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**Semi-annual oscillation (SAO) of the nighttime ionospheric D-region as detected
through ground-based VLF receivers**

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1 **Abstract**

2 Earth's middle and upper atmosphere exhibits several dominant large scale oscillations in many
3 measured parameters. One of these oscillations is the semi-annual oscillation (SAO). The SAO can
4 be detected in the ionospheric total electron content (TEC), the ionospheric transition height, the
5 wind regime in the mesosphere-lower-thermosphere (MLT), and in the MLT temperatures. In
6 addition, as we report for the first time in this study, the SAO is among the most dominant
7 oscillations in nighttime very low frequencies (VLF) narrow-band (NB) subionospheric
8 measurements. As VLF signals are reflected off the ionospheric D-region (at altitudes of ~ 65 km and
9 ~ 85 km, during the day and night, respectively), this implies that the upper part of the D-region is
10 experiencing this oscillation as well, through changes in the dominating electron or ion densities, or
11 by changes in the electron collision frequency, recombination rates, and attachment rates, all of
12 which could be driven by oscillatory MLT temperature changes. We conclude that the main source
13 of the SAO in the nighttime D-region is due to NO_x molecules transport from the lower levels of the
14 thermosphere, resulting in enhanced ionization and the creation of free electrons in the nighttime D-
15 region, thus modulating the SAO signature in VLF NB measurements. While the cause for the
16 observed SAO is still a subject of debate, this oscillation should be taken into account when
17 modeling the D-region in general and VLF wave propagation in particular.

18

1 **1. Introduction**

2 Earth's middle and upper atmosphere exhibit several dominant large scale oscillations in many
3 measured parameters. These oscillations can be found at all latitudes, from the equator to the mid and
4 high-latitudes. One of these oscillations is the semi-annual oscillation (SAO). Among different
5 parameters, the SAO can be detected in neutral atmospheric measures, e.g., the wind regime at the
6 mesosphere-lower-thermosphere (MLT) (e.g., Groves, 1972; Gregory and Manson, 1975; Lysenko et
7 al., 1994), MLT temperatures (e.g., Groves, 1972; Takahashi et al., 1995; Taylor et al., 2005; Huang
8 et al., 2006; Shepherd et al., 2006), as well as in concentrations of atmospheric species, such as
9 atomic oxygen at 80-115 km altitudes (e.g., Russell et al., 2004) and excited hydroxyl (OH^*)
10 molecules around 87 km (e.g., Takahashi et al., 1995; Marsh et al., 2006; Shepherd et al., 2006; Gao
11 et al., 2010). In addition, the charged part of the atmosphere (i.e., the ionosphere), experiences the
12 SAO, which was observed and derived in and from measurements of several parameters, e.g., the
13 ionospheric lower transition height at $\sim 180-260$ km (a level where atomic and molecular ion
14 concentrations become equal) (e.g., Lei et al., 2004), the electron and plasma density within the
15 daytime D, E, and F regions of the ionosphere (e.g., Lauter and Nitzsche, 1967; Bremer and Singer,
16 1977; Forbes et al., 2000; Peters and Entzian, 2015), and also ionospheric total electron content
17 (TEC) (e.g., Zhao et al., 2008; Opio et al., 2015).

18 Measurements of the D-region of the ionosphere, which lies at altitudes of $\sim 80-95$ km during
19 nighttime, and expands downwards to lower altitudes (~ 60 km) during daytime (mainly due to direct
20 solar extreme ultra-violet (EUV) and X-ray radiation) (Brasseur and Solomon, 2005; Inan et al.,
21 2010), are usually made using remote-sensing techniques, because these altitudes are too high for
22 weather balloons, and too low for in-situ measurements by satellites. In addition, remote-sensing
23 techniques usually do not possess the limited spatial and temporal coverage of rocket lofted
24 experiments (Rodger and McCormick, 2006). One of these remote-sensing techniques involves the
25 use of very low frequency (VLF) radio waves, spanning a frequency range of 3-30 kHz. These
26 waves, which are generated both by natural and man-made sources, propagate thousands of
27 kilometers within the Earth-ionosphere waveguide, reflected off the Earth's surface and inside the
28 ionosphere's D-region, while experiencing a very weak attenuation of ~ 2 dB/Mm (Barr et al., 2000;
29 Wait, 1957). Due to the significant difference in the D-region's characteristics (electron and ion
30 densities) between day and night (Hargreaves, 1995), the region's conductivity changes dramatically,
31 causing the D-region reflection height of VLF signals to change from as low as ~ 60 km during
32 daytime to ~ 85 km during nighttime (Hargreaves, 1995; Inan et al., 2010). Thus, as received VLF
33 signals inherently contain information of the ionosphere and its variability within the reflection

1 region (Inan et al., 2010; Rodger et al., 2012) these signals probe different altitudes during day and
2 night.

3 Because the D-region's formation and chemistry are tightly bounded to the neutral MLT (Brasseur
4 and Solomon, 2005), it is believed that the D-region is affected by the same forcings, experiencing
5 similar oscillations (e.g., Schmitter, 2011; Silber et al., 2013; Marshall and Snively, 2014). As far as
6 the current authors know, there are no previous works showing the SAO dominating the natural long-
7 term oscillations in the nighttime D-region, apart from the work of Toledo-Redondo et al. (2012),
8 who presented a SAO indication within the equatorial latitudes, by using space-based ELF-VLF data
9 from the DEMETER micro-satellite. In this paper, we present evidence of a strong SAO, detected in
10 the low and mid-latitude nighttime D-region, through ground-based VLF measurements in both
11 hemispheres.

