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On The Use of VLF Narrowband Measurements to Study the Lower Ionosphere and the Mesosphere-Lower-Thermosphere

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Abstract

The ionospheric D-region (~*60* km up to ~*95* km) and the corresponding neutral atmosphere, often refered to as the mesosphere-lower-thermosphere (MLT) are challenging and costly to probe in-situ. Therefore, remote sensing techniques have been developed over the years. One of these is based on VLF (Very low frequency, 3-30 kHz) electromagnetic waves generated by various natural and man-made sources. VLF waves propagate within the Earth-ionosphere waveguide, and are extremely sensitive to perturbations occurring in the D-region along their propagation path. Therefore, measurements of these signals serve as an inexpensive remote sensing technique for probing the lower ionosphere and the MLT region.

This paper reviews the use of VLF narrowband (NB) signals (generated by man-made transmitters) in the study of the D-region and the MLT for over 90 years. The fields of research span time scales from microseconds to decadal variability, and incorporate lightning-induced short term perturbations; extraterrestrial radiation bursts; energetic particle precipitation events; solar eclipses; lower atmospheric waves penetrating into the D-region; sudden stratospheric warming events; the annual oscillation; the solar cycle; and finally, the potential use of VLF NB measurements as an anthropogenic climate change monitoring technique.

1. Introduction

The atmosphere-ionosphere system has drawn plenty of attention in recent decades, due to the growing understanding of the important coupling between the neutral and charged parts of the atmosphere. In addition, the sensitivity and tight connection of this system to various atmospheric and space weather phenomena result in a growing interest in this large field of study. More and more evidence regarding the mutual influence of the atmosphere and the ionosphere are being published, thanks to an increasing number of satellites and other in-situ and remote sensing techniques. However, the spatial and temporal coverage of some of these techniques are still limited, thus influencing the reliability of the rapidly improving numerical models.

Measurements of the ionospheric D-region (*~60* km up to *~95* km) and the corresponding neutral atmosphere, often refered to as the mesosphere-lower-thermosphere (MLT), are particularly limited, as currently this altitude range cannot be measured in-situ, apart from scarce rocket experiments, which obtain very narrow spatial and temporal coverage. Weather balloons do not reach this altitude range and satellites fly above it. Therefore, remote sensing techniques are mostly used to study the region. The relative complexity of the D-region physics [*e.g., Hargreaves*, 1992; *Kelley*, 2009; *Pavlov*, 2014] demands a vast number of measurements, in order to properly model its chemistry and dynamics. Very low frequencies (VLF – 3-30 kHz) radio signal measurements are one of the techniques used to gather these required observations, performed either from the ground, from the air, or from space [*e.g., Kelly et al.*, 1981; *Barr et al.*, 2000; *Parrot et al.*, 2008].

VLF electromagnetic (EM) signals are generated in Earth's atmosphere both by natural and anthropogenic sources, dominated by lightning discharges and man-made VLF transmitters [*Barr et al.*, 2000]. The signals propagate within the Earth-ionosphere cavity, since the long wavelength of these waves causes them to be reflected off the base of the ionosphere, within the D-region, and because of Earth's high ground conductivity [*Budden*, 1988]. The D-region reflection height is related to the electron number density, and therefore depends on ionization sources, as well as dynamical and chemical forcing [*e.g., Thomson*, 1993; *Thomson and Clilverd*, 2000; *Thomson et al.*, 2007]. In addition, the strength of the VLF signal attenuation depends mainly on ionospheric and ground properties [*e.g., Wait*, 1957a, 1957b; *Hargreaves*, 1992; *Silber et al.*, 2015]. Nevertheless, the attenuation within this frequency range is considerably low, reaching $\sim 2\ dB \cdot Mm^{-1}$ [*e.g., Wait*, 1957a, 1958; *Croom*, 1964; *Barr*, 1971], thus allowing the generated signals to propagate thousands of kilometers within the cavity**.**

The low attenuation in conjunction with the relatively large skin depth in sea water [*e.g., Wheeler*, 1958; *Benhabiles et al.*, 1996; *Barr et al.*, 2000] enable the communication with submarines submerged in water. Therefore, several man-made powerful VLF transmitters, operated by a few military bodies are distributed around the globe at fixed positions. Each transmitter broadcasts at a constant power (for very long periods) and at an unchanging frequency band, as seen in the spectrogram presented in Figure 1. Consequently, measurements of the transmitted signal's amplitude and phase are widely performed around the world. This type of measurements is commonly referred to as narrowband (NB) measurements. Since the ground (mainly sea surface) conductivity is relatively constant in time and its variation effects on VLF signals are small [*e.g., Wait*, 1957b; *Watt*, 1967; *Hauser et al.*, 1969], changes in the recorded amplitude and phase of the received VLF NB signals reveal information on the D-region's properties and variations along the transmitter-receiver great circle propagation path (TRGCP), thus allowing the monitoring of the D-region. This monitoring can be performed at different time scales, and everywhere on Earth's surface, thus allowing these measurements to achieve high spatial resolution. These advantages together with the inexpensive costs of VLF receivers make VLF measurements in general and NB measurements in particular, a powerful tool to probe the D-region, in comparison with many other (expensive) spatially and temporally limited remote-sensing instruments. Therefore, VLF NB signals are widely used to study the D-region and its response to both dynamical and chemical forcing, originated both from inside and outside Earth's atmosphere. The coupling of the D-region with the neutral atmosphere grants an additional method to study this fascinating part of the atmosphere.

The purpose of this paper is to review the main fields of research, which exploit VLF NB measurements, in order to gather a better understanding of Earth's atmosphere and ionosphere. The sections of this paper progress from short to long time scales, with a separation into general topics. Section 2 deals with perturbations induced by lightning discharges; Section 3 focuses on the connection of VLF NB measurements to extraterrestrial forcing; Section 4 examines the use of VLF NB measurements to detect oscillations within the D-region, which are originated in the lower atmosphere; finally, Section 5 deals with the utilization of VLF NB measurements to study annual and long-term oscillations and changes in the MLT and the D-region.

2. Lightning-induced perturbations

VLF NB measurements can be used in order to detect and investigate lightning-induced perturbations. Lightning is known to influence the temperature and conductivity of the atmosphere above thunderstorms. These perturbations include lightning-induced electron precipitation effects (ref), and short-term VLF perturbations known as 'early' events (ref), which are possibly generated by the lightning quasi-electrostatic (QE) field (ref), a lightning discharge EM pulse (EMP) (ref), or by red sprites and other transient luminous events (TLEs) [*e.g., Pasko et al.*, 2012; *Siingh et al.*, 2015].

Figure 2 depicts lightning-induced perturbations in the NSY (Sicily, 45.9 kHz, 37.12º N, 14.43º E) transmitter signal amplitude on November 11, 2013. These perturbations were recorded by the TAU (Tel-Aviv University, Israel, 32.11º N, 34.80º E) VLF UltraMSK receiver [*see Clilverd et al.*, 2009] at 1 Hz integration frequency. As can be seen, these NB perturbations in the ~2 Mm TRGCP have recovery times of up to ~2 min. The bottom panel shows in red circles the associated lightning discharges' location, as detected by the World Wide Lightning Location Network (WWLLN)

[*Dowden et al.*, 2002; *Rodger et al.*, 2004], together with the TRGCP (yellow curve). These associated discharges were located up to ~185 km away from the TRGCP. The presented perturbations are hard to classify, due to the relatively low sampling frequency, the lack of phase data, and the absence of associated WWLLN detections in two of the presented events, which are possibly due to the limited WWLLN detection efficiency [*e.g., Rudlosky and Shea*, 2013]. The first two arguments will be discussed below. Nevertheless, for a more comprehensive discussion on this section's topics, see *Inan et al.* [2010], *Rodger* [2003], *Barr et al.* [2000] and references therein.

It should be mentioned that in addition to the phenomena described above, lightning discharges can also produce terrestrial gamma ray flashes (TGFs) [*e.g., Dwyer et al.*, 2012]. However, TGFs are studied using VLF broadband measurements (that record the entire VLF band) and not NB measurements, and therefore they are beyond the scope of this paper.

2.1. Lightning-induced electron precipitation

Lightning discharges occur in almost every location on Earth. Most of their generated EM energy is radiated within the VLF and ELF (extremely low frequencies) bands (peaking between 5-10 kHz) [*Cummer*, 1997; *Rakov and Uman*, 2003]. As mentioned earlier, these generated EM waves propagate within the Earth-ionosphere cavity. However, a small portion of the EM energy can penetrate through the ionosphere into the magnetosphere, and propagate as a whistler mode wave [*Helliwell*, 1965], where it can interact with radiation belt electrons, causing them to scatter and precipitate into altitude ranges of *60-120 km*. Subsequently, the incoming electron surge produces electron density enhancements in this altitude range [*Helliwell et al.*, 1973; *Inan et al.*, 2010].

Consequently, VLF NB measurements can monitor these temporary D-region electron density enhancements. The VLF signature of these events (often referred to as VLF 'Trimpi' events) commonly show an amplitude decrease together with a phase increase [*e.g., Inan and Carpenter*, 1987; *Clilverd et al.*, 1999b]. However, opposite behavior (which is probably initiated by modal interference of the VLF waves in the Earth-ionosphere cavity) was observed as well [*e.g., Dowden and Adams*, 1989]. Trimpi events can be powerful enough to cause up to ~6 dB amplitude difference and ~20º of phase shift [*e.g., Helliwell et al.*, 1973; *Dowden et al.*, 2001]. They have a typical VLF rise times of up to 1 s, which occur ~1-2 s after the whistler's parent lightning (that are usually observed as sharp pulses in VLF broadband measurements, known as 'sferics'). These perturbations may appear in regions distant from the parent lightning (due to the whistler wave's generation and propagation nature) , and have decay periods smaller than 100 s [*Rodger*, 2003; *Inan et al.*, 2010].

