A Method for Rapid Estimation of Moment Magnitude for Early Tsunami Warning Based on Coastal GPS Networks

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2 INTRODUCTION

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4 Recent great earthquakes of 26 December 2004 Sumatra, Indonesia (M_w 9.2), 26 February 5 2010 Maule, Chile (M_w 8.8), and 11 March 2011 Tohoku-oki, Japan (M_w 9.0) have, once 6 again, brought to focus the urgent need for early tsunami warning. These warnings mostly 7 rely on magnitude and location of an earthquake. A large/great magnitude, subduction zone 8 earthquake with rupture area extending up to the trench is potentially a tsunamigenic event. 9 The appropriate magnitudes for tsunami warning are those that are based on long-period 10 seismic waves (Abe 1979), *e.g.*, the moment magnitude, M_w (Kanamori 1977).

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12 Recently, W-phase (the long-period wave that arrives between P and S waves) has 13 been used to compute M_w (Kanamori and Rivera 2008; Hayes *et al.* 2009). This magnitude 14 can be determined in a relatively short time. For example, the first moment tensor solutions 15 of the Tohoku-oki earthquake, based on inversion of W-phase at teleseismic distances, 16 became available in 20 min (Duputel et al. 2011). For this reason, M_w computed from W-17 phase is especially useful for tsunami alert at distant sites. It is at local distances that early 18 tsunami warning becomes difficult. Even in this case, M_w based on inversion of W-phase 19 recorded at regional distances may be useful. Tests show that M_w of Mexican subduction 20 thrust earthquakes, based on W-phase recorded on broadband, regional seismograms, can be 21 estimated in ~ 7 min after the occurrence of the event (Pérez-Campos et al. 2010).

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23 In recent years there has been an increase in GPS stations along coastal region of 24 some subduction zones. Some of these stations are operated in continuous mode (either in 25 real-time high-rate mode or periodic low-rate download mode), and others in campaign 26 mode. Static displacement vectors obtained from GPS data alone or in conjunction with 27 seismograms, accelerograms, and tsunami waveforms have been used in many studies to 28 map slip on the fault. Since continuous GPS data provide displacement seismograms, it has 29 opened the new field of GPS seismology (e.g., Nikolaidis et al. 2001; Larson et al. 2003; 30 Bock et al. 2004; Miyazaki et al. 2004). Advantages of GPS-derived displacement 31 seismograms as compared to seismometer-derived seismograms are that they have low gain

and they provide reliable information until zero frequency. GPS displacement traces can also be combined with high-frequency accelerograms to produce very broadband strongmotion displacement seismograms (Nikolaidis et al., 2001; Emore *et al.*, 2007; Bock *et al.* 2011). With rapid advances in communication technology and analysis techniques, it is now possible to track the position of the GPS stations operating in real-time high-rate continuous mode with a latency of about a couple of seconds with cm-level accuracy (Genrich and Bock, 2006; Crowell *et al.* 2009). This makes it possible to use real-time

near-source GPS data for quick determination of M_w , useful for early tsunami alert in the

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region close to the source.

11 A few previous studies have discussed and/or demonstrated the potential of GPS 12 data for early tsunami warning. For example, Blewitt et al. (2007) analyzed data of the 13 2004 Sumatra earthquake from 38 GPS stations up to 7500 km from the epicenter. They 14 showed that by tracking the displacement field for 15 min after the origin time, the 15 estimated magnitude would have been M_w 9.0, indicating great tsunami hazard. The analysis 16 assumed that the epicenter and the focal mechanism were known. Since the heterogeneity 17 of slip on the fault has important effect on tsunami generation, Sobolev et al. (2007) 18 proposed an array of GPS stations perpendicular to the trench ("GPS shield") for early 19 tsunami warning in the Padang region of Sumatra. These authors also proposed deployment 20 of such arrays for other tsunamigenic active margins. Singh et al. (2008) explored the 21 feasibility of early tsunami warning based on coastal static displacement vectors and 22 proposed a simple method to estimate the length of the fault, approximate location of the 23 downdip edge of the rupture, and M_{w} . They partly validated the method using the static 24 displacement vectors reported for the earthquakes of 1995 Colima-Jalisco, Mexico ($M_w 8.0$) 25 and 2004 Sumatra, Indonesia (M_w 9.2). Melgar *et al.* (2011) have developed an algorithm 26 for real-time CMT determination of large earthquakes from near-source static displacement 27 field and have tested it by replaying the data of the 2003 Tokachi-oki, Japan (M_w 8.3) and 28 2010 El Mayor-Cucapah, Mexico $(M_w7.2)$ earthquakes. They conclude that a reliable 29 solution for these two earthquakes could have been found in 2-3 min. Rivera et al. (2011) 30 report that for the Tohoku-oki earthquake the inversion of W-phase recorded on GPS displacement seismograms at distances of 0.6° to 5.1° yields $M_{w}8.8-9.2$ and accurate fault 31

1 geometry. This solution would have been available in ~5 min. A disadvantage of the 2 methods proposed by Melgar *et al.* (2011) and Rivera *et al.* (2011) is that for great 3 earthquakes the point-source approximation may be grossly violated in the near-source 4 region, thus may lead to a biased solution. For their analysis of the Tohoku data, Rivera *et 5 al.* (2011) suggest using stations located at farther distances (> 3^0) in the *W*-phase 6 inversion.