12

13 **2. Instrumentation and methodology**

14 During this study, we used ground-based VLF narrow-band (NB) signals, which are generated by
15 VLF transmitters. These man-made transmitters are used nowadays worldwide primarily for
16 communication with military submarines (Clilverd et al., 2009; Rodger and McCormick, 2006).
17 However, they are extremely well suited to long-range remote-sensing of the D-region, because of
18 their high radiated power, their nearly continuous operation, and their fixed location and frequency
19 band (e.g., Clilverd et al., 2009; Inan et al., 2010). As the VLF signals travel from the transmitters to
20 the receivers along a great circle path (GCP), a time series of their recorded amplitude and phase
21 give an indication on the changes of the D-region along the GCP.

22 The NB signals were recorded at two VLF receiving stations. The first station is located at the Emilio
23 Segre' Observatory of the Israeli Cosmic Ray and Space Weather Center, at Mt. Hermon (MH), in
24 the north part of Israel ($33.18^{\circ}N$, $35.47^{\circ}E$). This VLF receiving system is part of the AWESOME
25 network (Cohen et al., 2010). The antenna is built from two orthogonal triangular loop antennas that
26 measure the two horizontal components of the VLF magnetic field. Each loop has a baseline of 2.6
27 m, and 1.3 m height, giving an area of approximately $1.69 m^2$ for each loop, and has a total number of
28 12 turns. The loop antenna impedance is $0.85 mH$ and 1Ω .

29 The second VLF receiving station, which is located in Dunedin (DN), New Zealand ($45.8^{\circ}S$,
30 $170.5^{\circ}E$), is part of the AARDDVARK network (Clilverd et al., 2009) and is operated by the
31 University of Otago. Its antenna measures the normal component of the VLF electric field, thus

1 making both of the stations' measurements equivalent. The VLF signal is recorded by an "OmniPAL"
2 narrowband VLF receiver (Dowden et al., 1998).

3 Global lightning activity responds to the Earth's surface air temperature on both the semiannual and
4 annual timescales, within the tropical belt and the mid to high latitudes, respectively (Williams,
5 1994). As most of the electro-magnetic (EM) energy generated by lightning discharges (termed
6 'sferics'), are radiated within the ELF (extremely low frequencies) and VLF bands (peaking between
7 *5-10 kHz*) (Cummer, 1997; Rakov and Uman, 2003), and due to the weak attenuation of VLF signals
8 (as mentioned above), sferics originating in the tropics and mid to high latitudes can be easily
9 detected at MH or DN. Although lightning pulses are very powerful, their duration is very short (up
10 to the order of 10^{-4} sec) (Rakov and Uman, 2003). Nevertheless, more than a thousand active
11 thunderstorm are present on average, at any given moment (Mezuman et al., 2014). Altogether,
12 global lightning activity generates significant VLF EM fields that can produce interference with the
13 NB measurements, inducing the lightning activity natural oscillations within the NB data. Therefore,
14 in order to investigate long-term oscillations in NB amplitudes, the background VLF noise, which is
15 created mainly by lightning discharges (Barr et al., 2000), should be removed.

16 The MH receiving system records the NB signals continuously. However, as discussed above, these
17 signals are potentially biased by the background VLF sferics. Thus, we decided to use the broad-
18 band (BB) data, which consists of signals of the whole VLF band, and is recorded in a synoptic
19 mode, i.e., each data file is a recording for one minute every *15* minutes. For the extraction of a NB
20 signal for a certain VLF transmitter frequency from the BB data, every minute of data was filtered
21 using a Parks-McClellan finite impulse response (FIR) band-pass filter, with a pass-band width of
22 *300 Hz*, where the VLF transmitter's central broadcast frequency lies at the middle of the pass-band.
23 In order to represent the noise at the transmitter's frequency, the original BB data was filtered again,
24 one time when the middle of the pass-band was *300 Hz* above the VLF transmitter's central broadcast
25 frequency, and another time when the middle of the pass-band was *300 Hz* below the VLF
26 transmitter's central broadcast frequency. The average of the two filtered noise time series was then
27 subtracted from the filtered NB signal time series, assuming that the average noise represents the
28 background noise in the frequency band of the transmitter's broadcast. Thus, we received the NB
29 data with a strongly reduced bias of the background lightning noise. We then average the resultant
30 NB signal every *10* seconds, thus receiving up to *24* values per hour made up of *10 s* averages across
31 the one in fifteen minute observation cadence.

32 Unlike the MH VLF receiver, the DN VLF receiving station operates the OmniPAL receiver, which
33 also records the NB signals continuously. During the data acquisition process, this receiver's

1 software uses a noise clipping algorithm, that reduces the effect of lightning impulses over the NB
2 data [see Dowden *et al.*, 1998]. Thus, there are no additional procedures needed in order to
3 investigate the NB received signals, which are being recorded at 1 Hz .

4 In this paper, we limit ourselves to observations from two VLF transmitters; with callsigns NSY
5 (38.00°N , 13.50°E , broadcasting at 45.9 kHz), and NWC (21.82°S , 114.17°E , broadcasting at 19.8
6 kHz). These transmitters were chosen as their operating power was constant along the datasets, their
7 signal-to-noise-ratio was relatively high, and their data were fairly complete. The NSY transmitter's
8 signal was recorded at MH between the years 2009-2012, while the NWC signal was recorded at DN
9 between the years 2005-2010 (Figure 1). We extracted from each time series of transmitter-receiver
10 amplitudes two datasets of the average amplitude during the midday and midnight hours, when the
11 solar elevation angle at the middle of the GCP during equinox was at its maximum and minimum,
12 respectively. It should be mentioned that atmospheric tides, which have time periods that are sub-
13 harmonics of a solar day (Oberheide *et al.*, 2003) can possibly affect our data. Therefore, a 24-h data
14 average should have been used in order to remove the tidal effect. However, as the ionosphere (and
15 as a result, the received NB signals) changes significantly between day and night, this was not
16 possible. As a result, we used one hour mean amplitudes and not the daytime or nighttime amplitudes
17 spanning those entire time periods, in order to reduce tidal averaging effects (assuming they exist) to
18 the minimum possible. In other words, because daytime and nighttime lengths change throughout the
19 year, in the case of complete daytime and nighttime averaging, a different window size of the tidal
20 oscillation would have been averaged every day.

