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Efficient simulation of prompt elasto-gravity signals (PEGS) based on a spherical self-gravitating earth model

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Reference: Zhang et al. (EPSL 2020), https://doi.org/10.1016/j.epsl.2020.116150

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What are PEGS signals?

How can we measure and model them?

Are they useful for early warning?

What are PEGS signals?

Fault rupture
(earthquake)

Elastic waves
$\boldsymbol{u}$

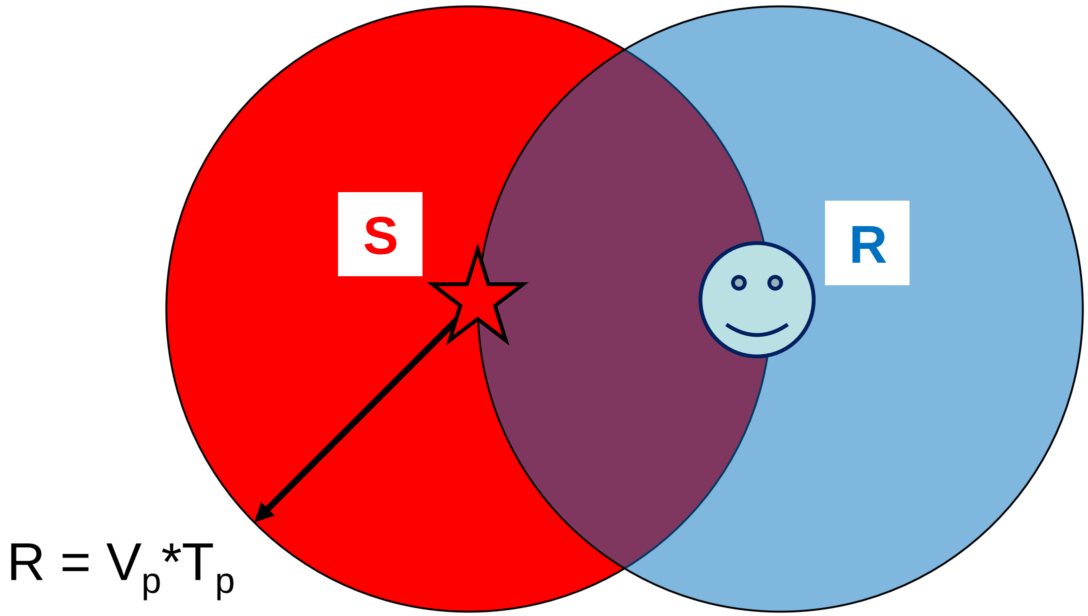
Mass redistribution
$\delta\rho = -\nabla \cdot (\rho \boldsymbol{u})$

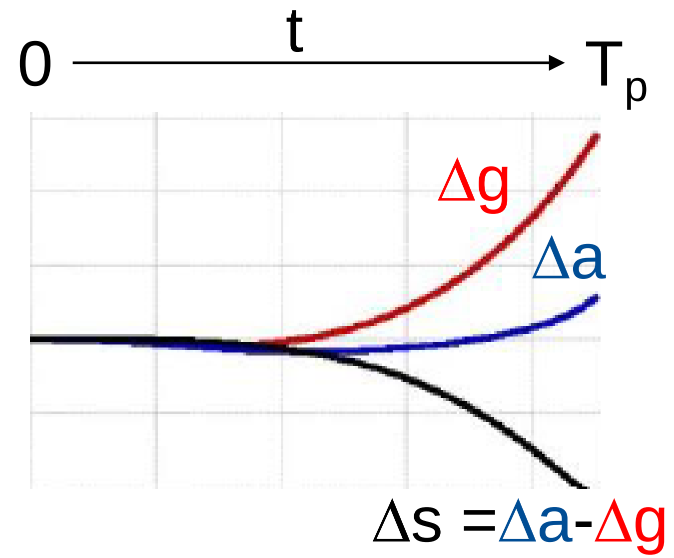
Elastogravitational coupling

Gravity change
$\nabla^2 \psi = -4\pi G \delta\rho$

Secondary sources
$\delta \boldsymbol{g} = \nabla \psi$

Propagates with speed of light!





What signals can arrive at receiver before t = T_p?

1. Gravity perturbation Δg due to mass redistribution within Volume S.
2. Ground motion Δa generated by secondary sources in Volume R.

How are they measured?

- ✓ Montagner et al.(2016) detected the PPEG signals of the 2011 Tohoku earthquake from records of a superconducting gravimeter (SG).
- ✓ Vallée et al. (2017) reported their more reliable detection of the same signals from low-noise records of 11 broadband seismometers (BB).
- ✓ Kimura et al. (2019) tried to verify the previous detections using array stacking of **SG (failed)**, **BB (successful)** and **tiltmeters (failed)**.
- ✓ Vallée and Juhel (2019) presented new detections for 5 other large earthquakes (Mw $\geq$ 8.5) with different focal mechanisms.
- ✓ …
- ✓ Future: **Use of GW detector** measuring gravity gradient (expected resolution $10^{-13}/s^2$) based on the general relativity theory are being developed in Japan (Juhel et al. 2018).

