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1 **Ionospheric Detection of Natural Hazards**

2

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4

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8

9 **Key-points:**

10

11 - Natural Hazards can generate perturbations in the atmosphere and ionosphere via
12 acoustic and gravity waves

13

14 - From ionospheric observations, it is possible to detect natural hazards and to determine
15 their source parameters

16

17 - The review presents recent advances in detection of ionospheric disturbances generated
18 by earthquakes, tsunamis and volcanic eruptions

19

20 **Abstract**

21 Natural Hazards (NH), such as earthquakes, tsunamis, volcanic eruptions, and severe
22 tropospheric weather events, generate acoustic and gravity waves that propagate upward and
23 cause perturbations in the atmosphere and ionosphere. The first NH-related ionospheric
24 disturbances were detected after the great 1964 Alaskan earthquake by ionosondes and
25 Doppler sounders. Since then, many other observations confirmed the responsiveness of the
26 ionosphere to natural hazards. Within the last two decades, outstanding progress has been
27 made in this area owing to the development of networks of ground-based dual-frequency
28 Global Navigation Satellite Systems (GNSS) receivers. The use of GNSS-sounding has
29 substantially enlarged our knowledge about the Solid Earth/ocean/atmosphere/ionosphere
30 coupling and NH-related ionospheric disturbances and their main features. Moreover, recent
31 results have demonstrated that it is possible to localize NH from their ionospheric signatures,
32 and also, if/when applicable, – to obtain the information about the NH source (i.e., the source
33 location and extension, and the source onset time). Although all these results were obtained in
34 retrospective studies, they have opened an exciting possibility for future ionosphere-based
35 detection and monitoring of NH in near-real time.

36 This article reviews the recent developments in the area of ionospheric detection of
37 earthquakes, tsunamis and volcanic eruptions, and it discusses the future perspectives for this
38 novel discipline.

39

40

41 **Plain Language Summary**

42 The ionosphere is the ionized region of the Earth’s atmosphere that is located between
43 ~60 to ~1000 km of altitude. The ionosphere is largely impacted by the solar and magnetic
44 activities, as well as by the neutral atmosphere. Besides these large-scale and global processes
45 influencing from above, the ionosphere can be more “locally” perturbed from below by
46 geophysical phenomena (e.g., earthquakes, tsunamis, volcanic eruptions, severe tropospheric
47 weather events), and by man-made events (e.g., explosions, rocket and missile launches, mine
48 blasts). From below, the disturbances arrive in the ionosphere as acoustic and gravity waves.
49 Upon their upward propagation, these waves grow in amplitude up to a million times owing
50 to the exponential decrease of the atmospheric density with height. Consequently, the
51 acoustic and gravity waves generated at the Earth’s surface may provoke significant
52 perturbations in the upper atmosphere and ionosphere. In the ionosphere, the perturbations
53 can be detected by ionospheric sounding tools, such as, Global Navigation Satellite System

54 (GNSS) receivers, ionosondes, airglow cameras, etc.

55

56

57 **1. Introduction**

58 Natural Hazards (NH) are naturally occurring phenomena of geological, hydrological or
59 meteorological origin that might have a negative impact on humans or on the environment.
60 Every year, NH affect and take the lives of thousands of people. Some NH of meteorological
61 origin (hurricanes, tornadoes, and floods) can be forecasted and monitored in near real-time.
62 However, most geological NH (earthquakes, volcanic eruptions, landslides and tsunamis)
63 remain difficult or even impossible to predict. Therefore, their rapid and timely detection
64 becomes the only way to prevent or to reduce the human loss. To detect these NH in near-real
65 time, many different multi-instrumental approaches and methods are already in use
66 throughout the world. For instance, the Seismic Monitor by the Incorporated Research
67 Institutions for Seismology (IRIS, <http://ds.iris.edu/seismon/>) allows users to monitor global
68 earthquakes in real time. For tsunami warnings and monitoring, there exist several regional
69 and global systems: the U.S. Tsunami Warning System (<https://www.tsunami.gov/>), the Japan
70 warning system by the Japan Meteorological Agency (<http://www.jma.go.jp/en/tsunami/>),
71 the German-Indonesian Tsunami Early Warning System for the Indian
72 Ocean (<https://www.gitews.org/>; Falck et al., 2010), etc. These systems largely rely on
73 “classic” geophysical datasets. However, despite numerous efforts, the classic methods still
74 fail to correctly estimate the magnitude of large earthquakes ($M_w > 8$) in real-time and,
75 therefore, they also fail to correctly estimate the tsunami potential. In response to this need, it
76 has recently been suggested that the ionosphere-based technique could, in future, present a
77 novel approach for NH-detection in near-real time (e.g., Savastano et al., 2017).

78 Huang et al. (2019) have recently provided an extended review on ionospheric detection
79 of explosive events. The current review article is exclusively targeted at ionospheric detection
80 of such natural hazards as earthquakes, tsunamis and volcanic eruptions. After a detailed
81 overview of the recent developments and main challenges in this branch of geophysics, this
82 article discusses future perspectives for this novel discipline.

83

84

85 **1.1 Ionosphere, a natural sensor for Geophysics**

86 The ionosphere is a part of the upper atmosphere that extends from ~60 to ~800km of
87 altitude, where the concentration of charged particles (i.e., free electrons and ions) is

88 increased. The total mass of the ionosphere is 10^{12} - 10^{13} times smaller than the mass of the
89 neutral atmosphere, and, at all altitudes, charged particles can be considered as a minor
90 constituent in the particle mix.

91 The ionosphere is a highly irregular and variable medium; its state is determined by
92 several competing large-scale impacts [Hargreaves, 1992]. First of all, the concentration of
93 charged particles varies depending on the level of solar activity. Second, the ionization also
94 depends on the day/night conditions, on the season and on latitudes (solar zenith angle
95 dependence). Third, the electron density distribution is dependent on the altitude, defining
96 the ionospheric layers: D (60-90 km of altitude), E (90-120 km), F (120-800 km). The
97 ionization maximum (hmF2) is reached within the F-layer, at the altitude of ~250-400 km.
98 Fourth, the ionosphere is tightly coupled with the neutral upper atmosphere (thermosphere).
99 The thermosphere serves as a supplier of particles that can be ionized, it determines the
100 speed of ionization and regulates the recombination processes. Fifth, the ionosphere can be
101 strongly perturbed by disturbances in the geomagnetic field, such as geomagnetic storms and
102 substorms. In addition, the magnetic field plays an important role in the propagation of
103 plasma perturbations. The ionized particles are not free to move horizontally, as they are
104 confined by the earth's magnetic field. As a result, any movement of the neutral air in the
105 meridional direction will blow ionization along the magnetic field.

106 All these features impact the strength of the Solid-Earth/atmosphere/ionosphere
107 coupling, as well as the propagation of NH-related disturbances in the ionosphere. The
108 perturbations are initially generated on the ground/ocean surface as acoustic and gravity
109 waves.

110

111

112 **1.2 Acoustic and gravity waves and their propagation in the atmosphere**

113 Impulsive forcing from the Earth's surface occurring due to earthquakes, explosions,
114 volcanic eruptions, tsunamis, etc., trigger atmospheric pressure waves. Depending on their
115 frequencies, these atmospheric waves can be distinguished as acoustic and gravity waves
116 (Figure 1; Blanc, 1985; Huang et al., 2019). The acoustic waves are characterized by
117 frequencies higher than the acoustic cut-off frequency (ω_a), i.e. higher than ~3.3 mHz. The
118 acoustic waves are longitudinal waves in which particle moves in the direction of the wave
119 propagation. The force restoring a displaced particle towards its original position is the
120 change of pressure associated with the compression of the medium [Hargreaves, 1992].
121 Therefore, acoustic waves travelling through a gas disturb the equilibrium state by

122 compressions and rarefactions [Zel'dovich and Raizer, 2002]. The acoustic waves propagate at
123 the speed of sound that is equal to ~ 330 m/s at the surface, and grows up to 800-1000 m/s at
124 the ionospheric altitude of 250-300 km. Therefore, it only takes $\sim 8-9$ minutes for the acoustic
125 waves to reach the ionosphere.

126 When the wavelength of a wave increases, gravity starts to act on the mass displacement
127 and becomes the restoring force [Hargreaves, 1992; Fritts and Alexander, 2003; 2012].
128 Gravity waves are characterized by frequencies below the Brunt-Väisälä frequency ω_b (i.e.,
129 typically ~ 2.9 mHz in the lower atmosphere). Because of the gravity action, the gravity waves
130 cannot propagate upward vertically but do so obliquely; their velocity is much lower than the
131 sound speed, and their group and phase vertical velocities are opposite-directed [Hines, 1960;
132 Huang et al., 2019]. The gravity waves reach the ionospheric altitudes in 45-60 minutes from
133 the time of their generation on the ground/ocean surface.

134 The sources of the acoustic waves are: earthquakes, Rayleigh surface waves, explosions,
135 including volcanic and nuclear explosions, mine blasts, rocket launches, etc. [e.g., Blanc, 1985;
136 Calais & Minster 1995; Calais et al., 1998; Pokhotelov et al., 1995; Afraimovich et al., 2001;
137 2013; Heki and Ping, 2005]. In turn, gravity waves are generated by tsunamis propagating on
138 the ocean surface, by volcanic eruptions, by severe tropospheric events, such as tropical
139 storms, typhoons, hurricanes, [e.g., Hines, 1960; Artru et al., 2004; Fritts and Alexander, 2003;
140 2012; Rolland et al., 2010; Occhipinti et al., 2013; Dautermann et al., 2009a,b; Nishioka et al.,
141 2013; Chou et al., 2017].

142 The propagation and evolution of the acoustic and gravity waves in the atmosphere is
143 greatly affected by the propagation medium and its properties. The exponential decrease of
144 the atmospheric density with the altitude leads to the growth of both acoustic and gravity
145 waves upon their upward propagation. The wave's amplitude is proportional to the
146 atmospheric density; however, in accordance with the energy conservation law, the amplitude
147 of the acoustic and gravity waves increases in order to maintain constant energy flux [Hines,
148 1960]. The resulting amplification factor reaches 10^4 at ~ 150 km of altitude (lower
149 atmosphere) and 10^5-10^6 at 350-400 km (upper atmosphere and ionosphere) [Blanc et al.,
150 2010; Lognonné et al., 2006].

151 Competing with the amplification, damping effects occur because of the wave energy
152 dissipation due to molecular viscosity and thermal conductivity [Hines, 1960; Blanc, 1985].
153 The rate of energy dissipation is proportional to the kinematic viscosity, which in turn is
154 almost inversely proportional to the atmospheric density. Consequently, the effect of energy
155 dissipation becomes more important at greater altitudes. Further, the dissipation tends to

156 stronger affect higher-frequency waves, and these waves are removed as the certain altitudes
157 are reached [Hines, 1960; Huang et al., 2019]. Thus, the wave amplitude reaches the
158 maximum before vanishing at the altitudes of ~110 km for periods of 1s and ~160 km for
159 periods of 10s [Blanc et al., 2010]. The damping is also affected by non-linearity, propagation
160 speed dispersion as well as by the magnetic field [Naugolnykh and Ostrovsky, 1998].

161 It should be noted that powerful sources such as explosions and large earthquakes can
162 generate intense pressure waves that transform into shock waves upon their propagation in
163 the atmosphere [e.g., Pokhotelov et al., 1995; Zettergen et al., 2017]. These waves therefore
164 propagate at velocities exceeding the speed of sound.

165
166

167 **1.3. Wave interaction with ionospheric plasma and magnetic field**

168 When acoustic and gravity waves reach the ionospheric altitudes, they induce
169 fluctuations in the ionospheric electron density through dynamical and photochemical
170 processes. The ionospheric photochemical balance can be significantly altered by neutral
171 waves at lower altitudes (below 250 km), i.e. in weakly ionized medium [George and Hooke,
172 1970]. In regions with higher electron concentration (typically above 250 km of altitude), we
173 can assume that the atmospheric wave interacts with the ionospheric plasma only through
174 momentum transfer by collisions of neutrals with electrons and ions, which respond by
175 moving along the magnetic field direction. The velocity of the charged particles follows
176 approximately $V=Vn*\cos(\alpha)$, where Vn is the neutral particle velocity and α the angle between
177 the acoustic wave propagation direction and the magnetic field line.

178 It should be noted that the neutral waves generate not only perturbations in the electron
179 density, but also perturbations in the electric and magnetic fields, because of the interaction
180 of charged particles through electromagnetic forces to maintain the ionosphere electrically
181 neutral [e.g., Occhipinti et al., 2008; Kherani et al., 2008; 2012; Zettergren and Snively, 2015].

182 The fluctuations of the ionospheric plasma provoked by acoustic and gravity waves are
183 commonly referred to as traveling ionospheric disturbances (TIDs).

184 Numerous instruments can be used for studying the ionosphere dynamics and for
185 detection of TIDs. Global Navigation Satellite Systems (GNSS) is the main instrument used for
186 observations of NH-related TIDs.

187
188

189 **1.4. Global Navigation Satellite Systems (GNSS) for Ionospheric Sounding**

190 GNSS offer a powerful and accessible tool allowing the monitoring of the ionosphere.
 191 Due to the dispersive nature of the ionosphere, signals from dual-frequency GNSS receivers
 192 make it possible to measure the total electron content (TEC) – the electron density integrated
 193 along the raypath from a satellite to a receiver (Hoffmann-Wellenhof et al., 2008):

$$TEC = \int_{raypath} N(d\vec{r})d\vec{r} \quad (1.1)$$

194
 195 As of today, operational GNSS comprise the American Global Positioning System (GPS),
 196 Russia's GLONASS, the European Union's Galileo, China's BeiDou, Japan's Quasi-Zenith
 197 Satellite System (QZSS), the Indian Regional Navigation Satellite System (IRNSS, or NAVIC).
 198 From GNSS signals transmitted at two distinct carrier-frequencies, f_1 and f_2 , it is possible to
 199 calculate the differential (relative) slant TEC:

$$TEC = \frac{1}{A} \cdot \frac{f_1^2 f_2^2}{f_1^2 - f_2^2} (L_1 \lambda_1 - L_2 \lambda_2 + const + nL) \quad (1.2)$$

200 where $A = 40.308 \text{ m}^3/\text{s}^2$, $L_1 \lambda_1$ and $L_2 \lambda_2$ are additional paths of the signal caused by the phase
 201 delay in the ionosphere, $const$ is unknown initial phase path, caused by the unknown number
 202 of total phase rotations along the raypath and nL are errors in determining the phase path.