21

22 **3. Results**

23 The midday and midnight one-hour-mean 30-days running average time series for the MH-NSY and
24 DN-NWC GCPs' deviation from the mean amplitude (of the entire time series) are shown in Figure 2
25 (black solid curves). As can be seen, all the time series exhibit a strong oscillatory behavior, with
26 higher amplitudes in the midnight data than in the midday data. The midday data in both GCPs show
27 a dominant oscillation with longer time periods than the midnight data in both GCPs. Examination of
28 the apparent time periods of the large oscillations shows that they appear to correspond to the annual
29 oscillation (AO) and SAO. Therefore, we wanted to fit these harmonics to the data and examine the
30 level of agreement with the different time series.

31 However, in addition to these oscillations, a trend can be seen in the time series; negative in both
32 DN-NWC time series, positive in the MH-NSY midnight data, but hard to determine in the MH-NSY

1 midday data, as the trend seems to shift from negative to positive around mid-2010. Examination of
 2 the datasets time span shows that the DN-NWC data was acquired during a period when solar
 3 activity was dropping towards a minimum, while the MH-NSY data was acquired when the Sun
 4 started to become active again, as part of its 11-year cycle. Thomson and Clilverd (2000) have
 5 showed a positive correlation between VLF amplitudes and solar activity. Therefore, as both of our
 6 datasets show a general positive correlation with solar activity, we may conclude that the trend is a
 7 result of solar activity. Nevertheless, because we did not have enough data to cover a full 11-year
 8 solar cycle, it was problematic to fit the data with an 11-year harmonic. Thus we decided that a linear
 9 fit to the data would be best. Therefore, the time series were fitted with curves, described by
 10 following equation:

$$11 \quad A_{fit}(t) = A_0 + St + A_{SAO} \cos\left[\frac{2\pi(t-t_{SAO})}{365.25}\right] + A_{AO} \cos\left[\frac{2\pi(t-t_{AO})}{182.625}\right] \quad (1)$$

12 Where A_{fit} is the fitted curve, t represents the time steps (in days), A_0 is the mean amplitude (which is
 13 equal to 0 in this case), S is the linear fit coefficient, A_{SAO} and A_{AO} are the fitted SAO and AO
 14 amplitudes (respectively), and t_{SAO} (t_{AO}) represents the SAO (AO) maximum time of year,
 15 respectively. Both the linear and the harmonic fits were made using a least squares method over all
 16 of the data points. The fitted curves are shown in Figure 2 (dashed red). As can be seen, these simple
 17 curves follow the VLF amplitude patterns fairly well. Pearson's correlation coefficients between the
 18 time series and the fitted curves were calculated and are shown at the bottom right of each panel. As
 19 the correlation coefficients span from values of 0.53 up to 0.84 (all statistically significant), we can
 20 deduce that the simple curve may explain from 28% up to ~70% of the midday and midnight long-
 21 term variability.

22 The simple model's parameters described in equation (1), can be investigated as well. Comparison of
 23 the two oscillation amplitudes (A_{SAO} and A_{AO}) shows that during midday A_{SAO} is three times weaker
 24 than A_{AO} , but during midnight it is stronger than A_{AO} by up to ~60%. Moreover, A_{SAO} appears to have
 25 a very strong peak to peak amplitude of 3.3 dB in MH-NSY and 4.2 dB in DN-NWC. By running the
 26 Long Wave Propagation Capability (LWPC) model (Ferguson, 1998), we found that these strong
 27 amplitudes are equivalent to h' (ionospheric base) change of no more than 1.8 km and 1.3 km in DN-
 28 NWC and MH-NSY, respectively, or 0.13 km^{-1} and 0.15 km^{-1} in β (electron density profile
 29 sharpness), respectively. By using the standard D-region electron number density profile (Wait and
 30 Spies, 1964),

$$31 \quad N_e(h) = 1.43 \cdot 10^7 \exp(-0.15h') \cdot \exp[(\beta - 0.15)(h - h')] \text{ cm}^{-3} \quad (2)$$

1 Where h is the altitude and N_e is the electron number density, it can be shown that the SAO alone is
2 equivalent to more than doubling of the electron number density at 85 km . We believe that the actual
3 solution for the ionospheric profile is a combination of changes in h' and β , but as we do not have
4 reliable phase measurements, the actual solution cannot be calculated at the moment.

5 By examining the t_{SAO} values for the midnight data, it is found that the SAO maxima occur up to a
6 month prior to Earth's winter and summer solstices (not shown). This is quite surprising, as we would
7 normally expect the maxima of a SAO-driven forcing to occur around equinox (e.g. Opio et al.,
8 2015; Taylor et al., 2005; Williams, 1994). The fitted curve for MH-NSY GCP data shows the SAO
9 maxima occurs during mid-November and mid-May, while the oscillation maximizes at the
10 beginning of December and June in the DN-NWC GCP data. Thus, a 16 days phase difference exists
11 between the two data sets.

12 In order to confirm our findings of the apparent dominating SAO in the nighttime NB measurements
13 (and possibly within the nighttime D-region), spectral analysis was performed. Because the data were
14 unevenly sampled (due to transmitter off-times, receiver malfunctions, etc) it was not possible to use
15 Fast Fourier Transforms (FFT), which demand constant time steps between samples. In addition, the
16 FFT calculated frequencies are directly determined by the dataset length and sampling rate, hence
17 there is no option to choose which exact frequencies to inspect. Moreover, a transformation of the
18 FFT frequencies into the oscillations time periods (via $T=1/f$) results in a very high time period
19 resolution at very low values (i.e., high frequencies) and very poor resolution at high values (i.e., low
20 frequencies, more likely to represent the SAO long-term oscillations). Therefore, we have analyzed
21 the midnight one-hour-mean data using the Lomb-Scargle (LS) periodogram (Lomb, 1976; Scargle,
22 1982), which results in spectral power of the data at user-determined frequencies (and hence, time
23 periods), and allows the spectral analysis of unevenly sampled data (Press and Rybicki, 1989).

