Variation in the D-region ionosphere after the 2015 Nepal earthquake using LF transmitter signals

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Abstract. We report variation in the D-region ionosphere after the 2015 Nepal earthquake using low-frequency (LF) transmitter signals. The Nepal earthquake (Mw 7.8) occurred at 6:11:26 UT on April 25, 2015. In this study, we used the BPC (China, 68.5 kHz)-Takine (TKN, Japan) LF radio wave propagation path, which has a great circle distance between the epicenter and the LF propagation path of 3,025 km. The observed periods of variation in the LF amplitude and phase were 100-300 s. The observed variation in the LF amplitude and phase was approximately +/-0.1 dB and +/-1°, respectively. The vertical velocity of the ground oscillation near the midpoint of the LF propagation path had a similar period to the LF waves. The Rayleigh wave spread concentrically from the epicenter to the LF path, and then the acoustic waves propagated vertically at the midpoint of the LF path from the Earth’s surface to the D-region ionosphere height. The maximum coherences between the LF amplitude and vertical seismic velocity, and between the LF phase and vertical seismic velocity were 0.90 (period: 146 s) and 0.77 (256 s), respectively, which were both significant at the 95% confidence level. The variation in the LF amplitude and phase was caused by acoustic waves excited by the Rayleigh wave.

Keywords: 2015 Nepal earthquake, acoustic waves, D-region ionosphere, LF transmitter signals

1. Introduction

After the occurrence of an earthquake, the electron density in the ionosphere varies due to acoustic and atmospheric gravity waves generated from ground motion or tsunamis (e.g., Astafyeva et al., 2013; Rozhnoi et al., 2014). Many studies of E- and F-region ionospheric disturbances associated with the 2011 earthquake off the Pacific coast of Tohoku, Japan, have been reported based on Global Positioning System (GPS) total electron content observations, ionosondes, high-frequency (HF) Doppler observations, Super Dual Auroral Radar Network (SuperDARN) Hokkaido HF radar data, and simulations (e.g., Liu et al., 2016; Matsumura et al., 2011; Ogawa et al., 2012; Saito et al.,
In the D-region response, the amplitude and phase of low-frequency (LF) transmitter signals oscillated over a period of ~100 s after the mainshock of the 2011 Tohoku earthquake (Ohya et al., 2018). Based on a one-dimensional simulation of the neutral atmosphere and the wave-hop method, the oscillations were found to be caused by acoustic waves excited by Rayleigh waves. However, the dependence of D-region variation on the magnitude of earthquakes, region of influence from the epicenter, and tsunami effects on the D-region ionosphere remains unsolved.

Neutral atmospheric waves such as acoustic and atmospheric gravity waves are important keys for understanding lithosphere-atmosphere-ionosphere coupling. Although it is difficult to observe motion of neutral particles at ionospheric height. Variation in electron density in the D-region ionosphere can be estimated sensitively using LF observations. In this study, we focused on the variation in the amplitude and phase of LF transmitter signals after the 2015 Nepal earthquake. Our results can contribute on revealing region of influence from the epicenter and the dependence of D-region variation on the magnitude of earthquakes.

2. Observations

The Nepal earthquake (Mw 7.8) occurred at 06:11:26 UT on April 25, 2015. Figure 1 shows location of the epicenter (geographic coordinates: 28.24°N, 84.75°E) of the earthquake (red star), a transmitter (light blue circle) of the BPC (China, 68.5 kHz, 34.63°N, 115.83°E), an LF receiver (purple circle) at Takine (TKN, Japan, 37.34°N, 140.67°E), a seismometer station (yellow triangle) of the Incorporated Research Institutions for Seismology (IRIS) at Incheon (INCN, Korea, 37.48°N, 126.62°E), and the midpoint of the BPC-TKN path (green triangle). The distance of the BPC-TKN path was 2,253 km; the INCN was close to the midpoint of this path. We observed the amplitude and phase of the LF transmitter signals using an active loop antenna (Wellbrook LFL1010) at TKN. The signal was amplified by a main amplifier and then digitized using a 16-bit A/D converter with a 200-kHz sampling frequency, which was synchronized with a GPS locked oven-controlled crystal oscillator. This oscillator has sufficient frequency stability to detect small phase variation in the received signals caused by changes in the radio reflection height in the D-region ionosphere. The amplitudes and phases of the 68.5-kHz component raw data were extracted and recorded on hard disks with a time resolution of 0.1 s.