203 TEC is measured in TEC units (TECU) which is equal to 10^{16} electrons/ m^2 . The accuracy
 204 of the differential TEC estimation from phase measurements is about 0.01-0.02 TECU (e.g.,
 205 Coster et al., 2013).

206 In order to normalize the TEC amplitude and to avoid effects of low elevation angles,
 207 the slant TEC can be converted to vertical TEC (VTEC) by applying the following formula
 208 known as mapping function:

$$VTEC = STEC * \cos \left[\arcsin \left(\frac{R_z}{R_z + H_{ion}} \cos \theta_s \right) \right] \quad (1.3)$$

210
 211 where R_z is the Earth's radius, H_{ion} is the altitude of the ionospheric thin shell, θ_s is the
 212 elevation angle.

213 Knowing the positions of GNSS satellites and those of ground-based GNSS-receivers,
 214 one can estimate the coordinates of TEC perturbations. Assuming that the ionosphere is a thin
 215 shell layer located at the altitude H_{ion} (Figure 2), the intersection point between the H_{ion} and
 216 line-of-sight (LOS) between a satellite and a receiver is called ionospheric piercing point
 217 (IPP). The IPP projection onto the Earth's surface is called a sub-ionospheric point (SIP).

218 These IPP and SIP allow tracking ionospheric TEC perturbations observed by the GNSS-
219 method.

220 It is important to note that the GNSS-sounding does not precisely determine the H_{ion} ,
221 but the altitude can be presumed from physical principles. It is generally assumed that GNSS-
222 detected perturbations are concentrated around the altitude of the maximum ionospheric
223 ionization ($hmF2$), i.e., in the ionospheric F-layer. The H_{ion} is usually taken between 250 and
224 400 km, depending on solar, geomagnetic, seasonal and diurnal conditions. It should be noted
225 that changing the H_{ion} will change the IPP and SIP coordinates (Figure 2). This H_{ion} ambiguity
226 is the main disadvantage of the GNSS-sounding method that is otherwise widely and
227 commonly used by the space science community to study the ionosphere. Below we
228 demonstrate the use of GNSS-sounding for detection of earthquakes, tsunamis and volcano
229 eruptions.

230

231

232 **2. Ionospheric response to earthquakes and tsunamis**

233 **2.1. Earthquakes & tsunamis. Terminology.**

234 An earthquake is a shaking of the ground that results from the sudden release of
235 energy in the lithosphere. The majority of earthquakes occur along faults - narrow zones
236 where rocks move in relation to one another. The earthquake energy is released in the form of
237 seismic waves. These waves can be of several types: "body waves" that propagate inside the
238 volume (longitudinal P-waves and transverse S-waves), and "surface waves" that propagate
239 along interfaces such as the surface of the Earth. Depending on the particle motion within the
240 surface waves, they can be of Rayleigh (waves rolling along the ground) or of Love
241 (horizontally polarized shear waves) type. As of today, the Rayleigh waves are the only type of
242 seismic waves whose signatures were evidently detected in the ionosphere. The horizontal
243 speed of the Rayleigh surface waves depends on their frequency and wavelength but typically
244 varies between 3.3 and 4 km/s.

245 Earthquakes are characterized by several parameters. At an earthquake's onset, the
246 rupture starts at the underground point called the hypocenter (or focus). The epicenter is a
247 projection of the hypocenter's position onto the Earth's surface. Depending on the epicenter's
248 position, earthquakes can be referred to as inland and submarine/undersea.

249 Earthquake's relative size is defined by a number called magnitude. Several magnitude
250 scales have been developed in the past. However, the moment magnitude (M_w) scale that is

251 directly related to the energy of an earthquake is the one commonly used by seismological
252 authorities and by scientific communities.

253 Because earthquakes result from a slip along a fault, they are also characterized by the
254 fault orientation (the focal mechanism), the fault dimensions, the amplitude and the
255 orientation of the slip motion along the interface. With respect to their surface effects,
256 earthquakes can be characterized by vertical and horizontal displacements.

257 There are three main types of focal mechanisms (Figure 3). Depending on the co-
258 seismic crustal motion at the surface this can be thrust (the upper block moves on top of the
259 lower one), normal (the upper block moves downward) or strike-slip (the blocks move
260 horizontally past one another). Hence, only thrust and normal type earthquakes, that are also
261 referred to as dip-slip earthquakes, are characterized by significant vertical crustal motion.

262 The amplitude of the co-seismic vertical crustal displacements is the most important
263 parameter for the generation of ionospheric perturbations. This displacement is related to the
264 magnitude of an earthquake and, therefore, other characteristics, such as the fault dimensions
265 and the type, contribute as well. These same parameters play a key role in the generation of
266 tsunamis.

267 A tsunami (from Japanese “harbor wave”) is an unusually large wave occurring in a
268 result of the displacement of a large volume of water. Tsunamis are usually produced by
269 submarine earthquakes, but can also be generated by landslides or underwater volcanic
270 eruptions. The tsunami propagation speed depends on the depth of the ocean/sea and can be
271 approximated by the following formula:

$$272 \quad v = \sqrt{g * d} \quad (2.1)$$

273 where g is the gravitational acceleration, and d is the depth of the sea/ocean.

274

275

276 **2.2. Earthquakes and their signatures in the atmosphere and ionosphere**

277 During earthquakes, vertical displacements of the ground or of the ocean floor induce
278 perturbations in the atmosphere and ionosphere (Figure 4). The ionospheric perturbations,
279 called co-seismic ionospheric disturbances (CSID), are usually detected ~8-9 minutes after an
280 earthquake. The Rayleigh surface waves generated by earthquakes, propagate along the
281 Earth’s surface and induce acoustic waves that ~8-9 min later can be registered in the
282 ionosphere, similarly to CSID generated by the co-seismic crustal piston-like motion (Figure
283 4).

284 The theory describes that CSIDs, being of acoustic origin, are N-shaped (Figure 5a), i.e.

285 an initial overpressure half-cycle with a relatively fast risetime and a slower pressure decay
286 followed by a half-cycle of rarefaction [e.g., Zel'dovich and Raizer, 2002; Naugolnykh and
287 Ostrovsky, 1998]. In reality, however, co-seismic ionospheric perturbations often have more
288 complex waveforms (Figure 5b, c), occurring due to a superposition of signals from several
289 sub-sources – co-seismic crustal uplift, the Rayleigh waves, tsunamis, etc.

290

291

292 **2.2.1 CSID and their main features. Ground-based observations**

293 The first observations of co-seismic atmospheric and ionospheric perturbations were
294 made after the great Mw 9.2 earthquake that occurred on 28 March 1964 in Alaska, USA.
295 Davies and Baker [1965] reported the first detection of ionospheric disturbances at Boulder
296 Doppler sounder station ~15 min after the Alaskan earthquake. The observed perturbations
297 were suggested to be the manifestation of long-period ducted acoustic-gravity waves emitted
298 into the ionosphere near the epicenter. Leonard and Barnes [1965] registered CSID generated
299 by the 1964 Alaskan earthquake at several ionosonde stations located in the continental USA
300 and Hawaii. Since those first observations, CSID are routinely detected by Doppler sounders
301 (e.g., Artru et al., 2004; Liu et al., 2006a,b; Chum et al., 2012; 2016), and more rarely – by the
302 ionosondes [e.g., Maruyama and Shinagawa, 2014; Maruyama et al., 2017].

303 The development of the GNSS and permanent networks of ground-based GNSS-receivers
304 has provided a new technique allowing continuous observations with better spatial coverage.
305 Calais and Minster (1995) were the first to use the ionospheric GPS-TEC data for detection of
306 ionospheric co-seismic perturbations over California after the 1994 Mw6.7 Northridge
307 earthquake. Since that pioneering work, the GPS/GNSS-sounding allowed better
308 understanding of the CSID and their properties. For instance, it is now demonstrated that
309 earthquakes with moment magnitudes $M_w > 6.8$ are very much probable to cause
310 perturbations in the ionosphere (Perevalova et al., 2014).

311 CSIDs can be detected in the vicinity of an earthquake's epicenter, but also up to 6000-
312 8000 km away from it. The near-field CSIDs represent ionospheric response to acoustic waves
313 generated directly by co-seismic vertical crustal displacements (e.g., Calais and Minster, 1995;
314 1998; Afraimovich et al., 2001; 2010a,b; Heki and Ping, 2005; Lognonné et al., 2006; Astafyeva
315 and Heki, 2009; Cahiyadi and Heki, 2013; Jin et al., 2015; Zettergren and Snively, 2015;
316 Komjathy et al., 2016; Li et al., 2019). The far-field CSIDs are usually associated with the
317 propagation of the Rayleigh surface waves (Ducic et al., 2003; Liu et al., 2006b; Astafyeva et al.,
318 2009; Rolland et al., 2011a; Kakinami et al., 2013), or of the air surface waves (Astafyeva and

319 Afraimovich, 2006; Liu et al., 2006a). In addition, large earthquakes can be accompanied by
320 the acoustic resonance signatures in the ionosphere (Choosakul et al., 2006; Liu et al., 2011;
321 Rolland et al., 2011b; Saito et al., 2011). Ionospheric perturbations generated by these
322 different sub-sources can be distinguished by their propagation speed. Thus, CSID associated
323 with the acoustic waves generated by the crustal motion propagate at 600-1000 m/s; CSID
324 generated by Rayleigh surface waves have speeds similar to those of the source waves on the
325 ground, i.e. 3.3-4 km/s. These different CSID components can be well seen in travel-time
326 diagrams (TTD, or hodocrones) that show the TEC amplitude with respect to the distance
327 from the epicenter and time (Figure 6). From TTD, first, one can verify whether the epicenter
328 is the source of the observed TEC perturbations. Second, the apparent horizontal speed can be
329 easily estimated from the TTD. Third, the TTD also allows tracking the distance of CSID
330 propagation. For instance, Figure 6a shows continuous observations of CSID as far as 1900 km
331 from the epicenter. One can also notice that the initial CSID propagating at ~ 1 km/s splits in
332 two perturbations at ~ 600 -700 km of the epicentral distance (Figure 6a). The velocities of the
333 “new” perturbations are 3 km/s and 0.6 km/s. This phenomenon was first observed by
334 Astafyeva et al. (2009) on the example of the M8.3 1994 Kuril earthquake and was called two-
335 mode propagation of CSID. The “splitting” of CSID into faster and slower modes was explained
336 by the difference in their propagation speed. The occurrence of multiple modes was later
337 confirmed for several other large earthquakes, including the great Tohoku-oki earthquake of
338 11 March 2011 (e.g., Liu et al., 2011; Rolland et al., 2011b; Galvan et al., 2012; Kakinami et al.,
339 2013; Jin et al., 2015), the M7.8 April 2015 Nepal earthquake (Reddy and Seemala, 2015;
340 Tulasi Ram et al., 2017) and the 2005 Northern California offshore earthquake (Jin, 2018).
341 Figure 6b shows multiple-mode CSID observed by satellite G15 after the Tohoku-oki
342 earthquake observed by Galvan et al [2012].

343 Besides the above-mentioned CSID components, some other types of co-seismic
344 ionospheric signatures can be distinguished. Thus, during large earthquakes, the initial near-
345 field TEC enhancement is often followed by a significant long-term depletion (e.g., Saito et al.,
346 2011; Tsugawa et al., 2011; Kakinami et al., 2012). Kakinami et al. (2012) analyzed TEC
347 response to three largest submarine earthquakes in Indonesia, Chile, and Japan, and found
348 that a TEC depletion occurred in all those cases just above the tsunami source area.
349 Consequently, Kakinami et al. (2012) attributed the observed depletion to “the downwelling
350 of the sea surface” due to the tsunami generation, and called it the “tsunamigenic ionospheric
351 hole” (Figure 5c-d). Astafyeva et al. (2013a), however, argued that such depletion represents
352 the negative half-phase of the N-wave, and showed that the “hole” can occur following any

353 large earthquake independently on the epicenter being located inland or undersea. These
354 latter conclusions were later supported by observations of TEC response to the inland Nepal
355 earthquake of 25 April 2015 (Kamogawa et al., 2015), and by modeling (Shinagawa et al.,
356 2013; Zettergren et al., 2017). Astafyeva et al. (2013a) also noticed that the strength and the
357 longevity of the ionospheric depletions scale with the magnitude of earthquakes (Figure 7c).
358 This observation indicates a close relationship between earthquakes' parameters and co-
359 seismic signatures in the ionosphere. More details can be found in the next Section.

360
361

362 **2.2.2 CSID and earthquake parameters**

363 Recently, it has been found out that the coupling between the ionosphere and the solid
364 earth is very tight. For instance, Astafyeva et al. (2013a) analyzed TEC response to 11
365 earthquakes of different magnitude and found that the amplitude of the near-field CSID grows
366 roughly linearly with the magnitude of an earthquake (Figure 7a), and with the magnitude of
367 the co-seismic uplift (Figure 7b). Cahiyadi and Heki (2015) later confirmed such dependence
368 for the magnitude of an earthquake. However, they have found no correlation between the
369 uplift and the amplitude of the ionospheric response. Therefore, this subject requires more
370 profound analysis.

371 The focal mechanism, a characteristic that describes the deformation in the source
372 region (Figure 3) is another earthquake parameter that may have a crucial impact on the CSID
373 waveform. Astafyeva and Heki (2009) analyzed TEC response to three earthquakes of
374 different focal mechanisms (thrust, normal and a combination of the two). They showed that
375 the waveform of the signal in the ionosphere repeated the initial crustal motion on the
376 ground: the thrust earthquake caused normal N-wave response in the ionospheric TEC,
377 normal fault earthquake cause inversed N-wave, while the TEC response to the mixed type
378 earthquake showed both direct and inversed N-waves. However, Rolland et al. (2013) argued
379 that such dependence might be an artifact caused by the magnetic field as shown in the next
380 Section.

