

Earthquake early warning using future generation gravity strainmeters

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Key Points:

- Future high-precision gravity strainmeters could record prompt gravity signals before the seismic waves arrival during an earthquake rupture
- Planned sensitivity is sufficient to observe gravity perturbations from earthquakes of magnitude larger than 7 at distances up to 1000 km
- Gravity-based warning system could perform faster detection and magnitude estimation of large earthquakes compared to conventional systems

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Abstract

Recent studies reported the observation of prompt elastogravity signals during the 2011 M9.1 Tohoku earthquake, recorded with broadband seismometers and gravimeter between the rupture onset and the arrival of the seismic waves. Here we show that to extend the range of magnitudes over which the gravity perturbations can be observed and reduce the time needed for their detection, high-precision gravity strainmeters under development could be used, such as torsion bars, superconducting gradiometers or strainmeters based on atom interferometers. These instruments measure the differential gravitational acceleration between two seismically isolated test masses, and are designed to observe signals around 0.1 Hz. We show that these instruments should be able to detect prompt gravity perturbations induced by earthquakes larger than M7, up to 1000 km from the earthquake centroid within P-waves travel time and up to 120 km within the first 10 seconds of rupture onset, provided a sensitivity in gravity strain of $10^{-15} \text{ Hz}^{-1/2}$ at 0.1 Hz can be achieved. The analysis involves simulations of the expected gravity strain signals based on an analytical model of gravity perturbations generated by fault rupture in a homogeneous half-space. As an immediate application, we discuss the possibility to improve current earthquake-early warning systems (EEWS). Our results suggest that, in comparison to conventional P-wave-based EEWS, a gravity-based warning system could perform faster detections of large off-shore subduction earthquakes (at least larger than M7.4). Gravity strainmeters could also perform earlier magnitude estimates, within the duration of the fault rupture, and therefore complement current tsunami warning systems.

1 Introduction

During an earthquake rupture, fault slip and the propagation of seismic waves redistribute masses within the Earth. The mass redistribution generates a dynamic long-range perturbation of the Earth's gravity, which propagates at speed-of-light and is thus recordable before the arrival of the direct seismic waves (Harms et al., 2015; Montagner et al., 2016; Vallée et al., 2017). These prompt gravity perturbations have been observed with broadband seismometers and superconducting gravimeter during the M9.1 Tohoku-oki earthquake (Montagner et al., 2016; Vallée et al., 2017). The potential contribution of such signals to tsunami early warning is substantial: such observations would indeed

50 have provided an early estimate of a magnitude greater than M9 within three minutes
51 of the earthquake origin time.

52 However, two factors hinder the observation of prompt elastogravity signals with
53 ground-based seismometers: the background seismic noise and a partial cancellation be-
54 tween the gravitational perturbation and its induced ground acceleration, whose differ-
55 ence is recorded by the instruments (Heaton, 2017; Vallée et al., 2017). Elastogravity sig-
56 nal detection based on individual seismometer records is thus limited to earthquakes with
57 magnitudes larger than 8.

58 One approach to improve earthquake monitoring capabilities is to overcome the lim-
59 itations associated with the use of ground-coupled seismometers and gravimeters, by mea-
60 suring the differential gravitational acceleration between two seismically isolated test masses.
61 This detector concept is known as a gravity strainmeter. Gravity strainmeters designed
62 to observe signals at 0.1 Hz, within the frequency range needed to detect earthquake-
63 related gravity changes, are being developed to observe gravitational waves (GW) sources
64 in the sub-Hz domain (Harms et al., 2013) and are briefly reviewed in Section 2. We note
65 that the sub-Hz instruments are much smaller, lighter than the instruments developed
66 for high-frequency (> 100 Hz) GW detection (1-m scale compared to 1-km scale). More-
67 over, GW detection has very stringent requirements, thus the sensitivity needed for earth-
68 quake detection should be achieved at an earlier stage of the instrument development.
69 In contrast to seismometers, gravity strainmeters implement sophisticated seismic iso-
70 lation schemes to measure differential displacements or rotations between test masses.
71 The differential measurement rejects partially the background seismic noise and the gravity-
72 induced inertial acceleration, which are similar for the two masses. Thus, for measur-
73 ing earthquake-induced gravity perturbations, gravity strainmeters may be considered
74 as a natural step toward improved sensitivities.

75 In this article we evaluate quantitatively the improvement of earthquake early warn-
76 ing systems (EEWS) that could be obtained, in principle, by using the gravity strain-
77 meters under development. Current EEWS are automatic systems formed by seismome-
78 ters and communication networks, intended to detect the occurrence of an earthquake
79 before the arrival of ground-shaking waves and to disseminate the information to the pop-
80 ulation (R. M. Allen, Gasparini, Kamigaichi, & Bose, 2009; Heaton, 1985). Conventional
81 EEWS rely on detecting the seismic P-waves, which travel at several km/s and are roughly

twice as fast as the usually stronger, more damaging S-waves. Since changes in gravity propagate at the speed of light, a gravity-based warning system could give a potential gain in the warning times, with respect to conventional EEWS. An expected consequence is a reduction of the blind zone of an EEWS, i.e. the area around the epicenter of an earthquake that cannot receive a warning before the arrival of the S-phase. We show that a gravity-based warning system could perform faster detections of large off-shore subduction earthquakes and early magnitude estimates, available as soon as the rupture stops.

2 High-precision gravity strainmeters

2.1 Detector concepts

Gravity strainmeters are instruments designed to measure gravity strain \mathbf{h} , which is the second time-integral of the spatial-gradient of gravity acceleration $\delta\mathbf{g}$:

$$\mathbf{h}(\mathbf{r}, t) = \int_0^t \int_0^{\tau'} \nabla \delta\mathbf{g}(\mathbf{r}, \tau) d\tau d\tau'. \quad (1)$$

Very sensitive gravity strainmeters have been developed in the context of GW detection. In their advanced configurations, laser-interferometric GW detectors LIGO (Abbott et al., 2009) and Virgo (Accadia et al., 2011) have designed strain sensitivities of $10^{-23} \text{ Hz}^{-1/2}$ between about 30 Hz and 2000 Hz. Due to their poor sensitivity in the sub-Hz region, advanced GW detectors cannot be used to measure gravity perturbations from earthquakes. In fact, in order to produce noticeable terrestrial gravity noises in the sensitive frequency band of advanced detectors, it has been shown that typical density perturbations have to be generated very close to the suspended test masses, i.e. within a few tens of meters (Driggers, Harms, & Adhikari, 2012; Harms, DeSalvo, Dorsher, & Mandic, 2009). This means that terrestrial gravity perturbations that will be measured in advanced detectors are likely to be of little interest in geophysics.

To access the sub-Hz region, which is very rich in GW sources, three concepts for 0.1 Hz gravity gradiometers are currently under development: superconducting gradiometers (SGG) (Moody, Paik, & Canavan, 2002; Paik et al., 2016), torsion-bar antennas (e.g. TOBA (Ando et al., 2010) or TorPeDO (McManus et al., 2017)) and atom-interferometric gradiometers (Geiger, 2017; Hohensee et al., 2011). In the following, we will refer to these detectors as GG10, i.e. gravity gradiometers with high-sensitivity for signals with periods around 10 seconds.

