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Dynamic seismic response of a stable intraplate region to a megathrust earthquake

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ABSTRACT

The 2011 M9.0 Tohoku-Oki megathrust earthquake produced strong ground motions with a peak ground acceleration of 1.52 cm/s ^{2} in the Korean Peninsula, inducing large dynamic-stress changes in the crust. Sixty-one triggered earthquakes with magnitudes of 0.5–2.5 including 17 unusual, spatially-clustered events were identified from continuous seismic record sections of dense seismic networks for 18 h following the megathrust earthquake. The triggered earthquakes occurred in regions of high seismicity. The earthquake occurrence frequency increased after the megathrust earthquake, keeping the Gutenberg-Richter *b* value invariant. The seismic occurrence rates decreased with time following a modified Omori law. The focal mechanism solutions of triggered earthquakes are consistent with those observed before the megathrust earthquake. An unusual earthquake swarm displays apparent migration with a speed of 28.6 ± 4.4 m/h. The changes in pore fluid pressure induced by strong seismic waves may have caused the triggering of earthquakes. The duration and strength of dynamic triggering may be highly dependent on the magnitude and distance of megathrust earthquake.

1. Introduction

Earthquakes can occur as a consequence of stress change in a medium. Coseismic and postseismic deformation produces a static stress change that depends on the fault geometry and distance (King et al., 1994; Stein, 1999; Kilb et al., 2002; Parsons et al., 2008). It was reported that a static stress change at a distance less than a few fault lengths can induce earthquakes, and a strong positive correlation between static stress change and post-event seismicity was found (Freed, 2005). The static stress change could be quantified through the change in a Coulomb failure function (e.g., Reasenberg and Simpson, 1992). The seismicity generally increases in a region of static-stress increase, and it decreases in a region of static-stress decrease (stress shadows) (e.g., Reasenberg and Simpson, 1992; Freed and Lin, 2001). Earthquake triggering caused by a static stress change is observed in various environments including plate boundaries, volcanic and geothermal areas and stable intraplate regions (e.g., Stein et al., 1992; Bohnenstiehl et al., 2002; Asano et al., 2011).

Strong ground motions can produce a large dynamic stress change in the medium (e.g., Tibi et al., 2003; Freed, 2005; Hill and Prejean, 2007), inducing subcritical crack growth (Atkinson, 1984), changes in medium properties (Johnson and Jia, 2005; Parsons, 2005), and pore-pressure change (Hill et al., 1993; Brodsky et al., 2003). The dynamic stress change is responsible for the remote triggering of earthquakes in regional and teleseismic distances in which the static stress change is not effective (e.g., Hill and Prejean, 2007; Gonzalez-Huizar et al., 2012). The dynamically triggered earthquakes have typically small magnitudes (Hill and Prejean, 2007; Parsons and Velasco, 2011). The dynamic triggering of earthquakes lasts for a few minutes to several weeks (Knopoff and Gardner, 1972; Hill and Prejean, 2007). The dynamic triggering is not only observed in active tectonic regions, but also in stable regions (e.g., Hill et al., 1993; Brodsky et al., 2000; Hough, 2001; Hough et al., 2003; Saccorotti et al., 2013; Tibi et al., 2003; Pankow et al., 2004; Miyazawa and Mori, 2005; West et al., 2005; Velasco et al., 2008; Jiang et al., 2010; Yukutake et al., 2013).

A megathrust earthquake produces large coseismic and postseismic lithospheric deformations, incorporating a large number of aftershocks (e.g., Takahashi, 2011; Toda et al., 2011; Lee and Hong, 2014). Large megathrust earthquakes triggered earthquakes over the globe in various tectonic environments (e.g., Hough et al., 2003; Gomberg et al., 2004; Velasco et al., 2008; Peng et al., 2010; Gonzalez-Huizar and Velasco, 2011; Wu et al., 2011; Miyazawa, 2011; Yukutake et al., 2011; Gonzalez-Huizar et al., 2012). The earthquake triggering temporally increases the seismicity, which is particularly crucial for the assessment of seismic hazard potentials in intraplate regions with long earthquake-recurrence intervals. We investigate the properties of dynamically triggered earthquakes around the Korean Peninsula after the 2011 M9.0 Tohoku-Oki earthquake.