Since the first report that connected between abrupt perturbations in VLF NB amplitude and phase measurements and lightning-induced whistler waves [*Helliwell et al.*, 1973], many VLF studies have tried to answer diverse questions regarding lightning-induced electron precipitation (LEP) events. Among different topics regarding the LEP, these questions were focused on the electron precipitation flux intensity based on the LEP driven VLF perturbation amplitudes [*e.g., Inan et al.*, 1985b; *Tolstoy et al.*, 1986; *Inan and Carpenter*, 1987; *Clilverd et al.*, 2004], the LEP event sources and mechanisms [*e.g., Inan et al.*, 1988b; *Burgess and Inan*, 1990; *Dowden and Adams*, 1990; *Poulsen et al.*, 1993; *Clilverd et al.*, 2004; *Peter and Inan*, 2004; *Gołkowski et al.*, 2014], and the precipitating electron energies [*e.g., Helliwell et al.*, 1973; *Lohrey and Kaiser*, 1979; *Inan et al.*, 1988a].

2.2. Early VLF events

With the improvement of VLF receivers over the years, higher temporal-resolution measurements were possible. This allowed researchers to distinguish between the relatively long delay of LEP events from the parent lightning and other VLF NB events with a very short delay (typically <50 ms) from the parent lightning [*Rodger*, 1999]. The first description of this type of event was given by *Armstrong* [1983]. They were later termed 'early' events by *Inan et al.* [1988b], due to their short time delay from the parent lightning. Moreover, these events are commonly referred to as 'early/fast' events due to their fast VLF NB amplitude rise times [*e.g., Inan et al.*, 1988b; *Corcuff*, 1998; *Moore* *et al.*, 2003]. Nevertheless, *Haldoupis et al.* [2004] observed 'early' events with long onset of up to ~2.5 s, granting them the term 'early/slow' events.

It is known that 'early' events are usually produced up to 50 km away from their associated cloud-ground (CG) lightning [*e.g., Inan et al.*, 1993; *Johnson et al.*, 1999; *Rodger*, 2003], although farther lightning discharges (up to a few hundred kilometers) can produce 'early' events as well [*e.g., Salut et al.*, 2013]. A lot of effort is still invested in explaining the origin of these disturbances. The main two mechanisms for the 'early/fast' events are the scattering of VLF signals from red sprite conducting columns [*e.g., Dowden et al.*, 1994, 1996; *Dowden and Rodger*, 1997; *Rodger*, 1999]; and D-region conductivity perturbations caused by QE field heating (that also initiates red sprites), which is generated by the associated lightning discharges [*e.g., Inan et al.*, 1995, 1996; *Barrington-Leigh et al.*, 2001]. The current explanation for 'early/slow' events is that the lightning EMP that also create photon emission patterns known as 'elves', and is generated by CG and intra-cloud (IC) lightning, produce ionization powerful enough to be detectible with VLF receivers [*e.g., Marshall et al.*, 2008, 2010; *Marshall and Inan*, 2010].

While several papers have shown one-to-one correlation between TLEs and VLF 'early' events [*e.g., Haldoupis et al.*, 2004, 2010], others did not observe these one-to-one correlations [*e.g., Marshall et al.*, 2006]. This might be due to the importance of simultaneous observations of VLF NB amplitude and phase. *Kotovsky and Moore* [2015] analyzed a few tens of 'early' events using both parameters. They concluded that some observed 'early' events can be deceptive when examining only one of the parameters (amplitude or phase), as was performed in many previous works. They proclaimed that the full analysis required for a correct characterization of 'early' events, i.e., whether they are 'fast' or 'slow', is essential for the understanding of the fundamental physical mechanisms behind 'early' events. Undoubtedly, 'early' events are still a topic of active research, and many aspects of these VLF NB perturbations are yet to be discovered.

2.3. Long recovery events

One of the new fields of 'early' event research is the study of 'long recovery early events' (LOREs). LOREs are 'early' events where the VLF NB amplitudes have a relatively long recovery period, which can exceed 30 min [*e.g., Haldoupis et al.*, 2012; *NaitAmor et al.*, 2013; *Gordillo-Vázquez et al.*, 2016]. Figure 3 illustrates an example of two consecutive LORE events in the NSY transmitter signal amplitude on November 25, 2013, recorded by the TAU VLF receiver. The total recovery time spans a staggering 31 min. The bottom panel shows in red circles the two associated lightning discharge locations, as detected by WWLLN. These associated discharges were very intense (based on the dozen WWLLN receivers that detected them), and were located ~150 and ~160 km away from the TRGCP,

LOREs were first classified by *Cotts and Inan* [2007], although observed earlier [*e.g., Inan et al.*, 1988b; *Dowden et al.*, 1997]. *Lehtinen and Inan* [2007] used a simplified D-region chemistry model [*Glukhov et al.*, 1992] and suggested that gigantic jets could be associated with LOREs, as they can generate enduring ionization of heavy negative ions below ~70 km. However, *Marshall et al.* [2014] used gigantic jets data recorded by the Imager of Sprites and Upper Atmospheric Lightnings (ISUAL) space-based instrument, and found that only 4 out of the 9 detected gigantic jets (which occurred within 100 km of the TRGCPs) were accompanied with VLF early events, none of which were LOREs, thus putting the gigantic jet mechanism in question.

Haldoupis et al. [2009] used the same model as *Lehtinen and Inan* [2007] and concluded that LOREs in VLF measurements can be initiated when the electron density enhancements at the higher altitudes of the D-region, i.e., near 85 km where VLF signals are reflected, are weak in comparison with the background electron density. *Haldoupis et al.* [2012] identified 10 LOREs in their data, from which 8 were associated with sprite-elve pairs, rather than with sprite or elve alone. They postulated that some sort of a coupling mechanism between the two TLEs (which are driven by the EMP and QE fields) is responsible for the LORE signature in VLF measurements. Analysis of nearly 50 detected LOREs made by *Salut et al.* [2012] has shown that LOREs can be observed when the associated lightning discharges are farther from the TRGCP than in other types of 'early' events. In addition, they noticed that ~87% of the events were triggered from lightning discharges above sea, as was also noted in other papers [*e.g., Haldoupis et al.*, 2012; *Kumar and Kumar*, 2013; *Schmitter*, 2014; *Kotovsky et al.*, 2016].

Similar to the observations made by *Salut et al.* [2012], *Haldoupis et al.* [2013] examined large VLF NB datasets and found that LOREs were generated when the associated lightning discharges were located up to 300 km from the TRGCP. In addition, they concluded that LORE occurrence strongly favors powerful CG discharges with peak currents above ±250 kA (which accounts for less than 0.5% of the total number of lightning discharges), thereby explaining some of the observations made by *Haldoupis et al.* [2012]. Nevertheless *Salut et al.* [2013] and *Gołkowski et al.* [2014] have shown that the induced LORE amplitudes were not directly related to the peak current strength and proximity to the TRGCP, and that in more than 75% of their occurrence, these VLF disturbances are produced by positive lightning discharges. Moreover, *NaitAmor et al.* [2013] have concluded that the occurrence of LOREs mainly depends on the modal structure of the VLF signal, the VLF NB scattering process, and the distance of the transmitter and receiver from the disturbed region, while the occurrence of TLEs and the lightning peak current only play a secondary role.

Recently, *Kotovsky and Moore* [2016] used a photochemical model in order to understand the mechanism behind LOREs. They concluded that LOREs occur when strong electron density enhancements (at altitude levels of 76 km or above) are regulated by slow recombination processes. These conclusions were supported by the kinetic modeling of *Gordillo-Vázquez et al.* [2016], who emphasized that the strong electron density enhancements needed for the LOREs should occur above ~79 km, where the electron lifetime is determined by recombination with $H^+(H_2O)_n$ and NO^+ ions. These two papers yet again reinforce the association of LOREs with high peak current lightning flashes. Nevertheless, *Kotovsky et al.*, [2016] argued that intense initial breakdown and fast first leaders are important properties of lightning discharges that produce LOREs, thus complicating the essence of the different LORE-producing mechanisms, and the relationship between them. While plenty of progress has been achieved during the last decade, the VLF NB LORE signature demands further investigation, together with other 'early' event aspects.

3. Space weather phenomena and eclipses

3.1. Extraterrestrial X-ray and gamma ray bursts

X-ray and gamma (γ) ray bursts from extraterrestrial sources are known to affect the ionosphere on time scales of seconds up to several hours. The effect of gamma ray bursts from cosmic origins, i.e., magnetars or neutron stars (also known as soft gamma repeaters - SGRs) on the ionosphere was first observed by *Fishman and Inan* [1988], and was based on VLF trans-equatorial NB measurements. They reported a disturbance in VLF amplitudes that occurred simultaneously with the gamma ray burst detection by the Prognoz-9 satellite. *Inan et al.* [1999] attributed huge periodic amplitude and phase perturbations (that reached 24 dB and 65º, respectively) to an SGR's massive periodic gamma ray burst. By using quantitative analysis, the authors concluded that these VLF periodic perturbations, which lasted for ~5 min, were mostly origninated in 3-10 keV low energy photons. *Inan et al.* [2007] observed even stronger perturbations of 26.5 dB of 328º, caused by a giant gamma-ray flare from an SGR, which lasted for an hour. These enormous perturbations corresponded to 3 orders of magnitude electron density enhancement at ~60 km, and even larger relative enhancements at lower stratospheric altitudes (down to 20 km). Other observations of (weaker) VLF NB amplitude and phase perturbations from around the world, triggered by gamma ray bursts from different SGRs have been reported as well [*e.g., Tanaka et al.*, 2008, 2010; *Chakrabarti et al.*, 2010a; *Mondal et al.*, 2012; *Raulin et al.*, 2014; *Nina et al.*, 2015; *Solovieva and Rozhnoi*, 2015].