109 All three concepts present novel solutions to the mitigation of seismic noise, which
 110 would otherwise exceed gravity signals by many orders of magnitude. The superconduct-
 111 ing gradiometer achieves seismic-noise reduction by common-mode rejection in the dif-
 112 ferential readout of test-mass positions relative to a common, stiff reference frame. Torsion-
 113 bar antennas can be engineered with very low torsion resonance frequency, which con-
 114 stitutes an efficient passive filter of rotational seismic displacement. The rejection of trans-
 115 lational displacement noise is obtained by reading-out the differential signal from two
 116 suspended bars (or a bar with respect to a suspended platform) (Shimoda, Aritomi, Shoda,
 117 Michimura, & Ando, 2018). Atom-interferometric gradiometers read out the displace-
 118 ment between freely falling ultracold atom clouds, which also provides partial immunity
 119 to seismic noise. In order to reduce the requirements of the seismic rejection, additional
 120 passive or active seismic-isolation techniques can be used (Winterflood, 2001).

121 The most sensitive instrument so far is the superconducting gradiometer with a strain
 122 sensitivity of about $10^{-10} \text{ Hz}^{-1/2}$ at 0.1 Hz (Moody et al., 2002). However, extensive gain
 123 in experience with these technologies has led to defining more ambitious strain-sensitivity
 124 targets: $10^{-15} \text{ Hz}^{-1/2}$ at 0.1 Hz (Ando et al., 2010; Hogan et al., 2011; Hohensee et al.,
 125 2011). It will be shown in section 3.2 that such design sensitivities are sufficient for the
 126 detection of prompt gravity perturbations from earthquakes (of magnitude $> \text{M}6.5$). It
 127 should be noted that all three concepts are also being considered as candidates for fu-
 128 ture, sub-Hz GW detectors with more ambitious sensitivity targets ($10^{-20} \text{ Hz}^{-1/2}$ at 0.1 Hz)
 129 (Harms et al., 2013; Paik et al., 2016).

130 2.2 Detector sensitivity models

131 The response of different types of gravity strainmeters to gravity-gradient fluctu-
 132 ations is not identical (Harms, 2015). The consequence is that instrumental noise spec-
 133 tra differ qualitatively between detector types. Current experimental efforts for all pro-
 134 totypes have the common gravity-strain sensitivity target of about $10^{-15} \text{ Hz}^{-1/2}$ at 0.1 Hz.
 135 Below 0.1 Hz, instrumental noise in all concepts rises steeply. The high-frequency noise
 136 spectra differ more strongly. While it is expected that instrumental noise of the super-
 137 conducting gradiometer keeps falling above 0.1 Hz (in units of gravity strain) (Moody
 138 et al., 2002), torsion-bar antennas have a flat noise spectrum above 0.1 Hz (Shoda et al.,
 139 2014), and atom-interferometric gradiometers reach their best sensitivity only within small
 140 frequency bands (Cheinet et al., 2008).

141 For the purpose of this paper, we will use simplified sensitivity models to repre-
 142 sent all GG10 concepts. The simplified approach chosen here is to assume that the sen-
 143 sitivity is proportional to $1/f^2$ at low frequencies, that signal contributions below 0.01 Hz
 144 are not considered (GG10 detectors are not designed for such low-frequency observations),
 145 and that instrumental noise at high frequencies is frequency-independent. To estimate
 146 the detection horizon of gravity strainmeters to earthquakes, four sensitivity models are
 147 tested: flat strain sensitivity of $10^{-15} \text{ Hz}^{-1/2}$ above 0.05 and 0.1 Hz (models 1 and 2,
 148 respectively), $10^{-14} \text{ Hz}^{-1/2}$ above 0.05 Hz (model 3) and $5 \times 10^{-17} \text{ Hz}^{-1/2}$ above 0.5 Hz
 149 (model 4). The resulting sensitivity curves are shown in Fig. 1, along with TOBA Phase
 150 III and SGG sensitivity curves. For the SGG sensitivity curve, a 20 kg mass and a 2 meters-
 151 long baseline are used, along with an energy resolution E_A of the superconducting quantum-
 152 interference device (SQUID) 10 times better than current commercial DC SQUID val-
 153 ues (Griggs, Moody, Norton, Paik, & Venkateswara, 2017).

154 **2.3 Local gravity noise**

155 GG10 detectors have different limiting noise sources and experimental challenges
 156 to reach the target of $10^{-15} \text{ Hz}^{-1/2}$ at 0.1 Hz, specific to each detector. A detailed de-
 157 scription of the contributions of various noise sources to the target sensitivity and the
 158 techniques to reduce them can be found in the references for each detector. But a grav-
 159 ity noise foreground is common to all the detectors: the local gravity noise (LGN). The
 160 LGN has several contributions: seismic LGN produced by density changes in the ground
 161 due to seismic waves; atmospheric LGN generated by density fluctuations in the atmo-
 162 sphere due to, for instance, infrasounds, temperature changes and turbulences; LGN as-
 163 sociated with human activity (Harms, 2015). This noise couples with the detector in a
 164 way completely equivalent to the earthquake signal: it is then impossible to shield the
 165 detector from it.

166 Provided that the detector is located at a sufficiently remote site to avoid transient
 167 contributions to gravity as could be produced by cars or trucks passing close to the de-
 168 tector, the LGN contributions that need to be mitigated further are of seismic and at-
 169 mospheric origin. One way to mitigate the LGN is to select seismically and atmospher-
 170 ically quiet sites. To some extent, this can be achieved by constructing the detector un-
 171 derground, but since seismic and sound waves have long wavelengths around 0.1 Hz, the