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8 In this paper, we follow the method proposed by Singh *et al.* (2008), make it less 9 subjective, and test it on the data of seven additional large and great earthquakes. An a 10 priori rough knowledge of the geometry of the plate interface and the extent of the 11 seismogenic zone is required. This information is available for most, if not all, subduction 12 zones. The earthquake is assumed to be an interface, shallow-dipping thrust event. It is 13 approximated by a rectangle. The location of the downdip edge of the fault, its length, L, is 14 estimated from the static displacement vectors. The width, W, of the rupture is 15 approximated from L and prior knowledge of the seismogenic zone. A uniform slip, D, on 16 the fault, consistent with the average observed horizontal displacement vectors over length L, is then computed, which leads to the estimation of the seismic moment, M_0 . The method 17 18 is ideal for near-source data, where the point-source approximation becomes tenuous and 19 casts doubt on the CMT solutions. Estimation of L in real time is useful in delineating the 20 region where the earthquake effects are likely to be most intense. The knowledge of the 21 location of downdip edge with respect to the coast is also important because a large/great 22 earthquake whose rupture area partly lies below the continent may have relatively enhanced 23 high-frequency radiation and may generate severe ground motions, causing damage to 24 engineering structures and loss of life. On the other hand, when the rupture area lies mostly 25 offshore, then the high-frequency radiation may be relatively depleted, and the earthquake 26 may not produce large, destructive ground motions. It may, however, have a higher 27 tsunamigenic potential.

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We test the method on the near-field static deformation reported for nine earthquakes (Table 1, Figure 1): 1995 Colima-Jalisco, Mexico (M_w 8.0); 2003 Tecomán, Mexico (M_w 7.3); 2003 Tokachi-oki, Japan mainshock (M_w 8.3) and its aftershock (M_w 7.3);

1 2004 Sumatra, Indonesia (M_w 9.2); 2005 Nias, Indonesia (M_w 8.6); 2010, Maule, Chile 2 $(M_w 8.8)$; 2011 Tohoku-oki, Japan mainshock $(M_w 9.1)$ and its aftershock $(M_w 7.9)$. For 3 several of these earthquakes, only a few data points near the coast above the rupture area 4 are available. As expected, the most extensive data are for the Tohoku-oki earthquake and 5 its large aftershock. We find that M_w of earthquakes, even when estimated from only a few displacement vectors, are within 0.3 of the values reported in the Global CMT (GCMT) 6 7 catalog (Figure 1). The estimated rupture lengths and the locations of the downdip edge of 8 the rupture with respect to the coast are also in rough agreement with those reported in 9 detailed studies of the events. The analysis is simple and suitable for real-time application, 10 and the results are remarkably robust.

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13 DOWNDIP EDGE OF THE FAULT, ITS LENGTH, AND M_w

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15 **METHODOLOGY**

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17 Our analysis is based on expressions given by Okada (1992) for surface displacement due 18 to a rectangular fault buried in a half space. It is convenient to introduce the coordinate 19 system used by Okada, which is shown in Figure 2. In the problems of our interest here, the 20 coast and the trench will be roughly parallel to x-axis, and y-axis will be perpendicular to 21 the coast. The white arrow on the rectangular fault indicates the direction of slip of the 22 hanging wall during interplate, thrust earthquakes. In our computations here, we will 23 assume pure thrust motion (*i.e.*, rake, λ , is 90°), a uniform slip (D) on the fault, rigidity (μ) 24 of 5×10^4 MPa, and a Poisson solid. Unless otherwise mentioned, the dip of the fault (δ) will 25 be taken as 15°.

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To illustrate how this simple model can be used to estimate critical source parameters for early tsunami warning, in Figure 3 we show theoretical surface displacements in the near-source region caused by a shallow-dipping, thrust earthquake of $M_w 8.4 (M_0 = 5.01 \times 10^{21} \text{ Nm})$ buried in a half space. The fault is approximated by a rectangle of width, *W*, of 80 km and dip, δ , of 15⁰. As mentioned earlier, the rake, λ , is taken as 90°. 1 The downdip edge of the fault is at a depth C of 25 km. [While the example is for 2 illustrative purposes only, we note that these parameters are reasonable for $M_w \ge 7.5$ 3 earthquakes along the Mexican subduction zone, from Jalisco to Tehuantepec, as revealed 4 by numerous studies on seismicity, and large earthquakes and their aftershocks (see, *e.g.*, 5 Singh et al. 1985; Suárez et al. 1990; Singh and Mortera 1991; Tichelaar and Ruff 1993; 6 Pacheco and Singh 2010)]. We have taken L = 320 km for the $M_w 8.4$ earthquake, consistent with the relation $M_w = \log A + 4.0$, where A is the rupture area in km². From the relation M_0 7 8 $= \mu LWD$, we obtain a uniform slip D of 4.9 m, the value used in the calculations shown in 9 Figure 3.

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From the frames on the right of Figure 3, we note that: (1) the hinge line for vertical displacement, U_z , is at a distance of ~13 km toward the trench from the projection of the deepest part of the fault. With respect to the hinge line U_z is negative towards the continent (y < 0) and positive towards the trench (y > 0). (2) While magnitude and polarity of U_z are very sensitive to the position of the observation point with respect to the hinge line, the horizontal displacement in the direction perpendicular to the strike of the fault (U_y) is much less so. (3) U_y falls off quickly beyond the edge of the horizontal projection of the fault.