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REPORT

Observations and modeling of the elastogravity signals preceding direct seismic waves

Martin Vallée[1,*], Jean Paul Ampuero[2], Kévin Juhel[1], Pascal Bernard[1], Jean-Paul Montagner[1], Matteo Barsuglia[3]
+ See all authors and affiliations

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Article Figures & Data Info & Metrics eLetters PDF

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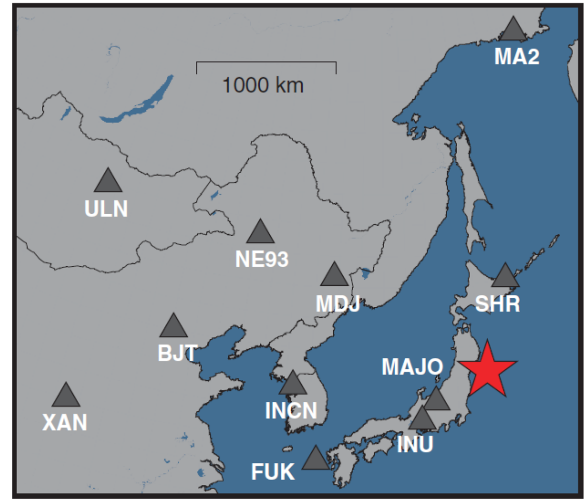
Gravity gets into the earthquake game

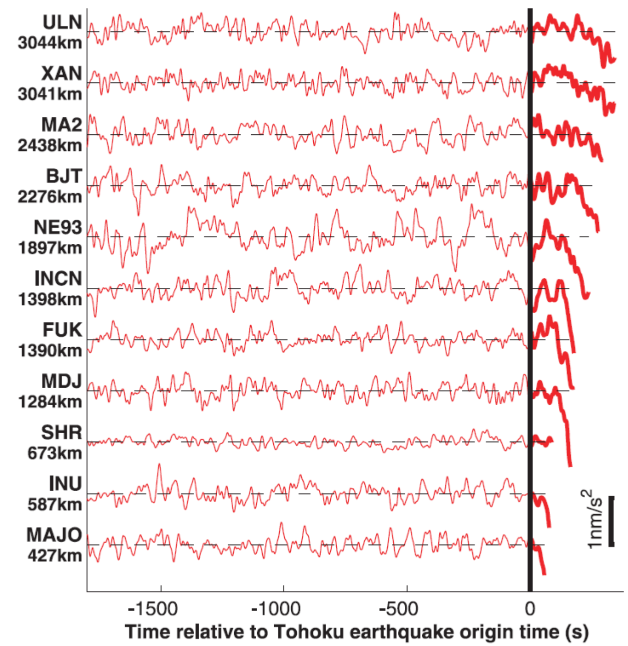
Earthquakes generate large movements of mass, which slightly change the gravitational field. Unlike the elastic waves that propagate from the earthquake, the gravity perturbations travel at the speed of light. Vallée *et al.* have finally observed these gravity perturbations in seismometer records from the great Tohoku earthquake in Japan in 2011. The signal would have allowed an accurate magnitude estimation in minutes, rather than hours, for this catastrophic earthquake.

Science, this issue p. 1164

Abstract

After an earthquake, the earliest deformation signals are not expected to be carried by the fastest (*P*) elastic waves but by the speed-of-light changes of the gravitational field. However, these perturbations are weak and, so far, their detection has not been accurate enough to fully





How are they modelled?

- ✓ Harms et al. (2015) calculated the transient PPEG gravity change based on a full-space model (only the gravity change).
- ✓ Harms et al. (2016) extended the full-space model to a half-space model (still only the gravity change).
- ✓ Heaton (2017) pointed out that all geophysical instruments as a spring-mass system response not only to the gravity change, but also to the inertial acceleration induced by it.
- ✓ Vallée et al. (2017) simulated the two effects successfully, but in a very primitive and time-consuming way:

Synthetic seismograms with reflectivity code → Discrete spatial density variations in Volume S → Transient gravity perturbation at station (Δg) and at discrete elements in Volume R as secondary single-force sources → Synthetic seismograms for all secondary sources at station (Δa) ⇒ ***Million of synthetics are needed***

A new approach
for simulating PPEG signals with the code

QSSP

by

Wang et al. (2017). **Complete synthetic seismograms based on a spherical self-gravitating Earth model with an atmosphere-ocean-mantle-core structure**. Geophysical Journal International.