3. Results and Discussion

Figure 2 shows (a) variation in BPC-TKN amplitude, (b) the wavelet spectrum of the BPC-TKN amplitude, (c) variation in the BPC-TKN phase, and (d) the wavelet spectrum of the BPC-TKN phase at 06:24–07:00 UT on April 25, 2015. The black and blue lines indicate the earliest arrival time of acoustic waves along the shortest path (over the BPC transmitter), and the arrival time at the midpoint of the LF propagation path, respectively (Figure 3). The white curves in Figures 2b and 2d indicate the 95% confidence level. When a mean background spectrum is assumed to be red noise, the
A distribution for the local wavelet power spectrum, \( \frac{|W_n(s)|^2}{\sigma^2} \), should be chi-square distributed with two degrees of freedom for the time series that the variance is \( \sigma^2 \), where \( |W_n(s)|^2 \) is the wavelet power spectrum, \( n \) is the localized time index, and \( s \) is the wavelet scale (Torrence and Compo, 1998). The confidence level is calculated as \( \frac{1}{2} P_k \chi^2_2 \), where \( P_k \) is the mean background spectrum at the Fourier frequency \( k \) smoothed using the wavelet function, \( \chi^2_2 \) is the chi-square distributed with two degrees of freedom. Here the confidence level of 95% for the \( \chi^2 \) is set. Peaks greater than 95% confidence level means that they are significant result at 5% level. The significance test successfully distinguished peaks greater than 95% confidence level from other random frequencies in the wavelet spectra. As shown in Figures 2b and 2d, both the LF amplitude and phase had a period of 100-300 s at 06:33 UT. From 06:33 to 06:48 UT, the LF amplitude and phase varied by +/- 0.1 dB and +/- 1°, respectively. The standard deviations of amplitude and phase during 06:00-07:00 UT on that day were +/- 0.17 dB and +/- 1.8°, respectively. The amplitude of the variation was similar to the background level and many isolated enhanced regions were observed in the wavelet spectrogram. However, fluctuations of 100-300 s in both amplitude and phase occurred simultaneously at arrival time of acoustic waves at the midpoint of the BPC-TKN path. Moreover, the periods of the variation were similar for both amplitude and phase, and both was above the 95% confidence level. Therefore, we concluded that this variation was caused by acoustic waves excited by Rayleigh waves. Figure 3 shows a schematic diagram of the acoustic wave propagation. Rayleigh waves are seismic surface waves: a point on the surface moves in an elliptical motion as Rayleigh wave propagates. Both vertical and horizontal components of motion of the point on the
surface are produced in the direction of wave propagation. The acoustic waves are longitudinal waves that result from an oscillation of pressure such ground motion. When an earthquake occurs, Rayleigh waves excited by the earthquake spread concentrically along the Earth's surface from the epicenter, and acoustic waves are then generated by the Rayleigh waves. The acoustic waves propagate vertically to the D-region height. In this study, we considered two cases: 1) the earliest arrival time when acoustic waves propagated vertically over the BPC transmitter, and 2) the arrival time when acoustic waves propagated over the midpoint of the LF propagation path. The horizontal distances from the epicenter to the BPC transmitter and from the epicenter to the midpoint of the LF path were 3,025 and 4,159 km, respectively (Figure 1). The propagation velocity of the Rayleigh waves was estimated to be 3.8 km/s based on the IRIS seismic data. The propagation times of the Rayleigh waves in cases (1) and (2) were calculated as 787 and 1,057 s, respectively.