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19 These characteristics may be used to estimate the location of downdip edge of the fault 20 with respect to the coast, the length of the rupture and M_w , from observed coastal static 21 displacements. We will assume that the dip of the interface, δ ; the location of the 22 seismically coupled part of the interface and seismogenic width, W_s ; and hence the 23 associated depth *C* (Figure 2) in the region are known from previous studies. Figure 3 24 suggests following steps to estimate the parameters useful for early tsunami alert:

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26 (1) Estimation of the location of the downdip edge of the fault from observed 27 subsidence or uplift of the station. For example, if U_z is negative along the coast 28 (subsidence), the surface projection of the edge of the fault cannot be much farther 29 inland than ~13 km from the coast. In such cases we can fix the downdip edge 30 below the coast. If U_z , on the other hand, is positive (uplift), then the fault 31 projection must be more than ~13 km inland. Here, *a priori* information on the

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downdip limit of the seismically coupled part of the interface provides a useful constraint. We note, however, that an error of ± 20 km in the selection of the downdip edge is possible. However, since U_y is roughly constant across the surface projection of the edge, this error is not significant.

6 (2) Estimation of the length L from the horizontal static displacement vectors. Due to 7 heterogeneity of slip on the fault, these vectors will neither be as parallel nor as 8 constant along the coast above the fault as seen in Figure 3. We, nevertheless, 9 expect U_x to be much smaller than U_y over a subduction thrust fault. This is 10 confirmed from Figures 4 to 11 which show observed the static vectors of eight of 11 the nine earthquakes studied here. Henceforth we will assume that $U_x = 0$ and U_y equals the amplitude U_h of the horizontal vector. We will define L to be equal to the 12 13 distance along the coast where $U_{v} \ge (U_{v})_{20}$. Here $(U_{v})_{20} = 0.2(U_{v})_{max}$. In fact, in the 14 estimation of L we will include all stations within ± 20 km of the surface projection 15 of the downdip edge where $U_{v} \ge (U_{v})_{20}$. Since U_{v} decreases very rapidly away from 16 the edges, the estimation of L is straightforward if there is sufficient number of 17 stations along the coast. For most of the earthquakes considered here, the data along the coast is sparse. In these cases, we take the last station with $U_v > (U_v)_{20}$ and the 18 19 adjacent one where $U_{y} < (U_{y})_{20}$, and use a linear interpolation to determine the point 20 where $U_y = (U_y)_{20}$. For the Sumatra 2004 earthquake all the available displacement 21 vectors are larger than $(U_{y})_{20}$ (Figure 8). Thus, these vectors can't be used to 22 estimate L. In our analysis of this earthquake, we take L defined by aftershocks and source inversion studies. 23

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(3) Estimation of the width of the fault, *W*. We note that $W \le W_s$. It seems reasonable to require that if $L > W_s$ then $W = W_s$, but if $L < W_s$ then W = L. The depth *C* is known for most subduction zones if $W = W_s$. For $W \le W_s$, we compute *C* from the location of downdip edge of the fault from the trench, and the dip δ . We now have all the elements to define the origin of the coordinate system in Figure 2.

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- 1 (4) Computation of $\langle U_y \rangle$, the average of the observed U_y values over L. We note that, 2 in general, the stations along the coast will not be along a straight line parallel to the 3 trench, *i.e.*, their locations will not be along y = constant. However, since U_y is not 4 very sensitive to y, we will assume that the stations fall on a y = constant line in the 5 estimation of $\langle U_y \rangle$. As in the estimation of L, we compute $\langle U_y \rangle$ including all 6 stations with $U_y \ge (U_y)_{20}$ within ± 20 km from the surface projection of the downdip 7 edge.
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9 (5) With the rectangular fault already defined, we compute the uniform slip *D* that will 10 produce U_y equal to observed $\langle U_h \rangle = \langle U_y \rangle$ along the line y = constant where the 11 stations are roughly located, and x = L/2. For the model in Figure 2, U_y is nearly the 12 same between 0 < x < L. The requirement that computed U_y be equal to the 13 observed $\langle U_y \rangle$ is for simplicity and, within the framework of the simple model, is 14 not important. Now that *L*, *W*, and *D* have been estimated, the seismic moment is 15 obtained from the relation $M_0 = \mu L WD$.

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18 TESTS ON OBSERVED DATA

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20 1. COLIMA-JALISCO, MEXICO EATHQUAKE OF 9 OCTOBER 1995

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22 The coseismic static displacement caused by this earthquake was obtained from campaign-23 mode GPS measurements carried out before and after the earthquake (Figure 4) (Melbourne 24 et al. 1997). We note that the vertical displacement, U_z , was negative along the coast. The 25 tide gauge record at Manzanillo also shows a subsidence (Ortiz et al. 2000). This indicates 26 that the rupture did not extend more than \sim 13 km inland from the coast. The horizontal 27 displacement rapidly decreases between stations CHAM and CHAC to the NW and 28 between CRIP and SJDL to the SE. From the criterion mentioned above, the estimated 29 rupture length, L, is 227 km. As mentioned earlier, the width, W_s , of the coupled interface 30 along the Mexican subduction zone that ruptures in great earthquakes is about 80 km. Since 31 in this case $L > W_s$ we take $W = W_s = 80$ km. These estimates are in reasonable agreement

1 with those obtained from a detailed aftershock study by Pacheco et al. (1997): rupture 2 reaching up to the coast, L = 170 km, and W = 70 km. From U_y at CHAM, PURI and CRIP, 3 we obtain an average horizontal displacement, $\langle U_{v} \rangle$, of 0.66 m. This observation, along 4 with L = 227 km, W = 80 km, C = 25 km, and assuming that the stations are located along y = 0, yields an average dislocation, D, of 1.85 m on the fault and, hence, $M_0 = 1.68 \times 10^{21}$ N m 5 $(M_w = 8.08)$. Assuming y = -10 km, gives the same M_0 . For y = 10 km, $M_0 = 1.59 \times 10^{21}$ Nm 6 $(M_w = 8.07)$. Taking W = 60 km, and y = 0 but keeping all other parameters the same, gives 7 $M_0 = 1.50 \times 10^{21}$ N m ($M_w = 8.05$). These tests demonstrate the insensitivity of the results to 8 9 uncertainty in some of the parameters. We note that the estimated values of M_0 are surprisingly close to $M_0 = 1.15 \times 10^{21}$ Nm ($M_w = 7.97$) reported in Global CMT catalog. 10

