with lightning observations. They found striking similarities with the spatial distribution of rainfall and lightning. Figure 5.5 shows the seasonal variation of CG for the entire island of Viti Levu. To find the correlation between rainfall and CG flashes, the rainfall in Suva during the year 2003 – 2004 is considered. According to the Fiji Meteorological Services, Suva is one of the areas in Viti Levu, which receives high rainfall. Although the rainfall data is only for Suva, it can be used to infer the seasonal distribution over the entire island. There
appears to be good association between rainfall and lightning during the summer months but not so much during the winter months.

5.4.4 Comparison of WWLLN and LIS data

![Diurnal variation of LIS Lightning flashes during 2003 – 2004.](image1)

**Figure 5.6:** Diurnal variation of LIS Lightning flashes during 2003 – 2004. (Total LIS flashes Viti Levu – 148, Vanua Levu – 67).

![Monthly variation of LIS lightning flashes during 2003 – 2004.](image2)

**Figure 5.7:** Monthly variation of LIS lightning flashes during 2003 – 2004. (Total LIS flashes Viti Levu – 148, Vanua Levu – 67).
Figures 5.6 and 5.7 show the diurnal and monthly variations of lightning flashes obtained from the LIS passes over Viti Levu and Vanua Levu during 2003 – 2004. The LIS data are available at http://thunder.msfc.nasa.gov/data/lisbrowse.html. The total flashes were compiled and sorted for the geographical grids of Viti Levu and Vanua Levu. The LIS does not provide continuous data for a given location and so is unsuitable for studying localized weather. However, the two years of observations can be considered to give an approximate large-scale pattern of total flashes. The diurnal and monthly variations were studied from the LIS data. The diurnal variation shows enhanced lightning in the afternoon which is same as those found from WWLLN data [see Figure 5.1]. On Viti Levu, except for the month of August, no flashes were recorded during the winter months. There were no flashes recorded by LIS in November while maximum flashes were recorded in December. The monthly variation shows no lightning flashes over Vanua Levu between April to November and the maximum flashes in January. On both the islands lightning was prominent during the summer months. A comparison with WWLLN detection shows similarities in the diurnal and seasonal pattern, but the recordings made by LIS are very little during the two years. Ramachandran et al. (2005) compared the LIS data with WWLLN data over Viti Levu. They used five years of the LIS data and found enhanced lightning activity between November to April, which is similar to the two years of CG flashes obtained in this work from the WWLLN. A detailed analysis of the LIS data for 2004 and the data from WWLLN revealed that none of the lightning events recorded were coincidental. During the months of April to July 2004, the LIS recorded zero flashes while WWLLN recorded a total of 761 CG flashes over the two islands. The pass of LIS on 27 December 2004 in the time interval 02:43:00 – 02:45:00 UT recorded the highest count of 68 flashes on Viti Levu. The WWLLN on the same day recorded zero CG flashes at that time interval. There were a total of eight other passes in the year 2004, which recorded a total of 27 flashes on Viti Levu. On Vanua Levu the LIS recorded a maximum flash of 30 on the pass on 20 January between 0:41:00 – 0:43:00 UT. WWLLN recorded zero CG flashes on the same time. The geographic location and the time of occurrence of lightning seen by the LIS show no correlation with that obtained from WWLLN. However, it is interesting to note the similarity of the LIS lightning in the pattern with the seasonal rainfall in Suva,
indicating the presence of active clouds. A particular interest is drawn for the month of August as it is showing a pronounced LIS detection over Viti Levu, correlating to the pronounced rainfall over Suva for the same month. Light and Jacobson (2002) reported that satellites do not see lower altitude events, as the optical scattering losses are more severe. This may be a plausible reason for the discrepancy of the LIS and WWLLN data. The LIS recorded IC flashes happening at higher altitudes where as WWLLN recorded CG flashes as intended at low altitudes.
Chapter 6