The acoustic velocity, $C_s$, was determined as follows (Chum et al., 2012):

$$C_s = \sqrt{\frac{\gamma RT_k}{M}}$$

where $\gamma (7/5)$ is the heat capacity ratio, $R$ (8.314 JK$^{-1}$ mol$^{-1}$) is the molar gas constant, $T_k$ is the atmospheric temperature (K), and $M (2.897 \times 10^{-2}$ kg/mol) is the molar mass of the atmosphere. We estimated the acoustic velocity from the ground to 70 km height (typical daytime reflection height of LF waves), using height profiles (0.1 km steps) atmospheric temperature obtained from the Naval Research Laboratory - Mass Spectrometer Incoherent Scatter (NRLMSISE-00) atmospheric model (Picone et al., 2002). We input parameters of UT, latitude, and longitude when and where acoustic waves were excited into the NRLMSISE-00 model. The vertical propagation time of acoustic waves from the
ground to 70 km height was 225 s.

Figure 4 shows (upper) the vertical velocity of the IRIS seismic waves at INCN and (lower) the wavelet spectrum of the vertical velocity during 06:10–07:10 UT on April 25, 2015. The frequency response was corrected. The seismic waves had a period of 100–200 s around 06:30 UT, which was similar to that of the LF waves.

Figure 5 shows (upper) the coherence between the LF amplitude and vertical seismic velocity at INCN, and (lower) the coherence between the LF phase and INCN seismic velocity. The red line indicates the 95% confidence level. Coherence, coh(ω), was defined as follows:

$$\text{coh}^2(\omega) = \frac{|S_{xy}(\omega)|^2}{S_{xx}(\omega)S_{yy}(\omega)} \quad \text{...............(2)}$$

where ω is the angular frequency; $S_{xy}(\omega)$ is the cross-spectra of two discrete time signals; $S_{xx}(\omega)$ and $S_{yy}(\omega)$ are the spectra of $x(t)$ and $y(t)$, respectively (Hino, 1977); and $x(t)$ and $y(t)$ are the variation in LF amplitude or phase and vertical seismic velocity at INCN, respectively. $|S_{xy}(\omega)|$ was calculated as follows:

$$|S_{xy}(\omega)| = \sqrt{K_{xy}^2(\omega) + Q_{xy}^2(\omega)} \quad \text{............................... (3)}$$
where $K_x(\omega)$ and $Q_x(\omega)$ are the co- and quad-spectra, respectively. The start and end times of the LF data used in this analysis were 06:28:18 UT and 07:02:26 UT, respectively. Those of the seismograph were 06:18:26 UT and 06:52:34 UT, respectively. The number of points in both LF and seismic data was 2048 (2048 s). These time windows were selected for occurrence time of each main variation. The highest coherence values between the LF amplitude and seismic data at INCN were 0.90 (at 146 s), and 0.67 (113 s). The highest coherence values between the LF phase and seismic data at INCN were 0.75 (204 s), 0.72 (128 s), 0.74 (146 s), and 0.77 (256 s). These coherence values were all significant at the 95% confidence level. These results show that there was common variation over a period of 100-200 s in both the LF wave variation and the ground motion. If the acoustic waves were generated from the ground motion, the acoustic waves would have the same period of 100-200 s. Based on a one-dimensional simulation of the neutral atmosphere, Ohya et al. (2018) reported that the same period of 100-200 s was seen in both the vertical velocity of the neutral atmosphere and the variation in D-region electron density after the mainshock of the 2011 Tohoku earthquake. In the case of the 2015 Nepal earthquake, the acoustic waves also would also become a mediator of the Earth's surface and the D-region ionosphere.

