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Strike-slip earthquakes can also be detected in the ionosphere

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A B S T R A C T

It is generally assumed that co-seismic ionospheric disturbances are generated by large vertical static displacements of the ground during an earthquake. Consequently, it is expected that co-seismic ionospheric disturbances are only observable after earthquakes with a significant dip–slip component. Therefore, earthquakes dominated by strike-slip motion, i.e. with very little vertical co-seismic component, are not expected to generate ionospheric perturbations. In this work, we use total electron content (TEC) measurements from ground-based GNSS-receivers to study ionospheric response to six recent largest strike-slip earthquakes: the Mw7.8 Kunlun earthquake of 14 November 2001, the Mw8.1 Macquarie earthquake of 23 December 2004, the Sumatra earthquake doublet, Mw8.6 and Mw8.2, of 11 April 2012, the Mw7.7 Balochistan earthquake of 24 September 2013 and the Mw 7.7 Scotia Sea earthquake of 17 November 2013. We show that large strike-slip earthquakes generate large ionospheric perturbations of amplitude comparable with those induced by dip–slip earthquakes of equivalent magnitude. We consider that in the absence of significant vertical static co-seismic displacements of the ground, other seismological parameters (primarily the magnitude of co-seismic horizontal displacements, seismic fault dimensions, seismic slip) may contribute in generation of large-amplitude ionospheric perturbations.

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1. Introduction

Vertical co-seismic crustal displacement is a standard feature of earthquakes dominated by reverse motion (mainly causing uplift at the surface) or normal motion (mainly causing subsidence at the surface). But the third type of characteristic fault motion, strike-slip motion (SSM) leads to predominant horizontal motion and thus generates much less vertical motion at the Earth's surface. Here we prefer the term "strike-slip motion" to "strike-slip fault" (SSF), as several of the events have their strike-slip motion on faults with dipping fault, such as the 2013, Mw7.7, Balochistan earthquake (Avouac et al., 2014; Jolivet et al., 2014).

It is generally assumed that co-seismic vertical motion of the ground can generate perturbations in the ionosphere, referred to as co-seismic ionospheric disturbances (CID): the sudden impulsive forcing from the ground or sea surface generates atmospheric pressure waves that propagate upward into the atmosphere and the ionosphere, where they are detectable by using ionospheric monitoring techniques (e.g., Calais and Minster, 1995; Afraimovich et al., 2001, 2010; Liu et al., 2010, 2011; Rolland et al., 2011a, 2011b, 2013; Astafyeva et al., 2013a, 2013b; Cahyadi and Heki, 2013). This piston-like vertical component of co-seismic crustal deformations appears to be the decisive forcing component, as it serves as a source of the primary acoustic waves. When the fault is overlain by a mass of water, the same vertical displacements of the ground/seafloor are responsible for the generation of tsunamis (under the hypothesis that water is incompressible). When the bathymetry above the fault is significantly inclined, the horizontal motion of the substrate can also contribute to the formation of the tsunami (Tanioka and Satake, 1996; Hooper et al., 2013). However, in most cases, this effect is much smaller than the strict vertical motion as it requires significant horizontal motion of large areas with steep bathymetry: horizontal motion of a seamount will only generate small and localized vertical motion not able to generate a tsunami.

The first work that paid attention to possible differences in the ionospheric response to earthquakes with different focal mechanisms, was a work by Astafyeva and Heki (2009): their analysis of the total electron content (TEC) response to three earthquakes with thrust and normal fault earthquakes showed that, in some cases, the polarity of co-seismic TEC perturbation matched the polarity of the ground motion. However, Astafyeva and Heki (2009) did not investigate earthquakes with the SSF focal mechanism.

To our knowledge, Perevalova et al. (2014) were the first to mention on the ionospheric response to SSF earthquakes. Using

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data from several IGS GPS-receivers located in Indonesia, they analyzed the far-field TEC response to three large SSF earthquakes. Their preliminary analysis showed a tendency of TEC response to SSF events to be of lower amplitude than that to earthquakes with the vertical component of the ground motion. Earlier, without emphasizing the focal mechanisms of earthquakes, Afraimovich et al. (2001) analyzed the near-field TEC response to two SSF earthquakes in Turkey in 1999. The TEC response to the both earthquakes in Turkey was N-shaped, as usually reported for CID. The first event, the Mw7.6 Izmit earthquake of 17 August 1999, was an SSF earthquake with latitudinal and meridional fault planes, and caused a TEC perturbation of 0.14 TECU in the near-field, which propagated with a horizontal velocity of 1.3 km/s. The second event, the Mw7.1 Düzce earthquake of 12 November 1999, was a normal fault quake with a significant strike-slip component. It produced a smaller TEC perturbation with near-field amplitude of 0.08 TECU, that propagated at 1.5 km/s (Afraimovich et al., 2001). One more past SSF example is the Mw7.9 Denali earthquake of 3 November 2002. While the near-field response seemed not to be detected due to the absence of GPS-receivers around the epicenter, Ducic et al. (2003) by use of data of Californian GPS Network observed TEC perturbations ∼3000 km away from the epicenter. Triggered by propagating Rayleigh surface waves, the CID had peak-to-peak amplitude of 0.05 TECU and propagated with a horizontal velocity of about 3.5 km/s.