381 With respect to the focal mechanism, it is of interest to study the ionospheric response
382 to strike-slip fault earthquakes (i.e., earthquakes with small vertical crustal motion) and that
383 do not generate large tsunamis. Astafyeva et al. (2014) analyzed TEC response to six large
384 strike-slip earthquakes that occurred in 2001-2013 and found that the response in the
385 ionospheric TEC was almost as strong as for perturbations caused by dip-slip earthquakes.
386 They also showed that the scaling law for the CSID amplitude increasing with the magnitude

387 is also true for strike-slip earthquakes. Contrary to those results, Cahiyadi and Heki (2015)
388 demonstrated that strike-slip earthquakes generate CSID with weaker amplitudes than dip-
389 slip earthquakes of similar magnitudes.

390 The observation of ionospheric response to strike-slip earthquakes remains one of the
391 open questions and is still difficult to explain. Astafyeva et al. (2014) suggested that in
392 absence of significant vertical static co-seismic displacements of the ground, other
393 seismological parameters (primarily the magnitude of co-seismic horizontal displacements,
394 seismic fault dimensions, seismic slip or dip angle) may contribute in the generation of large-
395 amplitude ionospheric perturbations. Future studies, especially modeling of the ionospheric
396 response to strike-slip fault earthquakes, will shed light on this mystery.

397

398

399 **2.2.3 CSID and the propagation medium parameters**

400 It is known that the amplitude and the waveform of CSID are largely controlled by
401 observational factors, such as the geometry of GNSS-sounding, geomagnetic field and the
402 background ionization [e.g., Calais and Minster, 1995; Heki and Ping, 1995; Afraimovich et al.,
403 2001; Rolland et al., 2013; Cahiyadi and Heki, 2015].

404 The amplitude of ionospheric perturbations detected by the GNSS-method, experiences
405 strong azimuthal dependence due to the integral character of transionospheric sounding.
406 When a perturbation's wave vector and satellite-to-receiver LOS are perpendicular to each
407 other, we observe perturbation of the largest amplitude. Otherwise, if the wave vector is
408 parallel to LOS, the integration of wavefronts cancels the wave phases and, therefore, results
409 in the minimum amplitude of perturbation (e.g., Afraimovich et al., 1998; 2001; Bagiya et al.,
410 2017a).

411 Further, as explained above, the magnetic field configuration at the epicentral area has a
412 major impact on the amplitude and waveform of CSID propagating in the ionospheric F -
413 region. Rolland et al. (2013) computed the ionospheric radiation pattern, and showed that the
414 magnetic field impact is somewhat different for the positive and negative sub-phases of CSID
415 (Figure 8). The positive part is attenuated for CSID propagating at angles exceeding $\pm 10^\circ$ to
416 the magnetic field lines, and significantly attenuated for angles larger than $\pm 30^\circ$. On the north
417 from the epicenter in the Northern Hemisphere (i.e. poleward from the epicenter), only the
418 weakened negative part of the signal can be seen. These modeling results are in line with
419 observations showing that CSID do not propagate far in the poleward or quasi-poleward
420 direction and fade out very fast (e.g., Otsuka et al., 2006; Astafyeva and Heki, 2009; Rolland et

421 al., 2013). Whereas, the equatorward propagation is favored, and CSID can be observed
422 farther from the source and they have larger amplitudes (Figure 8).

423 The background ionospheric parameters (i.e., the electron density, absolute vertical
424 TEC) also play a significant role in CSID amplitude. Hence, in order to compare amplitudes of
425 CSID generated by different earthquakes, normalization on the background ionospheric
426 parameters should be done [Cahiyadi and Heki, 2015].

427

428

429 **2.2.4 Space-borne observations of CSID**

430 The development of new space missions made it possible to detect atmospheric and
431 ionospheric perturbations directly in space. Thus, Garcia et al. (2013) analyzed data of neutral
432 mass density calculated from accelerometers onboard Gravity Field and Steady-State Ocean
433 Circulation Explorer (GOCE) spacecraft with orbital altitude ~ 250 km, and presented the first
434 in situ observations of a post-seismic infrasound wavefront. The atmospheric infrasound
435 signals due to the great Tohoku-oki earthquake of 11 March 2011 induced variations of air
436 density and vertical acceleration of the GOCE platform. Perturbations up to 11% in the air
437 density and 1.35×10^{-7} m/s² in vertical acceleration were reported. Yang et al. (2014) observed
438 co-seismic ionospheric TEC and atmospheric density perturbations at the altitude of ~ 450 km
439 by using data of Gravity Recovery and Space Experiment (GRACE) satellite. The perturbations
440 were generated by the Tohoku-oki earthquake in Japan but were detected over Alaska, ~ 8
441 minutes after the arrival of seismic and infrasound waves on the ground. Yang et al. (2014)
442 also reported sudden changes in GRACE acceleration following the Tohoku-oki earthquake.

443 Furthermore, space-borne observations shed light on the vertical scale of CSID, a feature
444 that remains unknown. Sun et al. (2016) used data from satellite mission FORMOSAT-
445 3/COSMIC and GPS radio-occultation technique, and, on the example of the Mw7.8 April 2015
446 Nepal earthquake, demonstrated that the earthquake-induced perturbations penetrate into
447 the ionosphere at supersonic speeds of approximately 800 m/s. The corresponding change in
448 the electron density structure reach 10% near the maximum ionization height (hmF2). The
449 vertical scale of the perturbation was found to be of 150 km, and the hmF2 was uplifted by
450 more than 30 km within 1 min. These results showed that earthquakes represent a significant
451 force that perturbs the vertical structure of the ionosphere.

452

453

454 **2.3. Tsunamis and their signatures in the atmosphere and ionosphere**

455 Tsunamis propagating along the ocean surface generate internal gravity waves that,
456 unlike the acoustic waves, propagate obliquely. Because of the low vertical velocity of ~ 50
457 m/s, the tsunami-generated gravity waves reach the ionospheric heights in ~ 45 -60 min after
458 their generation on the surface (Figure 4). The co-tsunami ionospheric disturbances (CTID)
459 are usually characterized as quasi-periodic structures with typical periods between 10 and 30
460 min (Figure 9a,b). CTID can be detected in the vicinity of earthquakes' epicenters soon after
461 they are generated, so that CTID can be seen in TTD together with co-seismic ionospheric
462 perturbations observed in the near-epicentral area (Figure 6b). Also, CTID can be observed in
463 open sea/ocean (by using GNSS data on islands), and/or upon their arrival on shores. CTID
464 match the period, velocity, and propagation direction of the tsunamis causing them.
465 Consequently, tracking CTID provides an opportunity to follow the propagation of tsunamis
466 and can potentially be used for near-real-time tsunami warning systems (e.g., Najita et al,
467 1974; Savastano et al., 2017).

468

469 **2.3.1 CTID and their main features. Ground-based observations**

470 Hines [1972] and Peltier&Hines [1976] were the first who theoretically predicted that
471 sea surface variations typical of tsunamis can seed internal gravity waves with amplitudes
472 strong enough to generate ionospheric signatures. The first GPS-detection of tsunami-driven
473 ionospheric perturbations belongs to Artru et al. (2005). The $M_w=8.2$ earthquake in Peru on
474 23 June 2001 generated a tsunami in the Pacific Ocean that reached the coast of Japan 22
475 hours later. Artru et al. (2005) used data from the Japanese Network of GPS-receivers
476 GEONET and observed small-scale TEC perturbations over Japan and up to 400 km offshore.

477 The giant tsunami of December 2004 in the Indian Ocean with maximum wave heights
478 of 10 m generated significant perturbations in the ionosphere that were observed by different
479 techniques and methods. Liu et al. (2006a) reported the detection of CTID registered by GPS-
480 receivers in the Indian Ocean area. The observed CTID with periods of 10-20 min, and
481 horizontal wavelengths of 120-240 km traveled away from the epicenter with an averaged
482 horizontal speed in the ionosphere of about 190 m/s. Das Gupta et al. (2006) also used data
483 from ground-based GPS-receivers and reported observations of CTID of a quite significant
484 amplitude of 1.5 to 2 TECU. The perturbations occurred within 45 minutes of the quake at
485 stations situated near the east coast of the Indian subcontinent. These ionospheric
486 observations showed good agreement between the CTID parameters and those of the tsunami,
487 providing strong evidence of the tsunami-ionosphere coupling. Subsequent studies by Rolland
488 et al. (2010) and Galvan et al. (2011) brought more observational evidences of the ionosphere

489 sensitivity to tsunami-generated gravity waves. By using data of GPS-receivers located in
490 Hawaii and in Japan, they showed tsunami-associated TEC fluctuations of the order of $\sim 1\%$ of
491 the background TEC value. Comparison with oceanic data from Deep-ocean Assessment and
492 Reporting of Tsunamis (DART) buoys showed similarity in the waveform and in the spectral
493 signatures of the ionospheric and oceanic data.

494 The great March 2011 Tohoku tsunami, with wave heights reaching up to 38 m in
495 Ofunato region of Japan, opened a new era in the ionospheric detection of tsunamis. The
496 GEONET, the World's most advanced network of GPS-receivers, allowed the first-time
497 detection of CTID in the near-field of the seismic source (e.g., Liu et al., 2011; Rolland et al.,
498 2011b; Tsugawa et al., 2011; Galvan et al., 2012; Komjathy et al., 2012; Occhipinti et al., 2013;
499 Jin et al., 2014). The Tohoku tsunami was also detected in the far-field, i.e., several thousand
500 kilometers from the epicenter, by using different techniques. Data from ground-based GPS –
501 receivers enabled ionospheric detection of the Tohoku tsunami near the Hawaiian Islands
502 (e.g., Makela et al., 2011; Occhipinti et al., 2011), and near the west coast of North America
503 (Crowley et al., 2016; Azeem et al., 2017). In addition to the GPS-TEC data, the Tohoku
504 tsunami was observed in the airglow layers at altitudes near 250 km, which was the first
505 detection of tsunami-related perturbations in the airglow (Makela et al., 2011; Figure 9c).
506 Images were taken by ground-based wide-angle camera system located at the top of the
507 Haleakala Volcano on Maui, Hawaii, and showed good correlation with GPS measurements of
508 the TEC from Hawaii GPS stations and the Jason-1 satellite. After that first-time registration,
509 the tsunami-induced airglow signatures were also observed following the 2012 Haida Gwaii,
510 the 2015 Illapel and the 2010 Chile earthquakes (Grawe and Makela, 2015; 2016). These
511 works demonstrated the utility of monitoring of the Earth's airglow layers for tsunami
512 detection and tracking.

513 Observations of the airglow response to the 2011 Tohoku tsunami revealed another
514 interesting phenomenon: the occurrence of signatures in magnetically conjugate regions.
515 Simulation results presented in Huba et al. [2015] indicated that gravity wave-induced
516 variations in the electric field can map to the conjugate hemisphere where they impact the
517 ionospheric TEC and 630 nm airglow emission. This result, however, needs to be confirmed
518 with further observational studies.

519 Finally, it is interesting to note that CTID can, in some cases, arrive before tsunamis
520 (Makela et al., 2011; Bagiya et al., 2017b). Bagiya et al (2017b) explained the generation of
521 such ahead of tsunami ionospheric disturbances by dissipation of transverse mode of the

522 primary acoustic-gravity waves generated by tsunamis. This can be useful to short-term
523 tsunami forecasts.

524

525 **2.3.2 Space-borne observations of CTID**

526 Besides ground-based instruments, tsunami-driven atmospheric and ionospheric
527 signatures were detected by space-based equipment. Garcia et al. (2014) used data from
528 GOCE satellite and observed the Tohoku tsunami-driven perturbations in the air density.
529 Coisson et al. (2016) presented the first observations of the tsunami-gravity waves by using
530 ionospheric radio-occultation data from COSMIC constellation. Yang et al. (2017) used data
531 from the Sounding of the Atmosphere using Broad-band Emission Radiometry (SABER)
532 spacecraft and observed significant perturbations of nightglow emissions in the mesosphere.
533 The perturbations were caused by tsunami waves of the 2015 Chile and 2011 Tohoku
534 earthquake/tsunami events.

535

536 Summarizing the above, great progress has been achieved recently in the research of
537 atmospheric and ionospheric response to earthquakes and tsunamis. These results can
538 already be used to track tsunamis propagating in the far-field and to predict their arrivals to
539 distant coasts. The next frontier question is – *can it be possible to use our knowledge on CSID
540 and CTID to trace tsunamis generated in the near-field from the source?* For a short-time
541 tsunami warning, one needs to know in near-real time the information about the seismic
542 source (i.e., the magnitude of the co-seismic vertical crustal displacements and/or the fault
543 dimensions). While this task still remains challenging for the “classic” seismology [LaBrequé
544 et al., 2017], it may be addressed in future to the Ionospheric Seismology – a new branch of
545 Geophysics.

546

547

548 **2.4. Ionospheric Detection of Earthquakes and Tsunamis. “Ionospheric Seismology”**

549 In earlier times, the term “Ionospheric Seismology” was used to describe a branch of
550 Geophysics studying the ionospheric response to earthquakes and to tsunami propagation.
551 Nowadays, the Ionospheric Seismology is becoming more of the use of ionospheric
552 measurements for seismological purposes: to localize the seismic source, to estimate the
553 dimensions of the seismic fault, or –in future- to estimate the magnitude of an earthquake or
554 the height of tsunami waves from ionospheric data and observations. All these tasks are yet
555 quite challenging but have a very important perspective for the future – the possibility to use

556 ionospheric measurements for near real-time tsunami warning systems (e.g., Najita et al.,
557 1974; Kamogawa et al., 2016; Savastano et al., 2017; Occhipinti, 2015).

558 With regard to the new definition of the Ionospheric Seismology, the first detection of
559 the seismic source from the ionosphere was done by Afraimovich et al. (2006). By using data
560 from ground-based GPS-receivers, they first analyzed CSID generated by the 2003 Tokachi-oki
561 earthquake in Japan. Then, assuming that the CSID propagate from a point source at a
562 constant speed as a spherical wave, Afraimovich et al. (2006) managed to determine the
563 position of the CSID source in the vicinity of the epicenter. This method was later successfully
564 tested for detection of the seismic source of several other large earthquakes that occurred in
565 Japan and in Sumatra in 1994 - 2006 (Afraimovich et al., 2006; Kiryushkin and Afraimovich,
566 2007; Astafyeva et al., 2009; Kiryushkin et al., 2011).