However, most of the reports on extraterrestrial radiation effect on the ionospheric D-region are originated in solar flares. During these flares, X-ray flux received at the Earth strongly intensifies within a few minutes and decays in up to several hours, while strongly affecting ionization processes in the ionosphere [*e.g., Mitra and Rowe*, 1972; *Hunsucker and Hargreaves*, 2002; *Khan et al.*, 2005; *Clilverd et al.*, 2009].

The effects of solar storms on VLF measurements and the ionosphere were discussed almost a century ago [*e.g., Pickard*, 1927; *Austin*, 1932; *Bailey and Thomson*, 1935], but the connection between VLF NB disturbances and X-ray bursts were observed only after the emergence of the satellite era and space-borne X-ray detectors [*e.g., Kreplin et al.*, 1962]. *Chilton et al.* [1965] were the first to directly correlate VLF NB sudden phase anomalies (SPA) and X-ray flux measured by satellites, although mutual occurrences were reported earlier [*e.g., Kreplin et al.*, 1962]. They concluded that VLF SPAs accompany (during small zenith angle conditions) almost all X-ray bursts in the 0.5-4 Å band. In addition, they used their observations in order to derive the effective recombination coefficient. These conclusions and applications were strengthened and supported in subsequent papers [*e.g., Jones*, 1971; *Deshpande et al.*, 1972; *Ananthakrishnan et al.*, 1973; *Bain and Hammond*, 1975; *Basak and Chakrabarti*, 2013].

Nevertheless, several studies were made in order to find X-ray flux thresholds needed for the generation of a VLF NB SPA [*e.g., Kaufmann and de Barros*, 1969; *Muraoka et al.*, 1977; *Pant*, 1993; *Khan et al.*, 2005]. Similarly, *Kaufmann et al.* [2002] studied numerous solar flare events that did not produce SPAs. *Raulin et al.* [2006] examined several hundred solar flare events and concluded that the probability of observing a SPA during weak solar flares (class C2 or lower) is higher during solar minimum, while stronger flares (higher X-ray flux) are independent of the solar activity conditions, implying that the D-region has different sensitivities to X-ray flux intensity, being more sensitive during solar minimum. *Pacini and Raulin* [2006] retrieved a quantitative minimum X-ray flux threshold (in the 0.5-2 Å range) needed in order to produce SPAs as a function of solar activity. In addition, *Raulin et al.* [2010] retrieved these thresholds for the 0.1-0.8 Å X-ray flux range, based on several hundred VLF NB SPA events.

The effect of a flare's X-ray burst on ionospheric parameters can also be inferred from VLF NB measurements. *Thomson and Clilverd* [2001] used VLF NB amplitudes together with the Naval Ocean Systems Center (NOSC) Long Wave Propagation Capability (LWPC) model [*Ferguson and Snyder*, 1987; *Ferguson*, 1989, 1998] and deduced 8 km and 0.06 km^{-1} reflection height (h') and electron density slope (β) changes, respectively, due to M5 class solar flares (which enhanced the amplitudes by ~8 dB). Using these two parameters (also known as the 'Wait' parameters), the vertical electron density profile can be evaluated [*e.g., Thomson and Clilverd*, 2001; *McRae and Thomson*, 2004; *Grubor et al.*, 2008] based on the work of *Wait and Spies* [1964]. *McRae and Thomson* [2004] studied an X5 solar flare using phase and amplitude measurements of long trans-equatorial TRGCPs. They found that VLF phase and reflection height changes do not have a linear proportion to the logarithm of the solar X-ray flux, in agreement with some studies [*e.g., Selvakumaran et al.*, 2015], in contradiction to others [*e.g., Pacini and Raulin*, 2006; *Singh et al.*, 2014; *Pandey et al.*, 2015]. By using the LWPC model they determined that such a strong flare can be explained by 13 km and 0.13 km^{-1} reflection height and electron density slope changes, respectively. Moreover, *Thomson et al.* [2005] used a similar methodology and inferred a staggering 17 km and 0.18 km^{-1} reflection height and electron density slope change, respectively, as a result of the great solar flare of November 4, 2003, which saturated the GOES satellite's X-ray detectors, and was strong enough to be detected on dawn (half lit) long TRGCPs. That flare's magnitude was approximated to be X45 [*Thomson et al.*, 2004], based on the conclusions of *McRae and Thomson* [2004]. Other studies have also examined the reflection height, electron density slope, and vertical profile modifications, due to various classes of solar flares, and obtained different parameter values, possibly due to different latitude range and lengths of the TRGCPs [*e.g., Kaufmann and Mendes*, 1968; *Kamada*, 1985; *Žigman et al.*, 2007; *Grubor et al.*, 2008; *Schmitter*, 2013; *Kolarski and Grubor*, 2014, 2015; *Singh et al.*, 2014; *Pandey* *et al.*, 2015; *Šulić et al.*, 2016]. These calculated values had a large variance, and a few of them [*e.g., Kamada*, 1985; *Grubor et al.*, 2008] reached magnitudes which were comparable to those deduced by *Thomson et al.* [2005].

3.2. Solar proton events

In addition to the solar flare X-ray radiation impacts on the D-region, solar proton events (SPEs), also known as polar cap absorption events and solar energetic particles (SEP) events, are also an active topic in VLF research. SPEs are generated when protons originating from the Sun are accelerated to relativistic energies by X-ray radiation produced during a solar flare or by the coronal mass ejection (CME) shock [*e.g., Krucker and Lin*, 2000]. These energetic particles can reach the Earth from a few minutes (relativistic particles) up to hours (lower energy particles) from the time of acceleration [*Shea and Smart*, 1990]. SPEs can persist up to several days, while affecting mainly the polar atmosphere (at magnetic latitudes higher than 60º), due to the partial guidance of these charged particles by the geomagnetic field [*Rodger et al.*, 2006].

Similar to X-ray bursts, VLF NB perturbations could be attributed to SPE intensity only after the deployment of proton detectors on-board satellites, although the connection between VLF perturbations and SPEs was speculated before [*e.g., Pierce*, 1956; *Bates*, 1962; *Bates and Albee*, 1965]. *Potemra et al.* [1967] were the first to combine VLF and satellite detector data. They found that the VLF NB phase deviation due to a SPE was comparable to 67% of the normal diurnal variation, and (by using a simplified exponential ionospheric model) that a flux of 150 relativistic protons per cm^2 can result in a 10 km change in the daytime reflection height. The consistency of their simplified model for SPE studies was further investigated by *Potemra et al.* [1970].

Potemra et al. [1969] used a similar methodology and derived the electron density profile during disturbed conditions. They also concluded that VLF NB measurements are more sensitive to ionization processes produced by SPEs than riometers (that measure the ionospheric high frequency absorption), as was also comprehended by other studies [*e.g., Crary and Diede*, 1969].

Westerlund et al. [1969] studied several SPEs using multiple TRGCPs, and concluded that a linear relationship exists between the VLF phase anomaly and the double logarithm of the proton flux. *Kossey et al.* [1983] found a linear correlation between the VLF reflection height and energetic proton flux with energies <20 MeV.

In addition to the linear relation found in their study, *Westerlund et al.* [1969] also found that the phase anomalies always advance during a SPE, with larger anomalies occurring with increasing TRGCP geomagnetic latitude, decreasing signal frequency, and decreasing ground conductivity. Furthermore, they have speculated that the proton precipitation is bounded to the geomagnetic latitude of 62.5º. Subsequently, *Mendes and Ananthakrishnan* [1972] calculated the relative parts of the examined TRGCPs affected by the SPEs (based on two events), as well as the VLF reflection height change (that reached an enormous 23 km in one of the TRGCPs), due to these events.

Beloglazov et al. [1990] examined a single SPE that occurred during 1984 using VLF NB, riometer, and satellite data, and concluded that the electron density at an altitude of 45 km steeply rose to 1000 km^{-1} during the event. In addition, they demonstrated the existence of two regions and mechanisms responsible for the ionization produced by the protons; the direct impact zone in the polar cap's center, and the day-evening auroral sector affected by SPE-driven precipitation of quasi-trapped magnetospheric electrons. However, other studies have indicated SPE events where the energetic electron precipitation (EEP) influence on ionization was insignificant in comparison with the energetic proton effect [*e.g., Clilverd et al.*, 2005]. *Mendes da Costa and Rizzo Piazza* [1995] also studied several SPEs and their subsequent energetic electron precipitation (EEP). They concluded that the associated nighttime VLF reflection height lowering and electron density slope decrease reached ~9 km and 0.24 km^{-1}, respectively.

Clilverd et al. [2005] examined VLF NB amplitude measurements during a SPE in November, 2001, and deduced that the observed disturbances in the VLF data are mainly produced by fluxes of protons with energies higher than 50 MeV. In addition, using the VLF data they showed that the utilization of the Sodankylä Ion Chemistry (SIC) model [*Verronen et al.*, 2002] in conjunction with the LWPC model can successfully reproduce the ionization variations (with the highest sensitivity at an altitude range of 50-60 km during a SPE). *Clilverd et al.* [2006] used the SIC and LWPC models as well to study several other SPEs and showed that different TRGCPs show different amplitude behavior (based on the TRGCP's length) during the same SPE. They advocated the methodology's capability to solve the inverse problem, i.e., the D-region behavior in response to a SPE, based on VLF NB measurements. Thus, this methodology also enabled the study of SPEs effect over the neutral atmosphere as well.

Consequently, *Verronen et al.* [2005] used this methodology together with satellite data in order to investigate mesospheric ozone depletion variations during SPEs. *Seppälä et al.* [2008] examined a series of strong SPEs which produced a 10 dB VLF amplitude decrease. They concluded that the energetic protons generated an enormous (more than 400%) increase in mesospheric odd nitrogen (NOx) and a significant decrease (at least 30%) in mesospheric ozone concentrations at 70º in both hemispheres, while much more moderate changes were produced in the stratosphere.