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12 2. TECOMÁN, COLIMA, MEXICO, EARTHQUAKE OF 22 JANUARY 2003

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14 The static displacements caused by this earthquake, retrieved from permanent and 15 campaign-mode GPS stations, are given by Schmitt et al. (2007) (Figure 5). Based on the 16 criterian above and the observed horizontal displacements, we estimate L = 92 km and 17 $\langle U_{y} \rangle = 0.12$ m (computed from U_{y} at UCOL, CRIP, and MIRA). We note that the sites 18 near the coast subsided, indicating that the rupture did not extend more than ~13 km inland 19 from the coast. Assuming W = 80 km, C = 25 km, and the stations to be located along y = 0, D = 0.382 m for $\langle U_{v} \rangle = 0.12$ m, which yields $M_{0} = 1.38 \times 10^{20}$ Nm ($M_{w} = 7.36$), close to the 20 value of $M_0 = 2.05 \times 10^{20}$ N m ($M_w = 7.47$) reported in the Global CMT catalog. 21

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The first few days of aftershocks of this earthquake define an area of ~ 60 x 60 km² (Singh *et al.* 2003), some what smaller than estimated here: L = 92 km, W = 80 km. With LW = 60 km, C = 25 km, D = 0.50 m for $\langle U_y \rangle = 0.12$ m, which gives $M_0 = 9.1 \times 20^{19}$ N m ($M_w 7.24$).

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From the inversion of the coseismic static displacement field, Schmitt *et al.* (2007) report L = 80 km, W = 65 km, C = 40 km, and $M_0 = 9.1 \times 10^{19}$ Nm ($M_w = 7.24$). With the same L, W, and C values, assuming y = 0 for coastal stations in the epicentral zone, and 1 $\langle U_y \rangle = 0.12$ m, we get $M_0 = 1.99 \times 10^{20}$ N m ($M_w = 7.47$). If C = 25 km is chosen, then $M_0 = 1.13 \times 10^{20}$ N m ($M_w = 7.30$).

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From joint inversion of near-source strong-motion and teleseismic body-wave data, Yagi *et al.* (2004) find L = 35 km, W = 75 km and $M_0 = 2.3 \times 10^{20}$ Nm ($M_w = 7.51$).

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It is not surprising that *L* and *W*, reported in the studies mentioned above, vary so much, since the criteria used in estimating them are not uniform and the methods employed differ. We do not expect such large differences for great subduction thrust earthquakes (for which L >> W and $W=W_s$). It is encouraging that $M_w7.36$ computed following our simple approach is close to $M_w7.47$ reported in the GCMT catalog. It is also within the range of the values reported in the detailed studies of Yagi *et al.* (2004) and Schmitt *et al.* (2007).

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14 3. TOKACHI-OKI, JAPAN, EARTHQUAKE OF 25 SEPTEMBER 2003, 15 MAINSHOCK

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17 Extensive GPS data, recorded by GEONET array which is operated by Geographical 18 Survey Institute (GSI) of Japan, are available for this earthquake (see, e.g., Larson and 19 Miyazaki 2008). These data have been used in several source inversion studies (e.g., 20 Koketsu et al. 2004; Miyazaki et al. 2004; Romano et al. 2010). Figure 6 shows all GPS 21 stations along the SE coast of Hokkaido with $U_{y} \ge (U_{y})_{20}$. $(U_{y})_{max}$ of 0.9 m occurred at 22 station 0015. We note a subsidence at these stations. The figure also includes station 0010 23 where U_{y} was less than $(U_{y})_{20}$ but the site was uplifted. From these data we surmise that the 24 slip on the plate interface occurred offshore, with the horizontal projection of the downdip 25 edge reaching the coast. In any case, it did not extend more than about 13 km inland. There 26 is some ambiguity in defining the SW limit of the fault due to the geography of Hokkaido. 27 In this case, we estimated the limit by linearly extrapolating the data at stations 0144 and 28 0142, and determining the point where $U_v = (U_v)_{20}$. The estimated L is 176 km and $\langle U_v \rangle$ is 29 0.61m. The distance of the coast from the trench is about 200 km. This requires us to 30 choose $W \le 176$ km. The dimension of the square in the figure is L = W = 176 km. The 31 depth of the interface below the coast is about 50 km (see Figure 1 in Koketsu et al. 2004).

1 We take the coast line to be at a distance of 20 km NW of the horizontal projection of the downdip edge of the rectangular fault (y = -20 km). Assuming $\delta = 15^{\circ}$, $\langle U_{y} \rangle = 0.61$ m 2 along the coast requires D = 1.95 m, which yields $M_0 = 3.02 \times 10^{21}$ Nm ($M_w = 8.25$). 3 Choosing a width W of 80 km yields D = 2.93 m and, hence, $M_0 = 2.06 \times 10^{21}$ Nm ($M_w =$ 4 8.14) which is nearly identical to the previous estimate. For comparison, the GCMT catalog 5 reports a focal mechanism characterized by $\varphi = 250^\circ$, $\delta = 11^\circ$, $\lambda = 132^\circ$, and $M_0 = 3.05 \times 10^{21}$ 6 Nm ($M_w = 8.26$). Although the fault plane defined by the square in Figure 6, $\varphi = 210^\circ$, $\delta =$ 7 8 15° , $\lambda = 90^{\circ}$, differs considerably from the one reported by GCMT, the seismic moments 9 are nearly the same. We note that our estimates of the source parameters are in reasonable 10 agreement with those from the inversion studies mentioned above.

