8. Ionospheric seismology and volcanology

1,2 Kosuke Heki

1Shanghai Astronomical Observatory, 80 Nandan Rd., Shanghai 200030, China
2Dept. Earth Planet. Sci., Hokkaido Univ., N10 W8, Kita-ku, Sapporo 060-0810, Japan

Abstract

Large earthquakes and volcanic eruptions disturb ionosphere in various ways, and we can observe them with global navigation satellite system (GNSS) receivers as changes in total electron content (TEC). In the first half, I discuss ionospheric disturbances related to earthquakes. They appear ~10 minutes after earthquakes as sudden TEC changes propagating with three different speeds. The disturbances caused by direct acoustic waves from the epicenters are useful to estimate moment magnitudes (Mw) of earthquakes. However, we should take care of the influences of geomagnetic fields and line-of-sight geometries on the observed amplitudes. In the second half, I discuss ionospheric TEC changes by large volcanic eruptions. They emerge in two distinct forms, either long-lasting harmonic oscillation or short impulsive changes. They are often associated with Plinian continuous eruptions and Vulcanian explosive eruptions, respectively. In both types, amplitudes of the disturbances normalized by background vertical TEC provides useful measures of the eruption intensities.

Keywords: earthquake, volcanic eruption, ionospheric disturbance, total electron content, GNSS

8-1. Introduction and observation history
Fault dislocation of a large earthquake causes vertical movement of the earth’s surface and excites atmospheric waves. Such waves are also excited by strong volcanic eruptions. They propagate upward and disturb the ionosphere. In typical cases, the disturbance starts shortly after earthquake/eruption when acoustic and internal gravity waves reach the ionospheric F region. This usually ends as a temporary phenomenon but could be followed by long-lasting disturbances.

Coseismic ionospheric disturbances were found first by Doppler sounding as the vertical oscillation of the ionosphere following the 1968 May Tokachi-oki earthquake (M7.9) (Yuen et al., 1969) and the 1982 March Urakawa-oki (M7.1) earthquake (Tanaka et al., 1984), both in northern Japan. Disturbances related to volcanic eruptions were also found using Doppler sounding shortly after the 1981 eruption of the St Helens volcano, North America (Ogawa et al., 1982). The 1991 eruption of the Pinatubo volcano, the Philippines, was one of the largest eruptions in the 20th century, and Cheng and Huang (1992) and Igarashi et al. (1994) detected traveling ionospheric disturbances in Taiwan and Japan, respectively. It was difficult then to understand the phenomena with such observations of limited temporal and spatial coverage.

Chances of detecting such signals have significantly increased over the last 30 years by the deployment and densification of continuous receiving stations of global navigation satellite systems (GNSS), such as the Global Positioning System (GPS). Because the primary purpose of GNSS networks is crustal deformation monitoring, they are densely deployed near tectonic plate boundaries. With the phase difference in microwave carriers in different frequencies from GNSS satellites, we can measure changes in ionospheric total electron content (TEC), number of electrons integrated along the line-of-sight connecting the receiver and the satellite. In this chapter, I review current studies of ionospheric seismology and volcanology based on the “GNSS-TEC” technique.

Coseismic ionospheric disturbance was detected first with GNSS for the 1994 Northridge earthquake (Mw6.7), California (Calais and Minster, 1995). A comprehensive study with a dense GNSS array for the 2003 Tokachi-oki earthquake (Mw8.0), Japan, revealed various properties of the near-field disturbances, e.g., propagating velocities and directivities (Heki and Ping, 2005). Although ionospheric
TEC changes caused by an artificial mine blast was already reported by Calais et al. (1998), those by a real volcanic eruption were first studied using a dense GNSS array by Heki (2006) for the 2004 September eruption of the Asama volcano, Central Japan (VEI 2).

Since then, numbers of new cases of such disturbances have been reported, as reviewed in Tanimoto et al. (2015), Jin et al. (2019), Astafyeva (2019), and Heki (2021). Here I present general methodology of the GNSS-TEC technique in Section 8-2. Then I discuss essential contribution of GNSS-TEC observations to seismology (Section 8-3), and volcanology (Section 8-4). There I mainly explore how we can infer earthquake magnitudes and volcanic eruption intensities using ionospheric disturbances. Here I do not discuss possible ionospheric changes before earthquakes, which is reviewed in Heki (2021).

8-2. GNSS-TEC Observations

Phase difference and TEC

Satellites of GPS, the oldest GNSS, transmit microwave signals in the two frequencies 1.57542 GHz (L1) and 1.22760 GHz (L2) from altitude of ~20,200 km. The signals propagate through ionosphere, dispersive media with frequency-dependent delays. The ionosphere ranges in height from ~60 km to beyond 1,000 km, and the largest ionization occurs in the F region, ~300 km in altitude. By tracking the differences between the L1 and L2 phases $\Delta(L1−L2)$, we can monitor the temporal changes of TEC along line-of-sight (slant TEC or STEC). The TEC changes are related to the phase difference changes as follows (in TEC unit, i.e., $10^{16}\text{el/m}^2$).

$$\Delta\text{STEC} = \frac{1}{40.308} f_1 f_2^2 (f_1^2−f_2^2) \Delta(L1−L2). \tag{1}$$

There, $f_1$ and $f_2$ show the frequencies of the two microwave carriers from GNSS, and the phase difference $\Delta(L1−L2)$ should be expressed in meters.

STEC shows U-shaped changes coming from the apparent movement of the satellites in the sky (and consequent changes of the incidence angles of line-of-sights with the ionosphere). Recent GNSS include geostationary satellites. In that case, STEC...
does not show such U-shaped changes.

