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A study of changes in apparent ionospheric reflection height within individual lightning flashes

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ABSTRACT

Ionospheric reflection heights estimated using the zero-to-zero and peak-to-peak methods to measure skywave delay relative to the ground wave were compared for 108 first and 124 subsequent strokes at distances greater than 100 km. For either metric there was a considerable decrease in average reflection height for subsequent strokes relative to first strokes. We showed that the observed difference cannot be explained by the difference in frequency content of first and subsequent return-stroke currents. Apparent changes in reflection height (estimated using the peak-to-peak method) within individual flashes for 54 daytime and 11 nighttime events at distances ranging from 50 km to 330 km were compared, and significant differences were found. For daytime conditions, the majority of the flashes showed either decrease (57%) or non-monotonic variation (39%) in reflection height with respect to the immediately preceding stroke. With respect to the first stroke, 91% of the flashes showed monotonic decrease in height. For nighttime flashes, patterns in reflection height changes with respect to the immediately preceding stroke were as follows: 46% no change, 27% monotonic decrease, and 27% non-monotonic variation. When changes were measured with respect to the first stroke, 54% of nighttime flashes showed monotonic decrease and 46% no change. Ionospheric reflection height tends to increase with return-stroke peak current. The observed daytime effects can be explained by (a) the dependence of EMP penetration depth on source intensity, which decreases with stroke order, (b) additional ionization associated with elves, or (c) combination of (a) and (b) above.

1. Introduction and literature review

The ionosphere is a weak plasma (less than 1% of the neutral atoms are ionized) that has a complex structure composed of three major regions or layers, the D, E, and F regions. The lowest layer, the D region, extends in height from about 40 to 90 km. Its typical electron density is of the order of $10^9\,\mathrm{m}^{-3}$ in the daytime and diminishes to a much lower value after sunset. The E region of the ionosphere extends between about 90 and 160 km. The electron density in this region typically has a value above $10^{11}\,\mathrm{m}^{-3}$ in the daytime. At night, the electron density in the E region is about two orders of magnitude lower. Above the E region is the F region which extends to a height of 1000 km or so. The peak F-region electron density has an average value of about $2\times10^{12}\,\mathrm{m}^{-3}$ during the day and $2\times10^{11}\,\mathrm{m}^{-3}$ at night. The ionospheric plasma appears to be opaque to electromagnetic waves of frequencies below the so-called plasma frequency (9 MHz for electron density of $10^{12}\,\mathrm{m}^{-3}$ and 285 kHz for electron density of $10^9\,\mathrm{m}^{-3}$). An electromagnetic wave of frequency below the plasma frequency may be absorbed or reflected by the ionospheric plasma depending upon the electron collision frequency with neutral atoms. A high collision frequency results in absorption of the incident electromagnetic energy, while a low collision frequency allows the electrons to reradiate in phase producing what is in essence a reflection of the incident electromagnetic energy. The absorption of high-frequency radio waves propagating in the ionosphere takes place mostly in the D region, where the product of the electron number density and the collision frequency reaches a maximum (Pavlov, 2014). Ionospheric reflection in the D-region occurs where the real and imaginary parts of the index of refraction squared are equal to each other. The theory of interaction of electromagnetic waves with ionospheric plasma is found in works of Ratcliffe (1959), Stix (1962), Spitzer (1962), Ginzburg (1970), Yeh and Liu (1972), Budden (1988), and Rakov and Uman (2003, Ch. 13).

Tropospheric thunderstorms have been reported to disturb the lower ionosphere, at altitudes of 65–90 km, by convective atmospheric gravity waves and by electric field changes produced by lightning discharges. Theoretical simulations suggest that, under

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nighttime conditions, lightning electric fields reduce electron density (via enhancing electron attachment to oxygen molecules) at altitudes 75–85 km and substantially increase electron density (due to ionization of nitrogen and oxygen) at 85–95 km (e.g., Taranenko et al., 1993). The rate of the density change depends on the amplitude and duration of the lightning electromagnetic pulse (EMP), and on the preceding (ambient) electron density profile (Shao et al., 2013). The opposite EMP effects above and below the 85-km level may cause a sharpening of the electron density profile near 85 km. More recent work (e.g., Marshall et al., 2010) shows that the altitude ranges of the effects often overlap and events might show only attachment or only ionization effects depending on source parameters.

Shao et al. (2013), using lightning VLF/LF signals to probe the ionospheric D region, observed that electron density in the nighttime D region (at 75–80 km) was reduced significantly above a small storm, and the extent of the reduction was closely related in time and space to the rate of lightning discharges, which seems to be in support of the EMP-enhanced electron attachment theory. No electron density increase at higher altitudes, predicted by the theory, was observed by them. Shao et al. (2014) noted that the static electric field/current effect may induce more electrons at the lower level of the nighttime D region and that the two competing processes may cause the D-region electron density distribution to become highly inhomogeneous in space and time.