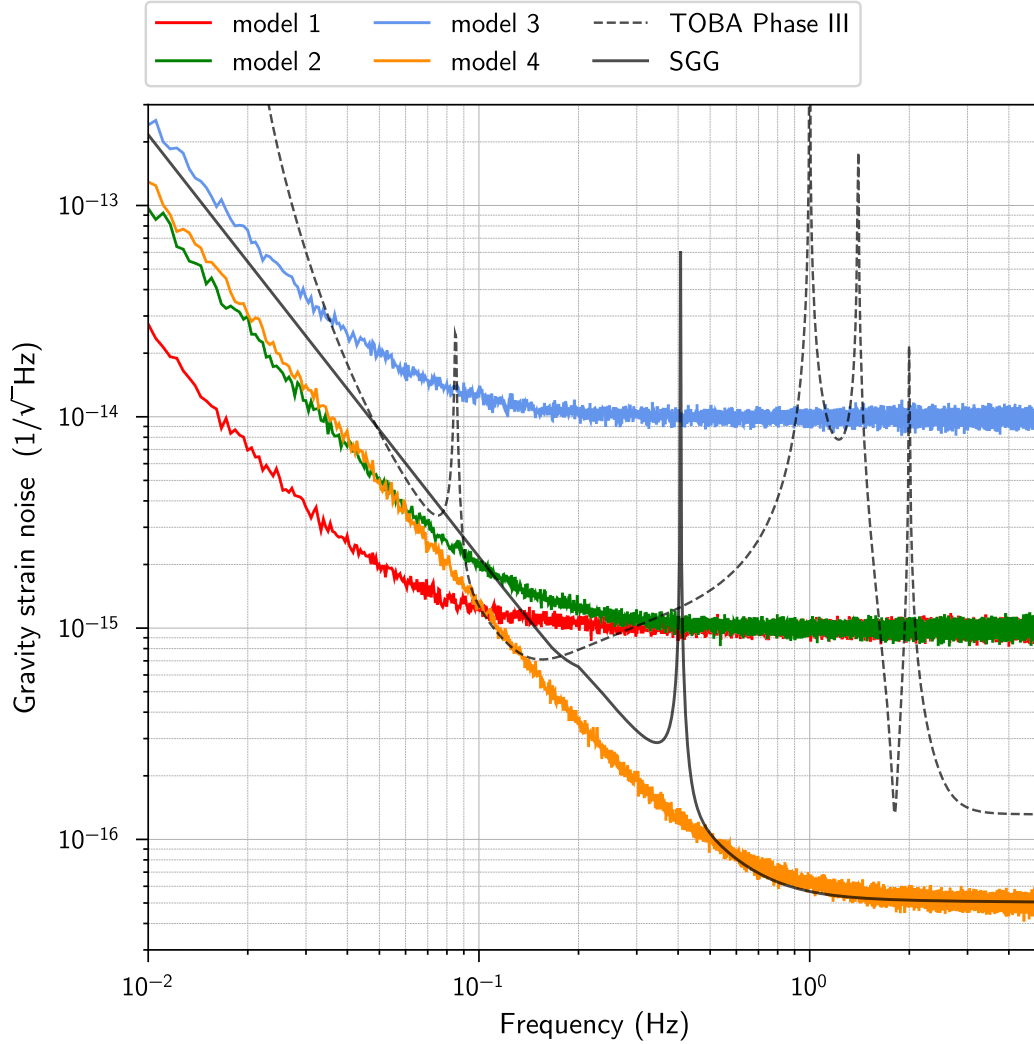


Figure 1. Simplified sensitivity models for gravity gradiometers designed for high-sensitivity around 10 seconds (GG10), along with TOBA Phase III and SGG sensitivity curves. Several noise floors (10^{-14} , 10^{-15} and 5×10^{-17} Hz^{-1/2}) and corner frequencies (0.05, 0.1 and 0.5 Hz) are considered to estimate the detection horizon of future GG10 detectors to earthquake ruptures.

172 associated gravity disturbances are only weakly suppressed underground for feasible de-
173 tector depths (Beker et al., 2011; Fiorucci, Harms, Barsuglia, Fiori, & Paoletti, 2018).

174 As an alternative solution to LGN mitigation, it was proposed to coherently sub-
175 tract LGN using data from arrays of environmental sensors, such as seismometers for the
176 seismic LGN and microphones for the infrasound atmospheric LGN (Cella, 2000; Harms
177 & Paik, 2015). The idea is to obtain sufficient information about local mass-density fluc-
178 tuations to calculate an accurate estimate of the associated LGN. This method exploits
179 correlations between environmental sensors and the GG10 detector by calculating a Wiener
180 filter whose output corresponds to the optimal (linear) estimate of the LGN (Driggers
181 et al., 2012).

182 Based on models for the infrasound atmospheric LGN from Fiorucci et al. (2018)
183 and the even smaller seismic LGN (see for instance Fig. 9 from Harms et al. (2013)), we
184 note that above a few mHz these LGN components should not affect significantly the
185 GG10 sensitivity required for the earthquake-related signal detection. This will be shown
186 in Section 3.2. Nevertheless, we point out that the use of average seismic and infrasound
187 spectra to estimate the LGN components can lead to underestimate the challenge as-
188 sociated with this noise. With this in mind, while work is ongoing to reduce further these
189 noise contributions, it is assumed for the remainder of the paper that all forms of LGN
190 lie below the instrumental noise at all frequencies.

191 **3 Detectability of prompt gravity strain perturbations**

192 **3.1 Optimal matched-filter detection and signal-to-noise ratio**

193 In order to assess the detectability of the gravity perturbation, we compute the signal-
194 to-noise ratio (SNR) obtained with each gravity strainmeter described in section 2.2. To
195 compute the SNR, we consider detection via optimal matched-filtering (Jaranowski &
196 Królak, 2012), which is based on the cross-correlation between the gravity data and a
197 template of the expected gravity strain perturbation. Template-matching techniques have
198 been widely used in modern seismology to detect earthquakes with low SNRs (Frank et
199 al., 2014; Gibbons & Ringdal, 2006; Shelly, Beroza, & Ide, 2007), and in astrophysics to
200 detect coalescing compact binaries (Bose, Dayanga, Ghosh, & Talukder, 2011; Pai, Dhu-
201 randhar, & Bose, 2001).

202 For simplicity, the detector noise is here assumed to be Gaussian and stationary.
 203 Before the matched-filter is applied, both the template and the detector data are passed
 204 through a whitening filter, in order to obtain an approximately frequency-independent
 205 detector noise spectral density. The whitening filter is a highpass filter (Butterworth, 2
 206 poles) whose corner frequency is the corner frequency of the considered instrument sen-
 207 sitivity model. Once the noise spectrum is uniformly distributed, the optimum filter is
 208 the time-reversed, whitened template. The SNR is then defined as the ratio between the
 209 output of the optimal matched-filter in the presence of a signal and the standard devi-
 210 ation of the output in the absence of a signal.

211 3.2 SNR at P-wave arrival time and 10 seconds after onset time

In order to compute the gravity signal templates for a class of earthquakes, a model for source time functions (STF) should be used. In this section we adopt a self-similar source model, which implies that the initial phase of a large-magnitude event is identical to that of a lower-magnitude event. While this universal rupture-initiation behavior is still debated (Colombelli, Zollo, Festa, & Picozzi, 2014; Meier, Ampuero, & Heaton, 2017; Meier, Heaton, & Clinton, 2016), such hypothesis is a classic assumption and represents the worst-case scenario for EEW. Here, the following self-similar model of seismic moment rate function \dot{M}_0 is employed:

$$\dot{M}_0(t) = a \frac{M_0}{T} (t/T)^2 \quad (2)$$

if $0 < t < T$, and

$$\dot{M}_0(t) = a \frac{M_0}{T} \left(1 - (t/T - 1)^2\right)^6 \quad (3)$$

212 if $T < t < 2T$, where T is the half-duration of the rupture and a a scalar. We adopt
 213 the empirical magnitude-duration relation $2T = (M_0 / 10^{16} \text{ N.m})^{1/3}$ (Houston, 2001).
 214 The shape of the first half of this source time function is that of a circular crack with
 215 constant rupture speed and uniform stress drop. The second half is a polynomial approx-
 216 imation to the stopping stage in the moment rate function of the circular crack model
 217 of Madariaga (1976).

218 Based on this source model, we compute gravity strain perturbations for various
 219 magnitude-distance pairs, induced by a dip-slip event with angles (strike, dip, rake) =
 220 (180°, 10°, 90°). We consider magnitudes ranging from 5 to 9.1, and epicentral distances
 221 ranging from 75 to 1100 km. Ten different azimuths are considered, ranging from 270

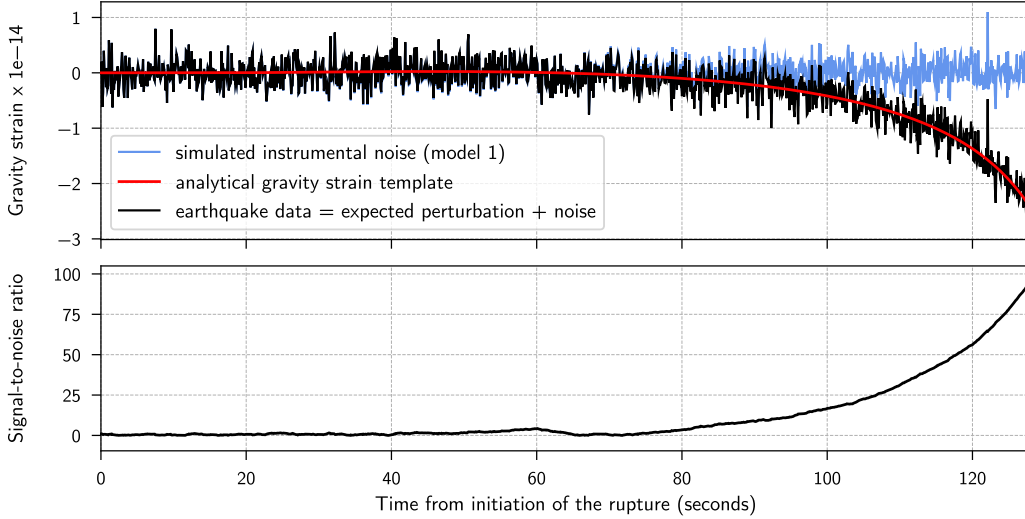


Figure 2. Earthquake data and SNR time series during a M7.5 earthquake. (top) Whitened gravity strain data (black curve, XZ component) recorded by a model-1 gravity strainmeter located 1000 km away from the epicenter of a M8 dip-slip earthquake. The recorded data is obtained as the sum of the whitened instrumental noise (blue curve) and the whitened gravity strain perturbation (red curve, Harms, 2016). (bottom) Corresponding signal-to-noise ratio, defined as the ratio between the data filtered with the time-reversed template, and the standard deviation of the noise filtered with the time-reversed template.. The time series are truncated at P-wave arrival time, 128 seconds after the earthquake onset time.