There are increasing number of ground stations with receivers capable of receiving multi-GNSS. Different systems transmit microwave signals in slightly different frequencies, which should be considered in converting the phase differences into TEC. Some GNSS also employ L5 signals (1.17645 GHz in GPS) with a frequency a little lower than L2. GNSS raw data files are available as RINEX (Receiver independent exchange format) files and can be downloaded from data centers run by, e.g., International GNSS Service (IGS), and University NAVSTAR Consortium (UNAVCO). Typical sampling interval at ground GNSS stations is 30 seconds. High-rate sampling (e.g., 1 second) is useful for ionospheric seismology/volcanology (e.g., Astafyeva et al., 2013), but is not indispensable. Examples shown in this chapter are all derived from regular 30-second sampling stations.

From STEC to VTEC

STEC time series of non-geostationary GNSS satellites show U-shaped long-term (hours) changes coming from the satellite motion, in addition to real spatial (e.g., latitudinal difference of ionization) and temporal (e.g., diurnal changes) variations. They also include station- and satellite-dependent biases. Conversion from biased STEC to absolute vertical TEC (VTEC) simplifies TEC behaviors and often makes interpretations easier. The conversion can be done in three steps.

We first remove phase ambiguities by aligning STEC derived from carrier phases with those derived by pseudo-ranges, which have larger noises but no phase ambiguities. Then we correct for satellite and receiver differential code biases (DCBs) available online from the header information of Global Ionospheric Map (GIM) files (Mannucci et al., 1998). If the receiver bias of a station of interest is not available, we can infer its value by minimizing the scatter of night-time VTEC at that station, the approach known as “minimum scalloping” (Rideout and Coster, 2006). At last, such bias-free STEC can be converted to absolute VTEC by multiplying with cosine of the incidence angle of line-of-sight with ionospheric F region.

The coordinates of the ionospheric piercing points (IPP) of line-of-sight are calculated assuming a thin ionosphere normally at altitude of 300 km, and the
trajectories of their ground projections (sub-ionospheric points, SIP) are plotted on the map to indicate the horizontal position of the observed ionosphere.

**Finding signals related to earthquakes and volcanic eruptions**

Ionospheric disturbances by earthquakes and volcanic eruptions are overprinted to other changes, which include regular diurnal changes, spatial changes due to movement of IPP (i.e., crossing of equatorial ionization anomalies), short-term (a few minutes) positive pulses caused by sporadic-E irregularities, sudden drops due to plasma bubbles, large- and medium scale traveling ionospheric disturbances (LSTID/MSTID), and so on.

Long-term changes can be removed by applying a high-pass filter, easily done by modeling the VTEC (or STEC) changes with polynomials of time (appropriate degree depends on the time window) and by extracting the departures from these reference curves. Wavelet transformation is also effective in extracting components of certain periods.

Disturbances related to earthquakes and volcanic explosions often appear ~10 minutes after the events and propagate in several different velocities. The keys to attribute certain TEC changes to earthquakes or eruptions would be (1) if they emerge at right times, and (2) if they propagate outward from the epicenters/volcanoes with prescribed velocities. Then, one would not wrongly interpret irrelevant ionospheric disturbances as those by earthquakes and volcanic eruptions. In other words, it is difficult to interpret the observation when we have just one station-satellite pair recording the TEC change.

**Multi-GNSS**

Most of the GNSS-TEC studies have been done using GPS, the first GNSS by America, and GLONASS, the second GNSS by Russia. Now, two other GNSS are operated, namely European Galileo, and Chinese BDS (Beidou Navigation Satellite System). In addition to them, regional satellite navigation systems are in operation in Japan as QZSS (Quasi-Zenith Satellite System) and in India as NAVIC (Navigation Indian Constellation). Stations equipped with multi-GNSS receivers have been
increasing.

BDS, QZSS, and NAVIC partly employ geostationary satellites. By using them, we can keep observing the same satellites continuously and obtain TEC time series without disruptions. This offers a unique opportunity in ionospheric seismology and volcanology. In a later section, I present an example of ionospheric volcanology taking advantage of the QZSS geostationary satellite.

Another benefit of modern GNSS is the addition of the new carrier frequency L5, to conventional L1 and L2. Its frequency is somewhat lower than L2, and the combination of L1-L5 offers a larger frequency difference than the L1-L2 pair. Now we have three different combinations of carrier phases, L1-L5, L2-L5, in addition to L1-L2. These three have different factor $f_1^2f_2^2/(f_1^2−f_2^2)$ in equation (1). Heki and Fujimoto (2022) compared TEC time series from these three pairs and separated real TEC changes from noises in GNSS receivers.

8-3. Ionospheric seismology

Three different atmospheric waves

An earthquake causes coseismic vertical crustal movement, which excites atmospheric waves (Figure 8-1). Such waves propagate upward and disturb the ionosphere. The disturbance usually starts ~10 minutes after an earthquake, when the acoustic wave reaches the ionospheric F region. Disturbances caused by direct acoustic waves from epicenters propagate in the sound velocity of the F region height (~0.8 km/s).

Vertical surface motions associated with the passage of the Rayleigh surface waves also excite acoustic waves (Figure 8-1). They make disturbances propagating much faster (~4 km/s) than direct acoustic waves. They decay less with distance and propagate worldwide. Ionospheric disturbances caused by the 2002 Denali earthquake (Mw7.9), Alaska, were detected in GNSS stations California (Ducic et al., 2003), and those by the 2004 Sumatra-Andaman earthquake (Mw9.2) have been detected in Japan (Heki, 2021). These acoustic wave origin disturbances have typical periods of ~4 minutes.
Large earthquakes and tsunamis also excite internal gravity waves with longer periods (Figure 8-1). Clear concentric wave fronts of internal gravity waves were observed after the 2011 Tohoku-oki earthquake (Mw9.0) (Tsugawa et al., 2011). Considering that they emerged above the Japan Sea, the waves would have been excited at the epicenter rather than by propagating tsunamis. Disturbances caused by internal gravity waves have typical periods of ~20 minutes and propagate with a speed 0.2-0.3 km/s.