3.3. Energetic electron precipitation

As mentioned earlier, transient events of magnetospheric EEP can be triggered by whistler mode waves. Man-made EM waves can induce electron precipitation as well [*e.g., Inan et al.*, 1985a, 2007b; *Tolstoy et al.*, 1986]. However, large and persistent EEP events at high geomagnetic latitudes of 55º-72º are generated as one of the main outcomes of different processes during geomagnetic storms, which are generated by perturbations or intensifications of solar activity (through solar flare radiation, SPEs, etc.) [*Thorne and Kennel*, 1971; *Rodger et al.*, 2007; *Andersson et al.*, 2012].

Potemra and Rosenberg [1973] presented the first direct correlation between VLF NB (nighttime phase) measurements and EEP, which was suggested by earlier studies as a source for observed VLF perturbations [*e.g., Belrose and Thomas*, 1968; *Doherty*, 1971], and concluded that VLF NB measurements are a useful tool, sensitive enough to detect low-energy electron precipitation. They used Bremsstrahlung X-ray riometer at L-shell value of L ≈ 4 as a proxy for the EEP and VLF NB TRGCPs located mainly in the region between L-shell values of L ≈ 2 and L ≈ 4. *Larsen et al.* [1977] studied VLF phase measurements from similar TRGCPs during an EEP on December, 1971. They inferred VLF reflection height changes of up to 10 km, and deduced the ion production rate profile.

Abdu et al. [1981] studied VLF NB phase perturbations from a low L-value TRGCP located in the South Atlantic magnetic anomaly (SAMA) region, and concluded that the disturbances were originated from EEP that generated enhanced ionization at an altitude range of 70-110 km. *Pinto et al.* [1990] analyzed VLF phase data from the same TRGCP and found that the VLF phase advances in the SAMA region's TRGCP are proportional to a logarithm function of the EEP flux, thus allowing the estimation of the precipitation flux based on VLF NB measurements.

Similarly, *Kikuchi and Evans* [1983] found a linear correlation between VLF phase advances of TRGCPs located in the auroral (60º-70º) geomagnetic latitudes and logarithm functions of the EEP flux. They advocated that precipitating electrons with energies above 300 keV are the main ionization source in the D-region which influences the VLF signals. However, *Cummer et al.* [1997] studied several EEP events using VLF NB amplitude data and suggested that the VLF perturbations are produced by electrons with a lower energy threshold of 100 keV.

Clilverd et al. [2006b] showed using VLF NB amplitudes and satellite detector data that an observed 10 min sudden EEP flux in both hemispheres (at L ≈ 5) accounted for no more than 10% of the (>2 MeV) geosynchronous electron loss flux. Nevertheless, they also suggested that continuous precipitation between L ≈ 4 and L ≈ 6 which was measured for more than 2 hours accounted for ~50% of total geosynchronous electron loss flux. *Clilverd et al.* [2010] used ~4.5 years of VLF NB amplitudes (of a TRGCP spanning between L ~ 3-7) and utilized the LWPC model, in order to calculate the precipitating electron flux. In addition, they observed up to 3 orders of magnitude variations in the electron flux during a geomagnetic storm. *Neal et al.* [2015] elaborated that study based on more than 9 years of VLF NB data and advocated that their methodology is at least one order of magnitude more sensitive to low precipitating electron flux than satellite measurements. This conclusion about VLF NB measurements as a sensitive technique for monitoring EEP events was strengthened by *Rodger et al.* [2012] who examined the sensitivity of a riometer, GPS-based VTEC (vertical total electron content), and VLF NB measurements to EEP events. They concluded that VLF measurements are the most sensitive technique (by up to several orders of magnitude) for detecting radiation belt EEP, and EEP events with energies above 200 keV.

Similar to SPE research, VLF NB measurements enables the examination of EEP effects on the neutral atmosphere. *Rodger et al.* [2007] analyzed midday and midnight VLF NB amplitude data from three TRGCPs at L > 2, together with satellite data and the SIC and LWPC models, in order to study the ionization changed during and after the major geomagnetic storm in September, 2005. They observed 2.4 dB midday increase and 14 dB midnight decrease in VLF amplitudes during relativistic energies electron precipitation (REP) at L ≈ 3. By utilizing the models, they calculated the >150 keV electron flux, which was later corrected by *Rodger et al.* [2010]. In addition, they found that the midday precipitation flux was ~20 times larger than the midnight flux during the 6 days following the storm, and thereby concluded the plasmaspheric hiss [*e.g., Thorne et al.*, 1973; *Bortnik et al.*, 2008] role as the driver for the electron loss in the inner zone of the outer radiation belts. Subsequently, *Rodger et al.* [2010] studied the effects of the same event on the neutral atmosphere's ozone, odd hydrogen (HOx), and NOx concentrations, by utilizing VLF NB measurements, and the LWPC and SIC models. They showed that the large increases of the HOx and NOx (>300%)

generated by the EEP did not lead to significant ozone depletion on that event, largely due to the relatively low latitude where the electrons were deposited.

3.4. Solar eclipse

In addition to their sensitivity to EEP events, VLF NB signal measurements are sensitive to changes in the D-region, caused by solar eclipses. The first measurements of VLF NB perturbations produced as a result of a solar eclipse were presented by *Bracewell* [1952]. He showed that the partial (30% obscurance) solar eclipse (PSE) of April 28, 1949 induced 35º phase difference, which were equivalent to 1 km VLF reflection height change.

Subsequently, VLF NB studies continued to investigate the influence of total solar eclipse (TSE) on VLF phase anomalies [*e.g., Decaux and Gabry*, 1964; *Albee and Bates*, 1965; *Kaufmann and Schaal*, 1968; *Hoy*, 1969]. *Crary and Schneible* [1965] examined the phase anomaly during the passage of a TSE through the 400 km TRGCP, and found that the large 144º anomaly was equivalent to ~11 km change in reflection height. These observations also allowed them to calculate the D-region recombination coefficients during the event.

Noonkester and Sailors [1971] used a VLF propagation and D-region aeronomy model to predict VLF phase change due to solar eclipse. However, they managed to accurately predict only one out of two events. *Lynn* [1981] concluded that the VLF phase response to the solar obscuration during a solar eclipse is non-linear, and that a 4 min delay exists between solar eclipse maximum and VLF response. However, other studies have noted other response delay values [*e.g., Mendes Da Costa et al.*, 1995], that reached up to 8 min [*e.g., Cheng et al.*, 1992].

Sen Gupta et al. [1980] observed a 1.4 dB VLF NB amplitude change due to a PSE (65% obscurance) passing through a 6 Mm TRGCP, from which a 3 km reflection height change was deduced, a value which was also obtained by other studies (that investigated different TRGCPs and eclipse events) [*e.g., Cheng et al.*, 1992; *Guha et al.*, 2010; *Pal et al.*, 2012]. Nevertheless, the reflection height change due to solar eclipse events can be quite variable [*e.g., Mendes Da Costa et al.*, 1995; *Chakrabarti et al.*, 2012].

The amplitude deviation as a result of a solar eclipse event may be variable as well, depending onthe TRGCPs length, location, and the geometry of the obscured region with regards to the TRGCP. *Clilverd et al.* [2001] studied the TSE of August 11, 1999 influence on VLF NB measurements from 19 TRGCPs, spanning from 90 km up to 14 Mm. They observed negative phase perturbations in all TRGCPs, while the amplitude changes were positive (negative) on TRGCPs shorter (longer) than 2 (10) Mm, respectively. However, somewhat different behavior were observed in subsequent studies [*e.g., Pal et al.*, 2012]. The typical perturbations were ~3 dB (in amplitude) and ~50º (in phase), while the deduced effective reflection height (electron density profile slope) changes due to the TSE reached 8 km (0.07 km^{-1}), respectively. These modifications were smaller as the TRGCP's length increased. This conclusion was strengthened by other studies [*e.g., Crary and Schneible*, 1965; *Hoy*, 1969; *Phanikumar et al.*, 2014], although very large phase deviation (115º) due to a TSE in a 13.3 Mm TRGCP was reported by *Kaufmann and Schaal* [1968]. Similar to *Clilverd et al.* [2001], *Guha et al.* [2010] examined amplitude changes during a partial passage (90% obscurance) of a TSE, though they deduce smaller changes in the two 'Wait' parameters, from which an 80% reduction in electron density at 71 km was concluded, similar to some studies [*e.g., Singh et al.*, 2012], but different from others [*e.g., Phanikumar et al.*, 2014].

Pal et al. [2012] studied the TSE of July 22, 2009 which was seen above India, and showed using the LWPC and VLF NB data from several TRGCPs that the assumption of 4 km ionospheric base (i.e., effective reflection height) increase in regions where the eclipse's totality passes through the TRGCP matches well with the observations. Similar to *Lynn* [1981], they also endorsed the conclusion that the ionospheric parameter changes are not linearly dependent on the solar obscuration percentage, in contradiction to some linearity found in other studies [*e.g., Clilverd et al.*, 2001; *Singh et al.*, 2012]. Recently, *Chakraborty et al.* [2016] used a D-region ion-chemistry model alongside the LWPC model, in order to improve the inferred ionospheric profile (and its 'Wait' parameters) during the passage of a TSE, which they compared to the results of *Pal et al.* [2012]

In addition to the gradual ionospheric variations during a solar eclipse passage through a TRGCP, gravity wave signatures were also observed during this type of events using VLF NB measurements. Turbulence effects in the D-region and VLF NB measurements during a solar eclipse event were already speculated a few decades ago by *Meisel et al.* [1976]. More recently, *Chernogor* [2010] reported an intensification of oscillations with time periods of 10-15 and 18 min based on spectral analysis of VLF NB amplitude and phase data during a solar eclipse. *Maurya et al.* [2014] described wave like signatures with periods of 16-40 min during TSE passage through TRGCPs, and 30-80 min oscillations in TRGCPs under PSE influence. They advocated that these oscillations arise from the sharp electron density gradient created in the obscured region, and gravity waves induced by the solar eclipse process. These findings were also supported by *Pal et al.* [2015], who deduced similar wave patterns during TSE passage through a few TRGCP.