222 to 360° such that half of the down-dip part of the radiation pattern is computed (the
 223 remaining half being inferred by symmetry).

224 We first compute vertical (Z) gravity perturbations δg_z in a half-space model, us-
 225 ing the analytical formulations developed by Harms (2016). The medium is defined by
 226 a P-wave velocity of 7.8 km/s and an S-wave velocity of 4.4 km/s, such that the P-wave
 227 arrival times for stations located 250 to 1100 kilometers away from the epicenter are glob-
 228 ally comparable to travel times obtained in a more realistic Earth model, such as PREM
 229 (Dziewonski & Anderson, 1981). The XZ and YZ components of the gravity gradient ten-
 230 sor $\nabla\delta g_z$ are then obtained by the finite difference of gravity perturbations computed
 231 at two close locations, aligned along the East-West (X-) and North-South (Y-) directions.
 232 Two integrations over time then lead to the associated gravity strain perturbations.

233 We add simulated instrumental noises, for the four different sensitivity models con-
 234 sidered in section 2.2. For each magnitude and distance, we then apply the optimal matched-
 235 filter (i.e the whitened template used to compute the synthetic earthquake data) with
 236 a finite duration time-window and normalize by the standard deviation of the matched-
 237 filter output in the absence of signal. The result is a set of continuous time series of SNRs,
 238 with values fluctuating around 1 during the absence of a signal. An example of (XZ) grav-
 239 ity strain data and corresponding SNR time series are shown in Fig. 2, for a model-1 grav-
 240 ity strainmeter located 1000 km away from the epicenter of a magnitude M7.5 earthquake,
 241 in the along-dip direction. At P-wave arrival time, 128 seconds after the earthquake on-
 242 set time, the SNR reaches ~ 100 .

243 For each epicentral distance and magnitude, the SNRs accumulated within the travel
 244 time of P-waves to the detector are shown in Fig. 3. Each point of the contour plot is
 245 the average SNR obtained for ten different azimuths, two components of the gravity strain
 246 tensor (XZ and YZ) and a hundred different realizations of the detector noise. As ex-
 247 pected, SNRs globally increase with increasing magnitudes. Two lobes of high SNRs are
 248 observed at high magnitudes, separated by a region of lower SNR where the gravity strain
 249 records reach a zero-crossing at P-wave arrival time. The zero-crossing is observed at longer
 250 epicentral distances for higher magnitudes, since higher magnitudes are associated to STF
 251 with longer half-duration (which acts as a low-pass filter on the gravity strain record).
 252 High SNRs (> 50) are reached within P-wave travel times for detector models with high-
 253 sensitivity around 0.1 Hz (models 1, 2 and 4). The highest ratios (>100) are measured
 254 with the instrument 1, which displays the highest sensitivity at low frequencies. For this
 255 model, SNR larger than 10 are reached for every earthquake of magnitude larger than
 256 7, up to 1000 kilometers from the epicenter. An improved high-frequency sensitivity (from
 257 model 2 to model 4) leads to slightly higher SNRs for epicentral distances below 300 km.

258 The SNRs of signals measured within the first 10 seconds of a fault rupture are shown
 259 in Fig. 4. The SNRs increase with decreasing epicentral distances, and saturate for event
 260 magnitudes greater than 6.5 as a consequence of self-similarity (the moment rate func-
 261 tions of earthquakes of larger magnitudes are identical in the initial 10 seconds). Accord-
 262 ingly, earthquake detection with SNR higher than 10 based on only 10 seconds of data
 263 would require next-stage detectors to be about 100 km or closer to the hypocenter, in-
 264 dependent of the event magnitude above M6.5 (models 1 and 4). Improved low-frequency

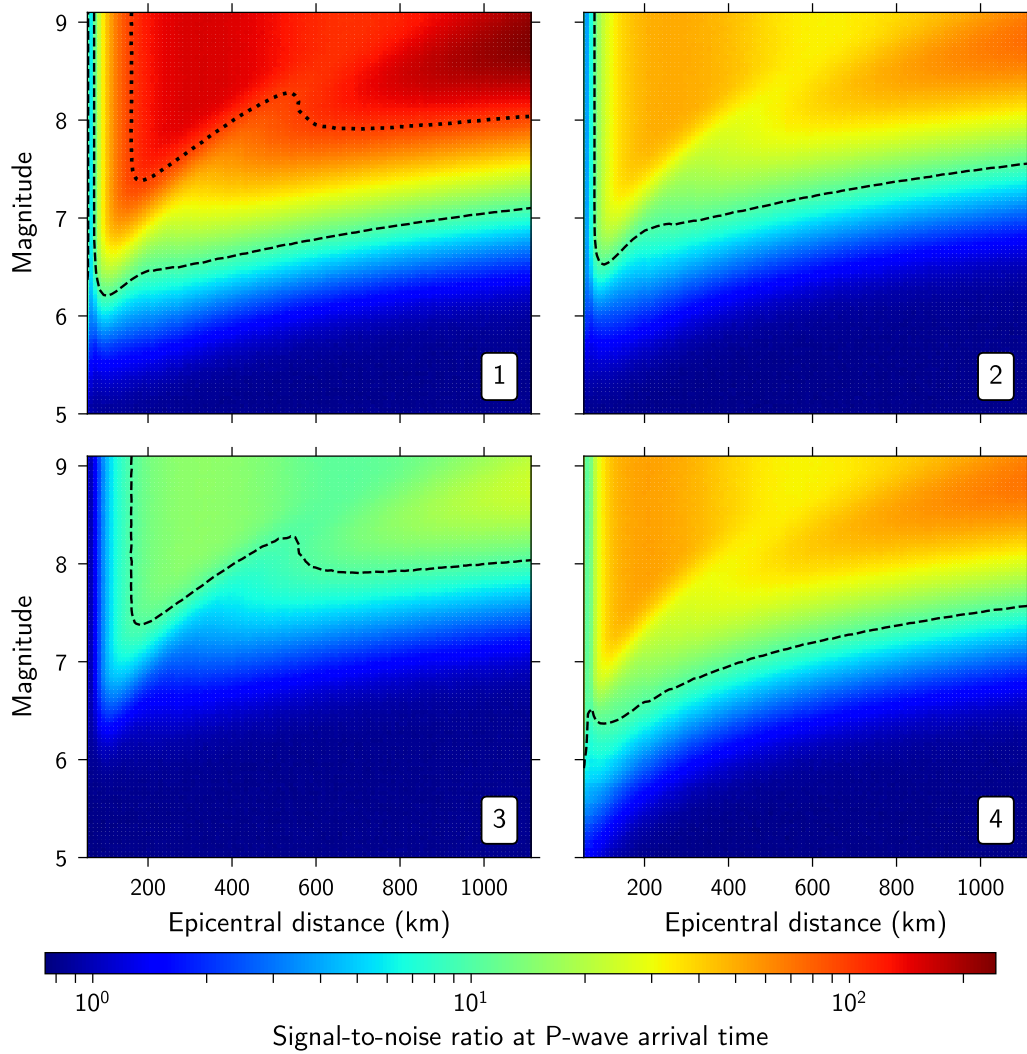


Figure 3. Signal-to-noise ratio accumulated within P-wave travel time to the detector as a function of event magnitude and distance of the detector. The sensitivity models 1 to 4 of the next-stage gravity strainmeter are used (see section 2.2). Contour lines are for SNR = 10 (dashed) and 100 (dotted).