Figure 8-1. Earthquakes excite three kinds of atmospheric waves that disturb ionosphere, i.e., (1) direct acoustic wave from the epicenter, (2) internal gravity wave propagating obliquely upward from the focal area and from propagating tsunami, and (3) acoustic wave excited by propagating Rayleigh surface wave. They are observed with GNSS as TEC changes by receiving microwave signals from satellites in different carrier frequencies. Inset shows the frequency dependent atmospheric filtering effect of acoustic waves after Blanc (1983).

Discriminating the three different waves

Figure 8-2 shows an example of the coseismic ionospheric disturbance in the 1994 Hokkaido-toho-oki earthquake (Mw8.3) (Astafyeva et al., 2009), where signatures of the three atmospheric wave (Figure 8-1) are visible. Figure 8-2a shows slant TEC time series of GPS satellite 6 (called as G06 here), whose SIP moves northward between stations and the epicenter (Figure 8-2b). TEC changes are characterized by the first strong positive
anomalies ~10 minutes after earthquake, followed by lower amplitude components with slower velocities. We can also recognize faint harmonic oscillation with a period ~4 minutes lasting for a half hour or so within ~500 km from the epicenter.

Figure 8-2. Coseismic ionospheric disturbance associated with the 1994 October Hokkaido-toho-oki earthquake (Mw 8.3) that occurred off the east coast of Hokkaido, Japan. (a) Slant TEC time series from 5 GNSS stations with satellite G06 are drawn as the residual from degree 4 best-fit polynomials. (b) SIP trajectories (ionospheric height assumed as 300 km) of G06 during the period shown in (a). Blue dots indicate SIPs at the occurrence time of the earthquake (~13.38UT). STEC were converted by wavelet transformation highlighting components with period 4 minutes (c) and 12 minutes (d) and drawn as the function of time and focal distance. We could see propagations of three different atmospheric waves, Rayleigh surface wave (RW), acoustic wave (AW), and internal gravity wave (IGW) as signatures with different slopes. IGW signals are visible only in (d).

Here, I perform wavelet transformation using a Mexican hut wavelet to extract certain period components (Heki and Ping, 2005), and show them as a function of time and focal distance in Figure 8-2c, d. By using a short period wavelet, we can get sharp signatures
of Rayleigh waves and harmonic oscillations propagating with a similar speed (Figure 8-2c). However, direct acoustic wave and internal gravity wave signatures become clearer by using a longer-period wavelet (Figure 8-2d). There, we can see components propagating by ~4, ~0.8, and ~0.25 km/s, corresponding to three atmospheric waves illustrated in Figure 8-1. Internal gravity wave signatures are difficult to recognize in the slant TEC time series (Figure 8-2a) but are clear after wavelet transformation (Figure 8-2d).

For earthquakes without dense GNSS networks, identification of all the three waves would be difficult. Nevertheless, the first pulses occurring ~10 minutes after earthquakes are easy to find for most large earthquakes with a few nearby GNSS stations. Although typical periods of disturbances due to acoustic waves are ~4 minutes, the periods may appear longer for very large earthquakes, e.g., 2004 Sumatra-Andaman earthquake, because the observed signals are made from acoustic waves excited by multiple segments of the faults with small time lags (Heki et al., 2006). Recently, Bagiya et al. (2023) demonstrated that interferences of such multiple acoustic sub-pulses may result in azimuthal dependence of waveforms and amplitudes of the disturbance signals.

Direct acoustic waves from epicenter

Here, I focus on the TEC disturbances made by direct acoustic waves from the epicenters due to coseismic vertical crustal movements (its signature cannot be separated from Rayleigh wave signatures in near fields, see Figure 8-2). Indeed, these disturbances have a potential of providing key information on Mw of earthquakes with their amplitudes relative to background VTEC (Cahyadi and Heki, 2015; Heki, 2021).

In Figure 8-3, I simulate the propagation of an acoustic wave with a period of 4 minutes. It propagates upward and is refracted gradually so that it partly propagates horizontally at ionospheric altitudes (ray tracing is illustrated in Figure 8-3c). Here I assume an N-shaped acoustic pulse using a simple function

\[ f(t) = -at \exp\left(\frac{-t^2}{2\sigma^2}\right). \]  

This function has a maximum and minimum at \( t = -\sigma \) and \( t = \sigma \), respectively, and I assume 60 seconds for \( \sigma \) (4 minutes in period). The parameter \( a \) represents the amplitude. This period is close to the acoustic cut-off (Figure 8-1 inset), and simple ray tracing involves errors by neglecting gravitational restoring forces. Nevertheless, rough discussions on the line-of-sight geometry dependence of the disturbance amplitudes would be valid.
TEC disturbance amplitudes are sensitive to two factors, geomagnetic fields, and line-of-sight geometry (Rolland et al., 2013). Even from stations at the same distance from the epicenter, differences in these factors may result in amplitude contrasts exceeding an order of magnitude. In the F region height, electrons can move only along geomagnetic fields, and this results in the suppression of electron density anomalies in regions where atmospheric particle motion is perpendicular to the field. In the northern hemisphere, this occurs to the north of epicenters (Figure 8-3c).