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2. Geology and seismicity

The Korean Peninsula is located on the far-eastern Eurasian plate, and belongs to a stable intraplate regime (Fig. 1). The current shapes of the Korean Peninsula and East Sea (Sea of Japan) were formed after consecutive tectonic evolutions of a continental collision between the North and South China blocks from the late Permian to the Jurassic and after continental rifting from the Oligocene to the mid-Miocene (Jolivet et al., 1994; Chough et al., 2000; Oh, 2006). The surface of the Korean Peninsula is composed of three Precambrian massifs (Nangrim, Gyeonggi, and Yeongnam), two Permo-Triassic orogenic belts (Imjingang and Okcheon), two Paleozoic sedimentary basins (Pyeongnam and Taebaeksan), and a Cretaceous volcanic-sedimentary basin (Gyeongsang) (Fig. 1(b)) (Choi, 1986; Chough et al., 2000; Oh, 2006; Ernst et al., 2007). The orogeny from the late Paleozoic to the Mesozoic formed NE-trending geological structures (Chough et al., 2000). The geological and tectonic structures present distinct seismic properties (Kang and Shin, 2006; Hong and Kang, 2009; Choi et al., 2009; Hong, 2010; Jo and Hong, 2013).

The crust of the inland peninsula and Yellow Sea presents the properties of continental crust with a thickness of 28–38 km (Pasyanos et al., 2006; Chang and Baag, 2006; Hong et al., 2008). On the other hand, the East Sea displays a transitional structure between continental and oceanic crusts (Hirata et al., 1989; Kim et al., 1998; Sato et al., 2006; Hong, 2010; Kulinich and Valitov, 2011). The Japan basin in the central East Sea is composed of oceanic crust with a thicknesses of 8.5–14 km. The tectonic setting around the Korean Peninsula forms the ambient stress field of ENE-directional compression and SSE-directional tension (Choi et al., 2012).

The seismicity is distributed over the peninsula with some regional localizations in the southern and northwestern peninsula and the offshore regions of the southeast and west coasts (Choi et al., 2012; Hong and Choi, 2012; Houng and Hong, 2013). The largest magnitude of events observed around the peninsula since 1978, the year in which national seismic monitoring began, was M_L5.3. The largest instrumentally recorded event occurred at the northwestern peninsula in 1952, the magnitude of which was M_S6.5 (Engdahl and Villasenor, 2002). Historical earthquake records suggest that there were several earthquakes with magnitudes of ~7.0 in the last 600 years (Houng and Hong, 2013). Strike-slip earthquakes are the most dominant around the peninsula. Normal-faulting earthquakes occur around the paleo-continental-collision region in the Yellow Sea (Hong and Choi, 2012). We observe thrust events in the paleo-rifting region off the east coast of the peninsula (Choi et al., 2012). The earthquakes concentrate at depths of 5–15 km, although they can reach up to 35 km (Chen and Molnar, 1983; Hong and Choi, 2012).

3. Data

The 11 March 2011 M9.0 Tohoku-Oki earthquake occurred at the convergent boundary between the Pacific and Okhotsk plates, rupturing a ~440-km-long and ~180-km-wide zone along the plate interface for ~150 s (Yagi and Fukahata, 2011). The coseismic slips reached 30–60 m at the hypocenter and trench (Lay et al., 2011; Tajima et al., 2013). The coseismic slip was large near the trench, which developed a huge tsunami. The peak ground acceleration in Kurihara city in Miyagi prefecture, southeastern Tohoku region reached ~3 g, which is equivalent to seismic intensity X in the modified Mercalli intensity scale (Hoshiba et al., 2011; Irikura and Kurahashi., 2012; Peng et al., 2012). The run-up height of the tsunami reached ~40 m at the nearest coast from the epicenter (Mori et al., 2011). A large number of aftershocks followed the megathrust earthquake around the rupture zone (e.g., Takahashi, 2011; Toda et al., 2011; Lee and Hong, 2014).