24 Figure 3 shows the LS periodogram of the midnight (unsmoothed) MH-NSY (top panel) and DN-
25 NWC (bottom panel) one-hour-mean VLF amplitude anomalies, with arbitrary power units (as a
26 result of the LS periodogram procedure). The dashed red line denotes 95% confidence, which was
27 calculated using the quantile function (Wilks, 2006). The inspected time periods range from 2 - 730
28 days, thus spanning from as short as the datasets Nyquist frequency up to two years. Examination of
29 the MH-NSY periodogram confirms that the SAO at ~ 180 days is by far the most dominant and
30 significant oscillation within the data. The second peak of the periodogram is of 343 days. We
31 ascribe this peak to the AO, while we assume that the reason for the difference in periodicity from
32 365 days is attributed to our dataset covering only four years, thus containing only four AOs, which
33 are not fully sampled, due to the data gaps (data coverage of $\sim 57\%$ along the time span). Secondary

1 peaks, which do not pass the 95% significance threshold, are seen at time periods of 47, 96, 137, and
2 212 days. Some of these oscillations might be higher harmonics of the SAO, but it is not possible to
3 explain them at the moment, leaving this topic for future studies.

4 The SAO appears at ~ 180 days to be even more pronounced and significant in the DN-NWC
5 periodogram and is the dominant oscillation within the midnight data. As can be seen in the lower
6 panel of Figure 3, the second-highest peak is of time period of 241 days (~ 8 months), an oscillation
7 which is quite unexpected, but does not appear in the MH-NSY data. The probable signature of the
8 AO seen in this periodogram is also statistically significant, peaking at 366 days. Here, the secondary
9 peaks which do not pass the 95% significance threshold are located at time periods of 152 and
10 toward 730 days (~ 2 years), the latter might be hinting of a very weak quasi-biennial oscillation
11 (QBO) affect.

12 As the nighttime measurements of VLF amplitude are highly variable, robustness tests were made to
13 examine our nighttime data and methodology, by removing 20% of the raw data's measured points.
14 Later, we added Gaussian noise into the raw data's measured points, with a standard deviation equal
15 to that of the raw data. Each of these tests was repeated for 100,000 iterations. The results showed
16 that in both datasets, the SAO and the solar cycle trend passed these robustness tests in 100% of the
17 runs, and therefore strengthen our findings. The AO passed the robustness tests in 100% of the runs
18 for the DN-NWC dataset, but did not prove to be statistically significant for the MH-NSY dataset, as
19 only 91% of the data removal runs and 58% of the noise addition runs kept this oscillation
20 statistically significant.

21

22 **4. Discussion**

23 In this study, we analyzed several years of VLF NB data received in both hemispheres, during
24 midday and midnight hours. The analysis shows that the AO dominates midday VLF amplitudes, and
25 the SAO is the strongest oscillation during the hour long period centered on the GCP midnight. Both
26 the SAO and the essential differences between daytime and nighttime dominating oscillations should
27 be explained. We believe that the sources for both of these observations are of chemical and
28 dynamical origin, which take place in the transport of species, and tidal forcing. These sources shall
29 now be discussed.

30 **4.1. D-region ions and dynamical transport of neutral species**

31 When analyzing Earth's ionosphere in general (and the D-region in particular), we can assume
32 electrical neutrality (Kelley, 2009):

$$n^+ = n^- + n_e \quad (3)$$

Where n^+ is the positive ion number density, n^- is the negative ion number density, and n_e is the electron number density. VLF radio signals interact and are reflected off the D-region mainly by electrons rather than ions, as a result of their much lower mass (Inan and Inan, 2000). Using equation (3), we can assume that the electron number density is determined primarily by the dominant ion densities. Therefore, we can investigate the dominant D-region's ions, their production, and distribution.

As mentioned in the Introduction section, VLF reflection height ranges from $\sim 60\text{-}70\text{ km}$ to $\sim 85\text{ km}$ during daytime and nighttime, respectively (Inan et al., 2010; Rodger and McCormick, 2006). Although both altitudes are part of the D-region, their chemical composition and dynamical processes are different and both are very complicated. The lower altitudes of the D-region (below $\sim 80\text{ km}$) are dominated by positive cluster ions (mostly of the type $H(H_2O)_n^+$), due to a relatively high neutral density in general, and water molecules in particular, which enable an effective three-body reaction (with O_2^+ and NO^+ ions), thus creating this type of ions (Glukhov et al., 1992; Goldberg and Aikin, 1971; Mitra, 1981; Narcisi and Bailey, 1965). Cluster ions have a rapid recombination rate, and many of the chemical reactions involving them are strongly temperature sensitive (Kelley, 2009; Pavlov, 2014). Therefore, the ion composition of this region should be quite variable with season and latitude, and sporadic changes associated with local temperature variations should be observable (Brasseur and Solomon, 2005).