Theory:

Equations of motion

(momentum equilibrium, Poisson's equation)

$$\begin{cases} \rho \dfrac{\partial^2 \boldsymbol{u}}{\partial t^2} = \nabla \cdot \boldsymbol{\sigma} + \rho \nabla (\psi - g u_r) + \rho g (\nabla \cdot \boldsymbol{u}) \boldsymbol{e}_r + \boldsymbol{f} \\ \nabla^2 \psi \ = \ 4\pi G \nabla \cdot (\rho \boldsymbol{u}) \end{cases}$$

Observables

(displacement, incremental gravity potential)

$$\boldsymbol{u} = \boldsymbol{u}(r, \theta, \varphi, t)$$

$$\psi = \psi(r, \theta, \varphi, t)$$

Source

(single forces, dislocations)

$$\boldsymbol{f} = \boldsymbol{f}(r, \theta, \varphi, t)$$

Earth model

(spherical, self-gravitating, elastic)

$$\rho = \rho(r), \ \ g = g(r),$$

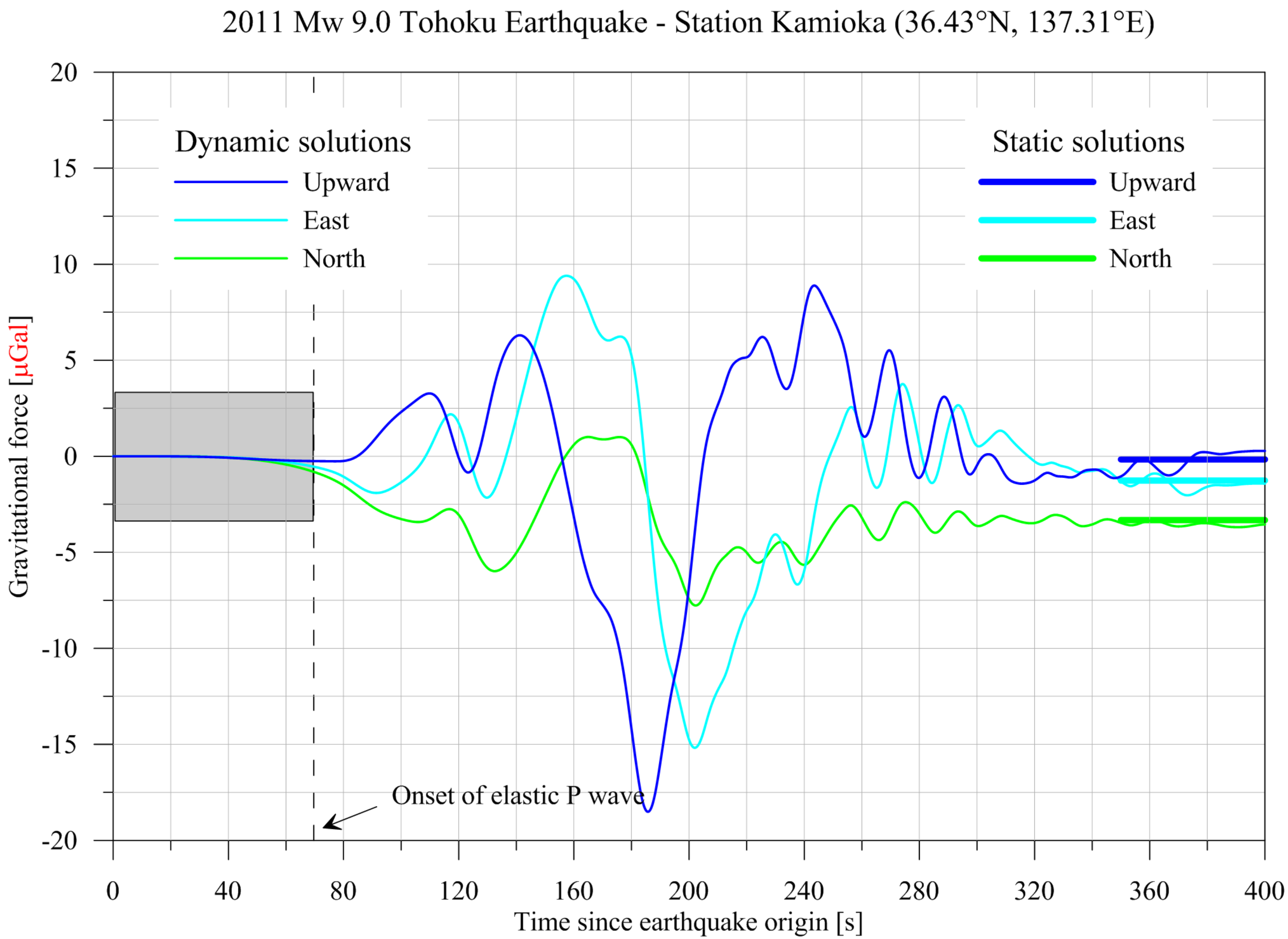
$$\lambda = \lambda(r), \ \ \mu = \mu(r).$$