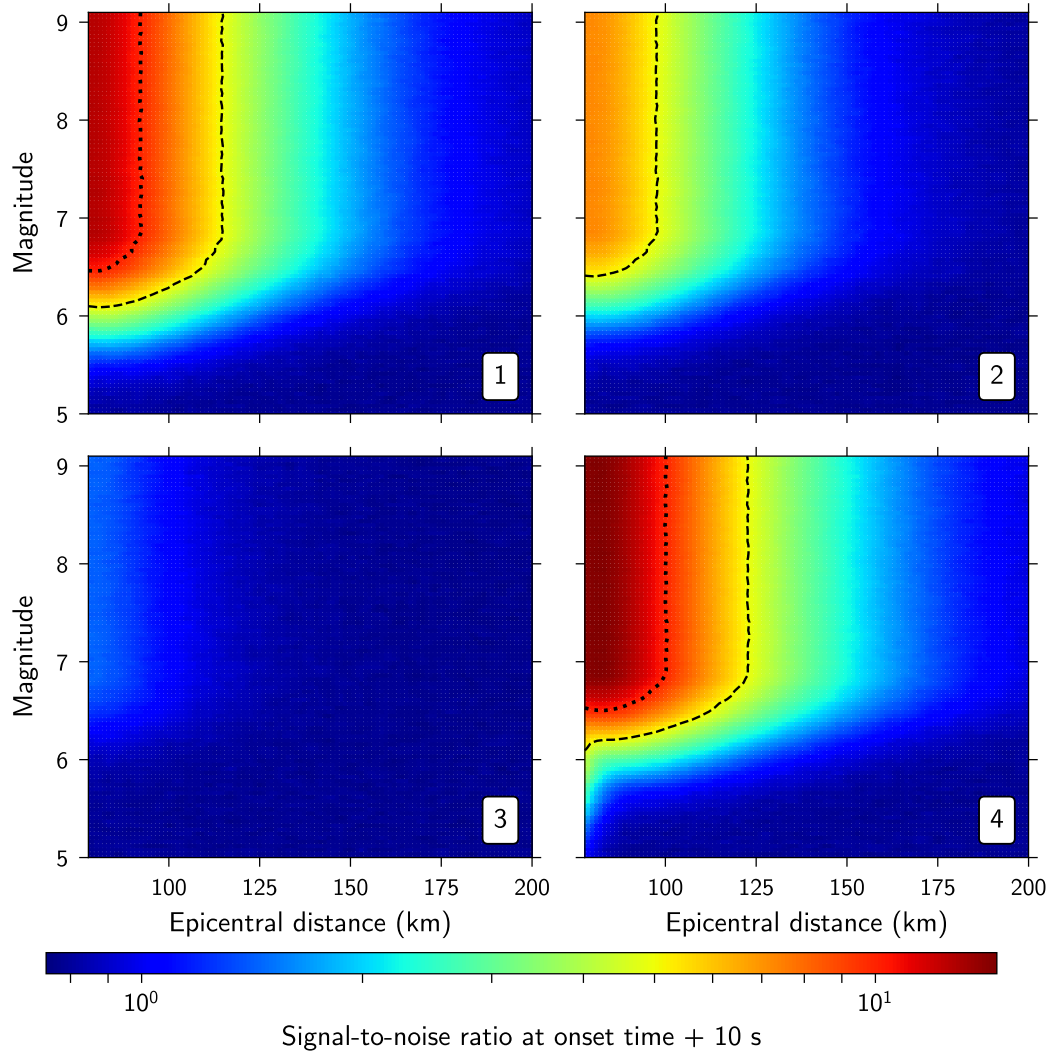


Figure 4. Signal-to-noise ratio accumulated within the first 10 seconds after onset time, as a function of event magnitude and distance of the detector. The sensitivity models 1 to 4 of the next-stage gravity strainmeter are used (see section 2.2). Contour lines are for SNR = 5 (dashed) and 10 (dotted).

265 sensitivity (from model 2 to model 1) and high-frequency sensitivity (from model 2 to
266 model 4) lead to improved SNRs.

267 While the detector corresponding to sensitivity model 3 (10^{-14} Hz $^{-1/2}$ above 0.1 Hz)
268 should detect events of magnitude M7.5 and above with SNR higher than 10 at P-wave
269 arrival time (see Fig. 3), its use for EEW purposes appears to be limited (see Fig. 4).
270 Prompt detections of earthquake-induced perturbations thus require the strain sensitiv-
271 ity target of 10^{-15} Hz $^{-1/2}$ at 0.1 Hz.

272 The magnitudes of the first seconds of real earthquakes can significantly exceed the
273 predictions of our self-similar source model (the 10-seconds SNRs saturate at magnitude
274 6.5). Thus, our source model can be considered as conservative for the estimation of SNR.

275 **4 Potential application to earthquake early warning**

276 The foregoing analysis shows that large earthquakes can induce significant grav-
277 ity strain perturbations at long distances. These perturbations are essentially instanta-
278 neous, compared to seismic wave propagation time-scales. This property opens new prospects
279 for the rapid estimate of earthquake source parameters and its application to the mit-
280 igation of earthquake and tsunami hazards. Here we discuss how this feature can be ex-
281 ploited to improve the capabilities of one of the most challenging applications in real-
282 time seismology: earthquake early warning. In particular, we demonstrate the potential
283 contribution to EEW of a regional network of gravity strainmeters.

284 **4.1 Earthquake early warning systems**

285 EEWS aim to detect the occurrence of an earthquake as soon as possible and broad-
286 cast an alert before the arrival of strong seismic shaking (R. M. Allen et al., 2009; Böse
287 et al., 2014). They are a crucial tool to reduce damage and injuries from earthquakes.
288 Existing EEWS are based on the difference of propagation speed between P-waves (fast,
289 compressional) and S-waves (slower, shear) and on the fact that the strongest, most dam-
290 aging ground motions are carried by S-waves and (even slower) surface waves.

291 On-site EEWS exploit the first few seconds of recorded ground motions (mainly
292 P-wave) to estimate the expected ground shaking at the recording site (Kanamori, 2005;
293 Zollo, Amoroso, Lancieri, Wu, & Kanamori, 2010). Estimates of the event location or
294 magnitude are not necessarily made, such that no regional warnings are issued. Network-

295 based EEWS use multiple stations closest to the epicenter to estimate the event loca-
296 tion, final magnitude and origin time before issuing a regional warning. Recent advances
297 include algorithms and systems to account for the finite rupture size of large-magnitude
298 events (e.g. Böse, Heaton, & Hauksson, 2012; Meng, Allen, & Ampuero, 2014; Minson,
299 Murray, Langbein, & Gomberg, 2014).