Figure 8-3. Upward propagation of an N-shaped acoustic pulse (with a period 4 minutes) in the atmosphere. The sound velocity structure of US Standard Atmosphere 1976 is assumed and the electron density obeys the Chapman distribution as a function of altitude. Electron density scale is arbitrary (actual intensity depends on $M_w$ and background TEC). Red and blue indicate positive and negative anomalies, respectively, and green indicates neutral. The geomagnetic field at Fukutoku-Okanoba, Japan, is assumed (Declination: -3.0 degree, Inclination: 34.5 degree). The figure shows three epochs, 10.0 (a1, a2, a3), 12.0 (b1, b2, b3), and 14.0 (c1, c2, c3) minutes after earthquake. We show horizontal cross sections at altitude 300 km (a1, b1, c1), east-west vertical cross sections (a2, b2, c2), and north-south vertical cross sections (a3, b3, c3). The disturbance front propagates outward with a speed ~0.8 km/s. The broad southward beam is caused by the interaction of the electron movements with the geomagnetic field. Actual amplitudes of the STEC disturbances strongly depend on the
incidence angles of the line-of-sight with the wavefront. TEC signatures at points P0 and P6 (white circles) are compared together with those from other points in Figure 8-4.

The other factor, geometry of line-of-sight and wavefront, matters when we observe electron density anomalies as TEC with line-of-sights penetrating them in various directions. Line-of-sights penetrating only positive parts of the anomalies would show positive TEC anomalies. However, those penetrating both positive and negative parts may show only small changes in TEC (see, e.g., Manta et al., 2020a). Figure 8-4 shows synthesized slant TEC time series at 7 different stations located along a circle around the epicenter with satellites in northern, eastern, and southern skies. The largest signal occurs when we observe a northern satellite at a southern GNSS station P6 (thick red curve in Figure 8-5a). If we observe a southern satellite at a northern station P0, the observed signal amplitude is reduced to ~1/4 (thick blue curve in Figure 8-5c). Arrival times of the TEC anomalies also depend on the geometry.

Knowing $M_w$ from amplitudes of disturbances

Cahyadi and Heki (2015) and Heki (2021) obtained empirical $M_w$ dependence of the coseismic ionospheric disturbance amplitudes, normalized by background VTEC. Rapid determination of $M_w$ of earthquakes using the coseismic ionospheric disturbance

Figure 8-4. Synthesized TEC time series for the case shown in Figure 8-3 from ground observing points P0-P6 along a circle with radius 270 km around the epicenter (d). (a), (b), and (c) show the time series when we observe GNSS satellites with elevations and azimuths (45, 0), (45, -90), and (45, 180), respectively. In all the cases, signals get strong when the satellites are in the direction of the epicenter. The top curve in (c) and the bottom curve in (a), drawn with thick lines, correspond to line-of-sights shown in Figure 8-3c.
amplitudes, ~10 minutes after earthquakes, may realize an effective tsunami early warning (e.g., Martire et al., 2023).

In Figure 8-5, I show normalized amplitudes of 28 earthquakes of $M_w$ from 6.6 to 9.2 compiled in Heki (2021). They show clear coseismic ionospheric disturbances by direct acoustic waves detected by GNSS-TEC. There, the satellite-station pairs showing maximum disturbance amplitudes of individual earthquakes are selected (i.e., pairs with closest geometry to the bottom curve of Figure 8-4a). They include two strike-slip earthquakes and two normal fault earthquakes, but all other examples are reverse fault earthquakes. Strike-slip earthquakes show somewhat weaker disturbances, but normal fault earthquakes do not show such a tendency (Cahyadi and Heki, 2015).

Heki (2021) considered that these amplitudes (unit: percent) obey a simple law,

$$\log_{10} \text{Amplitude} = a (M_w - 8.0) + b. \quad (3)$$

The offset $b$ is the common logarithm of the relative amplitude of a typical $M_w8$ event. The best-fit line inferred from all the earthquakes (dashed line in Figure 8-5) has the slope $a \sim 0.602$ and $b \sim 0.804$. The slope is close to 2/3, i.e., the amplitudes increase approximately by two orders of magnitude as $M_w$ increases by three.
Figure 8-5 Blue dots compare $M_w$ of the 30 earthquakes given in Heki et al. (2022), with their amplitudes of ionospheric disturbance caused by direct acoustic wave (AW) from epicenters. They are amplitudes of STEC changes and are normalized by background VTEC values. The dashed line indicates the best-fit line corresponding to Equation (3). Red triangles show amplitudes of internal gravity waves (IGW) by 4 large earthquakes (1994 Hokkaido-toho-oki, 2003 Tokachi-oki, 2010 Maule, and 2011 Tohoku-oki earthquakes). They obey a different scaling law.

Equation (3) enables us to infer $M_w$ by observing STEC oscillation relative to background VTEC, ~10 minutes after earthquakes. The data plotted in Figure 8-5 still show large scatters around the best-fit line. This is possibly because the ideal geometry (like Figure 8-4a, bottom curve) is not always realized due to insufficient GNSS station coverages. Calibration of the data considering such geometry factors would significantly improve this empirical $M_w$-amplitude relationship.

**Internal gravity wave signatures**

Figure 8-2 shows that internal gravity wave signature appeared for this $M_w$=8.3 earthquake, as a component propagating 0.2-0.3 km/s emerging ~40 minutes after earthquakes. Because we need to confirm its propagating velocity with time-distance plot...
like Figure 8-2, we can identify it only when earthquakes are sufficiently large and we have enough number of stations. This was the case for the 2003 Tokachi-oki (Mw8.0), 2010 Maule (Mw8.8), and 2011 Tohoku-oki (Mw9.0) earthquakes, in addition to the 1994 Hokkaido-toho-oki (Mw8.3) shown in Figure 8-2. I obtained their amplitudes relative to background VTEC. However, I did not correct for geomagnetic effect considering angles between the local geomagnetic field and the particle motions.

The Mw dependence of the ionospheric disturbance amplitudes of internal gravity wave is shown in Figure 8-5. The amplitudes are significantly smaller than those by direct acoustic waves. It is interesting that the slope is about twice as steep as acoustic waves. Their amplitudes might be sensitive not only to the area and amount of crustal uplift but also for the duration of faulting (Heki et al., 2022).