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12 4. TOKACHI-OKI, JAPAN EARTHQUAKE OF 25 SEPTEMBER 2003, 13 AFTERSHOCK

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15 A large aftershock followed the Tokachi-oki earthquake by about 78 min. The static

16 displacement field produced by the aftershock is given by Larson and Miyazaki (2008). The 17 displacement vectors in Figure 7 show a pattern similar to that of the mainshock. The maximum horizontal displacement, $(U_y)_{max}$, of 0.09 m occurs at station 0019. We follow the 18 same procedure as for the mainshock. The estimated L = W is 80 km and $\langle U_{v} \rangle$ is 0.050 m. 19 20 Similar to the mainshock, we take (a) the depth of the interface below the coast as 50 km, 21 and (b) the coast line to be at a distance of 20 km NW of the surface projection of the downdip edge of the fault (y = -20 km). In this case, D corresponding to $\langle U_y \rangle = 0.050$ m 22 at y = -20 km is 0.33 m. This yields $M_0 = 1.07 \times 10^{20}$ Nm ($M_w = 7.29$). Choice of y = -1023 24 km and -30 km results in almost identical seismic moment, once again demonstrating that 25 the uncertainty in the location of the coastal stations with respect to the downdip edge of 26 the fault is not important. 27

The GCMT catalog lists the focal mechanism as $\varphi = 208^{\circ}$, $\delta = 18^{\circ}$, $\lambda = 86^{\circ}$, and M_0 = 1.29×10^{20} Nm ($M_w = 7.34$). The fault plane, $\varphi = 210^{\circ}$, $\delta = 15^{\circ}$, $\lambda = 90^{\circ}$, and the estimated M_0 from the static field are almost identical to those reported by GCMT.

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3 The near-field static displacements for the 2004 earthquake were obtained from GPS 4 measurements carried out before and after the earthquake, in a campaign mode. Near- and 5 far-field geodetic data have been analyzed by themselves (e.g., Vigny et al. 2005; Banerjee 6 et al. 2005, 2007; Gahalaut et al. 2006; Rajendran et al. 2007) as well as in conjunction 7 with the seismic data (e.g., Subarya et al., 2006; Chlieh et al. 2007) to invert for the slip 8 distribution on the fault. For our test, we selected the near-field static deformation reported 9 in Gahalaut et al. (2006). These values have not been corrected for post-seismic slip, which 10 was small (Banerjee et al. 2007; V. Gahaluat, personal communication, 2008). Figure 8 11 shows the coseismic displacements. Note that the near-field data are available only between 7° and 14° N. The average amplitude of the horizontal vectors, $\langle U_y \rangle$, is 4.2 m. Since the 12 13 epicenter was located near 3° N, the length of the rupture cannot be estimated from the GPS 14 data. Furthermore, U_{v} is greater than $(U_{v})_{20}$ at all stations. Based on aftershocks and 15 numerous source studies, we assume that the rupture extended from 2° to 14° and that $\langle U_{\nu} \rangle$ 16 = 4.2 m over the entire fault. Although the rupture propagated along an arc, we 17 approximate the fault by a rectangle of length, L, of 1340 km. We take the dip of the fault, δ , as 15°, the width W as 150 km, and C as 50 km. These parameters are supported by 18 19 seismicity of the region (e.g., Engdahl et al. 2007). Finally, we assume that the static 20 displacements were measured at points above the deep edge of the fault (y = 0). This is not 21 true since the field observations and the geodetic data from GPS campaign mode (Figure 8) 22 demonstrate that some islands in the Andaman and Nicobar region were uplifted, while 23 others suffered subsidence. However, as mentioned earlier, the horizontal displacement is 24 not very sensitive to the exact location of the observation point with respect to the surface projection of downdip edge. Under these reasonable assumptions, an average dislocation, 25 26 D, of 11.9 m is needed to produce a horizontal displacement of 4.2 m at the surface. This vields a seismic moment M_0 of 1.14×10^{23} Nm ($M_w = 9.31$). The moment magnitude of the 27 28 Sumatra-Andaman earthquake has been controversial; the estimates, based on different data sets and techniques, vary between $M_w = 9.0$ and 9.3 (see Bilek *et al.* 2007 for a summary). 29 30 Our estimate is in the range of the values reported in studies based on detailed analysis of 31 the data.

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6. NIAS EARTHQUAKE OF 28 MARCH 2005