The higher altitudes of the D-region (above $\sim 80\text{ km}$) are dominated by NO^+ ions, mainly as a result of strong ionization by the solar Lyman- α line (121.6 nm). O_2^+ ions are also abundant in this region, and are created mostly by solar radiation in the $102.7\text{-}111.8\text{ nm}$ wavelengths. The three-body reactions that form cluster ions are less frequent in this region, due to the lower neutral density and the lack of water vapor, making cluster ions much less abundant (Mitra, 1981; Narcisi and Bailey, 1965). The Lyman- α radiation scattered by the hydrogen geocorona at the uppermost part of the atmosphere, is still a major ionization source during nighttime, though 2-3 orders of magnitude smaller than during daytime (Brasseur and Solomon, 2005). Although NO^+ and O_2^+ ion recombination rates are orders of magnitude slower than cluster ions (Kelley, 2009; Pavlov, 2014), the lifetime of ions in the D-region is short compared to the transport time scale, hence the ion concentrations are determined by a photochemical equilibrium between production and loss processes (Brasseur and Solomon, 2005). Nevertheless, in addition to the strong dependence on solar radiation variability, i.e., solar activity, solar zenith angle, etc., the production rates of NO^+ and O_2^+

1 ions are proportional to NO and O_2 neutral molecules, respectively (Pavlov, 2014). In the neutral
2 atmosphere, as we rise from ~ 60 km to ~ 90 km, the chemical lifetime of NO_x molecules increases, in
3 comparison with the typical constant for vertical exchanges K_{zz} . Thus, as we reach higher altitudes
4 within the D-region, which coincides with these altitude levels, the role of local dynamics (vertical
5 exchange) becomes more significant (Solomon et al., 1982a). In addition, the amplitudes of gravity
6 and planetary waves penetrating into the MLT (as a function of season (Lindzen, 1981)) grow
7 exponentially with altitude, as the ambient density drops (Ern et al., 2015; Smith, 2012). Therefore,
8 downward transport of NO molecules, which are created mainly in the lower thermosphere (Solomon
9 et al., 1982a, 1982b), can increase the D-region NO^+ concentrations, depending on the vertical wind
10 patterns (e.g., Clilverd et al., 2006).

11 Altogether, changes in neutral NO occurring by dynamical forcing will affect NO^+ ion concentrations
12 within the D-region, mainly at higher altitudes. At lower altitudes of the D-region, where daytime
13 VLF signals are reflected, we would expect chemical processes to dominate over NO dynamics, and
14 therefore a very strong signature of solar insolation changes (which are seen mainly in the AO),
15 together with relatively weak perturbations caused by other forcings and temperature changes are
16 observed (e.g., Schmitter, 2011; Silber et al., 2013). At higher altitudes within the D-region, where
17 nighttime VLF signals are reflected, dynamical processes are much more pronounced, thus
18 oscillations such as the SAO, which are driven by dynamical transport of important species as well as
19 dynamical forcing (e.g., gravity and planetary waves) are much stronger and thus more easily
20 detected.

21 There are additional factors that can be expected to increase the detected SAO amplitudes in NB
22 measurements. The first factor is the SAO of atomic oxygen in the MLT (e.g., Russell et al., 2004).
23 These atoms, which can also be transported from higher regions of the atmosphere, are important for
24 molecular ion chemistry, through numerous chemical reactions (Pavlov, 2014). Thus, they might
25 increase the SAO amplitude in the D-region, thus enhancing the measured oscillation in the received
26 VLF amplitudes. An additional affect comes from the MLT temperatures that, as mentioned in the
27 Introduction section, are also experiencing a SAO. MLT oscillatory temperature changes can
28 influence VLF received signals, by modifying the electron collision frequency, recombination, and
29 attachment rates (see the discussion in Silber et al., 2013), thus increasing the measured SAO in the
30 received NB amplitudes.

4.2. The VLF SAO phase and its comparison with satellite data

All of the above mentioned phenomena can explain the measured dominating SAO in the received VLF amplitude differences at midnight. However, the phase of the measured SAO may be determined by a number of affects. The observed difference between the SAO phase in DN-NWC and MH-NSY might originate in the phase differences between the phenomena as a function of latitude, or due to the changes of the phenomena's phases as a function of latitude and altitude (e.g., Groves, 1972; Lysenko et al., 1994; Russell et al., 2004; Taylor et al., 2005). In order to test our hypothesis, we decided to compare the phase of SAO in the VLF measurements, to the SAO phase detected in peak emission values of the OH^* $2.0 \mu m$ emission band, using the SABER instrument on-board the TIMED satellite (Mlynczak, 1997; Russell III et al., 1999). The OH^* airglow layer peaks around $87 km$ (Baker and Stair Jr, 1988), in the altitude vicinity of the VLF nighttime reflection height. We used the $2.0 \mu m$ data as it is a direct measure of the chemical reaction which creates the OH^* (Mlynczak, 1999; Mlynczak et al., 2013), and is tightly linked to atomic oxygen abundance as well as MLT dynamics (Gao et al., 2010; Marsh et al., 2006). A 60-day running-mean of the data was created, in order to obtain good local time coverage (Marsh et al., 2006). Only the SABER nighttime data (in a 1° zonal-mean resolution) was examined, as the daytime and nighttime OH^* airglow layers behave differently (Gao et al., 2015; Marsh et al., 2006; Smith, 2004). An examination of the 10 years (2002-2012) average peak emission is presented in Figure 4. An apparent SAO seems to dominate the OH^* emission pattern, especially in the equatorial and mid-latitudes, and a strong phase propagation of the oscillation towards the poles can be seen. It can also be seen that the OH^* emission does not show a constant SAO phase as a function of latitude (see also Gao et al., 2010), similar to the analyzed VLF data. We examined the OH^* data (1° width) in the latitudes that match the middle of the two transmitter-receiver GCPs, during the same time period as the VLF data (2005-2010 for DN-NWC, and 2009-2012 for MH-NSY), by using equation (1), and finding the SAO maximum phase. The results show that the SAO in the OH^* leads by 6 days the MH-NSY VLF data, and by 17 days the DN-NWC VLF data. The differences in the oscillation phase might be due to the long GCPs, which (unlike the data used from SABER measurements) are over 1° of latitude width. In addition, the OH^* emission was zonally averaged, and therefore it was less sensitive to local perturbations like the VLF data. Nevertheless, as the maxima are all in the same 30-day window, we note that both parameters are affected by similar dynamics and forcing. Moreover, according to our knowledge, the exact reason for this phase lag as a function of latitude, has not been investigated previously. As the dynamics of the MLT will affect multiple wind and wave fields as well as species concentrations, this topic should be investigated in future studies.