4. Waves originating in the lower atmospheric

The dynamics of the lower atmosphere cause it to function as a "factory" for pressure waves on various time scales, from a few tenths of a second (acoustic waves, also known as infrasound) up to several days (planetary waves). Each of these wave types has its own typical generating sources and mechanisms, wavelengths, time periods, and frequency range, as shown in Table 1. These waves can penetrate into the MLT, depending on their amplitude, as well as the other factors, e.g. the vertical wind profile (gravity and planetary waves), temperature profile (acoustic waves), etc. [*Fritts and Alexander*, 2003; *Blanc et al.*, 2010]. Under ideal conditions the amplitudes of the waves increase exponentially as the density drops [*e.g., Blanc et al.*, 2009], and hence waves that reach the MLT can trigger very large temperature fluctuations. In addition, these waves produce perturbations in the concentrations of atmospheric species [*Smith*, 2004]. These modifications of the MLT can induce changes in the electrical characteristics of the D-region, by affecting the ionization processes as well as the free electron collision frequency [*e.g., Forbes*, 1981]. Therefore, these variations in the neutral and charged atmosphere can be detected using VLF NB measurements.

Already several decades ago, lunar tides were detected with VLF NB measurements. *Brady and Crombie* [1963] analyzed one year of NB measurements, and deduced that the semi-diurnal lunar tide affects the D-region, producing 0.11 km difference in the VLF effective reflection height. However, *Bernhardt et al.* [1981] used a different methodology which involved the combination of phase and amplitude measurements and found 0.39 km reflection height change due to the semi-diurnal lunar tide. They claimed that the difference from *Brady and Crombie* [1963] emerged from the different TRGCP, as well as the longitudinal difference between the transmitter and the receiver, because the lunar tidal influence maximizes at the same local time. In addition, they observed a 0.26 dB amplitude perturbation caused as a result of the semi-diurnal lunar tide.

Similar to the atmospheric tide forcing on VLF measurements, planetary wave signatures were also observed in VLF analysis more than four decades ago. *Cavalieri et al.* [1974] and *Cavalier and Deland* [1975] found indications of travelling planetary waves in the D-region, by using VLF NB daytime phase measurements (ranging above 100º of longitude), and performing an auto-correlation analysis. Both of these studies concluded that VLF NB measurements are a useful tool to study the stratosphere-ionosphere coupling, and the upward propagation of waves into the ionosphere. Recently, *Schmitter* [2011] analyzed one and a half years of VLF NB amplitude data together with the LWPC model and found signatures of planetary wave activity as well, by calculating the difference in amplitude between midday and midnight, and comparing the results with satellite data. Spectral analysis of the data showed that a quasi 16-day oscillation was the most dominant oscillation in the data, especially during wintertime. This finding was supported by *Schmitter* [2012] and *Pal et al.* [2015].

The propagation of planetary waves into the mesosphere and their impact on VLF NB measurements has also been used in order to explain apparent precursors of earthquakes observed in VLF NB measurements, a growing field in VLF research [*e.g., Hayakawa*, 1996, 2011; *Hayakawa et al.*, 1996; *Molchanov and Hayakawa*, 1998; *Rozhnoi et al.*, 2004; *Pulinets and Boyarchuk*, 2005; *Sasmal and Chakrabarti*, 2009; *Chakrabarti et al.*, 2010; *Ray et al.*, 2011]. However, this field of study is still strongly debated [*e.g., Rodger et al.*, 1996; *Clilverd et al.*, 1999; *Cohen and Marshall*, 2012], and is outside the scope of this paper.

Unlike atmospheric tides and planetary waves, signatures of acoustic and gravity waves in VLF measurements were only reported in recent years, making this topic and methodology a new evolving field. *Nina and Čadež* [2013] presented a new methodology to detect acoustic and gravity wave signatures in the D-region via the analysis of VLF NB amplitudes at 90 min windows around sunrise and sunset, when the solar terminator travels along the TRGCP. Based on their analysis, they concluded that acoustic and gravity waves with periods shorter than 20 min are the main origin of the medium-scale traveling ionospheric disturbances in the D-region. *Rozhnoi et al.* [2014] examined VLF data (both phase and amplitude) during the passage of tropical cyclones under the TRGCP. They found negative anomalies in the VLF amplitude in 75% of their examined cases and a correlation with a few atmospheric parameters during some of these events. Similar to *Nina and Čadež* [2013], they spotted gravity wave signatures in the VLF data with apparent time periods of 7-16 min. However, signatures of 15-55 min oscillations were reported by them as well.

Marshall and Snively [2014] observed periodic fluctuations in VLF NB nighttime amplitude data (these fluctuations were not clearly detected in the phase data). Due to the short 1-4 min periodicity of the fluctuations, they attributed these oscillations to acoustic waves generated by convective and lightning activity in the region. They supported their conclusions by combining the output of a compressible fluid acoustic and gravity wave propagation model, together with an EM propagation model. The study also concluded that a large source (>100 km radially) is needed in order to produce perturbations of 0.5 dB, which were observed in the data.

Figure 4a portrays short period gravity wave signatures in the NSY transmitter signal amplitude, as measured by the TAU VLF receiver on October 22, 2013. These oscillations reached amplitude of ~0.25 dB and lasted for ~30 min. Figure 4b illustrates the Lomb-Scargle (LS) periodogram [*Lomb*, 1976; *Scargle*, 1982; *Press and Rybicki*, 1989] spectral analysis of the amplitude time series (black curve). The blue (red) dashed curves represent the 95% (99.9%) statistically significance thresholds, respectively. As seen, a 6 min oscillation dominates the time series. This periodicity matches high frequency gravity waves. Figure 4c shows lightning discharge location, as detected by the WWLLN receivers up to 3 hours prior to the identified gravity wave signatures (the discharge location colors are based on time of occurrence). A large thunderstorm occurred ~1750 km north of the NSY transmitter. Since intense thunderstorms may produce strong gravity wave signatures, and the occurrence time of the thunderstorm matches the time of the observed VLF signatures (based on gravity wave phase velocity), and because high frequency gravity waves can generally propagate large distances in a ducted mode [*Fritts and Alexander*, 2003], this thunderstorm in northern Europe might be the origin of the observed oscillations in the VLF NB data.

Similar to Figure 4, Figure 5 depicts acoustic wave signatures observed on September 9, 2013. Similar to the gravity wave example, the wave signatures had amplitudes of ~0.2 dB and continued for ~30 minutes. The LS periodogram clearly shows a dominating ~2.5 min oscillation. These low frequency acoustic waves can generally be produced by lightning as well as deep convection, and propagate over large distances, both horizontally and vertically with relatively low attenuation [*Blanc et al.*, 2010; *Evers and Haak*, 2010]. Examination of the WWLLN lightning detection map indicates, based on acoustic wave propagation velocity (i.e., the speed of sound), that the time of occurrence of the thunderstorm located ~700 km west of the NSY transmitter matches the time of the observed signatures, making it a plausible candidate for the source of the measured amplitude wave signatures.

5. Annual and long-term effects

5.1. The 'winter anomaly' and long-term meteorological effects

On time scales longer than several days, the D-region is disturbed, both on an instantaneous and oscillatory manners. One of the major phenomena that can affect the D-region on these time scales are sudden stratospheric warming (SSW) events. SSW results from the interaction of planetary waves with the zonal flow in the winter stratosphere. This interaction produces a breakdown of the winter polar vortex, and the reversal of the typical westerly stratospheric winds into easterly winds, accompanied by an increase in stratospheric temperatures [*Matsuno*, 1971; *Hsu*, 1980]. SSW events are able to enhance the D-region electron density by affecting the ionization rate through changing transport patterns and temperatures, thus forcing the D-region to behave more similar to its summer pattern [*e.g., Shapley and Beynon*, 1965; *Sechrist et al.*, 1969; *Offermann*, 1979; *Solomon et al.*, 1982; *Taubenheim*, 1983; *Garcia et al.*, 1987]. This disturbance of the D-region influences radio wave absorption measurements, where it is also called the 'winter anomaly'.

The winter anomaly has been extensively studied using VLF NB measurements. The connection between VLF measurements and SSW was first reported by *Belrose* [1967]. He described a change in VLF NB phase during February and March, 1952, and attributed it to the SSW that occurred in the course of that period. *Larsen* [1971] analyzed VLF NB daytime phase and amplitude changes during a SSW event, and found only a phase change (without an amplitude change). By using a full wave computer model he concluded that a 3 km increase in reflection height can explain the observations.

Cavalier and Deland [1975] have correlated VLF daytime phase measurements with SSW strength as well. Similarly, *Muraoka* [1979] analyzed VLF NB phase measurements for TRGCPs spanning ~100º of longitude, and found a change in phase during a winter anomaly event. He concluded that a planetary wave influence exists during both daytime and nighttime, at VLF nighttime reflection altitudes of *~80 km*. *Muraoka* [1983] studied the winter anomaly effect on VLF NB measurements and the D-region, by examining mode conversion effects, which occur at the day-to-night discontinuity of the D-region, when the terminator is migrating along the TRGCP. He concluded that the D-region electron density at 75-90 km is enhanced during SSW events, resulting in a notable lowering of the VLF nighttime reflection height. Subsequent studies [*e.g., Muraoka*, 1985; *Muraoka et al.*, 1986] strengthened these findings, while noting that the winter anomaly occurrence is associated with a mesospheric planetary wave with zonal wave number 1.