SHF Ionospheric Scintillation at Suva

6.1 Introduction

Scintillation refers to the rapid fluctuations in the phase and amplitude of radio signals. Ionospheric irregularities also called plasma bubbles are the irregular structures in the electron density of the ionosphere formed due to plasma instabilities and cause scintillations on trans-ionospheric radio signals. The plasma density in the F-region increases with altitude up to about 300 km and then decreases. This situation produces high-density plasma on top of low-density plasma and is very unstable (Gravitational Rayleigh Taylor, GRT instability). If the equilibrium is disturbed lumps of low-density plasma rise and cause density irregularities known as plasma bubbles. These bubbles break into irregular structures when they reach the topside ionosphere forming a system of ionospheric irregularities (Kuo and Chuo, 2001) and produce ionospheric scintillation patches in the pre-midnight period at stations situated off the magnetic equator. Scintillation phenomenon was initially studied using stellar sources and nowadays mainly using geostationary and GPS satellites.

The Equatorial Ionization Anomaly (EIA) region is the ± 20° wide belt of latitudes around the geomagnetic equator with enhanced electron density (Kersley and Chandra, 1984). Suva (Geomag. Lat. 21.2° S) lies just outside the boundary of the EIA and is expected to receive scintillations even during low solar cycle and at high frequency (GHz) signals. It is generally agreed that at post-sunset hours, plasma bubbles are generated in the bottom side of the F-region via the GRT mechanism. The plasma bubbles rise into the topside of the ionosphere through non-linear evolution of $E \times B$ drift
where \( E \) is the electric field caused by the neutral winds in the ionosphere and \( B \) is the Earth’s magnetic field strength (Woodman and LaHoze, 1976; Dabas, et al., 1992; Kuo and Chuo, 2001). The zonal neutral winds and conductivity gradient at the time of sunset interact to develop an enhanced eastward electric field on the dayside of the terminator and a westward electric field on the nightside (Basu et al., 2002). The enhanced eastward electric field causes the F-region to move upward lowering the bottomside density gradient, which triggers the GRT instability. This electric field also leads to a resurgence of the EIA in the post-sunset period. Plasma instabilities that occur at the post-sunset in the equatorial region of the ionosphere are collectively known as the Equatorial Spread-F (ESF). Studies have shown that a well-developed EIA is also one of the conditions conducive for the generation of ESF (Raghavarao et al., 1988; Alex et al., 1989; Jayachandran, et al., 1997). Scintillation occurrence at VHF due to Spread-F instability is a common nighttime phenomenon with a maximum in the pre-midnight period (Kumar and Gwal, 2000).

Daytime random scintillations are thought to occur because of the sporadic E layer. These are dense layers or patches of ionization in the E-region at heights of 100 – 120 km. Plasma instabilities occur in the E-region of the equatorial ionosphere, due to the electrojet current which flows in this region whenever conditions are favorable (Bhattacharyya, 1991). Aarons and Whitney (1968) reported possible association of daytime scintillations and the occurrence of sporadic E patches with horizontal extents from 300 – 600 km. McClure (1964) found that midday scintillations were produced by irregularities in the E-region. Hajkowicz (1977) using multiple satellite transmissions at 150 MHz reported an increase in scintillation activity between 14 - 18 hrs LT in the winter associating with increase in the occurrence of sporadic E. Satellite in-situ data have shown that large irregularity amplitudes in regions of high ambient ionization density in the nighttime equatorial F-region can cause GHz scintillations (Basu and Basu, 1976). Paul et al. (2002) observed equatorial ionospheric irregularities from amplitude scintillations measured from GPS and geostationary “Fleetsatcom” (244 MHz, 73° E) and “Inmarsat” (1.5 GHz, 65° E) signals. Fujita et al. (1982) studied the frequency
dependence of ionospheric scintillations from observations of 1.7 GHz ionospheric scintillations at a mid-latitude station. Dabas et al. (1992) studied equatorial plasma bubble dynamics using scintillation observations at 4 GHz in the Indian region. Satellite transmissions in the frequency range higher than the GPS are rarely available. It requires an elaborate receiving system. Therefore scintillation measurements in the GHz frequency remain very limited. However, the availability of global Positioning Satellites (GPS) has prompted many researchers to scintillation measurements in the 1.5 – 1.7 GHz range.