4.3. Tidal effects

Finally, we should point out one caveat in this study. Marsh et al. (2006) concluded that the observed SAO in OH^* emission is a result of the seasonal change in atmospheric tidal amplitudes, and is not caused by changes in diffusive transport, as was previously proposed. This might also be a significant effect in the VLF NB measurements, as tidal amplitudes also experience a SAO, driven by the strong shears in zonal mean zonal winds (McLandress, 2002a, 2002b). Tides tend to break at ~ 85 km (Lindzen, 1981) and like other large-scale atmospheric waves, are modifying MLT temperatures by ~ 5 K at mid-latitudes (Marsh et al., 2006). In addition, tides are able to generate several kilometers of atmospheric species transport from above and below (Marsh et al., 2006; Smith, 2004). As we mentioned in the Methodology, we average a constant (local-time) hour in order to reduce the tidal effect, as it was not possible to average 24-h of VLF data, due to the significant changes in the ionosphere between day and night. However, this procedure will not have completely solved the tidal effect problem for two reasons. Firstly, non-migrating tides have phases that are non-Sun-synchronous. As the phases of this type of tide do not follow the Sun's apparent motion in the sky, their total amplitude and phase during the midnight and midday one-hour-averaged could cause leakage of the tidal oscillation into the VLF data's long-term frequencies, known as tidal aliasing (Oberheide et al., 2003). Secondly, the migrating tides, which are Sun-synchronous, should have no effect over the data, when using a constant local time at a single station. However, as the transmitter and receiver are always located at different latitudes and longitudes, the migrating tides are in different stages of their phase along the GCP, and they may have an additional influence over the VLF received signals, amplifying the SAO.

5. Conclusion

A strong SAO was detected in the nighttime D-region using the amplitude of ground-based VLF NB signals. This oscillation dominates over all other long-term oscillations. We believe that the main source of the SAO is most likely to be due to NO_x molecules transported from upper levels of the atmosphere. This transport results in enhanced ionization and the creation of additional free electrons in the nighttime D-region, thus inducing the SAO signature on VLF NB amplitude measurements. Nevertheless, further research and analysis should be undertaken in order to confirm our conclusions. A good test would involve the use of both high-end chemistry and GCM models due to the complexity of the D-region, or by analysis of NO^+ measurements from space, which might be

1 acquired in the future using instruments such as NASA's Middle Atmosphere Sounder and Thermal
2 Emission Radiometer (MASTER) (Mlynczak et al., 2014).

3 In addition, as far the authors are aware, no current VLF wave propagation model (e.g., Ferguson,
4 1980) takes into account SAO-forcing of the D-region and hence the impact on received VLF
5 signals. As we have shown in this paper, the SAO influence over VLF signal attenuation is
6 significant, affecting the received signal amplitudes by several *dB*. VLF signal studies are an
7 important tool for understanding the D-region of the ionosphere, being low-cost, with high temporal
8 resolution, and potentially high spatial resolution (by using numerous receivers at many different
9 locations). Therefore, propagation models should take this oscillation into consideration, in order to
10 acquire better and more precise results, particularly over long time periods.