567 Liu et al. (2010) developed an ionospheric method based on the raytracing and the
568 beam-forming techniques, and successfully localized the epicenter of the 1999 Chi-Chi
569 earthquake in Taiwan from ionospheric measurements. Tsai et al. (2011) traced the epicenter
570 of the Tohoku-oki earthquake and the origin of the tsunami from ionospheric TEC
571 observations of co-seismic and co-tsunamic ionospheric perturbations. Their results for both
572 the tracked epicenter and the tsunami origin were close to those reported by the USGS. More
573 recently, Lee et al. (2018) localized the seismic source of the 2016 Kaikoura earthquake in
574 New Zealand by applying for the first time an ionosphere-based backprojection method. The
575 Mw7.8 Kaikoura earthquake resulted from a complex multi-fault rupture (e.g., Bagiya et al.,
576 2018), and Lee et al. (2018) estimated the source region located around the maximum uplift
577 area.

578 Therefore, nowadays it is certainly possible to determine the position of the seismic
579 source from the ionosphere. Going further forward, a more challenging task is to understand
580 the seismic source structure from the ionosphere. Several works have shown that this can be
581 achieved. Heki et al. (2006) used ionospheric TEC measurements and managed to reproduce
582 the rupture process of the M9.1 2004 Sumatra earthquake. Heki et al. (2006) also
583 demonstrated that the complex waveform of the observed co-seismic TEC variations was due
584 to the contribution of several sub-faults ruptured during the earthquake.

585 Astafyeva et al. (2011) introduced for the first time a method of ionospheric imagery of
586 seismic fault, the “seismo-ionospheric imagery”. They used high-rate 1Hz GPS-TEC data and,
587 on the example of the 2011 Tohoku-oki earthquake, obtained the ionospheric “imprint” of the
588 co-seismic uplift. It should be reminded that the seismic source generating CSID, is also
589 responsible for the generation of tsunamis; consequently, this method, once adapted for

590 systematic imagery, can be used for future ionosphere-based tsunami warning systems. One
591 of the biggest advantages of the seismo-ionospheric imagery is the timing, since the first CSID
592 occur in the ionosphere ~8 min after an earthquake. This means that by using the ionospheric
593 TEC data in quasi-real time, it is possible to obtain information about the location and
594 dimensions of the source of a tsunami within several minutes after a quake. For instance, in
595 the case of the Tohoku-oki earthquake, the first arrival of the tsunami was recorded between
596 26 and 35 minutes after the earthquake had struck, so that the 17 minutes of remaining time
597 could be used for ionosphere-based tsunami warning systems in near-real time.

598 To examine the possibility of the use of the ionosphere-based imagery systematically,
599 Astafyeva et al. (2013b) analyzed the geometry of GPS-sounding for multiple satellites during
600 the Tohoku-oki earthquake. They found that the seismo-ionospheric imagery is only possible
601 upon detection of the primary acoustic waves coming directly from the epicentral area.
602 Astafyeva et al. (2013b) also demonstrated that plotting standard snapshots of TEC over the
603 epicentral area is useful to retrieve the information about the seismic source structure. Thus,
604 for the case of the 2011 Tohoku-oki earthquake, two areas of enhanced TEC simultaneously
605 occurred at ~150 km eastward from the epicenter starting from 510 sec after the earthquake
606 (Figure 10b). The location of these two TEC enhancements corresponded to the location of
607 two segments of co-seismic crustal slip as it was shown by seismologists (e.g., Simons et al.,
608 2011; Bletery et al., 2014; Figure 10a). More recently, Thomas et al. (2018) and Astafyeva &
609 Shults (2019) demonstrated that the GNSS seismo-ionospheric imagery can work for
610 moderate earthquakes as well. Astafyeva and Shults (2019) also noted that lower resolution
611 data sampling, such as 15-sec or 30-sec, which so far is a standard resolution for GNSS data,
612 will, most likely, not work efficiently.

613 Kamogawa et al. (2016) suggested a method based on observations of “tsunami-
614 ionospheric hole” as potentially useful for near real-time tsunami warnings. Kamogawa et al.
615 (2016) found a quantitative relationship between the initial tsunami height and the TEC
616 depression rate caused by the hole from seven tsunamigenic earthquakes in Japan and Chile.
617 They showed that the percentage of TEC depression and initial tsunami height are correlated.
618 Another promising study was recently done by Rakoto et al. (2018), who performed for the
619 first time an inversion of the tsunami wave height from ionospheric TEC measurements.
620 Occhipinti et al. (2018) introduced for the first time so-called “ionospheric surface wave
621 magnitude”, a new way to characterize earthquakes by observing the signature of surface
622 Rayleigh waves in the ionosphere.

623 It is important to note that all those methods were proposed based on examples of past
624 earthquakes, and so far they can only potentially be used for ionospheric near-real-time
625 detection of earthquakes and for tsunami warnings. Going towards the real-time application,
626 Savastano et al. (2017) made the first preliminary feasibility demonstration for near real-time
627 ionosphere monitoring by GNSS-sounding. Their approach has been implemented at several
628 GNSS-receivers around the Pacific Ocean (<https://iono2la.gdgps.net>), and is aiming -in the
629 future- to detect CTIDs in near-real time.

630

631

632 **3. Ionospheric response to volcanic eruptions**

633 **3.1. Volcanic eruptions. Terminology**

634 A volcanic eruption is a sudden ejection of matter from the earth's interior through a
635 volcanic vent. Eruptions can be effusive, with outflows of hot lava, or explosive, throwing out
636 huge amounts of rock with volcanic ash and gases. The style and evolution of volcanic
637 eruptions are dictated by the fluid mechanics governing magma ascent (Gonnermann and
638 Manga, 2007).

639 The intensity of eruptions is usually estimated by the Volcanic Explosivity Index (VEI)
640 on the scale from 0 to 8. The VEI is a characteristic similar to a magnitude for earthquakes.
641 VEI=0 eruptions are continuous eruptions, while the recurrence of VEI=8 eruptions is 1 per
642 50,000 years. Up to date, there are only records of ionospheric response to explosive
643 eruptions with VEI from 2 to 6.

644 Volcanic eruptions are usually preceded by a series of small earthquakes, known as
645 seismic tremor. The tremor can last from several tens of minutes to several days.

646

647

648 **3.2. Volcanic eruptions and their signatures in the atmosphere and ionosphere.**

649 **Ground-based and space-borne observations**

650 Explosive volcanic eruptions can generate a broad spectrum of pressure oscillations,
651 from infrasonic to gravity waves (e.g., Kanamori, Mori, 1992; Kanamori et al., 1994). At the
652 near-ground level, eruptions generate infrasound waves (acoustic domain, below 20 Hz) that
653 can propagate for very long distances (e.g., Matoza et al., 2011; Fee and Matoza, 2013; Ripepe
654 et al., 2010). Whereas, lower frequency oscillations (between 0.97 and 3.5 mHz), typical of
655 atmospheric gravity waves are quite rare to observe and can only be captured in the vicinity
656 of eruptions (De Angelis et al., 2011).

657 Besides the near-surface perturbations, and similar to earthquakes and tsunamis,
658 volcanic eruptions can generate atmospheric and ionospheric perturbations (Figure 11). In
659 the atmosphere, the shock of the volcanic eruption can trigger mesospheric gravity waves (or
660 mesospheric airglow waves) as shown in Figure 12a for the 23 April 2015 eruption of Calbuco
661 volcano in Southern Chile. Such “night glow” can be emitted from a variety of gases located
662 near the mesopause (80-105 km).

663 In the ionosphere, volcanic eruptions generate so-called co-volcanic ionospheric
664 disturbances (CVID). The CVID are usually quasi-periodically shaped and occur ~ 10 to 45
665 minutes after the eruption onset (Figure 13). Up to now, only a few reports on detection of
666 CVID have been done, and our knowledge on CVID, their generation and evolution is not yet
667 sufficient.

668 Heki (2006) was the first who detected CVID in the ionospheric TEC. By using data from
669 the GEONET, Heki (2006) studied the ionospheric response to the 1 September 2004 Asama
670 volcano explosion in Japan. The moderate VEI=2 explosion caused a small-amplitude ~0.03-
671 0.16 TECU ionospheric response ~12 minutes after the beginning of the explosion. The
672 apparent velocity of the observed TEC perturbations was estimated to be ~1.1 km/s, which is
673 close to the speed of the acoustic and shock-acoustic waves in the ionosphere. The dense
674 coverage of the Japanese network of GPS-receivers GEONET allowed detecting the response
675 as far as 200 km away from the volcano. From the ionospheric GPS-TEC measurements Heki
676 (2006) estimated the energy of the Asama volcano explosion as of 2×10^{14} J.

677 Dautermann et al. (2009a) used data of GPS-receivers located around Guadeloupe,
678 Antigua and the Caribbean to analyze the ionospheric TEC response to the 13 July 2003 VEI=3
679 explosion of the Soufrière Hill Volcano in Montserrat, Lesser Antilles. They observed quasi-
680 periodic TEC oscillations with a period of about 12 min; the TEC response appeared ~18 min
681 after the explosion and lasted for about 40 min (Figure 13a). Overall, the co-volcanic TEC
682 perturbations were registered as far as ~700-800 km, and their propagation speed was
683 estimated to be ~550-650 m/s. Spectral analysis revealed peaks centered at 1 and 4 mHz,
684 which is consistent with the theory, suggesting both gravity and acoustic wave components.
685 Dautermann et al. (2009b) used a normal mode summation technique and showed that the 4
686 mHz peak can be explained by a single explosive source in the atmosphere. However, they
687 struggled to explain the quasi-periodic signal that occurred later. Similarly to Heki (2006),
688 Dautermann et al. (2009a) used TEC-observations to calculate the energy of the volcano
689 explosion: for the Soufrière Hill case it was estimated to be 1.53×10^{10} J.

690 Nakashima et al. (2016) analyzed the TEC response to the Kelud volcano explosion of 13
691 February 2014 in Java, Indonesia. The eruption lasted from ~16:01UT to ~18:00UT. By using
692 ground-based GNSS-receivers from several regional networks in Sumatra and Indonesia, they
693 observed TEC oscillations that commenced at ~16:25 UT and lasted until ~19:00UT. The
694 amplitude spectrum of the TEC data series showed peaks at 3.7 mHz and 4.8 mHz, suggesting
695 that the observed TEC oscillations were due to the coupling between the solid Earth and the
696 lower atmosphere known as the acoustic resonance.

697 The April 2015 eruptions of the Calbuco Volcano in southern Chile received more
698 attention from the scientific community. On 22 April 2015 the Calbuco volcano awoke after 43
699 years of inactivity. The first eruption began at 21:04 UT, lasted 90 minutes and generated a
700 gray ash plume that arose 15 km above the crater. A larger second event began at 04:00UT on
701 23 April, it lasted six hours, and generated an ash plume that rose higher than 15 km. Possibly
702 because of strong winds, the ash plume from this second eruption drifted northward,
703 northeastward, and eastward (Figure 12b). The VEI was estimated 4 for these two events
704 (<http://www.sernageomin.cl>; <http://www.onemi.cl>).

705 The ionospheric TEC response to the Calbuco eruptions was analyzed by several
706 research groups. Shults et al. (2016) used data of ground-based GNSS receivers and observed
707 quasi-periodic ionospheric TEC oscillations following the two eruptions. The TEC response
708 was registered ~15 minutes after the beginning of the first eruption (Figure 13b) and ~40
709 minutes after the second eruption (Figure 13c). Shults et al. (2016) explained such a time
710 delay in ionospheric responses by different source waves emitted by the eruptions. Most
711 likely, the first eruption, being the first in 43 years, was initiated by a shock acoustic wave,
712 and continued by the gravity waves generated by the ash emission. During the second
713 eruption, only ash plume was emitted, and produced such a late response in the ionosphere.
714 Shults et al. (2016) estimated the apparent velocities of the CVID to be in the range of 900-
715 1200 m/s.

716 Atmospheric-ionospheric disturbances generated by the April 2015 Calbuco eruptions
717 were also studied by X. Liu et al. (2017). By using data from 50 GPS stations, they observed
718 TEC disturbances with an amplitude of 0.1–0.4 TECU about one hour after the eruptions. The
719 disturbance lasted more than one hour for the first eruption and about half an hour for the
720 second eruption. The propagation velocities were around 800 m/s and 900 m/s for the two
721 eruptions, respectively. The spectrogram of the filtered TEC time series was centered at 3.7
722 mHz, which is in the frequency range of infrasonic waves.

723 It is interesting to note that, similar to earthquakes, the amplitude of CVID seems to
724 scale with the intensity of volcanic eruptions. Thus, a VEI=2 explosive eruption can generate
725 small TEC perturbations of about 0.1 TECU, while a VEI=4 eruption causes a stronger
726 response in the ionosphere when the TEC perturbations can reach 0.45 TECU. The relative
727 contribution of the CVID into the background TEC for VEI=2 eruptions is ~8%, and reaches
728 15-18% for VEI=4 eruptions (Shults et al., 2016).

729

730

731 **3.3. Ionospheric detection of volcanic eruptions. “Ionospheric Volcanology”**

732 Similarly to the Ionospheric Seismology discussed in P2.3, the term “Ionospheric
733 Volcanology” has been introduced by Shults et al. (2016) to define the use of ionospheric
734 measurements for the interests of volcanology. For instance, to detect an eruptive volcano
735 and/or to determine the time of an eruption’s onset from the ionosphere.

736 Shults et al. (2016) were the first and so far the only to localize an eruption from the
737 ionosphere. To do so, they used a method similar to that proposed by Afraimovich et al.
738 (2005) for the ionospheric detection of earthquakes. This method is based on the
739 approximation of a spherical wave propagating from a point source at a constant speed.

740 This ionosphere-based method allows not only to localize the source but also to estimate
741 the source onset time. Normally, the time of the eruption onset can be calculated from nearly-
742 located seismometers or/and from infrasound stations (e.g., Matoza et al., 2011; Caudron et
743 al., 2015). However, the absence of seismometers within 100 km from a volcano would not
744 allow to estimate the eruption time with high precision. Therefore, an ionosphere-based
745 method can be used in the areas with no seismometers installed.

746 The results from Shults et al. (2016) on the ionospheric localization for both Calbuco
747 eruptions are presented in Figure 14(a-b). One can see that for the first eruption, the source of
748 the ionospheric perturbation can be localized sufficiently well. The position of the eruption
749 was estimated within ± 1 -2 degrees of latitude and/or longitude. From their method, Shults et
750 al. (2016) calculated the time of the first eruption’s onset between 21.07UT and 21.29UT;
751 these ionosphere-based results were close to the onset time calculated from seismometers
752 (21.07UT) and to that from infrasound stations ($21.16\text{UT} \pm 0.12\text{h}$; Matoza et al., 2018).