The connection between VLF NB amplitude variations and meteorological effects were also studied on extensive data sets. *Correia et al.* [2011] associated the dynamics (mainly the 16-day wave) and several stratospheric parameters to VLF NB midday amplitudes using a 5-year dataset. They concluded that this connection is clearest during wintertime (at the receiver), and low solar activity periods. *Silber et al.* [2013] found a link between mesopause temperatures (altitudes of ~80-90 km) and VLF NB midday amplitudes in several datasets. They explained this connection by the free-electron production dependence on the ambient temperatures, and by modal interference effects caused by thermal contraction of atmospheric layers. By using the VLF measurements and performing a principle component analysis (PCA), they concluded that the variations in the incoming total solar irradiance accounts for ~72% of the mesopause temperature variations, while 28% of these variations are of other origin, most likely due to waves propagating from below. Recently, *Pal and Hobara* [2016], have correlated two years of VLF NB midnight amplitudes with stratospheric total column ozone and temperatures (at 31 km altitude), thus emphasizing the connection between the ionospheric base and the middle atmosphere. They concluded that each of these parameters can describe more than 33% of the VLF midnight amplitude variability.

5.2. Oscillations of an annual scale

As noted above, long datasets of VLF NB measurements can be examined, thus making them suitable to monitor oscillations of an annual scale. The annual cycle in VLF measurements was already measured during the 1920's through short-length (<1000 km) and trans-oceanic NB transmissions [*e.g., Espenschied et al.*, 1926; *Hollingworth*, 1926; *Ishii and Sakurazawa*, 1964]. However, the season of maximum amplitude in these works was not constant (the annual cycle maxima occurred in some TRGCPs during summer, while taking place during winter in others). This 180º shift can be explained using the waveguide 'mode theory' [*Jackson*, 1962; *Budden*, 1988; *Inan and Inan*, 2000], given that the total measured amplitude is the sum of several waveguide modes [*e.g., Rodger and McCormick*, 2006]. Nevertheless, it was broadly concluded by several studies that the effective daytime VLF reflection height is higher (lower) during winter (summer), respectively [*e.g., Bracewell et al.*, 1951; *Hargreaves and Roberts*, 1962; *Ferguson*, 1980].

Some papers focused on VLF NB signals on an annual time scale, while utilizing short-length TRGCPs, and using the waveguide 'ray theory'. 'Ray theory' presumes that (generally) in short distance propagation only the ground wave and the first (and sometimes second) sky-wave affect the resultant wave amplitude and phase, due to the intense attenuation of waves that are reflected several times from the ionosphere at very low incident angles (i.e., strong accumulating power loss) [*e.g., Laby et al.*, 1940; *Schonland et al.*, 1940; *Inan and Inan*, 2000]. These studies found an annual as well as semi-annual variation in daytime VLF reflection height, which was not observed during nighttime measurements of the same type [*e.g., Bracewell et al.*, 1951, 1954; *Bain et al.*, 1952; *Straker*, 1955]. Nevertheless, *Pintado et al.* [1987] have deduced an annual and a semi-annual oscillation in the day-night difference in reflection height, which were extracted from VLF NB amplitude and phase measurements. They concluded that the origin for these observations is in changes in the nighttime electron density due to the effect of neutral minor constituents in the mesosphere, probably atomic oxygen. However, recently *Silber et al.* [2016] observed a dominant (>3.3 dB from peak-to-peak) semi-annual oscillation in VLF NB midnight amplitudes of two long (>1900 km) TRGCPs (located in both hemispheres). Each of these datasets spanned over more than 4 years. By utilizing the LWPC model, they concluded that the peak-to-peak amplitude of the semi-annual oscillation is equivalent to ~1.5 km change in the nighttime VLF reflection height, and more than doubling of the electron density at an altitude of ~85 km. Unlike *Pintado et al.* [1987], they suggested that NOx transport from the lower thermosphere is the main driver of this oscillation in the D-region. The disagreement between *Silber et al.* [2016] and the studies that utilized the waveguide ray theory for short TRGCPs [*e.g., Straker*, 1955] with regards to the semi-annual oscillation during nighttime can be explained by the different methodologies, and the discrepancy in the probed region's altitude, due to the different transmitter-receiver distance, and hence the wave's incident angle [*e.g., Hargreaves and Roberts*, 1962; *Inan and Inan*, 2000; *Hunsucker and Hargreaves*, 2002].

5.3. Long-term oscillations and trends

In addition to the annual oscillations, it is well known for several decades that VLF NB measurements are suitable for monitoring the 11-year solar cycle of the D-region electron density, mainly by correlating the VLF measurements with active sunspot number [*e.g., Austin and Wymore*, 1928; *Austin et al.*, 1930; *Austin*, 1932; *Ishii and Sakurazawa*, 1964]. Moreover, *Thomson and Clilverd* [2000] analyzed 11 years of VLF NB midday amplitudes from several TRGCPs and found that 0.3 dB amplitude difference exists between solar minimum and maximum. They assumed that the greater attenuation rate during solar minimum arises from the combination of smaller solar Lyman-α flux and higher cosmic ray intensity, resulting in smaller electron density profile slope at the bottom of the D-region, followed by a stronger attenuation of VLF signals. In addition to monitoring the solar cycle during midday, *Raulin et al.* [2011] have demonstrated that the solar cycle can also be examined in phase effect measurements of the C-region, a transient reflecting layer located at an altitude range of ~63-69 km [*e.g., Sechrist*, 1968; *Rasmussen et al.*, 1980], which develops during sunrise, and produces a phase delay in VLF NB data for ~1-2 hours [*Kuntz et al.*, 1991; *Bertoni et al.*, 2013].

Finally, on time scale of several decades, VLF NB measurements might be suitable for monitoring climate change effects in the mesosphere and the D-region, as the intensified anthropogenic greenhouse gas (GHG) emissions are expected to influence not only the troposphere, but also the middle and upper atmosphere. *Roble and Dickinson* [1989] predicted that these regions of the atmosphere will experience a strong cooling, due to the increased radiative emissions by rising GHG concentrations. As a result of this cooling, thermal contraction of atmospheric layers (like some ionospheric layers) is expected to occur [*Laštovička et al.*, 2006]. Such a long-term trend of downward displacement of the D-region has been previously reported, though its amplitude is currently considered rather moderate [*e.g., Taubenheim et al.*, 1997; *Peters and Entzian*, 2015]. It should be mentioned that these studies were not performed using VLF methods. However, it can be assumed that long-term VLF NB studies will be reported in the future, due to their high sensitivity to D-region trends and perturbations.

6. Summary

VLF NB signals were already used during the first quarter of the 20th century for ionospheric studies [*e.g., Austin and Wymore*, 1928; *Bailey and Thomson*, 1935]. Nevertheless, this measurement technique is still relevant nowadays, with an increasing number of pertinent applications. The great sensitivity of the measurements to small perturbations, the relative inexpensive costs of VLF receivers, the minimum receiver maintenance requirements, and the straightforward employment of the antennas allow broad coverage monitoring of the D-region and MLT parts of the atmosphere. VLF NB measurements help in expanding the current knowledge of the D-region dynamical and chemical processes, while also granting the possibility of studying the chemistry and dynamics of the neutral atmosphere (in many occasions with the support of wave propagation and MLT chemistry models).

The high temporal resolution and the robustness of measurements enable the examination of various phenomena on different time scales, from the instantaneous lightning-induced short term perturbations to the long-term effect of anthropogenic driven climate change. In this paper, a summary of the different VLF NB based research fields was given. The ongoing studies, the countless questions answered in VLF studies and the numerous unsolved questions in these fields, which can be answered using VLF measurements, show the relevance of VLF NB research now and almost certainly in future.

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7. References

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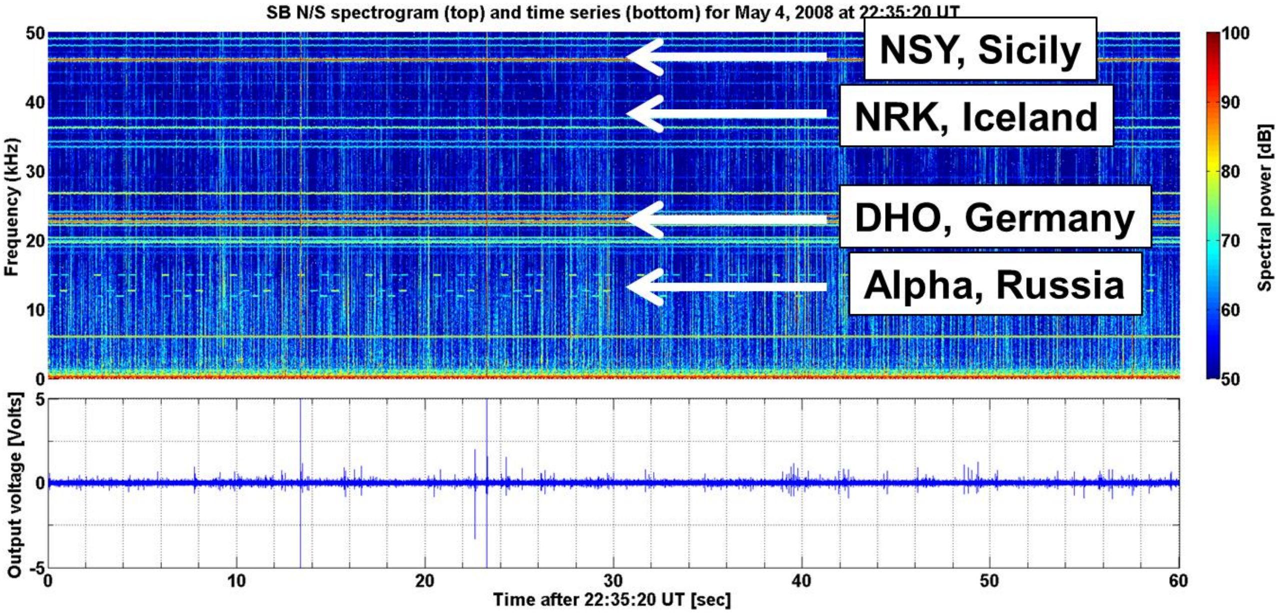

Figure 1: One minute measurement of the VLF band as received at the Sde-Boker receiver, Israel. (Top) spectrogram and (bottom) raw-waveform time series. Note the numerous vertical spikes (sferics) from lightning discharges, and the horizontal signals originating from man-made VLF transmitters (a few transmitter call-signs are marked in the spectrogram).