8-4. Ionospheric volcanology

Two types of ionospheric disturbances by volcanic eruptions

Dautermann et al. (2009a, b) reported GNSS-TEC observations of ionospheric disturbance caused by the 2003 eruption of the Soufrière Hills volcano, Montserrat, the Lesser Antilles. TEC changes due to volcanic eruptions was also observed by a dense GNSS network after the 2004 explosion of the Asama volcano, Central Japan (Heki, 2006). The TEC signatures of these two eruptions are very different. The former is characterized by ~4 mHz harmonic oscillation of TEC lasting for tens of minutes. The latter, on the other hand, appear as short pulses in TEC ~10 minutes after the eruption. Since these early reports, more than ten cases of ionospheric disturbances caused by volcanic eruptions have been published. These new cases were found to belong to either of these two types of TEC changes.

The first type is the long-lasting harmonic TEC oscillations (“Type 1” in Figure 8-6). The interference of upward and downward acoustic waves between the ground surface and the mesopause causes resonant oscillation of atmosphere in prescribed frequencies (Tahira, 1995). The frequency reflects the vertical atmospheric structure and have various overtones. Such “atmospheric modes” are excited typically by continuous Plinian-type volcanic eruptions.
Figure 8-6. Ionospheric disturbance caused by continuous (Type 1, left) and explosive (Type 2, right) volcanic eruptions can be detected by differential ionospheric delays of microwave signals of multiple carrier frequencies from GNSS satellites. Strong continuous eruptions sometimes excite atmospheric modes and long-term oscillatory disturbances in ionosphere (“Type 1” disturbance). For explosive eruptions, we often find short-term impulsive disturbances (“Type 2” disturbance) in ionosphere ~10 minutes after eruptions. The acoustic wave makes electron density anomalies (pairs of positive and negative anomalies as shown with red and blue colors in the figure) selectively on the southern side of the volcano (for northern hemisphere cases) due to interaction with geomagnetic fields (blue arrow).

Such atmospheric modes were first observed with seismometers. Acoustic resonance frequencies of 3.7 and 4.4 mHz were found in background free oscillations of the solid earth (Nishida et al., 2000). These frequency components were found to last >5 hours in seismometer records after the 1991 eruption of the Pinatubo volcano, the Philippines (Kanamori and Mori, 1992). Watada and Kanamori (2010) considered that the continuous Plinian eruption of the volcano excited atmospheric resonance, which caused harmonic ground oscillation. Such oscillations were found by GNSS-TEC during the 2003 eruption of the Soufrière Hills (Dautermann et al., 2009a, b), and new cases have been added to the literature since then.

The second type of disturbances (“Type-2” in Figure 8-6) occur ~10 minutes after volcanic explosions when acoustic wave pulses reach the ionospheric F region. They have periods of 1-2 minutes, and their records are characterized by short-term N-
shaped impulsive changes as Heki (2006) observed after the 2004 explosion of the Asama volcano.

Volcanic explosivity index (VEI) has been used to measure intensities of volcanic eruptions (Newhall and Self, 1982). For example, VEIs of the 2003 Soufrière Hills and the 2004 Asama eruptions are 3 and 2, respectively. In the previous section, I explored the way to use coseismic ionospheric disturbance amplitudes to infer earthquake magnitudes, here I discuss the feasibility to use ionospheric disturbances as a measure of volcanic eruption intensities. Manta et al. (2020b) proposed Ionospheric Volcanic Power Index (IVPI) and demonstrated its correlation with VEI. However, it is defined using the TEC signal power over a two-hours period and would not be appropriate for the short transient type-2 disturbances.

The 2022 Jan. 15, the Hunga-Tonga Hunga-Ha’apai submarine volcano, southern Pacific Ocean, erupted. It was the first VEI 6 volcanic eruption after GNSS became our tool, and its ionospheric signatures were beyond a simple categorization into two types. It caused ionospheric disturbances all over the world through the propagation of the Lamb wave that travelled round the world multiple times. The first wave was observed by GNSS receivers worldwide (Themens et al., 2022). In Japan, up to four passages of ionospheric disturbances have been recorded over several days (Heki, 2022). Astafyeva et al. (2022) revealed complex eruption sequence from near field TEC observations immediately after the eruption. One of the new remarkable findings would be the simultaneous propagation of conjugate anomalies in the northern and southern hemispheres connected by geomagnetic fields (Lin et al., 2022). Another finding is the identification of a unique disturbance that arrived in New Zealand prior to the Lamb wave and propagated ~0.6 km/s southwestward (Muafiry et al., 2023).

Type 1 disturbance

In addition to the 2003 eruption of the Soufrière Hills volcano, this type of disturbances has been found in Indonesia, Russia, Chile, and Japan. Harmonic TEC oscillations lasted ~20 minutes in the 2010 November 5 eruption of the Merapi volcano (VEI 4), central Java, Indonesia (Cahyadi et al., 2020). Similar oscillations lasted for ~ 2.5 hours during the February 2014 eruption of the Kelud volcano (VEI 4), in eastern
Java Island (Nakashima et al., 2016). They are both Plinian-type continuous eruptions. Shults et al. (2016) found two sequences of similar TEC oscillations in the 2015 April Plinian eruptions of the Calbuco volcano (VEI 4), Chile.

In the northern hemisphere, Shestakov et al. (2021) reported the TEC oscillations lasting for an hour during the 2009 eruption of the Sarychev Peak volcano (VEI 4), Kuril Islands, Russia. The Plinian eruption occurred on August 13, 2021, in a submarine volcano Fukutoku-Okanoba (VEI 4), Japan, and harmonic TEC oscillation has been observed for the first time using a geostationary satellite of QZSS (Heki and Fujimoto, 2022). Figure 8-7 shows four of these past cases, i.e., the 2014 Kelud (Indonesia), 2015 Calbuco (Chile), 2010 Merapi (Indonesia), and 2021 Fukutoku-Okanoba (Japan) eruptions.