11

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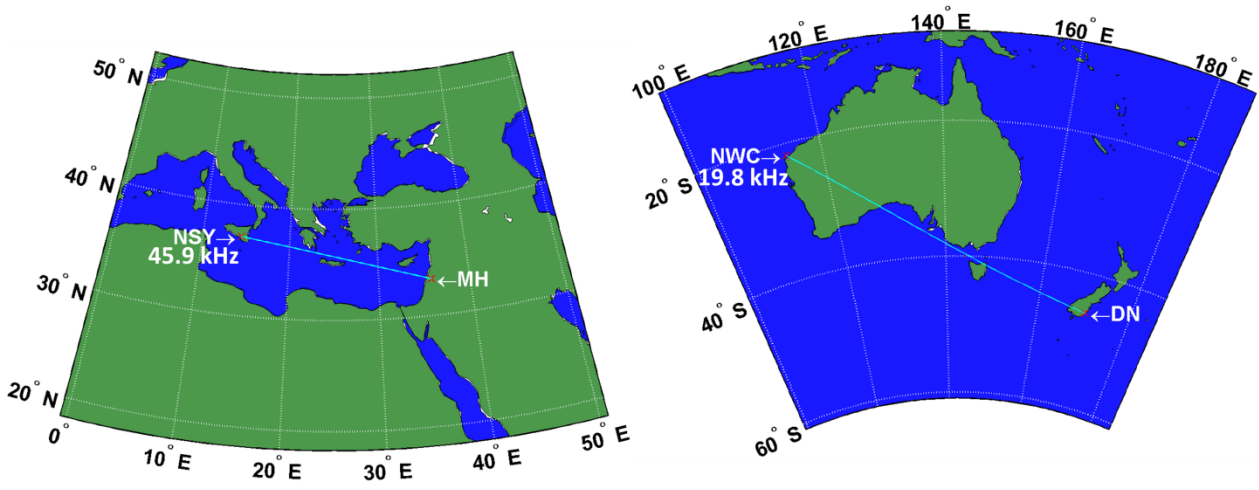
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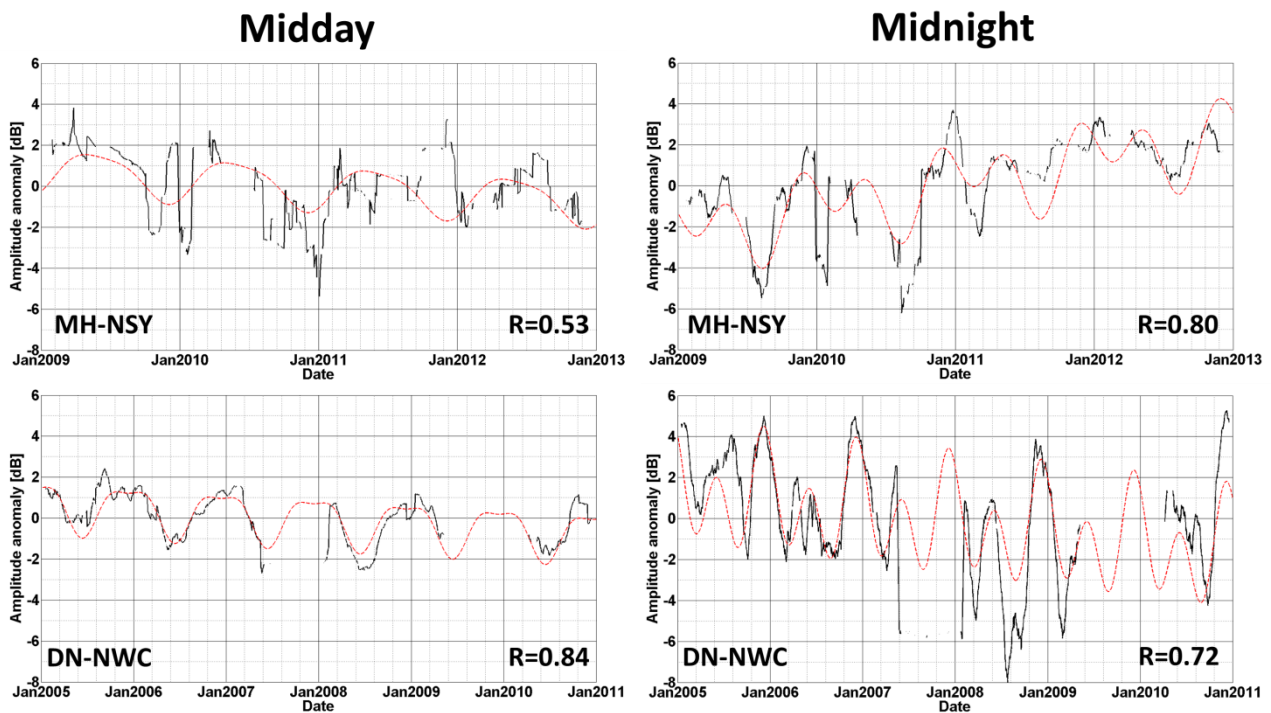
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2 Figure 1: MH-NSY (left panel) and DN-NWC (right panel) transmitter-receiver great circle paths,
 3 together with their corresponding frequencies.

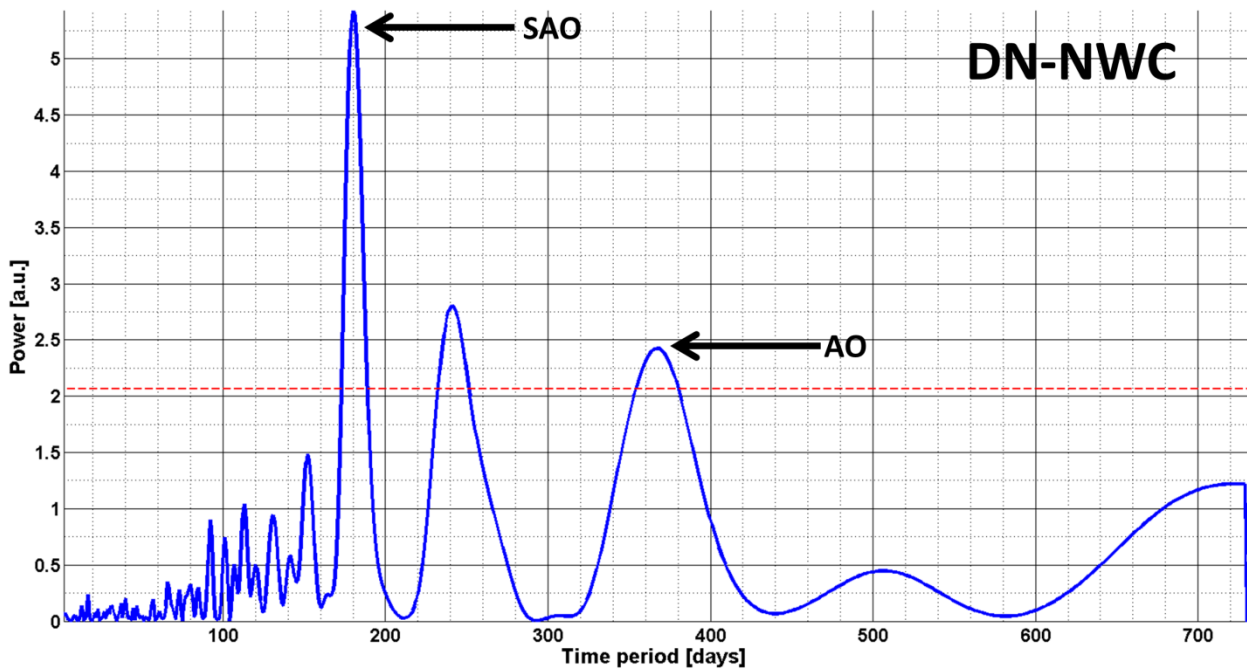
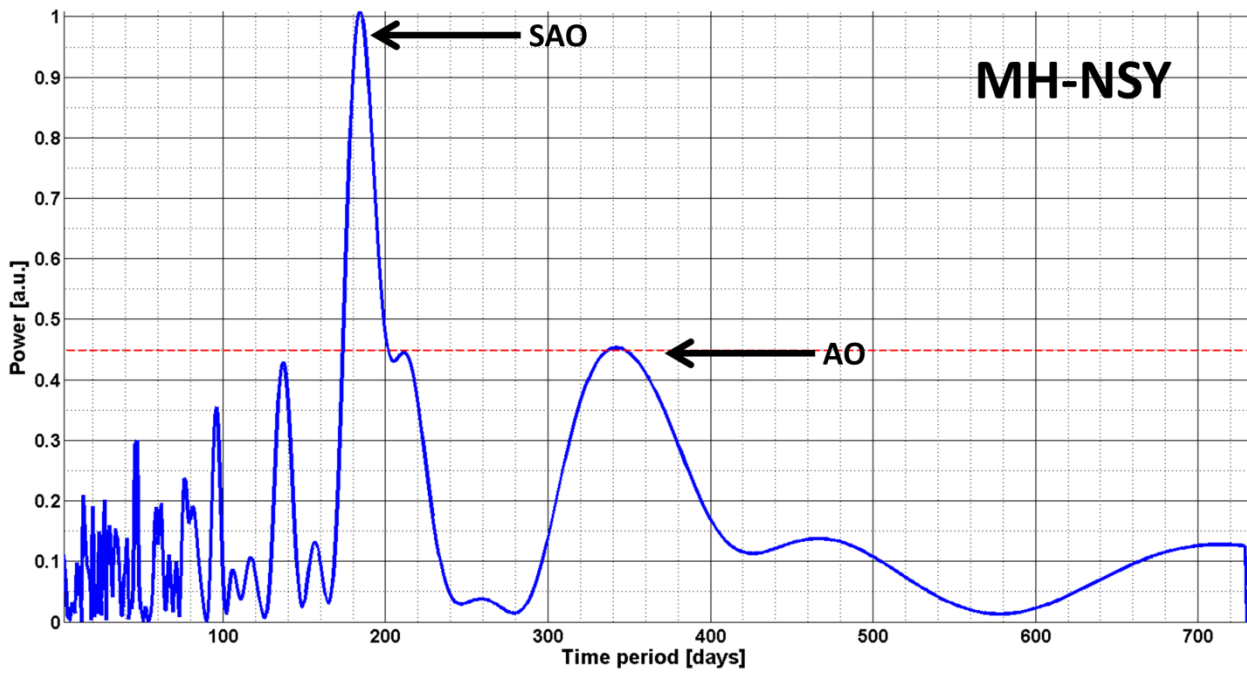
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2 Figure 2: Midday (left panels) and midnight (right panels) one-hour-mean 30-days running average
 3 time series of MH-NSY (top panels) and DN-NWC (bottom panels) transmitter-receiver GCPs'
 4 deviation from the mean amplitude (solid black curves). The dashed red curves show the
 5 combination of the SAO, AO, and linear fit to the data series (see equation (1)) A Pearson's
 6 correlation coefficients between the red and black curves is shown at the bottom right of each panel.