We used the wave-hop method to estimate the variation in reflection height due to the 2015 Nepal earthquake. The synthetic electric field strength $E$ was calculated as follows (Rec. ITU-R P.684-7, 2016):

$$ E = E_g + \sum_{K=1}^{10} E_{S_k} $$

(4)

where $K$ is the number of hops. The ground wave $E_g$ was calculated as follows:

$$ E_g = |E_g| \exp(jkG_r) $$

(5)

where $k$ is the wave number of the LF signals, $G_r$ is the great circle distance between the transmitter and receiver, and $|E_g|$ was calculated based on Rec. ITU-R P.368-7. (1992). The effective field strength of transmitted sky waves $E_{S_k}$ (mV/m) was calculated for a loop antenna as follows:

$$ E_{S_k} = \frac{600\sqrt{P_t \cos \psi}}{\sum_{L=1}^{K} (R_{C_{K,L}}) \prod_{L=1}^{K-1} R_{G_{K,L}} F_c R_t F_r \exp(-jk\sum_{L=1}^{K} P_{l_{K,L}})} $$

(6)

where $P_t$ is the radiated power (kW), $\psi$ is the angle of departure and arrival of the sky wave at the ground, $L$ is the number of reflection points from the transmitter, $R_c$ is the ionospheric reflection coefficient, $R_t$ is the complex reflection coefficient of the Earth's surface, $F_c$ is the ionospheric focusing factor, $F_t$ is the transmitting antenna factor, $F_r$ is the receiving antenna factor, $P_l$ is the path length of each sky wave, and $R_{C_{K,L}}$ is the ionospheric reflection coefficient of the K-hop sky wave at the Lth reflection point. In this study, $R_{C_{K,L}}$ was given by empirical functions of solar activity, solar zenith angle, ionospheric incidence angle at the reflection point, and frequency. $P_{l_{K,L}}$ was obtained geometrically with a one-hop ground range and ionospheric reflection height $R_{h_{K,L}}$. The magnetic field component was observed with a loop antenna, and the theoretical magnetic field $B$ in the atmosphere was calculated using the $E$ from Eqn. (4) as follows,
$B = \frac{E}{c}$  

where $c$ is the light velocity. We compared the observed and theoretical magnetic components using Eqn. (7). We used the electron density-height profiles of the International Reference Ionosphere (IRI) 2016 model. For more details, see Rec. ITU-R P.684-7. (2016) and Ohya et al. (2018).

Figure 6 shows (left) variation in the BPC-TKN amplitude and (right) the phase versus reflection height at 06:33 UT on April 25, 2015, based on the wave-hop method. The reference height was assumed to be 70 km.

4. Conclusions

We found variation of 100-300 s in the amplitude and phase of LF transmitter signals along the BPC-TKN path after the mainshock of the 2015 Nepal earthquake. This is the first report of coseismic D-region signatures based on long propagation observations of LF transmitter signals. Our conclusions can be summarized as follows:

1. Variation over a period of 100-300 s in the BPC-TKN amplitude and phase was observed at the arrival time of acoustic waves, which were generated by the Rayleigh waves on the Earth’s surface. The Rayleigh waves propagated in Earth’s surface for a long distance, and acoustic waves were continuously excited by the Rayleigh waves. For the cause of the LF variation, the acoustic waves were excited at the midpoint of the BPC-TKN path and propagated vertically to the D-region height (70 km).
2. coherence values between the LF amplitude or phase and vertical seismic velocity at INCN (near the midpoint of the BPC-TKN path) were significant (max: 0.90) over a period of 100-200 s.
3. The amplitude and phase of the LF waves varied by ±0.1 dB and ±1°, respectively, which corresponded to variation in the reflection height of ±40 m using the wave-hop method.
4. The 100-200 s period variation was originated from the propagation of Rayleigh waves that excited acoustic waves. The acoustic waves excited vertically upward at the midpoint of the BPC-TKN path reached the D-region height. The variation of the LF transmitter signal was caused by the variation of the ionosphere in the D-region due to the acoustic waves.

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