In contrast to nighttime conditions, no substantial electron density changes in response to lightning electric fields are predicted during daytime, since most of the electromagnetic energy goes into the excitation of the molecular levels lower than those of ionization and dissociative attachment (Taranenko et al., 1993). Haldoupis et al. (2013) argues that even a powerful EMP is very unlikely to generate ionization changes near the daytime lower VLF reflection heights (expected to be around 70 km), mainly because the electron mean free path there is too small in order for the electrons to gain sufficient energy to trigger impact ionization electron production. They, however, do not exclude entirely the possibility of momentary electron density depletions due to attachment at heights as low as 70 km by extremely powerful EMPs during daytime. Further, even though lightning electromagnetic fields should be heavily attenuated in the presence of elevated daytime D region electrical conductivities, a powerful EMP can reach the uppermost D region and produce ionization by impact there.

Ionospheric ionization by lightning occurs both by changes to the electron density, via ionization and attachment, and by changes to the collision frequency, via heating (i.e., changes to the electron mobility). Marshall (2014) showed that the dominant effect of lightning EMP in the upper atmosphere over short (of the order of milliseconds) time scales is collisional heating. It is likely to cause an increase of reflection height, which appears to be not supported by our observations.

According to Shao et al. (2013), ionosphere reflection of a VLF/LF signal cannot be considered a secular reflection off a definite height or layer in the lower ionosphere. Inside the ionosphere, an upward-propagating signal is gradually refracted and eventually bent backwards towards the Earth, acting as an apparent reflection. A higher-frequency signal is refracted more slowly than a lower-frequency signal and penetrates higher into the ionosphere. Therefore, the reflection is highly dispersive. For this reason, all the ionosphere reflection heights should be viewed as apparent or effective.

2. Objectives and structure of the paper

Haddad et al. (2012) found that the mean ionosphere reflection height for negative subsequent strokes was significantly lower

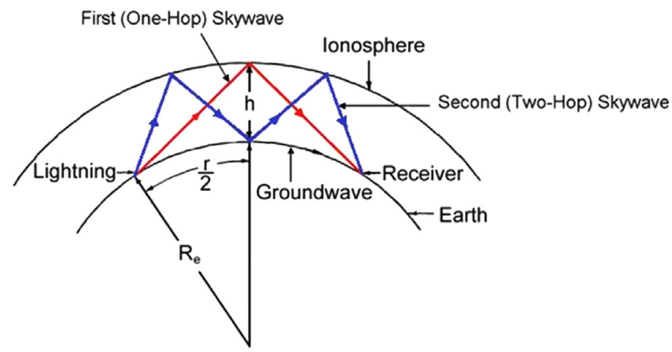

Fig. 1. Schematic representation of electromagnetic signal propagation in the Earth-ionosphere waveguide. Adapted from McDonald et al. (1979).

than for first ones. They computed the ionosphere reflection height, h_1, for the first skywave as (e.g., Laby et al., 1940):

$$h_1 = R_e\left[\cos\left(\frac{r}{2R_e}\right) - 1\right] + \sqrt{\left\{R_e^2\left[\cos^2\left(\frac{r}{2R_e}\right) - 1\right] + \left(\frac{ct_1 + r}{2}\right)^2\right\}} \tag{1}$$

where $R_e = 6367$ km is the mean radius of the Earth, r is the distance to the lightning channel (labeled D in the figures presenting data), t_1 is the difference in arrival times of the first skywave and the ground wave, and c is the speed of light in free space. The corresponding geometry is illustrated in Fig. 1. We will use Eq. (1) in this paper as well.

All field records examined by Haddad et al. (2012), were acquired under daytime conditions (between 12 noon and 8 p.m., local time) in May and June in Florida. Here, in Section 4.3, we will consider additional data acquired, in the same time period and in the same location, under nighttime conditions (between 8 p.m. and 6 a.m., local time) to see if Haddad et al. (2012)'s findings will hold. Examples of daytime and nighttime wideband electric field waveforms, recorded at the Lightning Observatory in Gainesville (LOG), Florida (Rakov et al., 2014), are shown in Fig. 2a and b, respectively. In order to make this paper consistent with previous pertinent publications, we use the atmospheric electricity sign convention (a downward-directed electric field change vector is considered as positive).

VLF electric field waveforms computed using the finite difference time domain (FDTD) technique for daytime ionosphere ($h' = 73$ km, $\beta = 0.40$ km^{-1}, where h' is often referred to as the effective reflection height and β is the steepness of the exponential electron density profile) are shown in Fig. 3. Note that the ground-wave signatures in Fig. 3 are bipolar, as expected (e.g., Rakov and Uman, 2003) at distances (>50 km) considered in this paper and that the second positive half-cycle occurs closer to the ground wave as distance increases, as expected for a skywave. Note also in Fig. 3 that at distances ranging from 100 to 400 km the skywaves have the same polarity as the ground wave. This is consistent with experimental data on nighttime ionopsheric reflections obtained within about 200 km (e.g., Smith et al. 1999, plate 3; Schonland et al. 1940; Jacobson et al. 2012, Figs. 10 and 11). Also given in Fig. 3 are the corresponding differences in arrival times of the first skywave and the ground wave, t_1 (measured using the peak-to-peak and zero-to-zero methods; discussed later in this section), and the ionosphere reflection heights, h_1, computed using Eq. (1). The time of flight in Fig. 3 is the expected t_1 which corresponds to a curved earth propagation model and expected reflection height of 73 km. Fig. 3 is our primary basis for both identification of skywaves and measurements of t_1 for daytime conditions. For