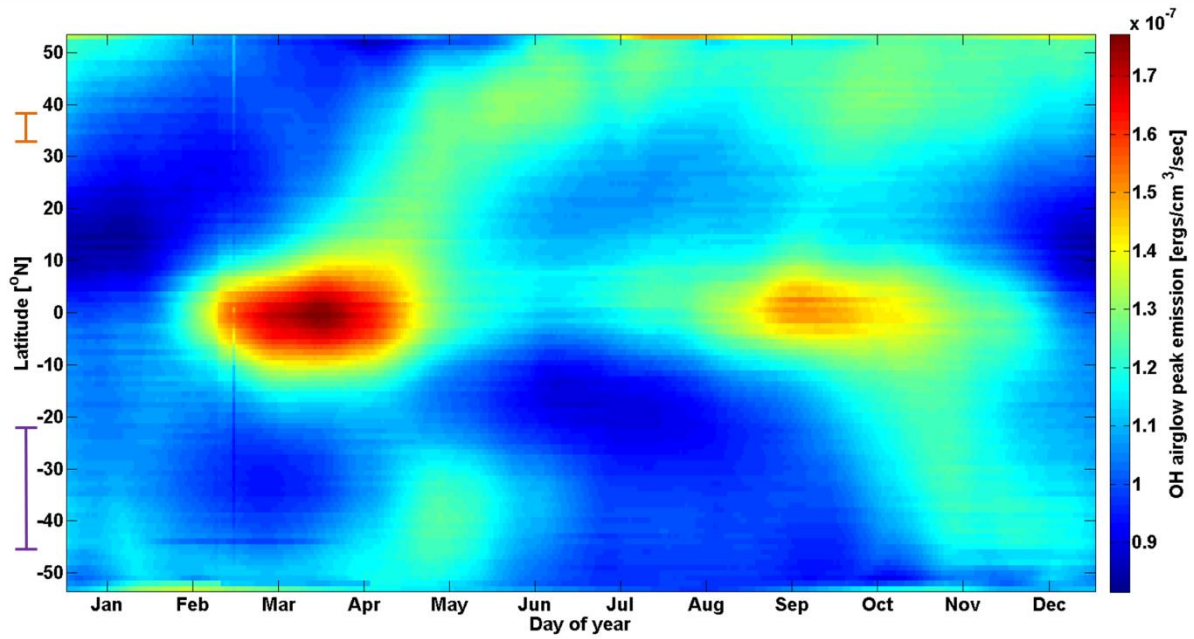
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2 Figure 3: Lomb-Scargle periodogram of the midnight MH-NSY (top) and DN-NWC (bottom) GCPs'
 3 one-hour-mean VLF amplitude anomalies in arbitrary power units. The dashed red line denotes the
 4 95% confidence level.

5



1

2 Figure 4: OH^* $2.0 \mu m$ peak emission zonal-mean 60-days running-mean, averaged over the years
 3 2002-2012.. The bars on the left denote the MH-NSY (orange) and DN-NWC (purple) GCP latitude
 4 ranges (see Figure 1).