Such harmonic TEC oscillations with a period ~4 minutes occur sometimes after large earthquakes, e.g., the 1994 Hokkaido-toho-oki earthquake (Figure 8-2c). Such oscillations were also found in the 2004 Sumatra-Andaman earthquake (Choosakul et al., 2009) and the 2011 Tohoku-oki earthquake (Rolland et al. 2011; Saito et al. 2011). However, they propagate outward from the epicenters in a Rayleigh surface wave speed (~4 km/s) possibly caused by efficient coupling of the Airy phase of the surface wave with atmosphere (Heki, 2021).
Figure 8-7. Examples of type-1 ionospheric disturbances, harmonic TEC oscillations, caused by the eruptions of the Kelud (Indonesia), Calbuco (Chile), Merapi (Indonesia), and Fukutoku-Okanoba (Japan) volcanoes in 2014, 2015, 2010, and 2021, respectively (Nakashima et al., 2015; Shults et al., 2016; Cahyadi et al., 2020; Heki and Fujimoto, 2022). (a, b) show the SIP trajectories for the first three cases, and (c) shows the stationary J07 SIP for the last case. (d) compares the time series of these cases over 1.4 hours after the onsets of the eruptions. For the 2015 Calbuco and 2010 Merapi cases, the onset times were 21:04 UT and 17:02 UT, respectively. The latter corresponds to the start of “phase 4” of the eruption (Cahyadi et al., 2020). For the 2014 Kelud and 2021 Fukutoku-Okanoba eruptions, exact onset times are unknown and were arbitrarily set as 16:12 UT and 5:03 UT, respectively.

On the other hand, wavefronts of the type-1 disturbances by volcanic eruptions propagate in an acoustic wave speed (~0.8 km/s) (Nakashima et al., 2015; Shults et al., 2016; Cahyadi et al., 2020; Heki and Fujimoto, 2022). Figure 8-7 gives two curves per eruption, observed by different station-satellite pairs. They slightly lag in time reflecting the outward propagation of acoustic wave (the SIP of the second pair is somewhat farther from volcanoes than the first pair).

The time lag between the eruption and the type-1 ionospheric disturbance is not clear because such continuous eruptions do not always have clear onsets (TEC does not show type-2 impulsive changes associated with the onsets, either). In the case of the 2010 Merapi eruption, the TEC oscillations emerged ~20 minutes after the “phase-4” Plinian eruption started (Cahyadi et al., 2020). The anomaly also started shortly before 20 minutes after eruption started (21:04 UT) in the Calbuco case (Shults et al., 2016). Thus, 20 minutes might be the approximate time for the continuous eruption to
excite significant atmospheric modes.

**Figure 8-8** (a) High pass filtered slant TEC time series at the Hahajima (0603) station observed using J07, the QZSS geostationary satellite (Figure 8-7c). Frequency components of the 4-hours data 5:20-9:20 UT (shown with a dashed line) obtained using the Blackman-Tukey method are shown in (b). We see three strong peaks (3.7, 4.4, and 5.4 mHz) and one weaker peak (4.8 mHz) that correspond to known atmospheric modes. (c) compares TEC signals from various satellites observed at the station 0603. Alphabets indicate the system (G: GPS, R: GLONASS, E: Galileo, J: QZSS) and the numbers indicate satellite numbers.

Long continuous TEC records enabled by geostationary satellites are suitable for studying their frequency spectra. Such records are also free from frequency shifts due to apparent movements of GNSS satellites in the sky. Heki and Fujimoto (2022) first obtained such data for the 2021 Fukutoku-Okanoba eruption. Figure 8-8c shows that the TEC oscillation was seen in nine satellites of various systems from the station 0603 equipped with a multi-GNSS receiver. I selected J07 (geostationary satellite of QZSS), whose slant TEC time series over 5:00-9:30 UT are shown in Figure 8-8a. We select the 4-hours data 5:20-9:20 and estimated their frequency components using the Blackman-Tukey method (Figure 8-8b).

Four frequency peaks shown in Figure 8-8b are close to the prescribed frequencies of 3.7, 4.4, 4.8, and 5.4 mHz (periods, 270, 227, 208, and 185 seconds), with the 4.8 mHz peak somehow weaker than the other three. The two frequencies, 3.7 and 4.4 mHz, are the atmospheric resonance frequencies detected by seismometers after the 1991 eruption of the Pinatubo volcano (e.g., Kanamori and Mori, 1992). The higher
two frequencies are their overtones (Watada and Kanamori, 2010). The power shows a sharp drop for frequencies lower than 3.7 mHz possibly because this frequency is close to the acoustic cut-off of the atmospheric filter (Blanc, 1985).

Figure 8-8a also suggests gradual decay of the harmonic oscillation in TEC. Increasing solar zenith angle during this period (local afternoon) also causes natural decay of the signal. However, the observed decay exceeds such a natural decay and rather reflects the decay of the Plinian eruption itself. Heki and Fujimoto (2022) used a new approach to compare fluctuations in three frequency pairs (L1-L2, L1-L5, L2-L5) and quantified real decay of ionospheric electron density fluctuations.

Figure 8-9 A new value from Fukutoku-Okanoba has been added to Cahyadi et al. (2020). The figure compares the products of the ionospheric disturbance amplitudes and the durations, with the total volume of the deposits inferred from geological approaches. For Fukutoku-Okanoba, I assumed the amplitude 3% of the background VTEC continued for 2 hours. Total volume of deposits is inferred assuming the pumice density same as water from the total weight of deposit of 3-10 x 10^{11} kg (Geol. Surv. Japan, 2021).