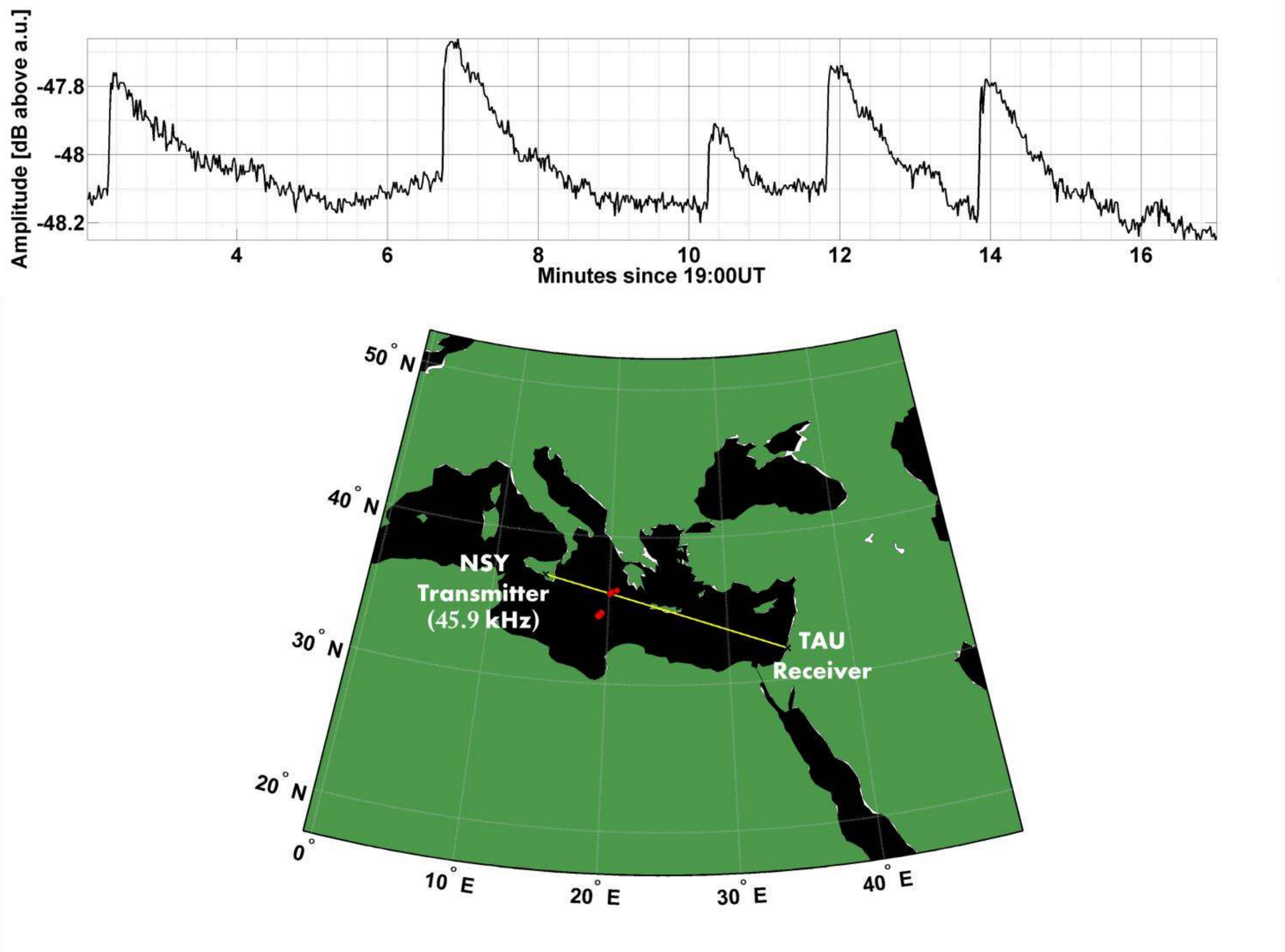

Figure 2: (Top) Lightning-induced perturbations in the NSY (Sicily) transmitter signal amplitude, as measured by the Tel-Aviv University (Israel) VLF receiver on November 11, 2013, (bottom) associated lightning discharges' location (red circles), as detected by the WWLLN receivers. The yellow curve depicts the TRGCP.

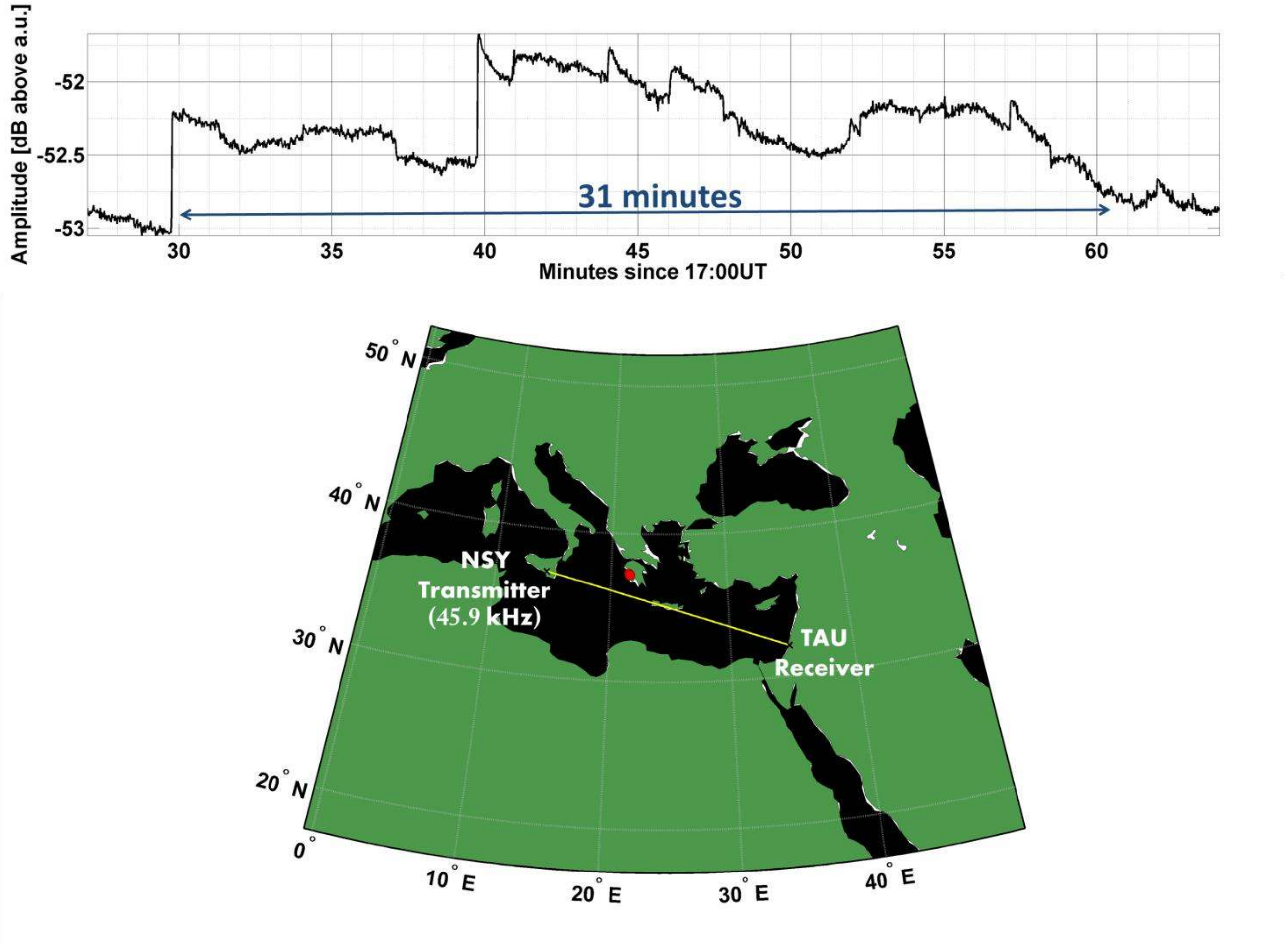

Figure 3: Same as Figure 2, but for 'long recovery early events' on November 25, 2013.