300 The finite speed of seismic waves, the density of a seismic network, signal trans-
301 mission delays, the minimal number of stations and signal duration required to estimate
302 earthquake magnitude impose together a minimum on the warning time and on the dis-
303 tance that can be reached by an early warning. This minimum distance is the distance
304 travelled by the S-waves at the time the warning is issued and defines the “blind zone”
305 of an EEWS. Seismic travel times can be longer than 10 seconds for offshore or deep earth-
306 quakes: if an EEWS could be based on other geophysical signals with negligible travel
307 time, valuable seconds could be gained. The instantaneous character of earthquake-induced
308 gravity perturbations in principle offers an opportunity for such a gain.

309 Every saved second can have an important impact in terms of life preservation and
310 earthquake mitigation, since advanced warning enables the launching of automatic pre-
311 ventions systems and the implementation of safety procedures (R. Allen, 2013). Short though
312 it may seem, a warning time of a few seconds is longer than the response time of auto-
313 mated control systems that mitigate the impact of strong shaking on industrial facili-
314 ties and infrastructure, such as Japanese bullet trains (Nakamura & Saita, 2007) or on
315 gas and chemical supply systems and high-precision equipment in a Japanese semicon-
316 ductor factory (Takamatsu, 2009). A few seconds gain in warning time can also enable
317 or enhance significantly personal protective actions (e.g. most people can perform ”Duck,
318 Cover and Hold On” within 10 seconds after warning) (Porter, 2016). Moreover, the ra-
319 dius of the blind zone would be reduced. Considering that earthquake intensity is much
320 higher at closer distance, this could result in a significant potential reduction of dam-
321 age and life loss.

322 **4.2 Real-time event detection with a gravity-based EEWS**

323 ***4.2.1 Template matched-filtering and event detection***

A gravity-based EEWS must first detect in real-time the initiation of an earthquake,
based on a few seconds of continuous recordings by a regional network of gravity strain-

meters. We propose to assess the real-time likelihood of an earthquake rupture with a network-based matched-filter approach. Our template matched-filter relies on the computation of an average likelihood ratio (LR), basically a multi-channel SNR, among all pairs of sensors and components, computed in a running window as:

$$\text{LR}(t) = \frac{\sum_i^{N_i} \sum_j^{N_j} \sum_n^N h_{ij}(n\Delta t) \hat{s}_{ij}(t - n\Delta t)}{\sqrt{\sum_i^{N_i} \sum_j^{N_j} \sum_n^N h_{ij}(n\Delta t)^2}}. \quad (4)$$

324 h and \hat{s} are respectively the pre-whitened gravity strain template and pre-whitened, nor-
 325 malized continuous gravity strain record. \hat{s} is normalized such that its root mean square
 326 is unity in the absence of signal. i and j are indices for the sensors and components, N_i
 327 and N_j their corresponding numbers, while N stands for the number of samples in the
 328 sliding window. We note that, in contrast to what is classically used in seismology, there
 329 is no moveout in equation (4) to describe the differential arrival times on each station,
 330 since the earthquake-induced gravity perturbations propagate almost instantaneously
 331 to each gravity strain sensor.

332 Detection is based on a threshold applied to the LR values, and can be achieved
 333 by communicating data in real-time to a central processor or by implementing a more
 334 distributed communication scheme. The threshold is set by the end-user, depending on
 335 his tolerance level on false alarm rates and missed event rates.

336 ***4.2.2 Template database and event parameter estimation***

337 A regional EEWS requires not only earthquake detection, but also an estimate of
 338 the source location, origin time, fault mechanism and seismic moment. This requires a
 339 library of signal templates, which is here composed of a collection of analytical gravity
 340 strain Green's functions, computed between the sensor locations and a source location
 341 grid. The gravity strain Green's functions are later convolved with a catalog of moment
 342 rate source time functions. We drop here the self-similar source model, as we are inter-
 343 ested in the detection of every event, including those that do not respect self-similarity.
 344 Thus, a whole range of source half-durations is considered for a given template final mo-
 345 ment (set to 1 N.m). Two types of onset are considered, with moment rate STF grow-
 346 ing either linearly (isosceles triangular STF) or quadratically with time (STF described
 347 in equations (2) and (3)).

348 Source parameters and their uncertainties are then estimated in real-time as the
 349 parameters corresponding to the template with the highest LR, and could be provided
 350 to the EEWS decision module with regular updates every second or so. The use of a pre-
 351 computed template database reduces computational times for real-time operation.

An estimate of the earthquake magnitude can be obtained from the scaling factor α between the optimal template \mathbf{t} (whose seismic moment is prescribed by the user) and the actual recorded gravity strain \mathbf{s} . The factor α is the least-squares estimate of the scaling factor between \mathbf{t} and \mathbf{s} , averaged over the sensors and components:

$$\alpha(t) = \frac{1}{N_i} \frac{1}{N_j} \sum_i^{N_i} \sum_j^{N_j} \frac{\sum_n^N t_{ij}(n\Delta t) s_{ij}(t - n\Delta t)}{\sum_n^N t_{ij}(n\Delta t)^2}. \quad (5)$$

352 The estimated earthquake moment is α times the template moment.

353 Templates are correlated fully or by parts, such that we monitor the growth of the
 354 rupture as soon as it begins. We point out that in comparison to conventional EEW sys-
 355 tems which estimate the final magnitude of an earthquake, our proposed system estimates
 356 the instantaneous magnitude, i.e. the seismic moment released up to the time the esti-
 357 mate is made. In principle, at stations that are located beyond the P-wave front at the
 358 earthquake end-time, rupture arrest can be diagnosed, and the final magnitude estimated.

359 *4.2.3 Rupture scenario*

360 To demonstrate the potential of a gravity-based EEWS, we focus on large off-shore
 361 subduction earthquakes. We thus consider in this subsection rupture scenarii of the 2011
 362 M9.1 Tohoku-oki earthquake and its M7.3 foreshock, as it would have been recorded by
 363 networks of gravity strain sensors. Two different networks of three gravity strainmeters
 364 are considered, one close to the ruptured areas (~ 250 km from the epicenter), the other
 365 at regional distances (~ 1100 km from the epicenter).

366 Gravity strain data is obtained through the sum of the analytical perturbations and
 367 the simulated instrumental noises for two components of the gravity strain tensor, as al-
 368 ready performed in section 3. We choose here to consider the intermediate detector sen-
 369 sitivity model 1, with a gravity-strain sensitivity of 10^{-15} Hz $^{-1/2}$ at 0.1 Hz.

370 We use the Global Centroid Moment Tensor (GCMT, Ekström, Nettles, & Dziewoński,
 371 2012) parameters for the epicenter coordinates ((37.52°N, 143.05°E) for the Tohoku event,
 372 (38.56°N, 142.78°E) for its foreshock) and fault geometries ((strike/dip/rake) = (203°/10°/88°)

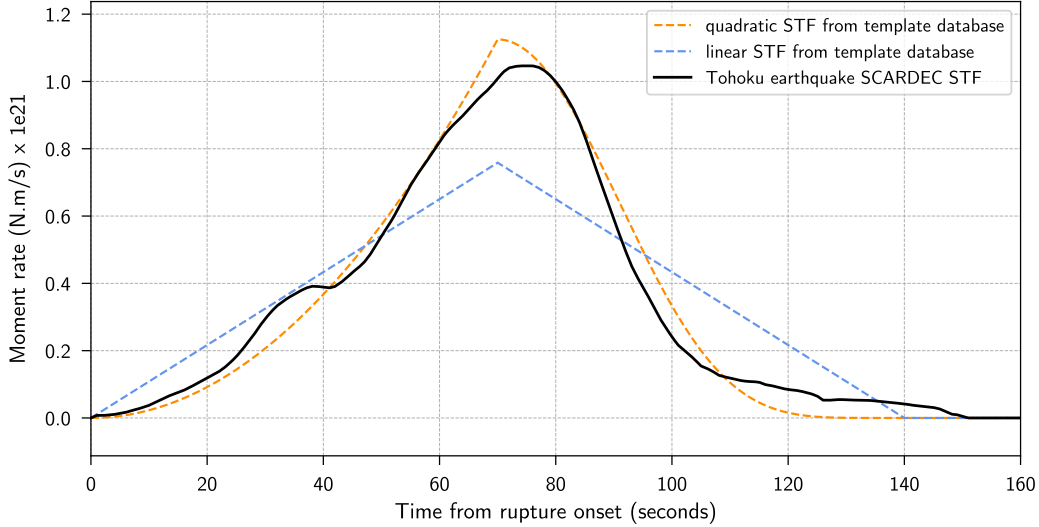


Figure 5. SCARDEC solution for the moment-rate STF of the M9.1 Tohoku earthquake (solid line), along with the closest templates (dashed lines). STF templates grow either linearly (blue line) or quadratically (orange line) with time. STFs are normalized to 1 N.m in the template database, but are here scaled to the event final moment for plotting purposes.