753 For the second eruption, the ionosphere-based localization method showed the
754 eruption located on the northeast from the volcano (Figure 14b). Shults et al. (2016)
755 suggested that such a discrepancy occurred, most likely, because of the significant impact of
756 horizontal atmospheric winds on vertical propagation of CVID. Otherwise, it can be due to the

757 volcanic plume (and, therefore, the atmospheric source) drift toward north-east as satellite
758 images show (Figure 12b). The ionospherically-estimated source time varied from 4.78 to
759 4.98UT, which is about 45-55 minutes behind the onset time estimated from the
760 seismometers and from the infrasound stations. Shults et al. (2016) explained this delay by
761 “the late occurrence and registration of the CVID after the second eruption”, which was 60
762 min (Figure 13), as compared to 15 min in the first eruption.

763 Therefore, ionospheric measurements can help to detect an eruptive volcano and to
764 estimate the eruption onset. Also, they provide indirect information on the position and drift
765 of the volcanic ash plume. However, more work is needed to improve the existing methods
766 and to obtain more precise results.

767

768

769 **4. Future perspectives**

770 Over the past two decades, we have substantially enlarged our knowledge about
771 ionospheric signatures of NH. Together with recent advances in the ionospheric remote
772 sensing, this opens new exciting perspectives for future applications of ionosphere-based
773 methods for detection of NH. For instance, the seismo-ionospheric imagery by using GNSS
774 data with high temporal and spatial resolution could provide information on the seismic
775 source extent. Furthermore, in the near future, we expect to be able to make an inversion of
776 seismic source parameters from ionospheric observational data. The latter will provide
777 information on the co-seismic crustal uplift, and will bring improvements for tsunami
778 warnings systems. In addition, future satellite missions with on-board airglow cameras will
779 help to track tsunami propagation over the entire oceans before they hit the coasts.

780 However, it should be noted that the development of such ionosphere-based methods
781 still requires better understanding of the Solid Earth/ocean/atmosphere/ionosphere
782 coupling. Along with new observational facilities, the development of new simulation tools is
783 necessary to further advance in this branch of Geophysics. Models help understand the
784 coupling, interpret the observations and forecast the ionospheric response. Within the past
785 two decades, numerous simulation techniques have been developed. For instance, ray-tracing
786 is the simplest, although not the most precise, way to reproduce near-field ionospheric
787 signatures of earthquakes and volcanic eruptions [Calais et al., 1998; Heki and Ping, 2005;
788 Heki et al., 2006; Rolland et al., 2013; Dautermann et al., 2009a; Thomas et al., 2018]. To
789 model CSID propagation in the far field and the ionospheric signatures of Rayleigh surface
790 waves and tsunamis, normal mode summation technique shows promising results [Lognonné

791 et al., 1998; Artru et al., 2001; 2004; Rolland et al., 2011a; Coisson et al., 2015; Rakoto et al.,
792 2018]. Further, to reproduce co-seismic and co-tsunamic signatures with better accuracy,
793 especially those generated by giant earthquakes, more sophisticated physics-based models
794 have been recently developed [e.g., Occhipinti et al. 2006, 2008; Krasnov et al., 2007; Kherani
795 et al., 2009; 2012; Shinagawa et al., 2013; Huba et al., 2015; Chum et al., 2016; Brissaud et al.,
796 2016; 2017; Meng et al., 2015; 2018; Zettergren and Snively, 2015; Zettergen et al., 2017]. For
797 instance, taking into account non-linear processes upon wave propagation in the atmosphere
798 allowed to reproduce the co-seismic ionospheric hole [Zettergen et al., 2017; Shinagawa et al.,
799 2013].

800 It should be pointed out that the development of accurate modeling tools is also crucial
801 for the future application of seismo-ionospheric methods for other planets than Earth, which
802 is another exciting perspective for this branch of geophysics. Seismometers are the
803 traditional instruments used for studying the seismic and volcanic activities of a planet.
804 Whereas, the newly introduced atmospheric/ionospheric seismology/volcanology via remote
805 atmospheric/ionospheric monitoring is a novel way to explore the planetary seismicity and to
806 learn about the formation and evolution of a planet. Addressing these fundamental questions
807 remains so far only a dream for humanity. In November 2018, Mars has finally got the first
808 seismometer with the planetary mission InSight (<https://mars.nasa.gov/insight/>), so that we
809 can expect new scientific discoveries concerning the Red Planet's interior quite soon.
810 However, our knowledge about another telluric planet -Venus- is still quite poor. The
811 situation is complicated by the adverse surface conditions on Venus, with extremely high
812 atmospheric temperature and pressure. The latter makes it impossible to place a seismometer
813 on the surface for an extended period of time. Therefore, for Venus, the remote monitoring of
814 atmospheric perturbations can be the only way to detect signs of seismic/volcanic activity
815 and to unveil the biggest mysteries of the Universe.

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823 can be accessed from the Geospatial Information Authority of Japan (GSI) database via the
824 following service: http://datahouse1.gsi.go.jp/terras/terras_english.html. Data shown in

825 Figure 13b-c are available from the SOPAC (<http://sopac.ucsd.edu>), UNAVCO
826 (<http://unavco.org>), and RAMSAC (<http://www.ign.gob.ar>) online data services. No other
827 data were used in this review. This is IPGP contribution 4086.

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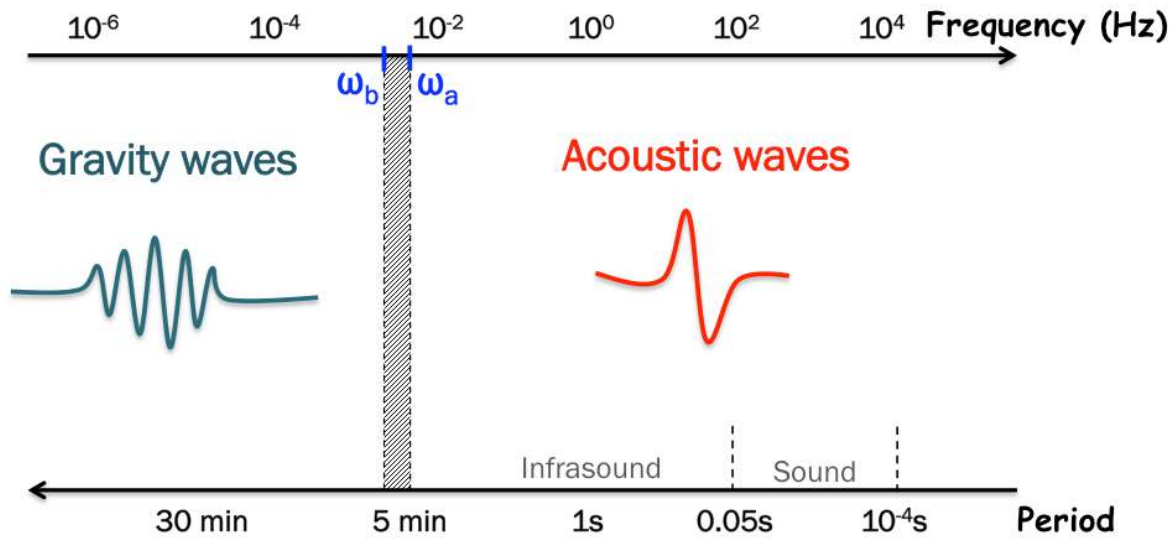
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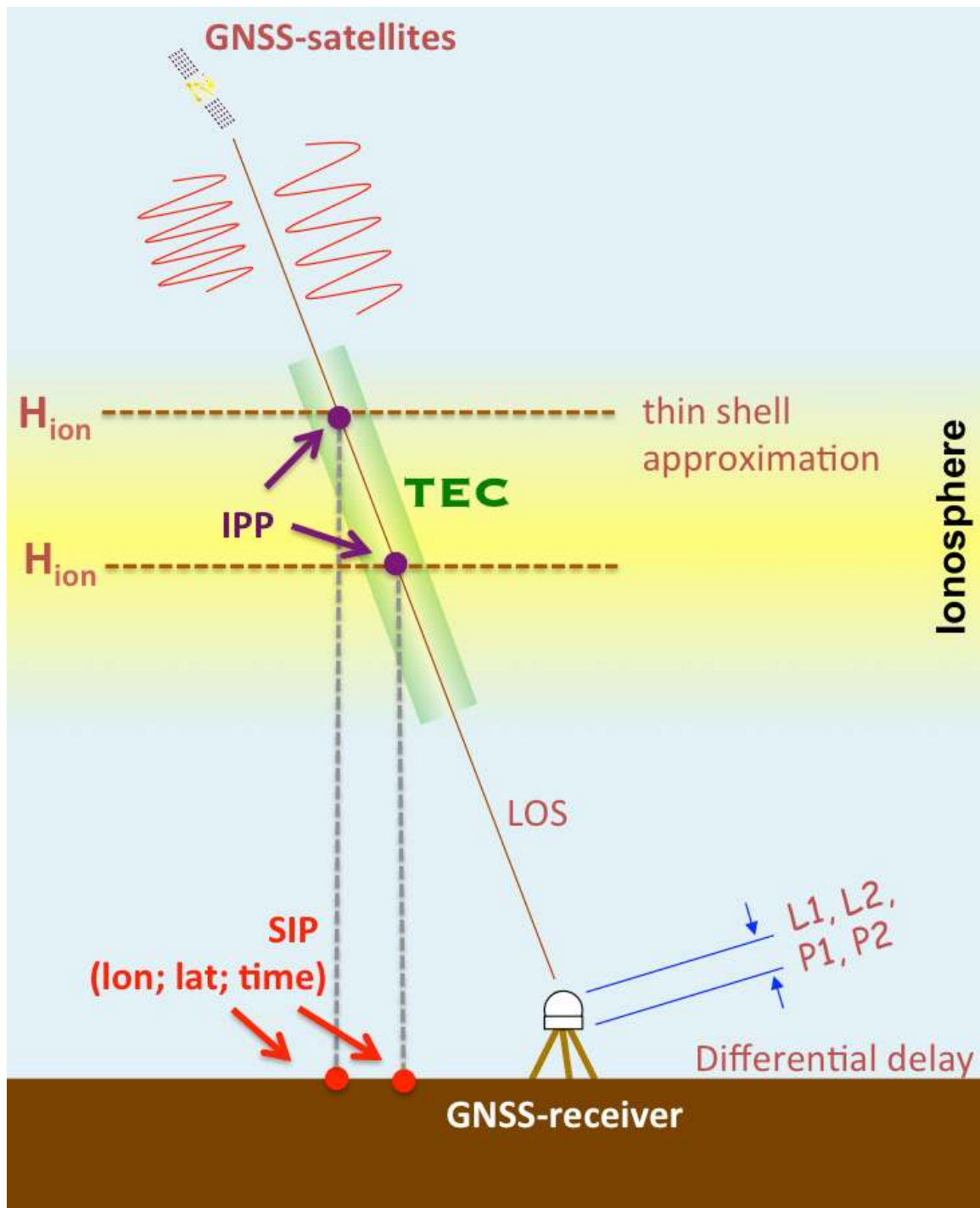
1263 **Figures**



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1265 **Figure 1.** Frequency and period ranges for acoustic and gravity waves. The frequency ω_a is
 1266 commonly called “the acoustic cut-off frequency” indicating that acoustic waves with $\omega > \omega_a$
 1267 propagate through the atmosphere. The frequency ω_b is the Brunt-Väissälä frequency (or
 1268 buoyancy frequency). Waves with frequencies below the ω_b are the gravity waves. In the
 1269 lower atmosphere, the acoustic cut-off frequency is typically 3.3 mHz and ω_b is 2.9 mHz.
 1270 Waves with frequencies $\omega_b < \omega < \omega_a$ (shaded rectangle) are called evanescent waves and can
 1271 only propagate horizontally.

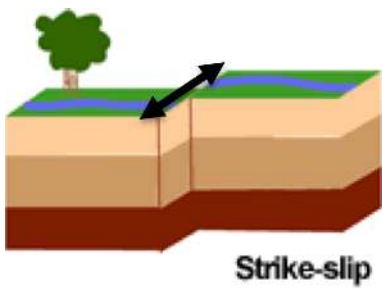
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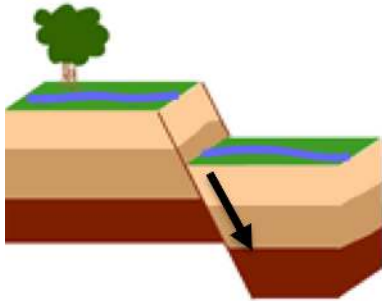
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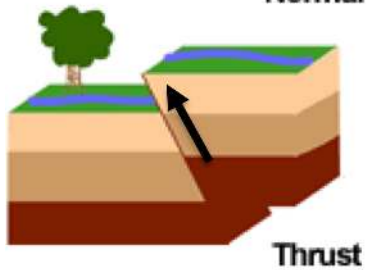
1275 **Figure 2.** GNSS-sounding of the ionosphere. LOS = line-of-sight between a GNSS satellite and
 1276 a GNSS receiver; IPP=Ionospheric Piercing Point, the intersection point between the LOS and
 1277 the altitude of the ionosphere (H_{ion}) within a thin shell approximation; SIP = the projection of
 1278 the IPP on the Earth's surface. Knowing the coordinates of a satellite and a receiver one can
 1279 calculate the coordinates of the SIP. Note that changing the altitude H_{ion} will change the
 1280 coordinates of the SIP.