The Korean Peninsula is located at distances of 1100 to 1500 km (about 3 times the fault length) from the epicenter of the 2011 Tohoku-Oki earthquake (Fig. 1). The peninsula was dislocated to the epicenter by 2–5 cm (Baek et al., 2012; Kim and Bae, 2012; Zhou et al., 2012). The groundwater level changed abruptly by ~110 cm (Lee and Woo, 2012; Lee et al., 2013). Seismic stations are densely distributed over the Korean Peninsula (Fig. 1). Strong seismic waves from the megathrust earthquake are observed in the Korean Peninsula. Continuous seismic waveform record sections for 2 h before and 18 h after the Tohoku-Oki earthquake were collected from 135 stations in the Korean Peninsula (Fig. 2). The average inter-station distance is 20.3 km. The seismic stations are composed of 41 broadband and 35 short-period velocity seismometers and 59 accelerometers. The sampling rate of record sections is 100 Hz.

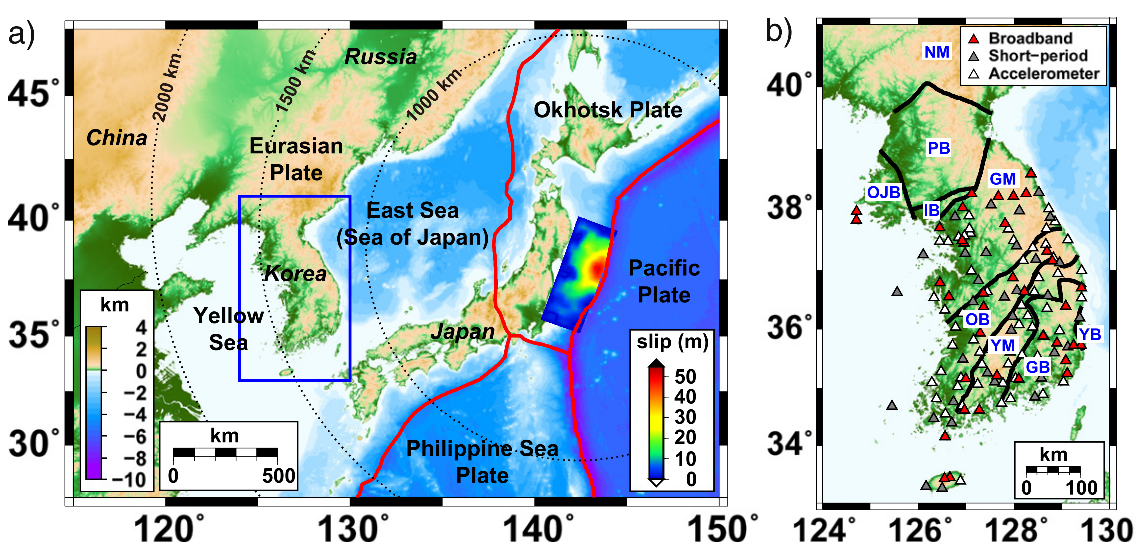

Fig. 1. (a) Map of the region around the eastern Eurasian plate. A source slip model of the 2011 M9.0 Tohoku-Oki earthquake (Yagi and Fukahata, 2011) and plate boundaries are presented. (b) An enlarged map of the region around the Korean Peninsula with major geological structures (solid lines) and seismic stations (triangles). The stations are equipped with velocity seismometers (broadband or short-period), or accelerometers. The major geological provinces include Gyeonggi massif (GM), Gyeongsang basin (GB), Imjingang belt (IB), Nangnim massif (NM), Okcheon belt (OB), Ongjin basin (OJB), Pyeongnam basin (PB), Yeonil basin (YB), and Yeongnam massif (YM).