373 for the Tohoku event, ($189^\circ/12^\circ/78^\circ$) for its foreshock), and the source time functions
 374 from the SCARDEC catalog (Vallée, Charléty, Ferreira, Delouis, & Vergoz, 2011).

375 In order to build a template database, gravity strain Green’s functions are com-
 376 puted analytically for three different dip-slip fault mechanisms, which approximate the
 377 subduction context near the Japan Trench (strike = 180° , 190° or 200° , dip = 10° and
 378 rake = 90° , depth 20 km). Two types of STF onset (linear and quadratic) and source
 379 half-durations ranging from 2 to 90 seconds are considered. It should be noted that the
 380 actual source solutions for the Tohoku event and its foreshock are not in the present tem-
 381 plate database. Examples from the STF database, along with SCARDEC solution for
 382 the Tohoku earthquake are plotted in Fig. 5. Additional priors based on the geometry
 383 of major active faults in the area, and a catalog of moment rate functions from past earth-
 384 quakes could complement the template database in the future.

385 As an example, the gravity strain recordings 15 seconds after the Tohoku earth-
 386 quake onset time, along with the corresponding optimal template are shown for one rup-
 387 ture scenario in Fig. 6. The earthquake would have been detected with a high LR (> 50)
 388 and accurate estimates for the epicenter location, onset time and released magnitude,

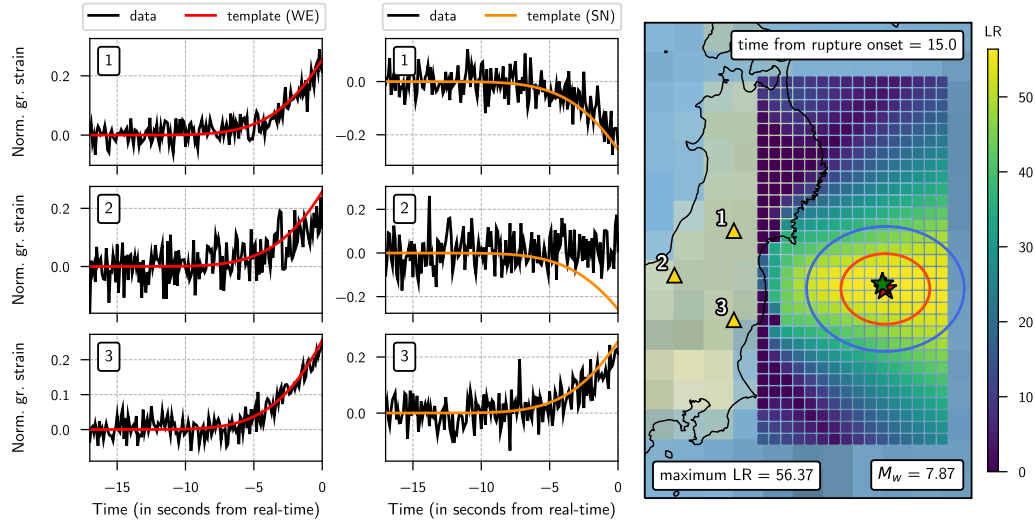


Figure 6. Offline real-time detection, source location and magnitude estimation of the M9.1 Tohoku earthquake, 15 seconds after onset time. (left) Gravity strain recording (black lines), along with the best-fitting gravity strain template (red and orange lines). Both gravity strain record and template are normalized by their respective root mean square, for plotting purposes. (right) The red and green stars represent the earthquake epicenter location and its estimated location, respectively. 15 seconds after onset time, P- (blue) and S-wavefronts (red) have not yet reached the coasts, while a M7.9 earthquake is detected with a high likelihood ratio (> 50), close to the actual M7.7 magnitude.

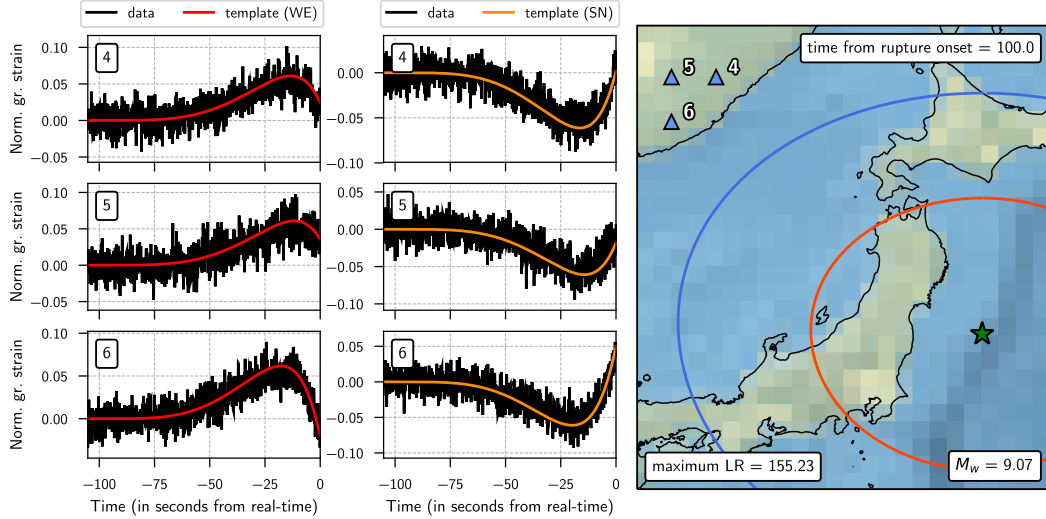


Figure 7. Offline real-time magnitude estimation of the M9.1 Tohoku earthquake, 100 seconds after onset time. (left) Gravity strain recording (black lines), along with the best-fitting gravity strain template (red and orange lines). Both gravity strain record and template are normalized by their respective root mean square, for plotting purposes. (right) The epicenter location is supposed to be known. 100 seconds after onset time, a $> M9$ earthquake is detected.

389 five seconds before the P-waves even hit the Japanese coastlines. The optimal template
 390 corresponds to a rupture still ongoing, while the lower bound magnitude estimate of the
 391 rupture ($> M8.7$) is high enough to issue a regional warning.

392 The seismic P-wavefront reaches sensors considered in Fig. 6 before the Tohoku earth-
 393 quake end-time, such that an estimate of the earthquake final magnitude cannot be made
 394 based on these sensors. A network of gravity strainmeters located further away from the
 395 rupture area (beyond ~ 1000 km) could monitor the whole rupture. This is the case for
 396 the second network we considered. See for instance the gravity strain recordings and cor-
 397 responding optimal templates 100 seconds after onset time in Fig. 7.