4 Four continuous GPS stations were operating in the epicentral zone of this earthquake 5 (Konca et al. 2007). These stations are shown in Figure 9 along with horizontal and vertical 6 static displacements (from Table 1 of Konca et al. 2007). The horizontal displacements at 7 stations LEWK and PSMK are less than 20% of the maximum at LHWA. From the 8 criterion laid out earlier, L = 372 km and $\langle U_{y} \rangle = 3.44$ m. Static displacements at BSIM and 9 LEWK, which are clearly up, show that the rupture propagated further downdip than the 10 vertical projection of these stations on the plate interface. Uplift/subsidence was mapped in 11 the epicentral zone from coral micro-atoll measurements (Konca et al. 2007). If we make 12 use of this data, then the estimated rupture area, outlined by the smaller rectangle in Figure 13 9, is given by L = 372 km and W = 135 km. If we ignore the information provided by the 14 coral measurements, then the rupture area could be extended up to the coast so that L = 37215 km and W = 215 km (larger rectangle in Figure 9). With W = 135 km, C = 40 km, $\langle U_{y} \rangle =$ 3.44 m at y = 40 km requires slip, D, on the fault of 6.99 m, giving $M_0 = 1.76 \times 10^{22}$ Nm (M_w) 16 = 8.76). With W = 215 km, C = 60 km, and y = 125 km, the estimated M_0 is 2.08×10^{22} Nm 17 $(M_w = 8.81)$. The normal-mode data and GPS data tightly constrain δ between 8° and 10° 18 (Konca *et al.* 2007). With $\delta = 10^{\circ}$ we obtain almost identical M_0 as for $\delta = 15^{\circ}$. We note 19 20 that our estimate of M_w is relatively insensitive to reasonable choices of the dip and the 21 width. For comparison, M_0 reported in the Global CMT catalog and by Konca *et al.* (2007) are 1.01×10^{22} Nm ($M_w = 8.60$) and $1.0-1.24 \times 10^{22}$ Nm ($M_w = 8.60-8.66$), respectively. 22

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24 7. MAULE, CHILE, EARTHQUAKE OF 27 FEBRUARY 2010

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Figure 10 shows coseismic static displacement vectors associated with this earthquake (from Vigny *et al.* 2011). These vectors were obtained from GPS stations operating in continuous and as well as in campaign mode. We note that the stations are concentrated between -35° N and -38° N. With the criteria mentioned earlier, we estimate *L* as 545 km and $\langle U_y \rangle$ is 3.32 m. The vertical displacement is up along the coast but becomes negative towards the continent. Thus, we take the surface projection of the downdip edge of the fault to extend up to the stations where subsidence occurs (Figure 10). Following Tichelaar and Ruff (1993), we take *C* as 50 km. Since $L \gg W_s$, we take $W = W_s \sim 140$ km. With these parameters, D = 10.13, 9.93, and 8.59 m for y = -35, 0, and 35 km, respectively. The corresponding seismic moments are 3.86×10^{22} Nm ($M_w = 8.99$), 3.79×10^{22} N m ($M_w = 8.99$), and 3.28×10^{22} N m ($M_w = 8.94$). The seismic moment listed in the Global CMT catalog is 1.86×10^{22} Nm ($M_w = 8.78$).

7

8 Our estimates of the dimension of the source, its location, and the seismic moment 9 agree well with those reported in detailed source studies (*e.g.*, Vigny *et al.* 2011; Moreno *et* 10 *al.* 2010; Lorito *et al.* 2011; Delouis *et al.* 2010).

11

8. TOHOKU-OKI, JAPAN, EARTHQUAKE OF 11 MARCH 2011, MAINSHOCK 13

14 The unexpectedly large and disastrous earthquake of Tohoku-oki is the best recorded 15 earthquake by a GPS array (~ 1200 stations of the GEONET). The Tohoku coast is ~ 200 16 km from the trench. Figure 11 illustrates the coseismic static displacement vectors at 17 stations located ≤ 250 km from the trench where $U_{y} \geq (U_{y})_{20}$. These data were provided by 18 ARIA group of JPL and Caltech. The original GEONET data were given to Caltech by 19 Geospatial Information Authority (GSI) of Japan. We note that the vertical displacement is 20 down at all stations whose vectors are shown in the figure, indicating that the rupture 21 occurred offshore; if the rupture did extend inland it could not have been much more than ~ 22 10 km.

23

Based on the criteria initially laid out, we find L = 373 km and $\langle U_y \rangle = 2.17$ m. 24 25 Along this margin the seismogenic width, W_s , and the maximum depth of the seismicallycoupled interface, C, are ~ 200 km and 50 km, respectively (e.g., Hasegawa et al. 1994; 26 Igarashi et al. 2001). Thus, $W = W_s = 200$ km. With these parameters, $\langle U_y \rangle = 2.17$ m at y =27 -10 km, requires D = 5.62 m which yields $M_0 = 2.11 \times 10^{22}$ Nm ($M_w = 8.82$). The Global 28 CMT catalog reports $M_0 = 5.31 \times 10^{22}$ Nm ($M_w = 9.08$). Our gross estimates of the source 29 30 parameters, which could in principle have been obtained in < 5 min, are in accordance 31 with those obtained in formal studies (e.g., Simons et al. 2011; Ide et al. 2011).

9. TOHOKU-OKI, JAPAN, EARTHQUAKE OF 11 MARCH 2011, AFTERSHOCK

4 The largest aftershock of the Tohoku-oki earthquake occurred at 06:16 GMT about 28 min 5 later, extending the rupture area of the mainshock towards SSW (see, e.g., Simons et al. 6 2011). The static displacement field caused by this earthquake is not available but the 7 displacement vectors are plotted in Figure 1 of Simons et al. (2011). We extracted the 8 relevant parameters from an examination of this figure: L = 150 km, $(U_y)_{max} = 0.50$ m, $\langle U_{y} \rangle = 0.30$ m. Taking L = W = 150 km, $\delta = 15^{\circ}$, C = 50 km, and y = -10 km, yields D = 0.009 1.05 m, and $M_0 = 1.18 \times 10^{21}$ Nm ($M_w = 7.98$), close to $M_0 = 8.48 \times 10^{21}$ Nm ($M_w = 7.88$) 10 reported in the GCMT catalog. We note that the estimation of M_0 changes by less than 20% 11 12 if y varies between -30 to 30 km.