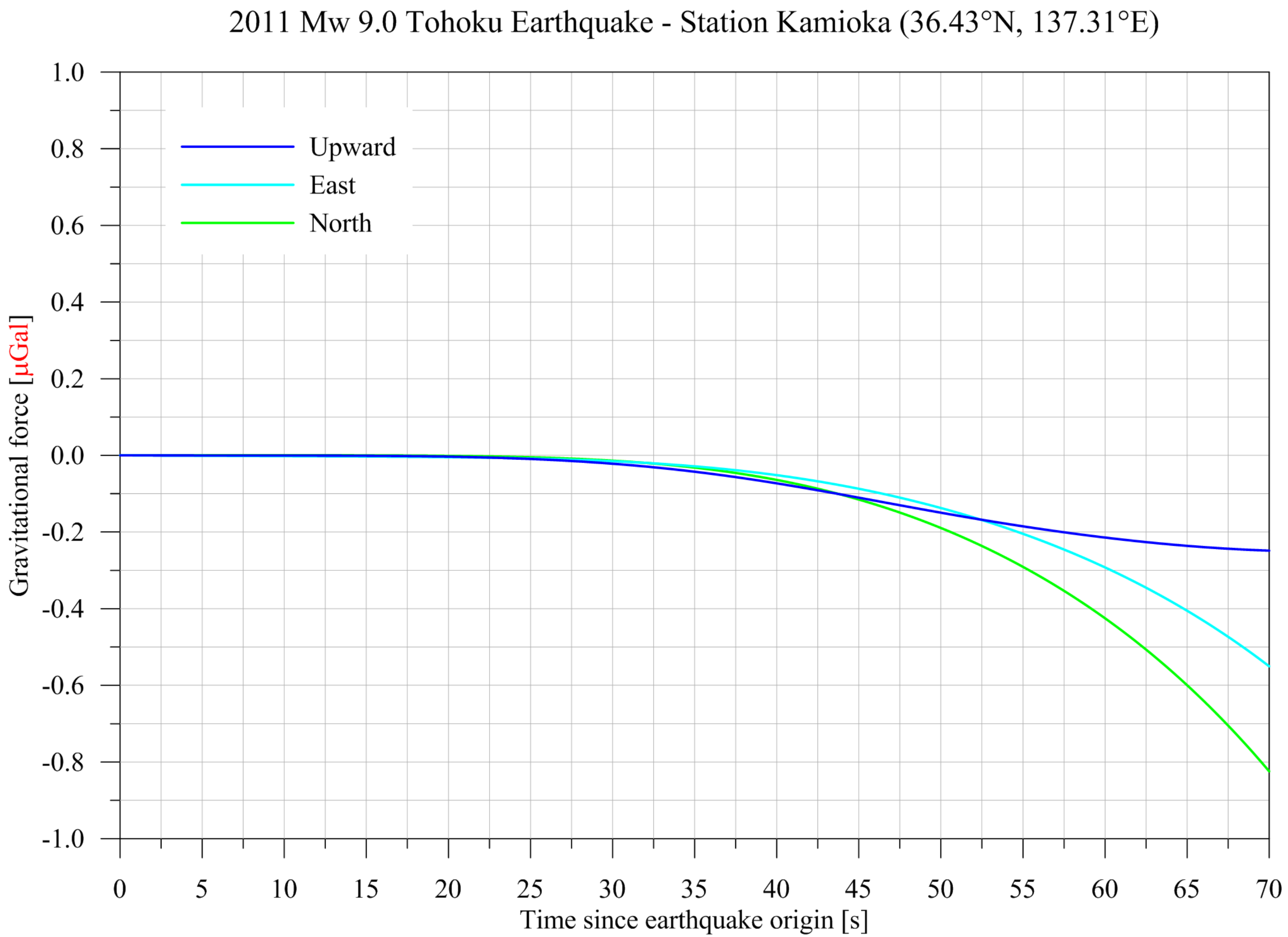
qssp2017-tohoku.inp* - Microsoft Visual Studio (Administrator)

```
#
#   RECEIVER PARAMETERS
#   ===================
# 1. select output observables (1/0 = yes/no)
#    Note: the gravitation defined here is the space-based gravitational attraction force, and the gravity is
#          the response of ground-based gravimeter which includes the gravitational attraction, the inertial
#          force and the free-air gradient effect (all positve downwards).
# 2. output file name
# 3. output time window [sec] (<= Green's function time window)
# 4. selection of order of Butterworth bandpass filter (if <= 0, then no filtering), lower
#    and upper corner frequencies (<= cut-off frequency defined above)
# 5. lower and upper slowness cut-off [s/km] (slowness band-pass filter)
# 6. number of receiver
# 7. list of the station parameters
#    Format:
#    Lat      Lon     Name      Time_reduction
#    [deg]   [deg]             [sec]
#    (Note: Time_reduction = start time of the time window)
#-------------------------------------------------------------------------------------------------------------
# disp | velo | acce | strain | strn_rate | stress | strs_rate | rotation | rot_rate | gravitation | gravity
#-------------------------------------------------------------------------------------------------------------
  1      1      1      0        0           0        0           0          0          1             1
  'tohoku'
  500.0
  3   0.00   0.20
  0.00   0.40
  1
  36.43   137.31  'Kam'  0.0
```