Strike-slip



Normal

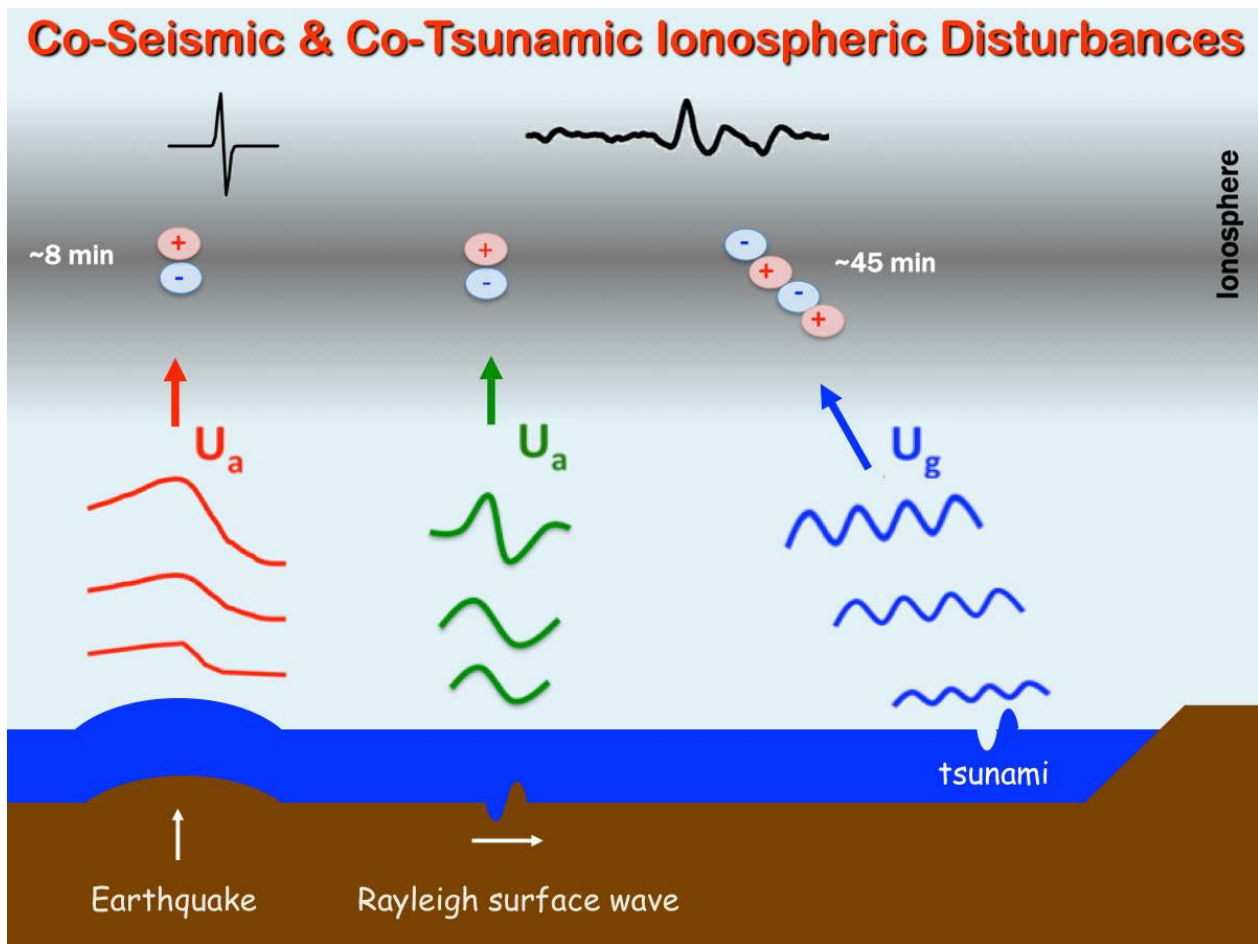


Thrust

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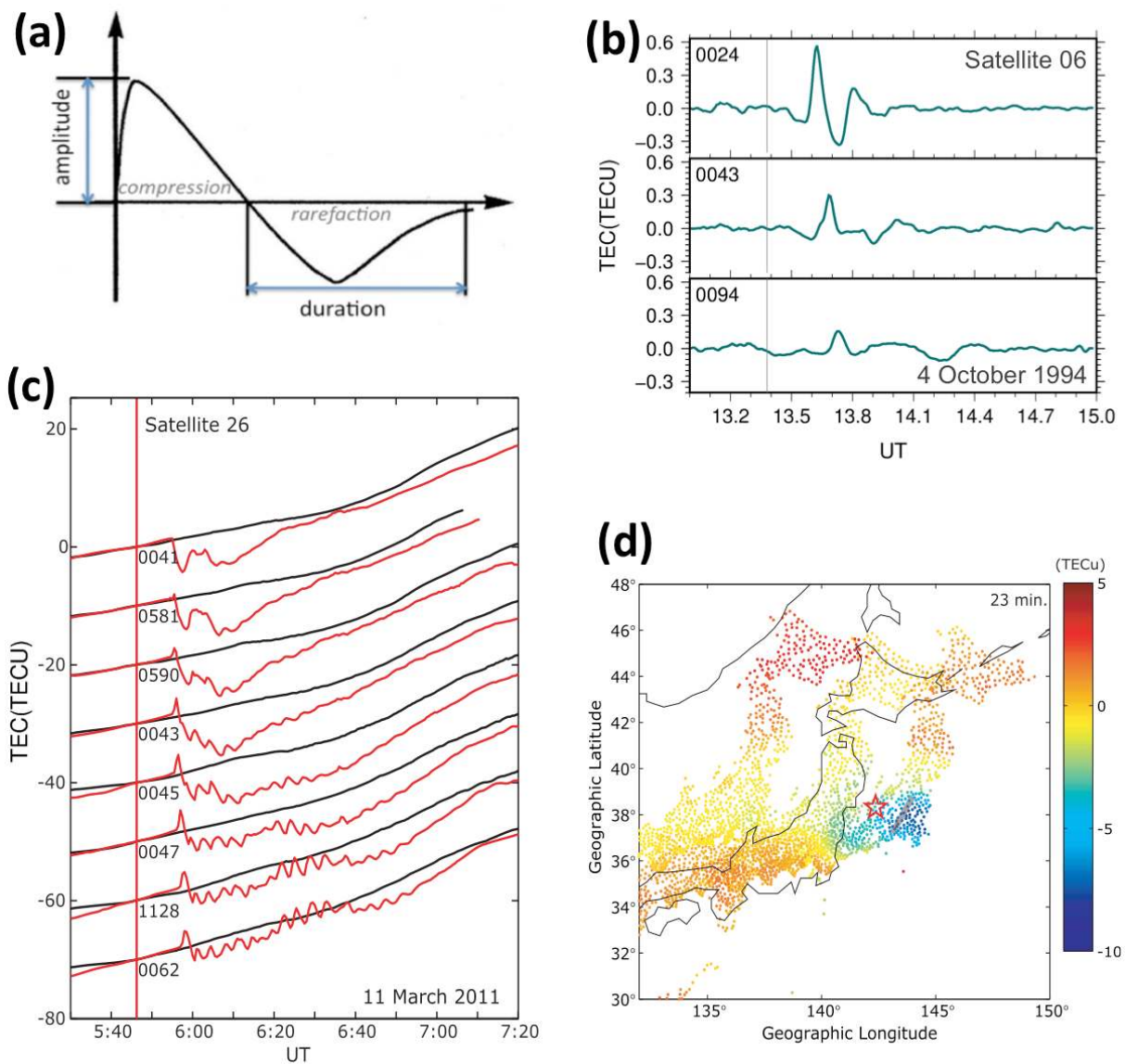
1282 **Figure 3.** Schematic view of the three main types of focal mechanisms: strike-slip (mostly
1283 horizontal motion), normal (mostly subsidence) and thrust (mostly uplift). Black arrows show
1284 the main direction of co-seismic crustal motion. [Image credit: U.S. Geological Survey,
1285 earthquake.usgs.gov.]

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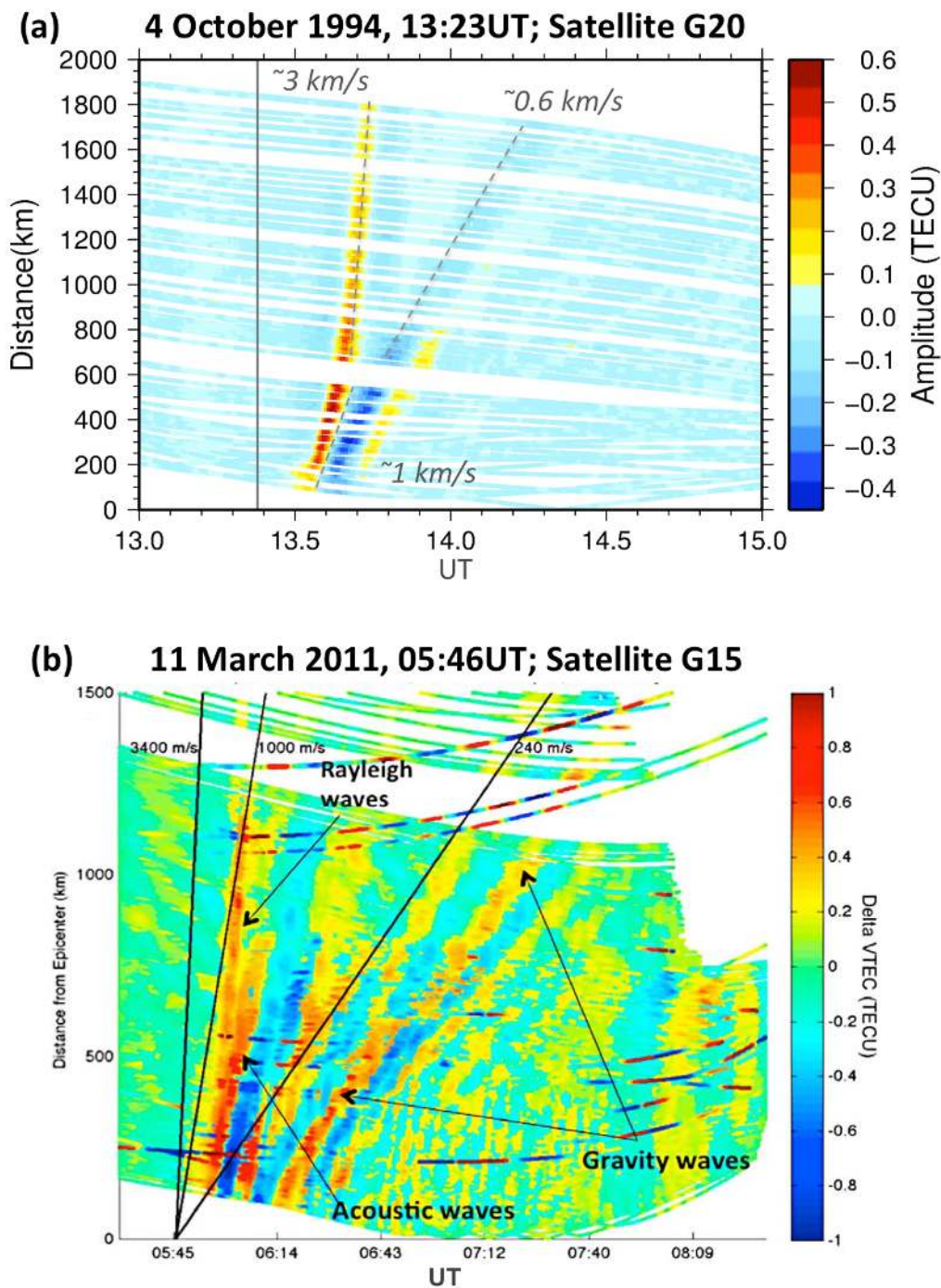
1288 **Figure 4.** Scheme showing the dynamic coupling between co-seismic crustal motion, Rayleigh
 1289 surface waves, tsunami propagation, and the ionosphere. The acoustic waves (U_a) generated
 1290 by an earthquake and by Rayleigh surface waves propagate upward and reach the ionosphere
 1291 in ~ 8 min. Tsunamis generate gravity waves (U_g) that propagate obliquely and reach the
 1292 ionosphere in ~ 45 -60 min. Both acoustic and gravity waves are amplified with the altitude
 1293 because of the exponential decrease of the atmospheric density with height.



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1295 **Figure 5.** Typical waveforms of ionospheric perturbations generated by earthquakes: **(a)**
 1296 theoretical N-wave produced by a piston-like motion due to sudden crustal motion [from
 1297 Astafyeva et al., 2013]; **(b)** example of CSID recorded by GPS-receivers 0024, 0043 and 0094
 1298 after the Mw=8.3 Kuril earthquake of 4 October 1994. Time of earthquake 13.38UT is shown
 1299 by a gray thin vertical line; **(c)** TEC time series for the day of the Tohoku-oki earthquake (red
 1300 curves) and for the reference day (black curve). A vertical solid line indicates the time of the
 1301 mainshock at 5:46UT. Each TEC curve is drawn with shifting by 10 TECU. One can clearly see a
 1302 strong TEC decrease as compared to the reference values; **(d)** Occurrence of the TEC
 1303 depletion (ionospheric hole) detected over the epicentral area 23 min after the Tohoku-oki
 1304 earthquake [Panels (c-d) are from Kakinami et al., 2012].