13

14 **REAL-TIME APPLICATION**

15

16 We recapitulate the steps involved in the method which clearly show how it would be 17 implemented in a real-time environment.

18 As mentioned earlier, for most subduction zones, the dip, δ , and the width, W_s , of 19 the coupled part of the interface, as well as the maximum depth of its downdip edge, C, is 20 known *a priori*. We are assuming that the GPS stations are located along the coast roughly 21 parallel to the trench. Once the system detects displacement vectors exceeding a threshold 22 level at more than a certain pre-established number of contiguous stations, the process gets 23 triggered. The strike of the fault is computed so that it is perpendicular, on an average, to 24 the recorded horizontal displacement vectors. The polarity of the vertical component of the 25 displacement vector fixes the surface projection of the downdip edge of the fault. This line 26 will be parallel to the strike. Next, the length, L, of the fault is estimated from the coastal 27 horizontal displacement vectors using the criterion outlined above and the average 28 horizontal displacement, $\langle U_h \rangle$, is computed over L. The width W is obtained from the 29 criterion: $W=W_s$ if $L > W_s$, otherwise W=L. Now the coordinate system, and the location 30 and size of the rectangular fault in the Okada's model are set. With respect to this 31 coordinate system, the locations of the GPS stations are known. A line parallel to the

surface projection of the downdip edge of the fault (y = c in Okada's coordinate system, where *c* is a constant) is determined so it minimizes the distance to the coastal stations. As a final step, Okada's theoretical expressions are used to compute a uniform dislocation, *D*, on the fault which gives $U_y = \langle U_h \rangle$ at y = c and x = L/2 (Figure 2), and the seismic moment is obtained from the relation $M_0 = \mu LWD$.

- 6
- 7

8 DISCUSSION AND CONCLUSIONS

9

Figure 1, inset, compares M_w estimated in this study with the corresponding M_w reported in 10 11 the Global CMT catalog. The magnitudes are within ± 0.3 of each other and the average 12 difference is 0.15. We conclude that for large and great earthquakes our proposed simple 13 analysis of near-source static displacement vectors at stations along the coast parallel to the 14 trench yields robust and reliable estimate of M_w and, in the process, generates useful 15 byproducts such as the length of the fault, L, and an approximate location of the surface projection of its downdip edge. These byproducts may be potentially very helpful in 16 17 delineating the area where severe ground motion and tsunami might be expected. In this 18 sense, they may be more useful than the centroid location of the earthquake provided by the 19 CMT inversion. The method requires a rough knowledge of the geometry and some details 20 of the seismically-coupled segment of the plate interface. This information is available for 21 most subduction zones.

22 Advances in communication technology and analysis techniques now permit 23 tracking of the position of GPS sites with a latency of ~ 2 s (e.g., Bock et al., 2011). It 24 follows that the time after the origin that it would take for the near-source static 25 displacement vector to be available for analysis at a GPS station would be ~ (S-wave travel 26 time + duration of the source time function). As an example, let us consider the 2011 27 Tohoku-oki earthquake. The duration of source time function was ~ 160 s (Ide *et al.* 2011). 28 Thus, the static displacement vectors at coastal GPS stations located \leq 500 km from the 29 hypocenter would have been available in < 5 min. As our analysis is simple, the estimate 30 M_w would have been available immediately afterwards. CMT inversion based on near-31 source static displacement field or W-phase would have taken comparable, or slightly more,

time. An advantage of the proposed method is that it does not suffer from the limitation imposed by the point-source approximation to analysis near-source data of great earthquakes.

4 The method can be customized for each segment of a given subduction zone so that 5 the selected parameters closely reflect the available knowledge for that segment.

6 We have assumed that the coastal static displacement vectors are associated with 7 shallow, thrust earthquakes. In some cases, they may be a result of outer rise normal-8 faulting earthquake (e.g., 1933 Sanriku, Japan, M_w 8.4-8.6; 2007 Kuril, Russia, M_w 8.1). 9 Large earthquakes also occur in the subducted plate near the coast (e.g., 1997 Michoacán, 10 Mexico, M_w 7.1; 1999 Oaxaca, Mexico, M_w 7.4). The near-source static displacement vectors 11 associated with such earthquakes will differ from those caused by shallow thrust events. 12 This possibility must be contemplated in implementing the proposed method in real-time 13 application of GPS data.

- 14
- 15

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17

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 Table 1. Earthquake parameters.

Event	Region	Date	Latitude	Longitude	Depth	$\mathbf{M_0}^*$	M_w^*	φ*	δ*	λ*	$\mathbf{M_0}^+$	$\mathbf{M}_{\mathbf{w}}^{+}$	\mathbf{L}^+	\mathbf{W}^{+}
Number	_	Time	0	0	km	N m					N m		km	km
1	Colima-	1995/10/09	19.34	-104.80	15.0	1.15E+21	7.97	302	9	92	1.68E+21	8.08	227	80
	Jalisco	15:35:28.8												
	Mexico													
2	Tecomán,	2003/01/22	18.86	-103.90	26.0	2.05E+20	7.47	308	12	110	1.38E+20	7.36	92	80
	Colima	02:06:48.9												
	Mexico													
3	Tokachi-oki,	2003/09/25	42.21	143.84	28.2	3.05E+21	8.26	250	11	132	3.02E+21	8.25	176	165
	Japan	19:50:38.2												
	Mainshock													
4	Tokachi-oki,	2003/09/25	41.75	143.62	47.3	1.29E+20	7.34	208	18	86	1.07E+20	7.29	80	119
	Japan	21:08:19.5												
	Aftershock													
5	Sumatra-	2004/12/26	3.09	94.26	28.6	3.95E+22	8.99	329	8	110	1.14E+23	9.31	1340	150
	Andaman	01:01:09.0												
6	Nias,	2003/03/28	1.67	97.07	25.8	1.05E+22	8.61	333	8	118	2.08E+22	8.81	372	215
	Indonesia	16:10:31.5												
7	Maule, Chile	2010/02/27	-35.98	-73.15	23.2	1.86E+22	8.78	19	18	116	3.86E+22	8.99	545	140
		06:35:14.5												
8	Tohoku-oki,	2011/03/11	37.52	143.05	20.0	5.31E+22	9.08	203	10	88	2.11E+22	8.82	373	200
	Japan	05:47:32.8												
	Mainshock				• • • •			100						
9	Tohoku-oki,	2011/03/11	35.92	141.38	29.0	8.48E+20	7.89	199	17	84	1.18E+21	7.98	150	150
	Japan	06:15:58.7												
	Aftershock													