In this chapter the morphological characteristics of ionospheric scintillation on 3.925 GHz beacon from geostationary satellite Intelsat (701) stationed at 180° E have been studied. The period of measurement falls in the low solar activity (average monthly sunspot number, SSN = 42.0).

6.2 Data Analysis

The amplitude of the beacon signal was recorded at a temporal resolution of 1 s during December 2003 to June 2004 using LabVIEW data acquisition software. The temporal analysis of scintillation can be broadly divided as daytime (06 – 18), pre-midnight period (18 – 00) and post-midnight period (00 – 06). The diurnal occurrence pattern is analyzed at 15 min intervals. Scintillation index \( S_4 \) is a parameter, which is indicative of the effects of the plasma irregularities. It is defined as (Briggs and Parkin, 1963)

\[
S_4 = \frac{1}{\bar{R}^2} \left[ \left( R^2 - \bar{R}^2 \right)^2 \right]^{1/2}
\]

where \( R \) is the instantaneous amplitude of the sampled data and \( \bar{R} \) represents the average over the 60 s period. The \( S_4 \) index can be used to classify the strength of scintillation. For the purpose of this work scintillations are classified as weak (\( S_4 \leq 0.25 \)), moderate (\( 0.25 < S_4 \leq 0.5 \)) and strong (\( S_4 > 0.5 \)) scintillations. The scintillation data are collected over an interval of 60 s for computing the scintillation index \( S_4 \).
6.3 Results and Discussion

The ionospheric scintillation occurrence on 3.925 GHz at Suva during December 2003 – June 2004 is found to be rare and sporadic. The onset of scintillations is abrupt with $S_4$ mostly less than 0.3. Scintillations with the $S_4$ index greater than 0.1 are considered for this analysis. Scintillations mostly occur in short duration patches. Figure 6.1 shows the scintillation occurrence pattern at Suva. The overall percentage occurrence of the scintillations at this frequency is highest during the daytime. From the total, 83.4 % of patches occurred during the daytime (756 patches). The nighttime scintillations are less frequent (150 patches) but stronger than the daytime. The nighttime occurrence is further divided into the pre-midnight and the post-midnight periods. The occurrence of patches in the pre-midnight period (9.7 %) is a little higher than that in the post-midnight period (6.8 %). The relative occurrence of weak, moderate and strong scintillations are shown in figure 6.2.
Figure 6.2: Relative occurrence of weak, moderate and strong scintillations.