QSSP output seismograms

$$\Delta \boldsymbol{g}$$

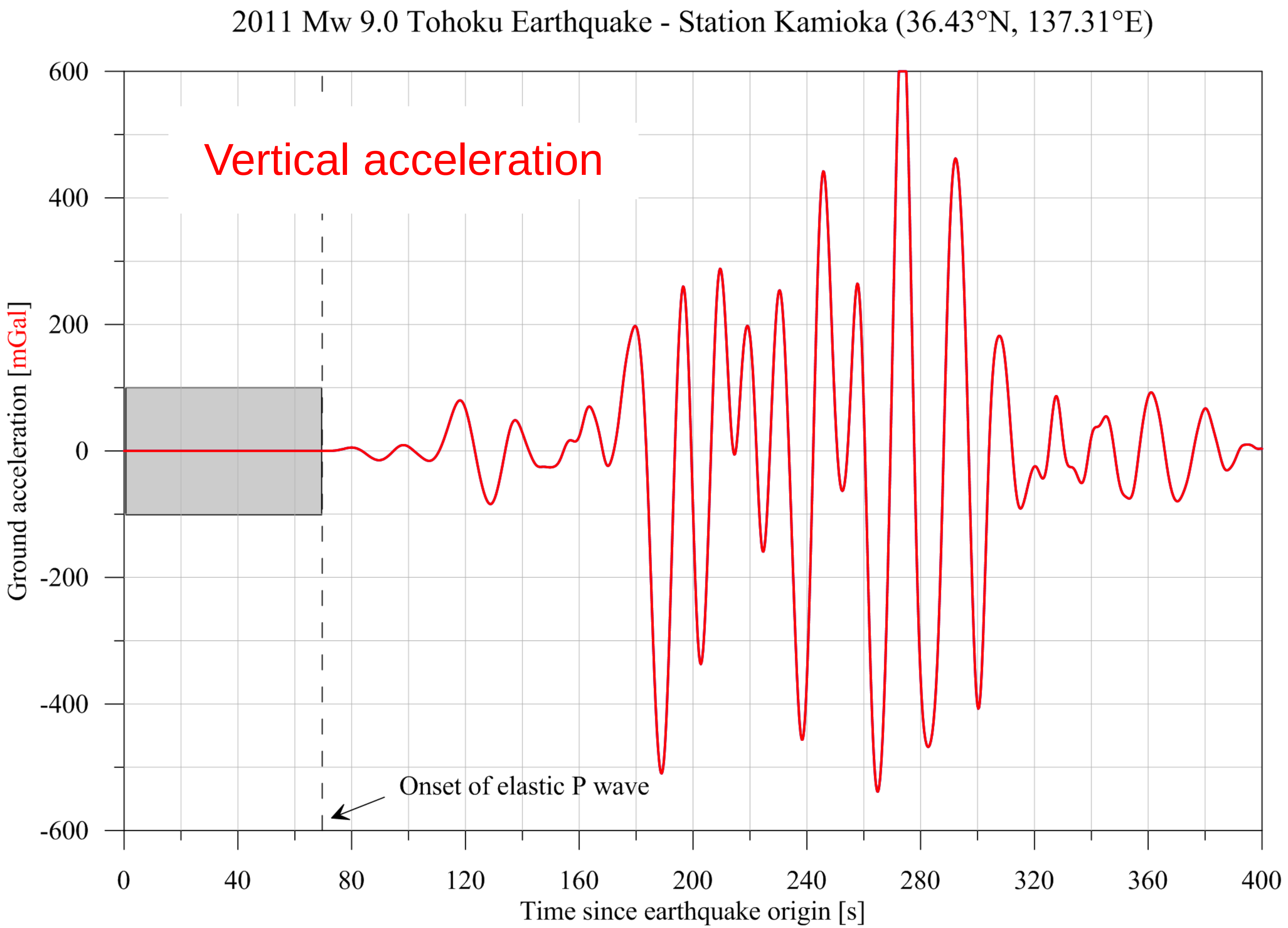

2011 Mw 9.0 Tohoku Earthquake - Station Kamioka (36.43°N, 137.31°E)
Dynamic solutions
Upward
East
North
Static solutions
Upward
East
North
Onset of elastic P wave
Gravitational force [µGal]
Time since earthquake origin [s]
20
15
10
5
0
-5
-10
-15
-20
0
40
80
120
160
200
240
280
320
360
400