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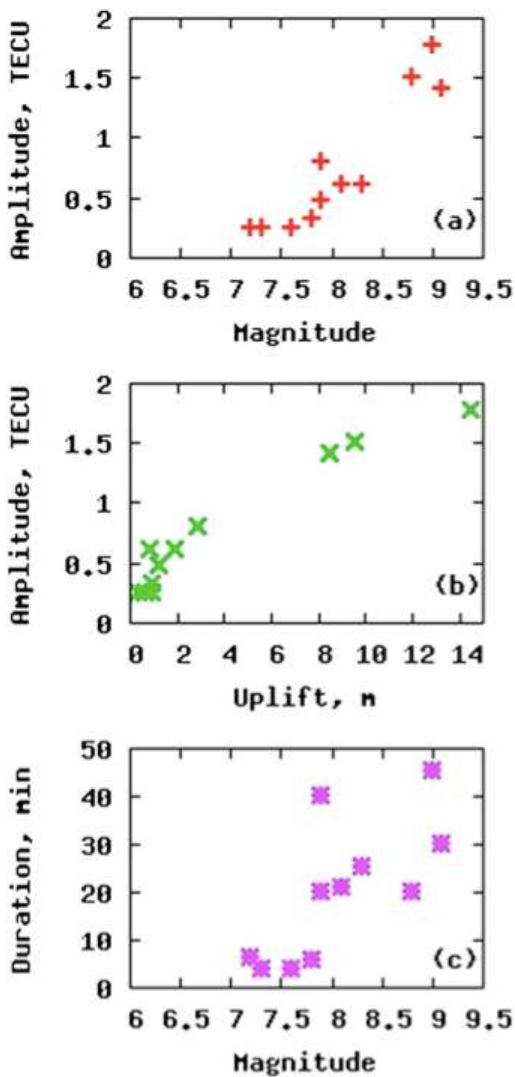
1307 **Figure 6. (a)** Travel-time diagram (TTD) for the case of co-seismic ionospheric perturbations
 1308 following the M8.3 earthquake that occurred on 4 October 1994 in the vicinity of the Kuril
 1309 Islands. The TEC observations are from GPS-satellite G20 and 100 ground-based GPS-
 1310 receivers of the Japanese Network GEONET. The color shows the TEC amplitude, the
 1311 corresponding color scale is shown on the right. In the near-field, we observe one CID with
 1312 propagation speed of ~ 1 km/s, starting from 600 km away from the epicenter, the CID splits
 1313 into two modes, the one (fast mode) continues to propagate at ~ 3 km/s, whereas the other
 1314 (slower mode) goes at 600 m/s; **(b)** TTD showing the occurrence of multi-mode ionospheric
 1315 perturbations observed after the 11 March 2011 Mw=9.1 Tohoku-oki earthquake. The TEC

1316 observations are from satellite G15 and ~1200 GPS-receivers of the GEONET. From the
1317 estimated velocities, we conclude that after the Tohoku-oki earthquake we observe Rayleigh
1318 waves, acoustic waves, gravity waves due to the tsunami. The velocities of these components
1319 are indicated on the panel [Panel (b) is adapted from Galvan et al., 2012]

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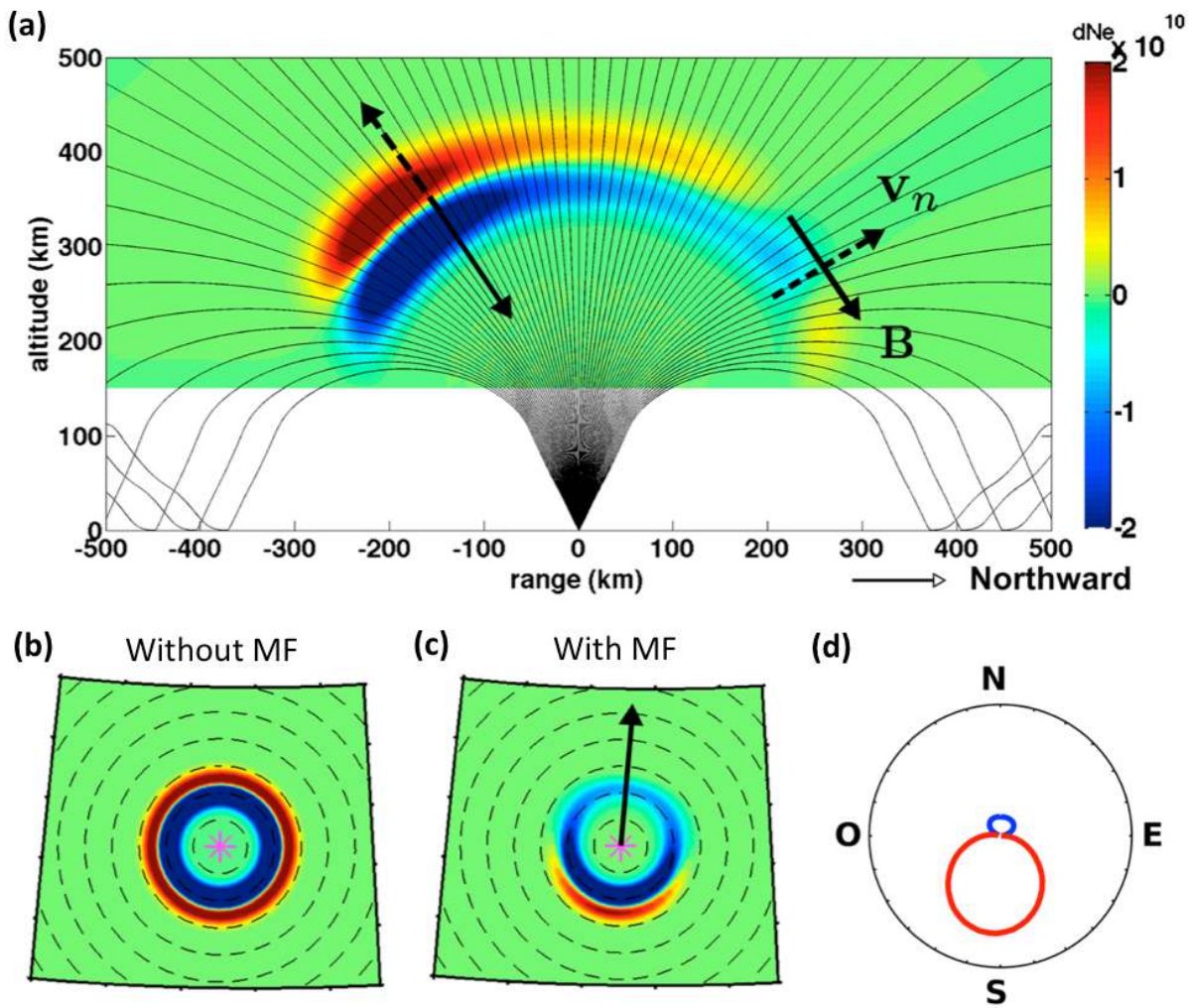


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1324 **Figure 7. (a-b)** Dependence of the amplitude of the near-field CSID on the magnitude of
1325 earthquakes (a) and on the magnitude of co-seismic uplift (b). **(c)** Duration of the negative
1326 phase of the TEC response versus EQ's magnitude. [Adapted from Astafyeva et al., 2013a]

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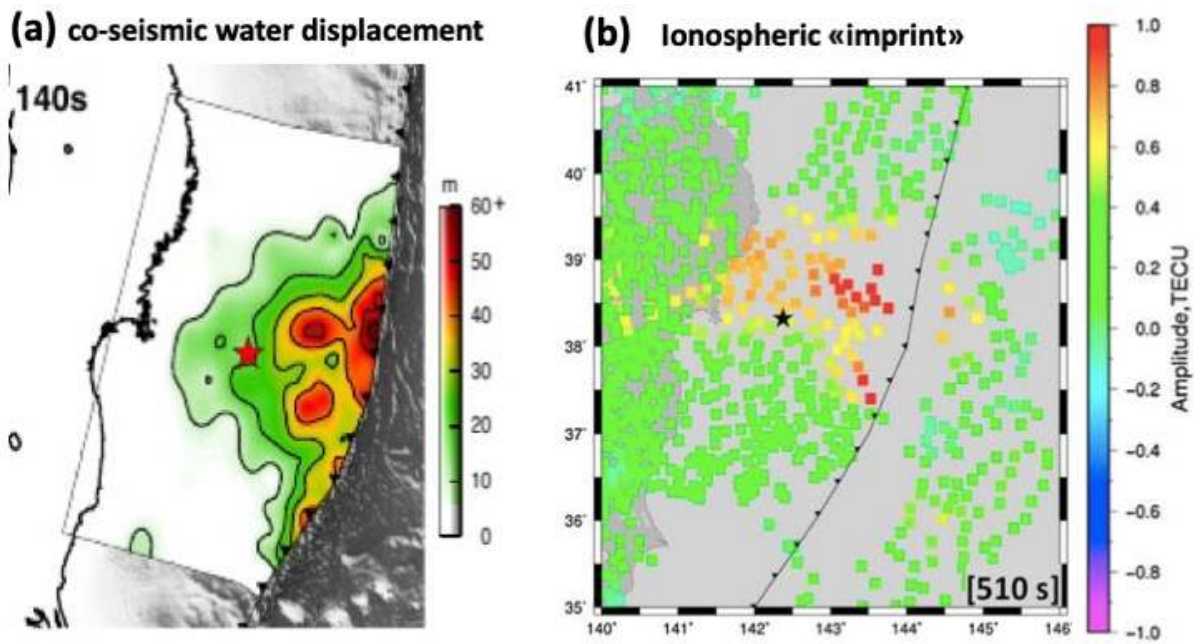
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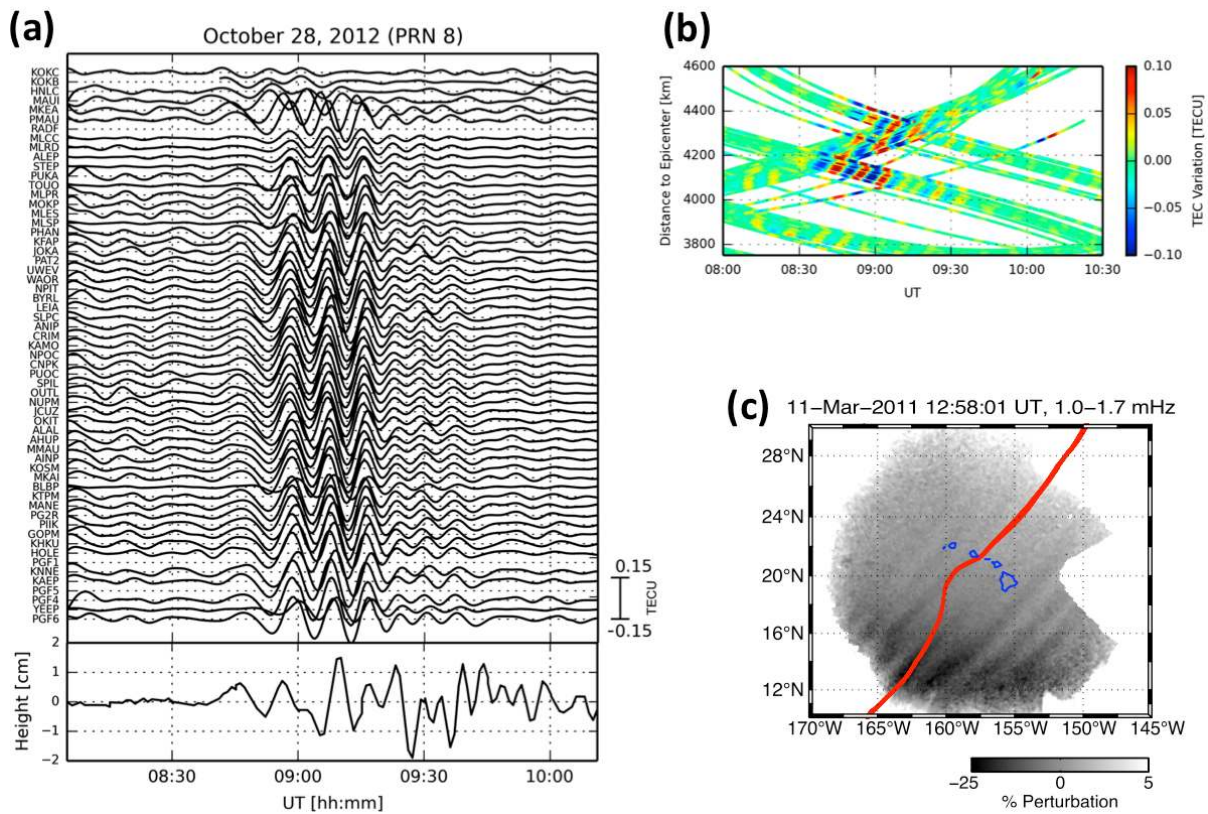
1330 **Figure 8.** Impact of the geomagnetic field on the evolution of co-seismic ionospheric
 1331 disturbances: **(a)** in vertical plane - electron density perturbation (dNe) 13 minutes after the
 1332 earthquake onset time. Solid black arrow indicates the magnetic field vector **B**, dotted arrow
 1333 shows the neutral perturbation velocity; **(b-c)** in horizontal plane - electron density
 1334 perturbation at 280 km height without **(b)** and with the geomagnetic field effect **(c)**. The
 1335 epicenter is located at the purple star, and the dashed circles show epicentral distance isolines
 1336 at 100 km intervals; **(d)** Ionospheric radiation pattern, computed at 280 km height and 200
 1337 km of epicentral distance. Blue is negative polarity, and red is positive polarity. It should be
 1338 noted that this example is valid for the Northern Hemisphere. In the Southern Hemisphere,
 1339 the magnetic field will “favor” the northward propagation. [Adapted from Rolland et al., 2013,
 1340 modified]

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 1343 **Figure 9. (a)** Model of co-seismic displacements of the water column occurred due to the
 1344 Tohoku-oki earthquake of 11 March 2011. Color shows the amplitude of the coseismic-slip,
 1345 red star – the epicenter. [From Bletery et al., 2014]; **(b)** TEC perturbations above the near-
 1346 epicentral region of the 11 March 2011 Tohoku-oki earthquake, a snapshot at 510 sec after
 1347 the earthquake. Color indicates the TEC value, and the color scale is shown on the right. Black
 1348 star indicates the epicenter of the Tohoku-oki earthquake [From Astafyeva et al., 2013b].