*Parameters from the GCMT catalog. + Parameters from this study.

Figure Captions

Figure 1. Location of the nine large/great earthquakes $(7.3 \le M_w \le 9.2)$ studied in this paper. The numbers are keyed to Table 1. The inset shows the magnitude comparison between the GCMT catalog and those obtained in this study. The solid line represents a one-to-one relationship; the dashed-dotted lines represent the ± 0.3 unit band. The average difference is 0.15.

Figure 2. Geometry of the rectangular fault and the coordinate system used by Okada (1992). In the problem of interest here, positive *y* is toward the trench and the coast line is assumed to be along *y* = constant. Dip (δ), maximum depth (*C*), of the seismically-coupled interface, and maximum seismogenic width ($W = W_s$) is roughly known for all subduction zones.

Figure 3. Displacement field of an earthquake of $M_w = 8.4$, calculated from Okada's (1992) model. (Left) Profile along the fault with y = 10 km (y = 0 corresponds to surface projection of fault's downdip edge), showing vertical displacement, U_z , and horizontal displacements, U_x and U_y . (Right) Profile across the surface projection of the downdip edge and x = L/2 (L = 320 km $M_w = 8.4$), showing U_z and U_y . Note that the hinge line of U_z is y = 13 km and U_y is constant around y = 0.

Figura 4. Static displacement vectors caused by 1985 Colima-Jalisco, Mexico earthquake (modified from Melbourne *et al.* 1997). Solid circles indicate coastal stations (for this earthquake ≤ 150 km away from the trench) with $U_h \geq (U_h)_{20}$; the solid gray circles indicate stations within 150 km from the trench but with $U_h < (U_h)_{20}$ which are useful in constraining the limit of the fault. All other stations are shown by while circles. Average horizontal displacement $\langle U_h \rangle$ is computed from U_h at stations indicated by solid circles. $(U_h)_{20}$ is shown by the black dashed line. U_h is in cm. In our interpretation $U_h = U_y$ (Figure 2) for all earthquakes. The horizontal and vertical displacement vectors are shown by dark gray and light gray arrows, respectively. Star shows the epicenter. The dashed-dotted gray lines denote the limits of the rupture estimated accordingly to the criteria described in the text. The dashed rectangle is the estimated rupture area (see text).

Figure 5. Static displacement caused by 2003 Tecomán, Colima, Mexico earthquake (modified from Schmitt *et al.* 2007). Station ≤ 125 km away from the trench are considered as coastal stations. Symbols are the same as in Figure 4.

Figure 6. Static displacement vectors caused by 2003 Tokachi-oki, Japan earthquake, mainshock (data from Larson and Miyazaki 2008). Station \leq 260 km away from the trench are considered as coastal stations. Symbols are the same as in Figure 4.

Figure 7. Static displacement vectors caused by 2003 Tokachi-oki, Japan earthquake, aftershock (data from Larson and Miyazaki 2008). Station \leq 260 km away from the trench are considered as coastal stations. Symbols are the same as in Figure 4.

Figure 8. Static displacement vectors caused by 2004 Sumatra-Andaman earthquake (data from Gahalaut *et al.* 2006). All stations available are considered coastal stations. In this case, due to limited areal extent of GPS stations, the near-field static displacement vectors do not provide a constraint on *L*. Curved dashed rupture area is based on aftershocks and source inversion studies (see text). Symbols are the same as in Figure 4.

Figure 9. Static displacement vectors resulting from 2005 Nias earthquake (data from Konca *et al.* 2007). Station ≤ 150 km away from the trench are considered as coastal stations. Symbols are the same as in Figure 4. Uplift/subsidence mapped in the epicentral zone from coral micro-atoll measurements constrains the rupture area to the smaller rectangle (see text). In the absence this data, the static displacement vectors from GPS would allow the larger rupture area.

Figure 10. Static displacement vectors caused by 2010 Maule, Chile earthquake (data from Vigny *et al.* 2011). Station \leq 200 km away from the trench are considered as coastal stations. Symbols are the same as in Figure 4. Note that in this case the vertical

displacement hinge line is clearly inland and hence the surface projection of the downdip edge of the fault is well constrained.

Figure 11. Static displacement vectors caused by 2011 Tohoku, Japan earthquake, mainshock. Preliminary GPS displacements, provided by ARIA group of JPL and Caltech. The original GEONET data were given to Caltech by *Geospatial Information Authority* (GSI) of Japan. Station ≤ 250 km away from the trench are considered as coastal stations. Only displacement vectors with $U_h \geq (U_h)_{20}$ are plotted. Symbols are the same as in Figure 4.





















Figure 6



Figure 7