2011 Mw 9.0 Tohoku Earthquake - Station Kamioka (36.43°N, 137.31°E)
Upward
East
North
Gravitational force [μGal]
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
-0.6
-0.8
-1.0
0
5
10
15
20
25
30
35
40
45
50
55
60
65
70
Time since earthquake origin [s]

QSSP output seismograms

$$\Delta \boldsymbol{a} = \ddot{\boldsymbol{u}}$$



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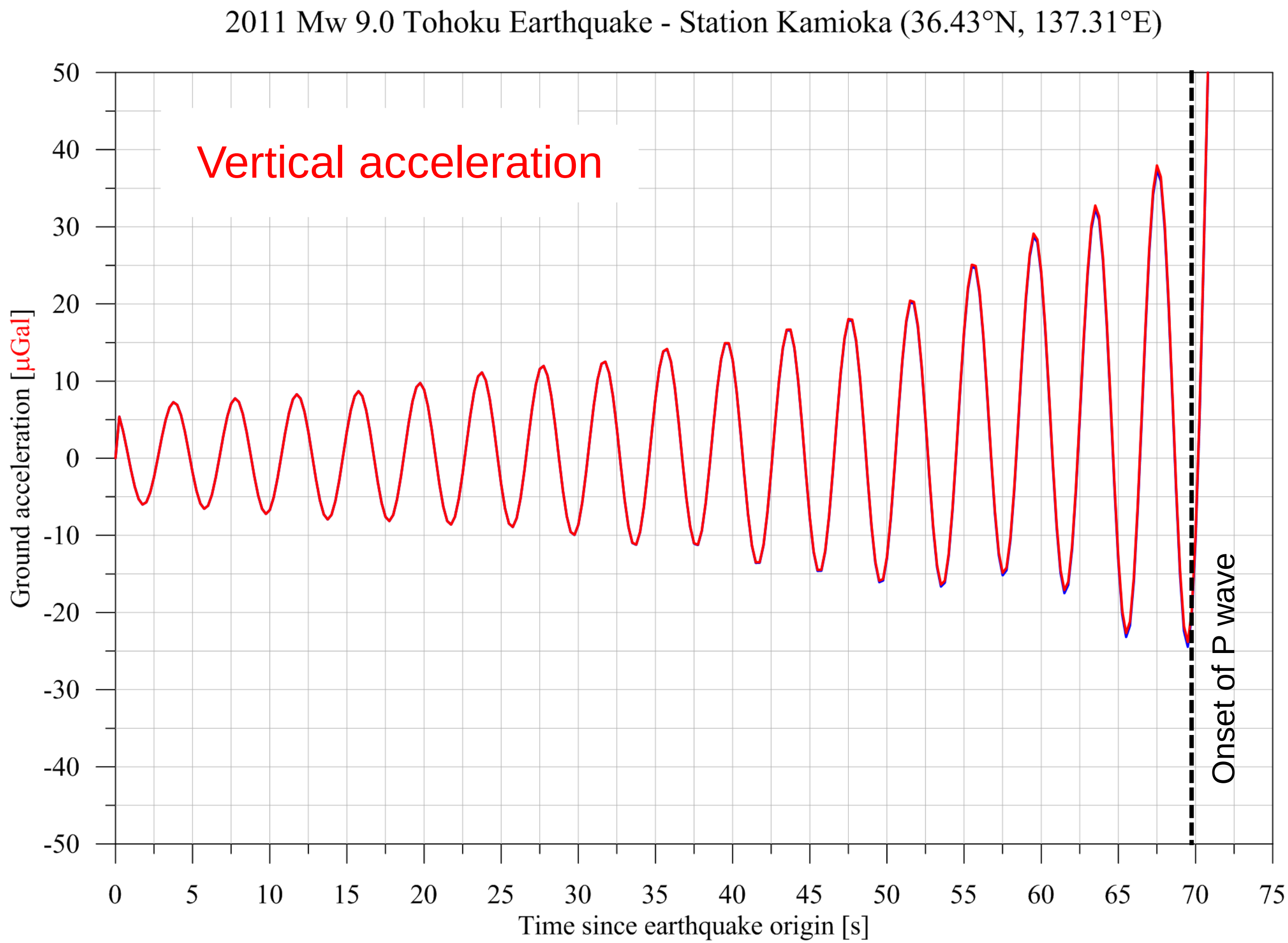