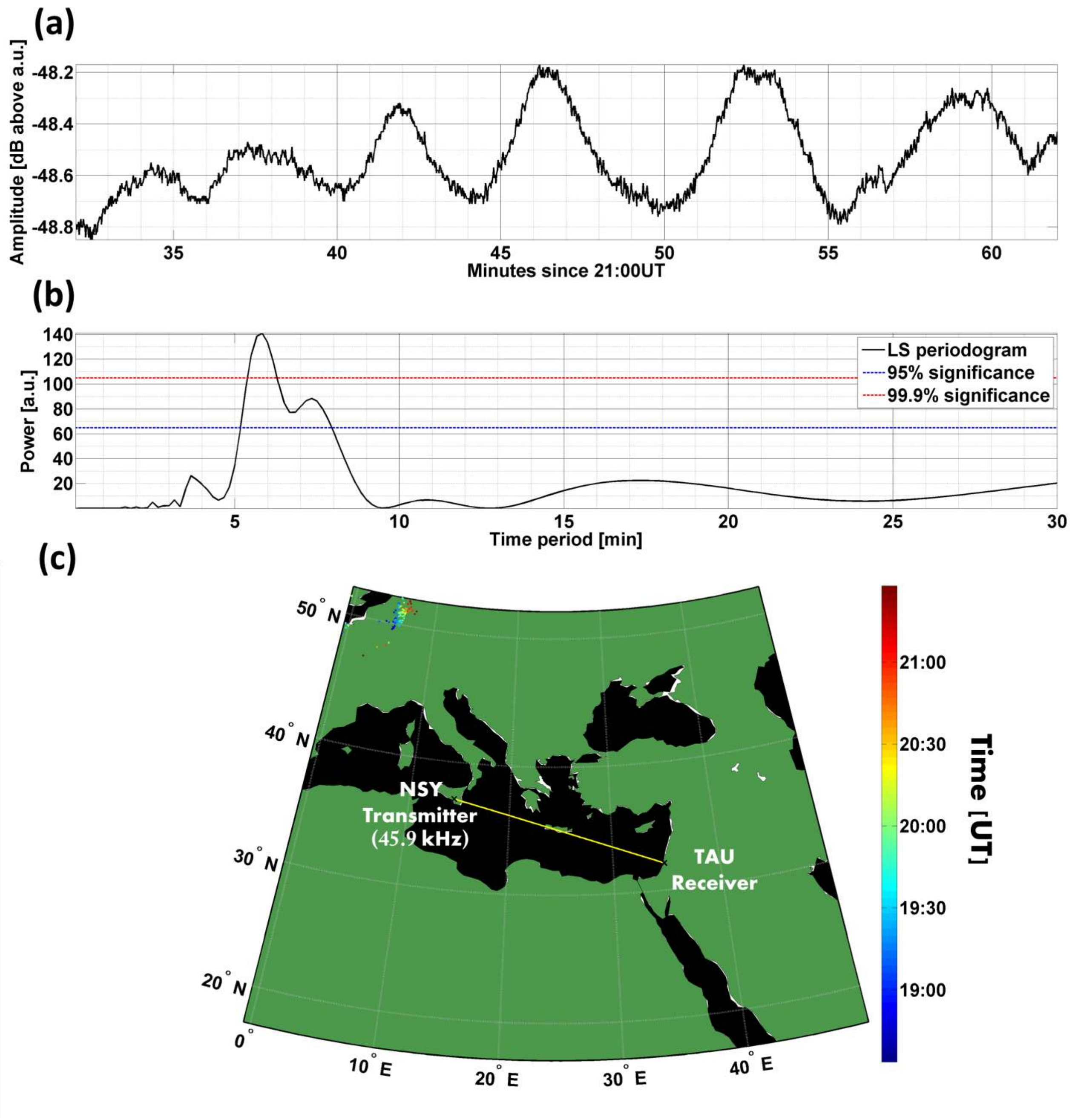

Figure 4: (a) short period gravity wave signatures in the NSY (Sicily) transmitter signal amplitude, as measured by the Tel-Aviv University (Israel) VLF receiver on October 22, 2013, (b) Lomb-Scargle periodogram of the amplitude time series (black curve). The blue (red) dashed curves represent the 95% (99.9%) statistically significance thresholds, respectively (c) lightning discharges' location, as detected by WWLLN up to 3 hours prior to the identified gravity wave signatures (the discharge location colors are based on time of occurrence).

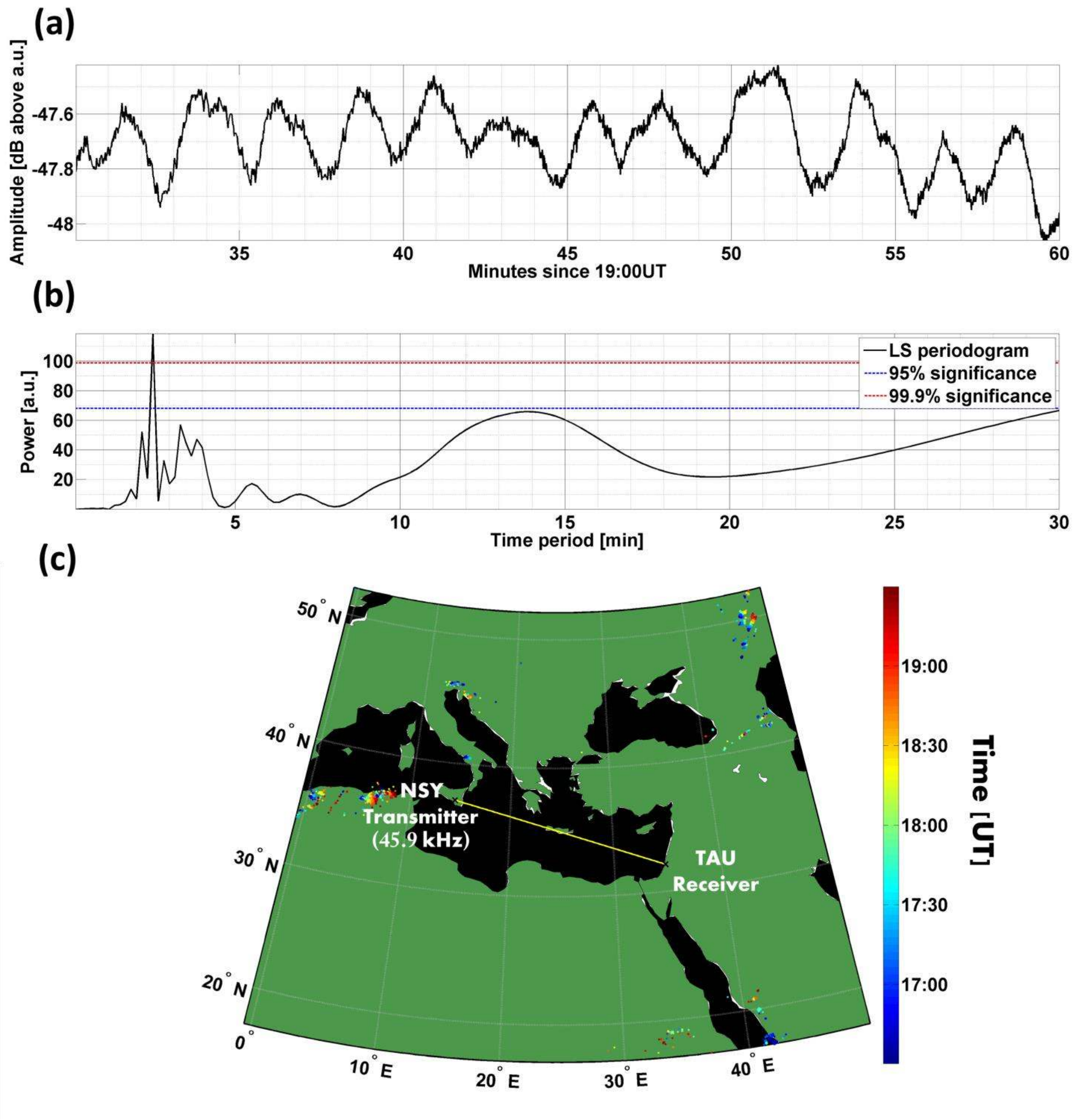

Figure 5: Same as Figure 4, but for acoustic wave signatures on September 9, 2013.

Table 1: Pressure waves in the atmosphere.

Wave type	Frequency range*	Time period	Horizontal wavelength scales	Typical sources	References
Acoustic waves / Infrasound (infrasonic waves)	f_a – *20 Hz*, f_a – acoustic cut-off frequency – typically *3.3 mHz*	*50 ms – ~5 min*	meters – *10s* of thousands of kilometers	Volcanic eruptions, earthquakes, ocean swells, thunderstorms, bolides, man-made explosions and rockets.	[*Francis*, 1975; *Drob et al.*, 2003; *Blanc et al.*, 2010; *Evers and Haak*, 2010; *ReVelle*, 2010]
Gravity waves	f_b - f_c, f_b – Brunt-Väisälä frequency – typically *2.9 mHz* f_c – Coriolis frequency (Coriolis parameter)	*~6 min* – (f_c^{-1}), f_c^{-1} equals to *12 h* at the poles	few kilometers – thousands of kilometers	Flow over topographic terrain or thermal "obstacles", convection (by thunderstorms, weather fronts, etc.), wave-wave interactions, and geostrophic adjustments.	[*Hines*, 1960; *Fritts and Alexander*, 2003; *Blanc et al.*, 2010]
Atmospheric tides		*hours – 24* hours (harmonics of a full solar day, for both the migrating and non-migrating components)	thousands of kilometers - *10s* of thousands of kilometers ($2\pi R_e$ at the equator, where R_e – Earth's radius)	Tropospheric and stratospheric solar insolation absorption by H_2O and O_3 molecules generates the migrating component (an additional O_2 and N_2 absorption effect exists within the thermosphere), solar insolation absorption by water in clouds and weather systems generates the non-migrating component. In addition, gravitational effect from the Sun and the moon.	[*Lindzen and Chapman*, 1969; *Forbes and Garrett*, 1979; *Forbes*, 1982, 1995]
Planetary waves		Few days – few weeks	thousands of kilometers - *10s* of thousands of kilometers	Earth's rotation (Coriolis effect) together with topographic, thermal, or convective obstacles.	[*Forbes*, 1995; *Holton and Hakim*, 2004]

***Not given for atmospheric tides and planetary waves as their frequencies (in Hz) are extremely low, thus it is more convenient to examine only their time periods.**