There were a total of 645 patches (71.2 %) with $S_d \leq 0.25$ and 261 out of the 906 patches calculated $S_d$ index in the range $0.25 < S_d \leq 0.5$. Strong scintillations $S_d > 0.5$ were not observed. The scintillation patches are also categorized according to their duration of occurrence. Scintillations occurred in the patches of durations 1 to 10 min. Figure 6.3 shows the percentage occurrence of scintillation patches versus the duration of patches. The patch duration of scintillations are mostly similar at the different periods, as seen in figure 6.2a – c. The scintillation patches with duration 1 – 4 min occur most often, 5 – 7 min less, and 8 – 10 min the least. The scintillations greater than 10 min are not observed.
Figure 6.3: Percentage occurrence of scintillation ($S_4 > 0.1$) versus patch duration. a. Daytime. b. Pre-midnight period. c. Post-midnight period.
Figure 6.4: Selected scintillation patches observed at Suva. a. Scintillation patch duration 13 min in the daytime. b. 2 min scintillation patches in the pre-midnight period. c. 2 min scintillation patches in the post-midnight period. d. Typical scintillation patch with 5 min duration in the pre-midnight period.
Figure 6.4: Selected scintillation patches observed at Suva. e. Periodic scintillation patches in the daytime period. f. Periodic scintillations in the daytime period. g. Typical midday scintillation patches. h. Short duration scintillation patch with quite periods.
The post-midnight period shows higher percentage occurrence of scintillations with durations of 7 to 10 min than that in daytime and pre-midnight period. The scintillations observed mostly appear as several short duration patches interspersed with quite intervals. Figure 6.4 depicts typical examples of scintillation patches observed. The scintillations observed at the station are mostly weak with $S_4$ index $\leq 0.25$. The $S_4$ index varies from mostly 0.1 to 0.3. Two min scintillation patches are very common in all the three categories (daytime, post and pre-midnight) and are shown in figure 6.4b. These are mostly weak scintillations with few moderate ones with $S_4$ index no more than 0.3. In figure 6.4c, a post-midnight weak scintillation patch of 4 min duration is shown. The figure 6.4d shows a typical scintillation patch observed in the pre-midnight period. This is a 5 min patch with a moderate strength with maximum $S_4 = 0.33$. Similar types were rarely observed. Scintillations shown in figure 6.4e and f can be characterized as periodic scintillations. Such scintillations are generally weak and occur mostly during the daytime. Figure 6.4g and h show typical examples of scintillations observed in the midday. Scintillations observed during daytime are weak and of short duration.

The diurnal variation of scintillation occurrence in percentage for the period December 2003 – June 2004 is shown in figure 6.5. The occurrence is analyzed at 15 min intervals. The scintillation occurrence is very less with maximum of about 2 % at 09 hrs LT and 17 hrs LT. The occurrence is large in the daytime than that in the nighttime, which is opposite to the scintillations at VHF during high solar activity period. Both day and nighttime scintillations are weak and rarely moderate. It is generally agreed that, after the local sunset, the bottom side of the F-region over the magnetic equator is subjected to GRT mechanism. Moderate scintillations observed in the nighttime on 3 March 2004 shown in figure 6.4d is a typical example of scintillation associated with ESF irregularity. The magnitudes of scintillations decrease with increasing frequency (Basu et al., 2002).
Figure 6.5: Diurnal occurrence of scintillation with $S_4 > 0.1$ at Suva.

Scintillations at VHF are intense and almost saturated ($S_4 > 0.6$), whereas at the L band the intensity is moderate, $S_4 \approx 0.3$ (Rama Rao et al., 2005). The frequency of transmission used in this work comes in the C band (3.925 GHz) and scintillation observed are mostly weak ($S_4 \leq 0.25$) 71.2 %. Moderate scintillations are very few (28.8 %) and last few minutes while strong scintillations were not observed.

To explain the occurrence of scintillations observed at our station we look at the studies conducted by others researchers. Studies have shown strong dependence of scintillations on solar activity. At a frequency of 150 MHz the scintillation activity declines by about 40 % from the period of sunspot maximum to the period of sunspot minimum (Hajkowicz and Dearden, 1988). Kumar and Gwal (2000) studied VHF scintillation near the equatorial anomaly crest and found an increase in scintillation activity with increase in sunspot numbers. Dabas et al. (1992) studied equatorial plasma bubble dynamics using a 4 GHz telemetry signals from two geostationary satellites INSAT-1B (74 ° E) and INSAT-1C (94 ° E) at Sikandarabad (Geog. Lat. 28.48 ° N, dip 42.0 °) and Chenglepet (Geog. Lat. 12.07 ° N, dip10.5 °) low latitude stations in the Indian region. During low solar activity weak scintillations were observed at Sikandarabad in the summer. It is suggested that the scintillation producing irregularities in the GHz band
at Sikandarabad have their origin at the magnetic equator during equinoctial periods of high solar activity. The occurrence probability of scintillation at low latitude station is less than that at equatorial station (Dabas et al., 1992).