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Our new idea:

$$\begin{cases} \rho \dfrac{\partial^2 \boldsymbol{u}}{\partial t^2} = \nabla \cdot \boldsymbol{\sigma} + \rho \nabla (\psi - g u_r) + \rho g (\nabla \cdot \boldsymbol{u}) \boldsymbol{e}_r + \boldsymbol{f} \\ \nabla^2 \psi \ = \ 4\pi G \nabla \cdot (\rho \boldsymbol{u}) \end{cases} \rightarrow \boldsymbol{u}$$

$$\begin{cases} \rho \dfrac{\partial^2 \boldsymbol{u}}{\partial t^2} = \nabla \cdot \boldsymbol{\sigma} + \rho \nabla (\psi - g u_r) + \rho g (\nabla \cdot \boldsymbol{u}) \boldsymbol{e}_r + \boldsymbol{f} \\ \nabla^2 \psi \ = \ \alpha \cdot 4\pi G \nabla \cdot (\rho \boldsymbol{u}) \end{cases} \rightarrow \boldsymbol{u}_\alpha$$

$\boldsymbol{u}_\alpha$ includes the same elastic waves and the same static gravity effects as $\boldsymbol{u}$, but the coupling effect by factor α smaller/larger than $\boldsymbol{u}$.

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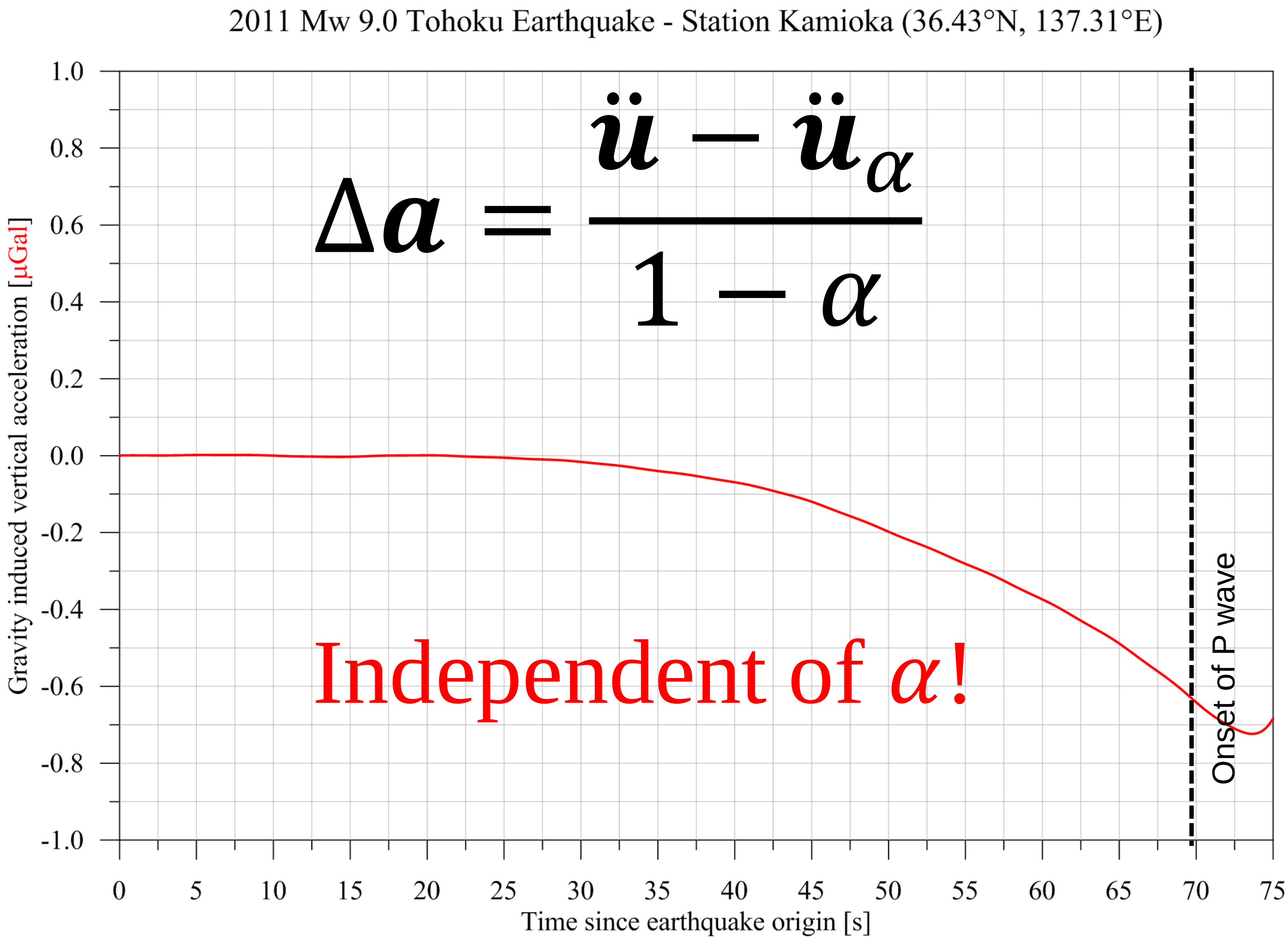