398 A hundred different rupture scenarii for the M9.1 Tohoku earthquake and its M7.3
 399 foreshock have been simulated, with a different random realization of instrument noise
 400 computed for each scenario. The earthquake magnitudes are estimated in real-time by
 401 the two networks, the corresponding results are displayed in Fig. 8. The input seismic
 402 moment functions from the SCARDEC database are accurately retrieved.

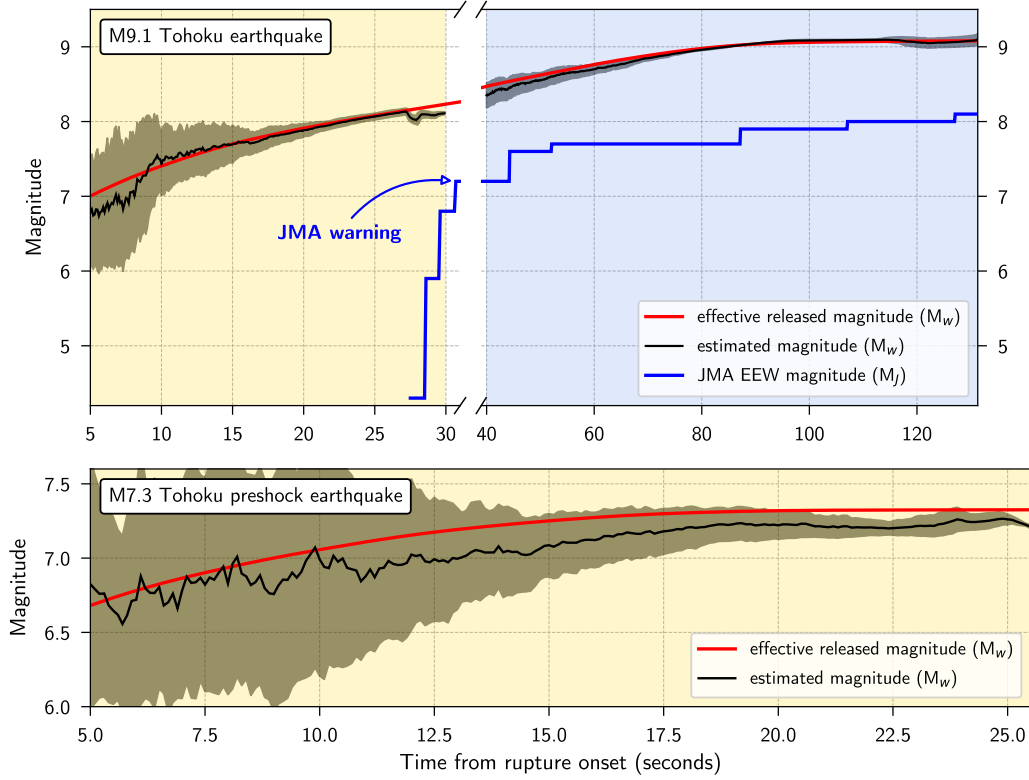


Figure 8. Rupture scenarii of the M9.1 Tohoku earthquake (top) and its M7.3 foreshock (bottom). The red lines represent the released magnitude (STF from the SCARDEC database), while the thick black lines represent the average estimated magnitudes over 100 iterations (bounded with the ± 1 standard deviation curves). The yellow panels indicate estimations provided by the network deployed close to the ruptured areas, and the blue panel estimations from the network at regional distance. Estimations are provided up to the seismic waves arrival at the network. The thick blue line represents the EEW magnitudes issued in real-time during the Tohoku earthquake by the JMA (Hoshiya & Ozaki, 2014).

403 To appreciate the magnitude uncertainties, we compute the standard deviation of
 404 the 100 estimated magnitudes from the random realizations. Based on the gravity strain
 405 data recorded by the local network, the LR exceeds 50 only 15 seconds after the Tohoku
 406 onset time, with a corresponding $M7.70 \pm 0.26$ magnitude estimate (the actual released
 407 moment is M7.7). For the foreshock, LR exceeds 50 only 18 seconds after the onset time,
 408 with a corresponding $M7.21 \pm 0.14$ magnitude estimate (actual released moment M7.31).
 409 In both cases, within 10 seconds of the rupture onset (at least 10 seconds before P-waves
 410 arrived to the coast) it could have been determined that the earthquake magnitude was
 411 likely to exceed M7. According to the SCARDEC STF, the Tohoku event released mo-
 412 ment exceeded the M9 magnitude 83 seconds after its onset time. Our magnitude esti-
 413 mate based on gravity-strain data recorded by the regional network is, in this idealized
 414 exercise, in perfect agreement, as it reaches at that time $M8.99 \pm 0.01$.

415 Estimates of seismic moment M_0 based on early P- and S-wave signals are usually
 416 highly uncertain and underestimated (see for instance JMA warnings issued from seis-
 417 mic data in Fig. 8). Robust estimates of M_0 for large earthquakes based on W-phases
 418 (Kanamori & Rivera, 2008) can be obtained after at least 20 minutes (Duputel et al.,
 419 2011). Seismogeodetic methods based on high-rate GPS data can, in principle, provide
 420 rapid source parameter estimates within a few tens of seconds of the earthquake initi-
 421 ation (Crowell et al., 2016; Ruhl, Melgar, Grapenthin, & Allen, 2017). A robust estimate
 422 of M_0 derived from gravity signals might be obtained earlier, and fully stable at the end
 423 of the rupture, which could significantly enhance tsunami warning systems in the near-
 424 source region.

425 5 Conclusion

426 We have established key quantitative results regarding the capabilities that future
 427 gravity strainmeters could achieve for earthquake source characterization, based on earthquake-
 428 induced prompt gravity signals before direct seismic wave arrivals. We computed earthquake-
 429 induced prompt gravity strain signals with an analytical model in a homogeneous half-
 430 space, and tested various gravity strainmeters sensitivity models. Considering the planned
 431 sensitivities of about $10^{-15} \text{ Hz}^{-1/2}$ at 0.1 Hz for the sub-Hz gravity strainmeters under
 432 development for GW detection, we have demonstrated that prompt perturbations induced
 433 by earthquakes larger than M7 can be observed with a single gravity detector at distances

434 shorter than 1000 km from the epicenter within the travel time of P-waves (SNR > 10),
435 and up to 120 km within 10 seconds of the earthquake onset time (SNR > 5).

436 Since gravity field fluctuations propagate essentially instantaneously in compari-
437 son with seismic waves, a very promising application of this study is the improvement
438 of the performance of EEWS. We demonstrate that a potential benefit of a gravity-based
439 EEWS compared to seismic-based EEWS is earlier warning for large off-shore subduc-
440 tion earthquakes. Moreover, one of the current issues in EEWS is the estimation of the
441 magnitude of the event, especially for mega earthquakes. Our simulations illustrate how
442 gravity strainmeters could accelerate the estimation of the magnitude of mega earthquakes,
443 by providing robust magnitude estimates within the duration of the fault rupture. We
444 therefore propose that gravity strain data has the potential to complement other geo-
445 physical data in the future, to enhance tsunami warning systems.

446 While the foregoing discussion presents key elements of a gravity-based EEWS, a
447 thorough assessment of the feasibility of this concept requires further developments: grav-
448 ity signal predictions in more realistic earthquake scenarios that incorporate the effects
449 of finite rupture size, Earth heterogeneities, attenuation and scattering, the analysis of
450 the inverse problem of location and magnitude estimation to determine the optimal sen-
451 sor network geometry, the assessment of the impact of non-stationary noise, and a cost-
452 benefit analysis of its integration with existing EEWS. The most difficult challenge will
453 be the development of gravity gradiometers with strain sensitivity of about $10^{-15} \text{ Hz}^{-1/2}$
454 at 0.1 Hz.

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