Daytime scintillations at our station are predominant over the nighttime scintillations. Daytime scintillations have been reported to be associated with Sporadic-E irregularities (Rastogi and Iyer, 1976; Das Gupta and Kersley, 1976; Rastogi et al., 1991). The daytime scintillation activity observed here shows two peaks; one at 09 hrs and the other at 17 hrs LT. Rastogi et al. (1991) observed daytime scintillations on VHF radio wave signal at a number of stations at equatorial latitudes in the Indian region. Their measurements showed a similar pattern in the daytime variation with scintillation activity starting after 06 hrs LT and peaking around noon then decaying by 18 hr LT. The daytime scintillation activity decreased rapidly with increase in the latitude. Hajkowicz and Dearden (1988) showed an increase in daytime scintillations during the years of sunspot minimum. The periodic type scintillations observed in figure 6.4e and f observed during the daytime is highly sporadic. Hajkowicz and Dearden (1988) showed two peaks in the occurrence of quasi-periodic (QP) scintillations, predominantly in the summer: in the late morning (08 – 10 hrs LT) and in the pre-midnight period (20 – 22 hrs LT). There was an increase in QP scintillations with decrease in the sunspot number. Although the type of scintillations observed in figure 6.4e and f were very few, they are believed to be non-symmetrical QP scintillations and QP scintillations superimposed on random scintillations. There occurrences were highly sporadic and mostly during the daytime period. There is however, very little information available on the occurrence of daytime scintillations at frequencies in the C band. Our results show that at the C band at 3.925 GHz the scintillation occurrence is very rare and sporadic with maximum occurrence during the daytime. It is due to a) the transmission frequency is very high at 3.925 GHz, b) the observing station is situated off the EIA belt and c) the observations fall in the low solar activity period.
Chapter 7
Summary and Conclusions

7.1 Summary
The instrumentation used in this study is primarily for lightning detection, setup in collaboration with WWLLN that uses the difference in TOGA of sferics received at pairs of stations to detect the location of lightning strikes. This is a relatively new and advanced system, which provides precise timing measurements and high-resolution spectrograms. With the lightning data provided by WWLLN, a statistical analysis of lightning occurrence over Viti Levu and Vanua Levu for the year 2003 and 2004 have been presented. Lightning measurements in the tropical areas are very limited and this study is the first of its kind for the Fiji islands. However conclusive predictions can be made only after a long period of statistical analysis.

The experimental setup was extended to study the characteristics of sferics produced by lightning. The tweek sferics, generated from lightning discharges have been used to determine the nighttime variation of ionospheric reflection heights. The total propagation distance of tweeks from lightning discharges have been computed. The attenuation factors of tweeks propagating in the Earth/sea-ionosphere waveguide have been calculated. Whistler sferics received at the station have been used to compute the magnetospheric plasma parameters. A possible propagation mechanism of the low latitude whistler obtained in this work is explained. Amplitude measurements of VLF transmitter signals namely NWC, Australia and NPM, Hawaii have also been used to study the effects of propagation path and the daytime ionosphere.

The measurements of 3.925 GHz beacon signal from Intelsat (701) were used to study the morphological features of scintillation at Suva, Fiji, a low latitude station
in the South Pacific region. The $S_4$ spectral index was computed to determine the strength of the scintillation patches observed.

### 7.2 Conclusions

The significant findings from ELF-VLF and SHF scintillation observations and analysis in this thesis can be concluded as follows:

- Tweek sferics occur in the nighttime between 18 – 06 hrs LT. No tweeks were observed during the daytime.
- First and second harmonic tweeks occur more often with cutoff frequencies around 1.8 kHz and 3.5 kHz respectively. The frequency of occurrence of tweeks decreases with increase in harmonic however tweeks of up to the 6th harmonic have been observed. The dispersed portion also reduces with increase in harmonics. The higher harmonics occur more often in the post-midnight period than that in the pre-midnight period.
- The ionospheric reflection height varies in the range 78 – 95 km during the nighttime.
- Tweeks have propagated distances in the range 600 – 20000 km in the waveguide formed by the Earth or sea and the lower region of the ionosphere.
- The finite conductivities of the Earth/sea-ionospheric waveguide reflection boundaries introduce attenuation to the ELF-VLF signals propagating within it. From the theoretical analysis, it has been found that the waveguide formed by the Earth surface and the lower ionosphere offers more attenuation than the waveguide formed by the sea surface and the lower ionosphere. The attenuation in the nighttime is much less than the attenuation in the daytime, which is the main reason for the occurrence of tweek sferics during the nighttime only.
- Whistler occurrence at the station is very rare. A whistler observed at the station was analyzed and found to have propagated in the mixed mode path. The various parameters calculated from the whistler analysis are as follows: corrected $f_n = 162.093$ kHz, $f_{He} = 405.234$ kHz, $L = 1.3$, $B_0 = 2.305$ $\mu$ T, and $B_{eq} = 1.069$ $\mu$ T.
The signal amplitudes of NWC (19.8 kHz) and NPM (21.4 kHz) are reduced considerably during sunrise (2 - 7 dBμV) and sunset (4 - 8 dBμV). These reductions may be due to the variations in the structure of the lower ionosphere during path transition from night to day and vice versa or modal conversions.

Diurnal variation of lightning occurrence over Viti Levu and Vanua Levu shows maximum between 14 – 16 hrs LT and a minimum between 07 – 08 hrs LT. The seasonal pattern shows maximum occurrence during summer months and minimum occurrence during June and July months.

The spatial distribution of lightning shows a maximum flash density of 0.8 flashes km\(^{-2}\) per year over Viti Levu and 0.35 flashes km\(^{-2}\) per year over Vanua Levu. The Lightning activity is more prominent in the western parts of both the islands.

The rainfall over Suva is found to correlate well with lightning occurrence over Viti Levu for the summer months, but shows no correlation for the winter months.

The lightning recorded by LIS showed no matches with WWLLN detections made over Viti Levu and Vanua Levu during 2003 and 2004. This discrepancy is because the LIS mainly records IC flashes which are clearly seen above the cloud heights whereas WWLLN records CG flashes.

Scintillation activity at 3.925 GHz at Suva is very low. The patches observed were weak with \(S_4\) index \(\leq 0.25\). The scintillations observed were generally of short duration (< 10 min) patches.

Scintillations were more pronounced in the daytime as compared to those in the nighttime.

The weak, short duration and pronounced daytime occurrence are mainly due to a) the high frequency (3.925 GHz) used, b) the observing station situated off the EIA belt and c) the period of observation falling in low solar activity.
7.3 Suggestions for Future Work

- Much of the work in VLF studies involved analyzing tweek sferics from the spectrograms received. Tweeks have been used to estimate height of the ionospheric reflecting layer. This could be extended to study the electron density at reflection heights and electron density profiles of the E-region. Radio atmospherics could be used to detect thunderstorms in the South Pacific region and predict severe weather hazards.

- The combination of WWLLN lightning data and tweek observations can be used to test the quality of the tweek range estimation technique, by contrasting the location of the lightning most likely to have produced the tweek.

- Whistlers are one of the most cost effective tools to study upper atmosphere. The long duration of data recording may result in many traces of whistler which will significantly contribute to understanding magnetospheric dynamics and the propagation mechanism of low-latitude whistlers.

- The VLF transmitter signals were manually recorded using a spectrum analyzer. With a continuous recording system, VLF phase and amplitude measurements can be used to model the D-region ionosphere for a variety of sub-ionospheric paths.

- Scintillation studies in the C-band are very limited especially in the South Pacific region. Longer-term scintillation measurements at this frequency at our station will be useful in the planning for future satellite links. Daytime scintillation reports in the GHz range are very limited. The occurrence of daytime scintillation in this work forms a good basis for further investigations on irregularity formations in the daytime ionosphere.
References