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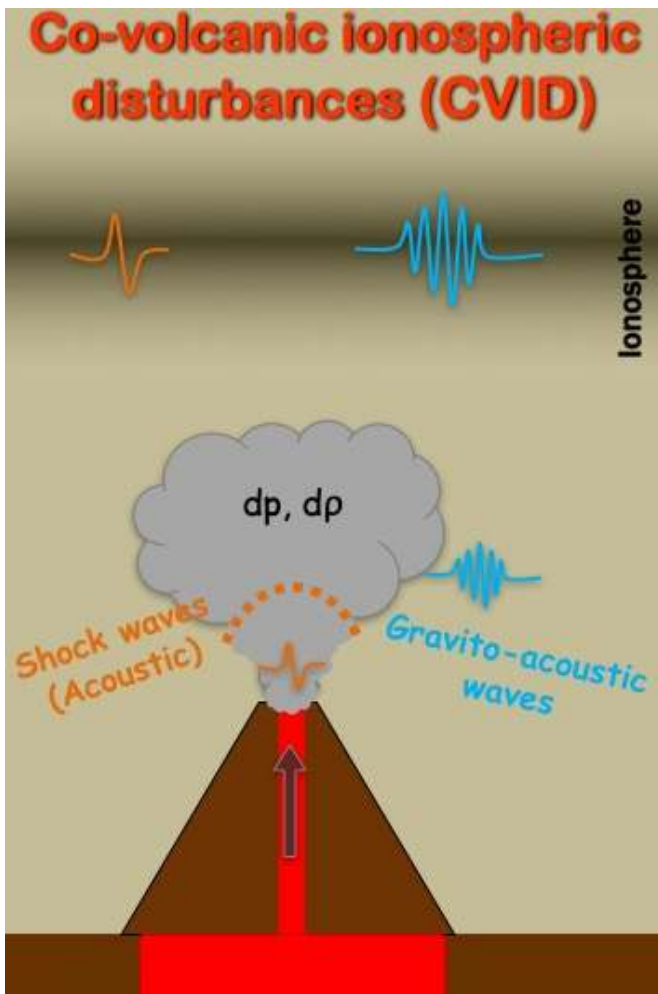
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Figure 10. Examples of co-tsunami ionospheric and atmospheric disturbances: **(a)** Ionospheric TEC-response to the propagation of the Haida Gwaii tsunami of 28 October 2012 as measured by GPS-satellite PRN08 (top). Names of GPS-receivers are indicated on the left. The tsunami was generated by the earthquake with an epicenter near Moresby Island, Canada, and here was detected over the Hawaii Islands (~4000 km away). Comparison of the TEC with tidal gauge from DART station 51407 (bottom) suggests the tsunami as the origin of the perturbation. **(b)** Travel-time diagram showing quasi-periodic TEC oscillations in response to the Haida Gwaii tsunami propagation in Hawaii. [Panels (a) and (b) are adapted from Grawe and Makela, 2015]; **(c)** The 11 March 2011 Tohoku tsunami-driven signatures in 630 nm airglow over Hawaii. Images are processed using filtering with passband 0.1-1.7 mHz to highlight the 14.2-min period waves. The red line indicates the tsunami location at the time of the image. [Adapted from Makela et al., 2011]

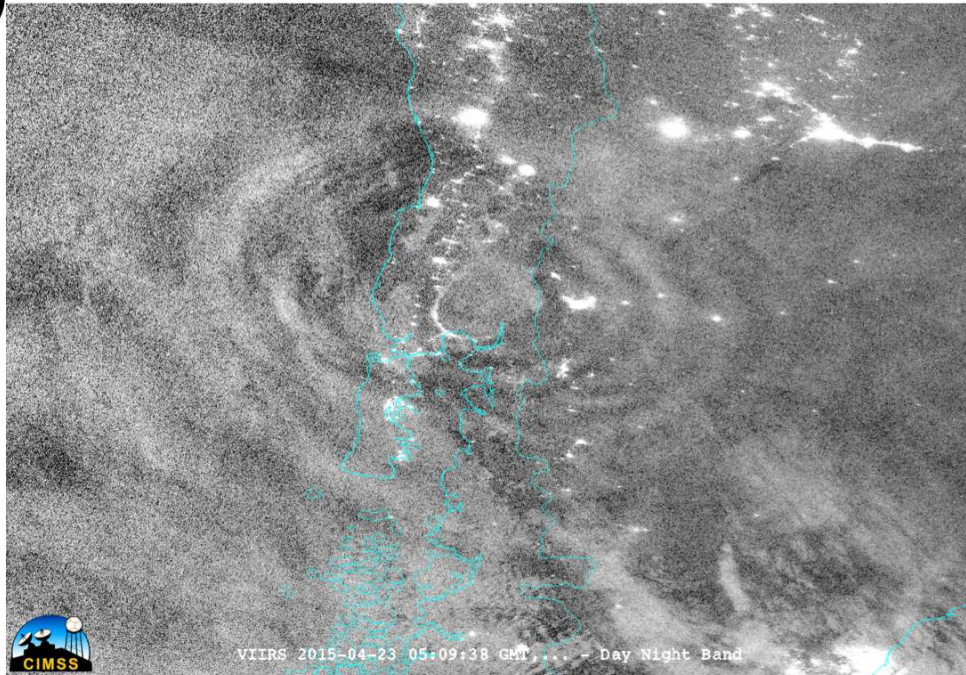


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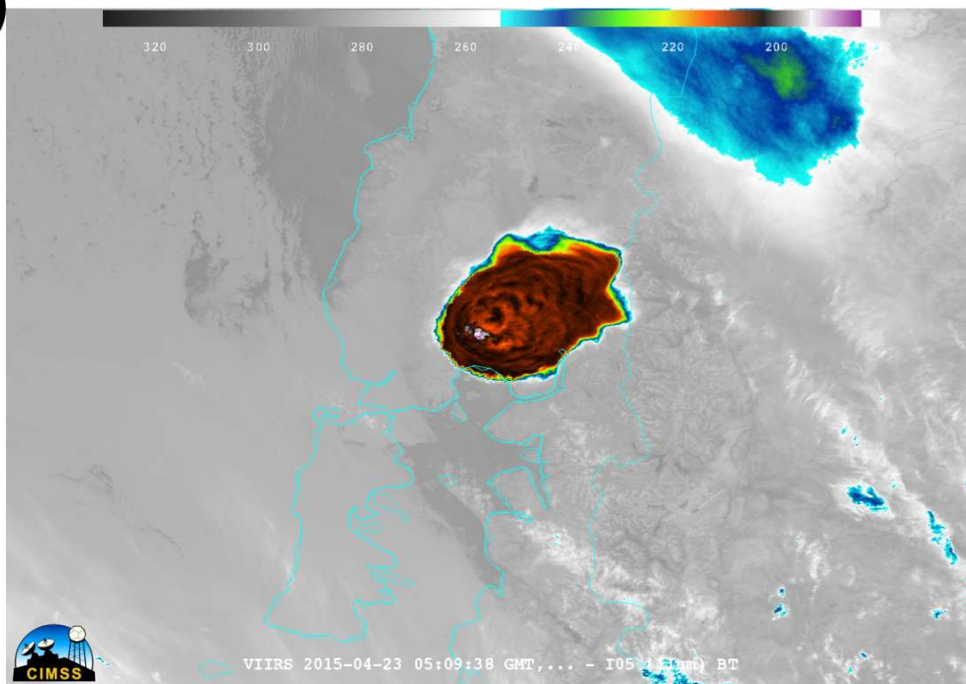
1368 **Figure 11.** Scheme showing the generation of ionospheric disturbances by a volcanic
 1369 eruption. First, shock-acoustic waves may be generated, they arrive in the ionosphere at
 1370 starting from ~8-9 min after the beginning of an eruption. Second, with eruption and sudden
 1371 changes in the local pressure (dp) and density ($d\rho$), gravito-acoustic waves are generated. The
 1372 latter waves usually reach the ionospheric altitudes ~45 min after the eruption.

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(a)



(b)



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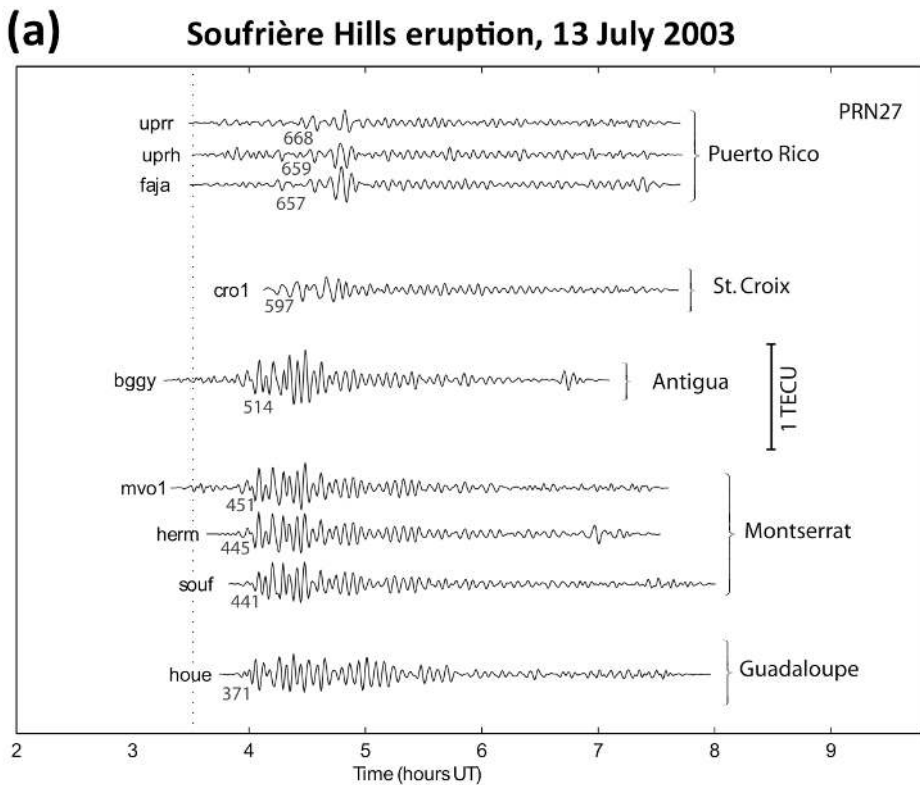
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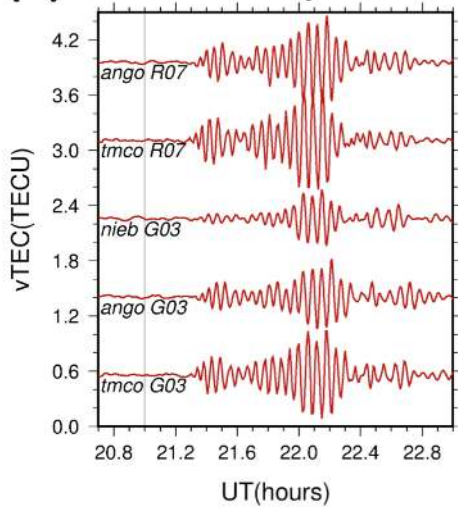
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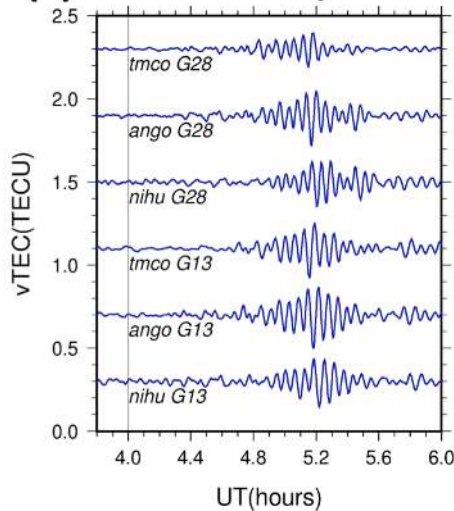
Figure 12. (a) Mesospheric gravity waves associated with the Calbuco Volcano Eruption of 23 April 2015. The image taken by the Suomi NPP VIIRS Day/Night Band imagery (Miller et al., 2012) at 5:09UT, i.e. one hour after the eruption onset [Image credit: CIMSS, University of Wisconsin – Madison]; **(b)** Infrared imagery from VIIRS 11.45 μm showing the volcanic plume drifting north-eastward from the volcano [Image credit: CIMSS, University of Wisconsin – Madison; <http://cimss.ssec.wisc.edu/goes/blog/archives/18174>].



(b) Calbuco eruption #1



(c) Calbuco eruption #2



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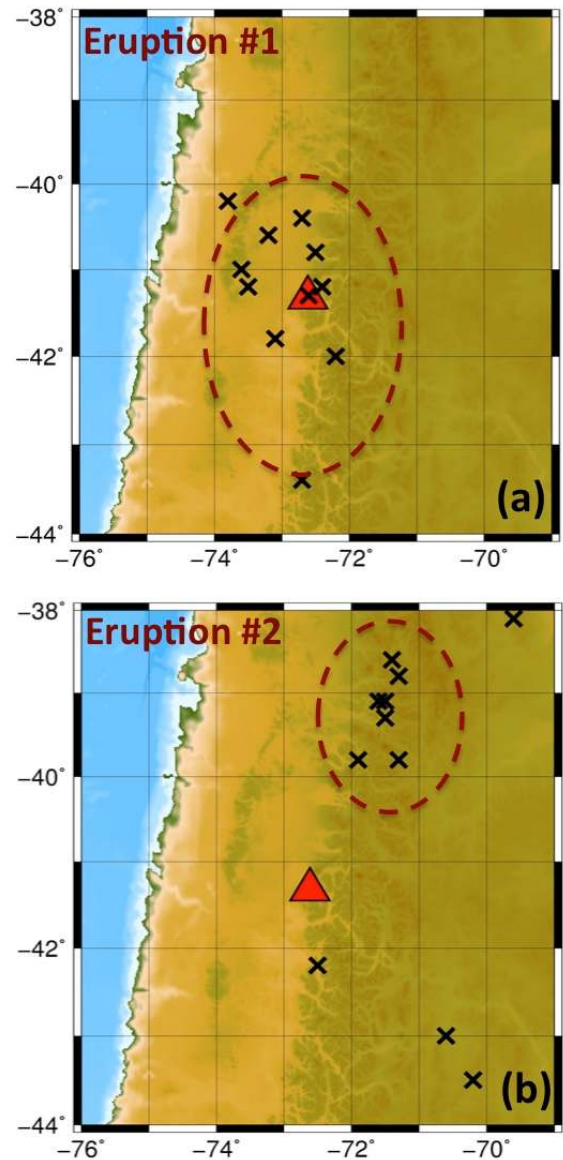
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Figure 13. Examples of co-volcanic ionospheric disturbances: **(a)** TEC response to the Soufrière Hill volcano explosion of 13 July 2003 as registered by GPS-satellite G27 (PRN27). The TEC signals are filtered between 2.2 and 8 mHz, the names of GPS-stations are marked on the left of the TEC data series, numbers show the distance in km from the volcano. Vertical dotted line indicates the time of the eruption. [Adapted from Dautermann et al., 2009a]; **(b, c)** Ionospheric TEC response to the Calbuco volcano eruptions of 22 April 2015 (b) and of 23 April 2015 (c). The vertical gray line shows the first eruption onset at 21:04UT and the second eruption at 4:00UT. Names of GNSS-stations and of GNSS-satellites are shown next to each curve.



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 1393 **Figure 14.** Localization of the eruptions of the Calbuco volcano from the ionosphere: **(a)** – for
 1394 the first eruption of 22 April 2015, **(b)** – for the second eruption of 23 April 2015. Red triangle
 1395 shows the position of the volcano, black crosses – solutions for ionospheric detection of the
 1396 eruptions from ionospheric TEC measurements. (The localization results are from Shults et al.,
 1397 2016).