Hence, despite a few observations, the specific analysis of the ionospheric near-field TEC response to SSF events has not been conducted yet. In this study, we analyze co-seismic ionospheric variations after the largest SSF quakes of 2001–2013, and in particular we investigate the compatibility of these observations with the assumption that CID are mostly generated by piston-like motion of areas where vertical static displacements are maximum (as suggested by Heki and Ping, 2005; Astafyeva and Heki, 2009; Cahyadi and Heki, 2013): if piston-like motion linked to static uplifts is the main source of CID, then SFF should generate much smaller CID for SSF events. Consequently, we pay special attention to the seismological parameters controlling the vertical and horizontal crustal displacements, such as the scaling laws relating the fault dimension to the average slip amplitude and the magnitude of the earthquake (Wells and Coppersmith, 1994). This work has important implications for the development of ionospheric seismology, but since tsunamis are triggered primarily by the vertical static displacement of the seafloor, it is also critical for the use of ionospheric measurements in tsunami early-warning system (Najita et al., 1974; Astafyeva et al., 2011, 2013b; Makela et al., 2011; Occhipinti et al., 2013).

2. GNSS-sounding of the ionosphere. Data processing

Ground-based GNSS (Global Navigation Satellite Systems such as GPS, Glonass or Galileo) observations offer a powerful method for remote sensing of the ionosphere. By computing the differential phases of code and carrier phase measurements recorded by the ground-based dual-frequency GNSS receivers, it is possible to calculate the ionospheric TEC. Methods to compute TEC have been described in detail in a number of papers (e.g., Afraimovich et al., 2001; Heki and Ping, 2005, and references therein). For convenience, TEC is usually measured in TEC units with 1 TECU = 10^{16} electrons m^{-2}.

Since the TEC is an integral parameter, the observed ionospheric disturbance accounts for a large range of altitudes. However, it is generally assumed that the main contribution to TEC variations appears around the height of the maximum of the ionosphere ionization (F2 layer). This allows us to consider the ionosphere as a thin layer located at the H_{ion} height of the ionospheric F2 layer. TEC then represents a point of intersection of a line-of-sight with this thin layer. We represent the propagation of CID by sub-ionospheric points (SIP), that is the projection of the ionospheric piercing points at the Earth's surface. In this paper we assumed H_{ion} at 300 km of altitude. Then, because low elevation angles tend to enlarge the horizontal extent of the ionospheric region, we used only data with elevations higher than 10°. We converted the slant TEC to vertical TEC by using Klobuchar's formula (Klobuchar, 1986). Finally, to eliminate variations of the regular ionosphere, we first smooth the initial TEC data series running a moving average over time windows of 3–4 min and then remove linear trends by applying a moving average with time window of 15–18 min. This procedure works as a band-pass filter to extract variations with periods 3–18 min.

3. Strike-slip fault earthquakes and ionospheric TEC response

For our analysis, we have chosen six large shallow SSF earthquakes with permanent GPS-stations installed ∼1000 km around the epicenter (Table 1, Fig. 1). Below we describe their main seismological characteristics and the associated ionosphere response. To study ionospheric response to earthquakes, we use GPS-measurements of TEC derived from data of the ground-based GPS-network GeoNet located in New Zealand (data are available from ftp://ftp.geonet.org.nz), the SUGAR network in Sumatra, Indonesia (ftp://eos.ntu.edu.sg), as well as IGS stations LHAS, LHAZ, JASK and YIBL (ftp://garner.ucsd.edu) and station KEPA which has been recently installed in South Georgia Island (www.unavco.org).

3.1. The November 2001, Mw7.8, Kunlun earthquake, China

The first event, the 2001 Kunlun earthquake, also known as the 2001 Kokoxili earthquake, occurred on 14 November 2001 at 09:26:55 UT, with an epicenter near Kokoxili, in the Qinghai province of China (Figs. 1 and 2a). With a magnitude of Mw7.8 (USGS-NEIC), this earthquake was associated with the longest surface rupture ever recorded on land, ∼450 km (Klinger et al., 2005). The rupture began on a relatively small strike-slip fault segment at the western end of the Kunlun fault in the region of the mountain Buka Daban Feng. The rupture propagated to the east via an extensional stepover before following the main strand of the Kunlun fault. The region of co-seismic deformation is unusually large for an earthquake of magnitude Mw7.8, with significant faulting being observed up to 60 km from the main rupture trace (Liu and Haselwimmer, 2006). An analysis of the propagation speed indicated that the rupture propagated at a normal velocity along the original segment, but increased in velocity to above the S-wave velocity after the jump across the extensional stepover and continued at that speed until propagation stopped (Bouchon and Vallee, 2003).

The computed co-seismic surface displacements due to the Kunlun earthquake are shown in Fig. 2b. Using the aforementioned seismic source parameters (also shown in Table 1), rigidity of 40 GPa, and modeling the fault as a dislocation in an elastic half-space (Okada, 1992), we estimate the maximum uplift at the source of 0.6 m, and maximum of the horizontal displacements of 0.44 m (arrows in Fig. 2b).

Unfortunately, during the time of the earthquake, no GPS-receivers were installed in the vicinity of the epicenter. The closest stations LHAS and LHAZ were located within several hundreds of meters from each other and at ∼700 km on the south from the epicenter (Fig. 2a). For station LHAZ, co-seismic perturbations were recorded in data of satellites PRN27 and PRN31 (Fig. 1c), at 635 and 622 km from the epicenter, respectively. For station LHAS, only measurements of satellite PRN31 were available and showed CID signature. The amplitude of the TEC response was quite large, it reached 1 TECU for PRN31 and 0.6 TECU for measurements of

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