Cahyadi et al. (2020) compared the TEC oscillation amplitudes relative to background TEC for the three cases, 2010 Merapi, 2014 Kelud and 2015 Calbuco. They suggested that such relative TEC oscillation amplitudes are proportional to the mass eruption rates (MER), and the products of such amplitudes and the duration scale with the total amount of the ejecta. Figure 8-9 compares the total amount of the ejecta volume by the three eruptions listed in Cahyadi et al. (2020) and a new case from the 2021 Fukutoku-Okanoba eruption (Heki and Fujimoto, 2022).

For the amplitude of TEC oscillations in Fukutoku-Okanoba, the geometry factor has
been corrected, i.e., we would have observed ~4 times as strong oscillation if we had a
GNSS station to the south of the volcano (Figure 8-4). After correction, the TEC
oscillation amplitude becomes ~3% of the background VTEC, suggesting the MER at the
peak time (~5:20 UT) may have reached $5 \times 10^7$ kg s$^{-1}$. If such a strong oscillation
continued for 2 hours, the product of the amplitude and the duration would become
consistent with the total amount of ejecta, inferred as $3-10 \times 10^{11}$ kg by Geological Survey
of Japan (2021).

In future, ionospheric disturbances may provide useful information in near real time.
TEC can be monitored from GNSS stations hundreds of kilometers away from the volcano
(Figure 8-7). The amplitudes of the harmonic TEC oscillation would offer a rough
estimate of MER of the ongoing eruption within minutes. After the eruptions, we could
infer the total volume of deposits from the products of the oscillation amplitudes and the
durations. This would be useful where geological approaches are difficult.

Type 2 disturbance

The type 2 ionospheric disturbances (Figure 8-6) emerge ~10 minutes after explosive
volcanic eruptions when acoustic wave pulses reach the ionospheric F region. Although
they resemble to coseismic ionospheric disturbances caused by direct acoustic waves,
their periods (1-2 minutes) are significantly shorter than those related to earthquakes (~4
minutes). Here, I show five past examples of the type 2 disturbances, all observed in Japan,
and discuss their volcanological implications.

Figure 8-10 illustrates upward propagation of an N-shaped acoustic pulse. I assumed
the function given as equation (2) (the disturbance period is changed to 80 seconds, i.e.,
$\sigma = 20$ sec) and repeated the same calculation as Figure 8-2. Because the geomagnetic
field in the northern hemisphere mid-latitude region was assumed, we see southward
directivity. The disturbance front propagates outward with a speed of ~0.8 km/s. Like the
earthquake cases, amplitudes of the STEC disturbances depend on the incidence angles
of the line-of-sight with the wavefront. A stronger eruption would make a stronger
signature in ionosphere under the same geometry. Hence, GNSS-TEC measurements
would help us quantify the explosion intensities.

Intensity of a volcanic explosion has been studied by near-field atmospheric pressure
changes associated with the passages of airwaves generated by the eruptions (e.g., Matoza
et al., 2019). However, geometric settings of such sensors and volcanoes differ from
volcano to volcano, and such airwaves amplitudes cannot be the universal index. Here, I
propose new type-2 ionospheric disturbance amplitudes as a new index for explosion
intensities. For this purpose, we compare ionospheric TEC responses to five recent

Figure 8-10. Same as Figure 8-2, but the disturbance period was reduced to 80 seconds. The geomagnetic field near the Sakurajima Volcano (Declination: -7.2 degree, Inclination: 45.7 degree) was assumed. The figure shows three epochs, 11.0 (left), 12.5 (middle), and 14.0 (right) minutes after explosion.

Figure 8-11 compares the disturbances associated with five eruptions from Cahyadi et al. (2021). The first case is the 2004 September 11 11:02 UT eruption (VEI 2) of the Asama volcano, central Japan, whose ionospheric disturbances were reported in Heki (2006). The second case is the 2009 October 3 07:45 UT eruption (VEI 2) of the Sakurajima volcano, Kyushu. Plume reached the height ~3,000 m above the caldera rim by this eruption.

The third and the fourth cases are the two VEI 2 explosive eruptions on Jan. 31 22:54 UT, and on Feb. 11, 02:36 UT, 2011 of the Shin-Moe Volcano, Kyushu. In these eruptions, the plume reached the height of ~2,000 and ~2,500 m above the caldera rim, and the atmospheric pressure changes were 458.5 and 244.3 Pa at a sensor ~2.6 km southwestward, respectively (Japan Meteorological Agency, 2013).

The last example is the Kuchinoerabu-jima volcano, ~100 km to the south of Kyushu. A VEI 3 eruption occurred on 29 May 2015 (00:59 UT). The plume height was ~9,000
m, and pyroclastic flow reached the ocean. This example shows faint harmonic oscillations after the initial N-shape disturbance, suggesting this was a hybrid (of type-1 and -2) eruption.

Figure 8-11. Geometry of volcanoes (large stars), SIP tracks (gray curves) and SIP positions at the time of the eruptions (small yellow stars), and GNSS stations (squares) for five Vulcanian eruptions in Japan (left and right). The middle panel shows the STEC time series for the three station-satellite pairs for each of the five examples. Long-period changes were removed from STEC by subtracting the best-fit polynomials (degrees 7-9) over 45 minutes periods. Small disturbances can be seen ~10 minutes after the eruptions. Rewritten after Cahyadi et al. (2021).

The raw amplitudes of the five cases are given with red squares in Figure 8-12. I used background VTEC to normalize such amplitudes following the cases for coseismic ionospheric disturbances (Figure 8-5). Dark blue squares in Figure 8-12 express the TEC amplitudes relative to VTEC at the time and location of eruptions from GIM.

Another important factor is the geometry of line-of-sight with the acoustic wavefront (Figure 8-4). Different line-of-sight geometry with the wavefronts may result in significant differences in disturbance amplitudes. Such geometric factors have been corrected for the five eruption cases by (1) calculating time series with synthesized data using the real geometry of these cases and the “good geometry” (observing a northern satellite from a point 270 km to the south of volcanoes), and (2) converting the observed amplitudes to those under the good geometry. Light blue squares in Figure 8-12 show such corrected disturbance amplitudes.
Figure 8-12. Comparison of the STEC changes (shown in Figure 8-11) in absolute amplitudes (red). The yellow circles show VTEC values at the time and place of the eruptions calculated using GIM, and dark blue squares indicate disturbance amplitudes relative to the background VTEC. These amplitudes were further corrected for geometry factors and shown in light blue squares. For the two eruptions of the Shin-Moe volcano, we compare amplitudes of atmospheric pressure changes detected by the same sensor ~2.6 km from the volcano.

Explosion intensities can be studied also by measuring amplitudes of airwaves (atmospheric pressure changes). However, different distance of the ground sensors from the volcanoes and different topographic and vegetation conditions makes it difficult to compare such intensities for different volcanoes. In contrast, TEC changes focus on upward propagating acoustic wave, and inter-volcano comparison is possible.

Figure 8-12 includes two cases of atmospheric pressure changes for the January 31 and February 11 explosions of the Shin-Moe volcano detected using the same sensor (Japan Meteorological Agency, 2013). These two eruptions show similar amplitudes of STEC changes. However, considering the background VTEC and line-of-sight geometry, the January eruption is about twice as large as the February eruption. This agrees with the ratio of the pressure changes for these two eruptions (458.5 Pa and 244.3 Pa for the January and February eruptions, respectively).

This “within.volcano” comparison provides a small piece of support for the validity of using ionospheric disturbances to measure the explosion intensities. Deployment of infrasound sensors would remain useful because TEC changes can be detected only for strong volcanic explosions occurring when number of ionospheric electrons are sufficient.
(e.g., during daytimes). In fact, there were two explosions of the Shin-Moe volcano (Feb. 120:25 UT, and Feb. 13 20:07 UT) with stronger airwaves than the February 11 02:36 UT eruption. However, their ionospheric disturbances were not found because of small background VTEC early in the morning (Cahyadi et al., 2021).

Figure 8-11 suggests that TEC changes by the five different volcanic explosions have similar periods (~80 seconds), suggesting its origin in the atmospheric properties. Figure 8-1 inset shows the atmospheric attenuation of acoustic waves with various periods at different altitudes (Blanc, 1983). There, 1.3 minutes corresponds to the shortest period of the airwaves that reach the ionospheric F region (~300 km) without large attenuations. Surface infrasound records includesh stronger powers in periods shorter than 1.3 minutes for volcanic explosions. However, such components decay before disturbing the ionosphere, resulting in the 1.3 minutes component outstanding in the TEC records.

Despite the search outside Japan, it has been unsuccessful to add more clear type-2 cases due to the lack of GNSS stations in appropriate places or to the insufficient intensities of the explosions. One exception is the TEC signatures made by a human-induced explosion in 2020 August in Lebanon (Kundu et al., 2021). I expect that new cases of ionospheric disturbances caused by volcanic explosions will be reported by intensive searching efforts in the world.

Acknowledgements

The author is supported by Chinese Academy of Sciences, President’s International Fellowship Initiative (Grant number 2022VEA0014). He thanks those who produce daily observation data of permanent GNSS networks all over the world. This chapter includes materials from PhD theses of M. N. Cahyadi, Yuki Nakashima, and MSc thesis of Ai Matsushita at Hokkaido University.

References


Heki, K. (2022), Ionospheric signatures of repeated passages of atmospheric waves by the 2022 Jan.


Matoza, R., Fee, D., Green, D., and Mialle, P., (2019), Volcano infrasound and the international
monitoring system in *Infrasound Monitoring for Atmospheric Studies*, edited by Le Pichon, A., Blanc,

Muafiry, I. N., D. D. Wijaya, I. Meilano, and K. Heki, (2023), Diverse ionospheric disturbances by the
2022 Hunga Tonga-Hunga Ha‘apai eruption observed by a dense GNSS array in New Zealand, *J.
Geophys. Res. Space Physics*, under revision.

Atmospheric resonant oscillations by the 2014 eruption of the Kelud volcano, Indonesia, observed
with the ionospheric total electron contents and seismic signals, *Earth Planet. Sci. Lett.*, 434, 112–116,

Newhall, C. G. and Self, S., (1982), The volcanic explosivity index (VEI) – an estimate of explosive

Nishida, K., Kobayashi, and N., Fukao, Y. (2000), Resonant oscillation between the solid earth and

Ogawa, T., Kumagai, H., Sinno, K. (1982), Ionospheric disturbances over Japan due to the 18 May

Rideout, W. and A. Coster (2006), Automated GPS processing for global total electron content data,

Rolland, L.M., P. Lognonné, E. Astafyeva, E. A. Kherani, N. Kobayashi, M. Mann, and H. Munekane
(2011), The resonant response of the ionosphere imaged after the 2011 off the Pacific coast of Tohoku

and Cappa, F., (2013), Discriminating the tectonic and non-tectonic contributions in the ionospheric

Goi, and N. Choosakul (2011), Acoustic resonance and plasma depletion detected by GPS total
electron content observation after the 2011 off the Pacific coast of Tohoku Earthquake, *Earth Planets

Shestakov, N., Orlyakovskiy, A., Perevalova, N., Titkov, N., Chebrov, D., Ohzono, M., and Takahashi,
H., (2021), Investigation of ionospheric response to June 2009 Sarychev Peak Volcano eruption,

Shults, K., Astafyeva, E., and Adourian, S., (2016), Ionospheric detection and localization of volcano
eruptions on the example of the April 2015 Calbuco events. *J. Geophys. Res. Space Physics*, 121,


Doppler observations of acoustic wave exhibited by the Urakawa-Oki earthquake on 21 March 1982.


Tanimoto, T., K. Heki, and J. Artru-Lambin (2015), Interaction of Solid Earth, Atmosphere, and