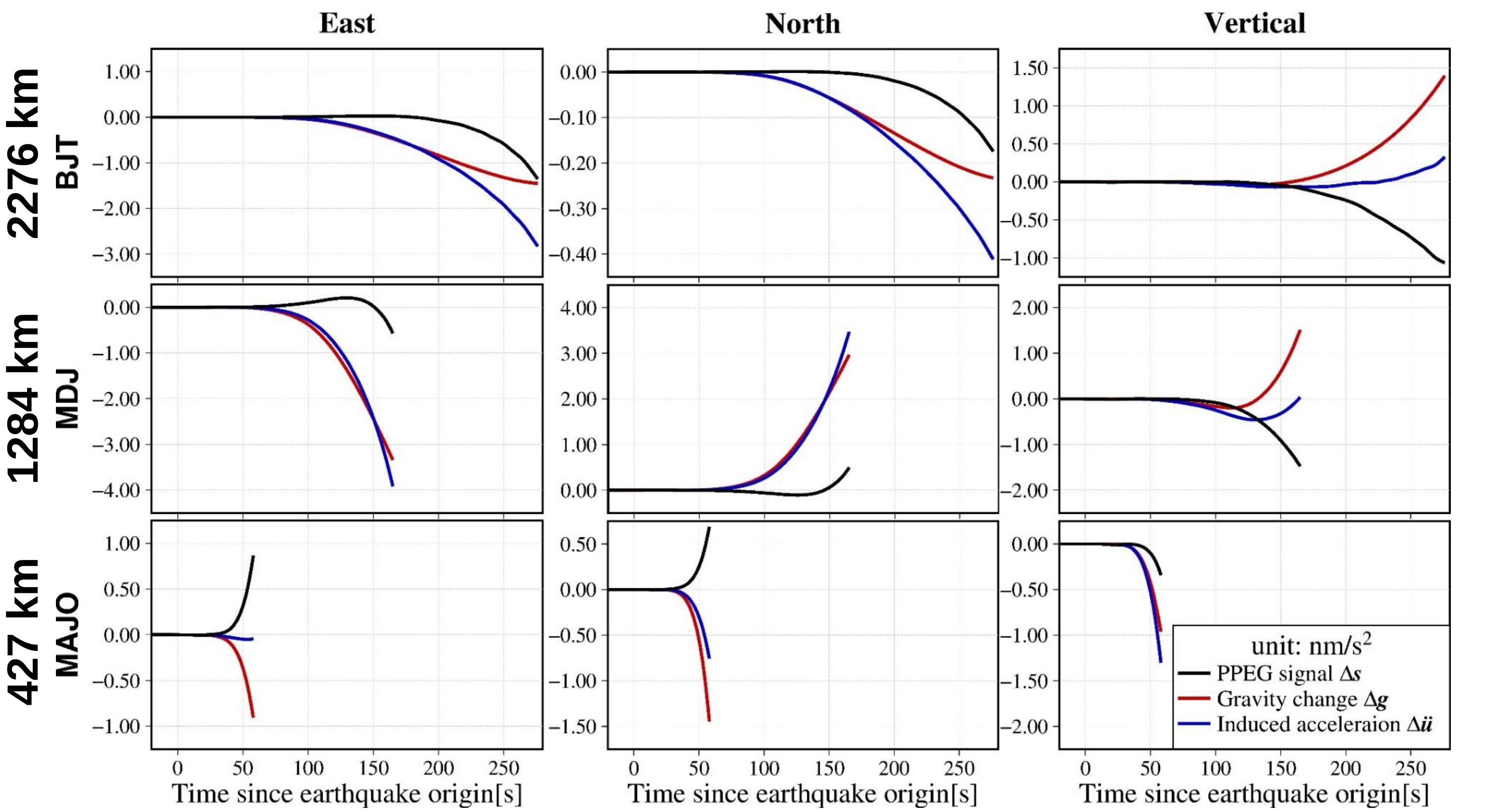
$$\Delta \boldsymbol{a} = \frac{\ddot{\boldsymbol{u}} - \ddot{\boldsymbol{u}}_\alpha}{1 - \alpha}$$

Independent of α!

Application to the 2011 Tohoku earthquake

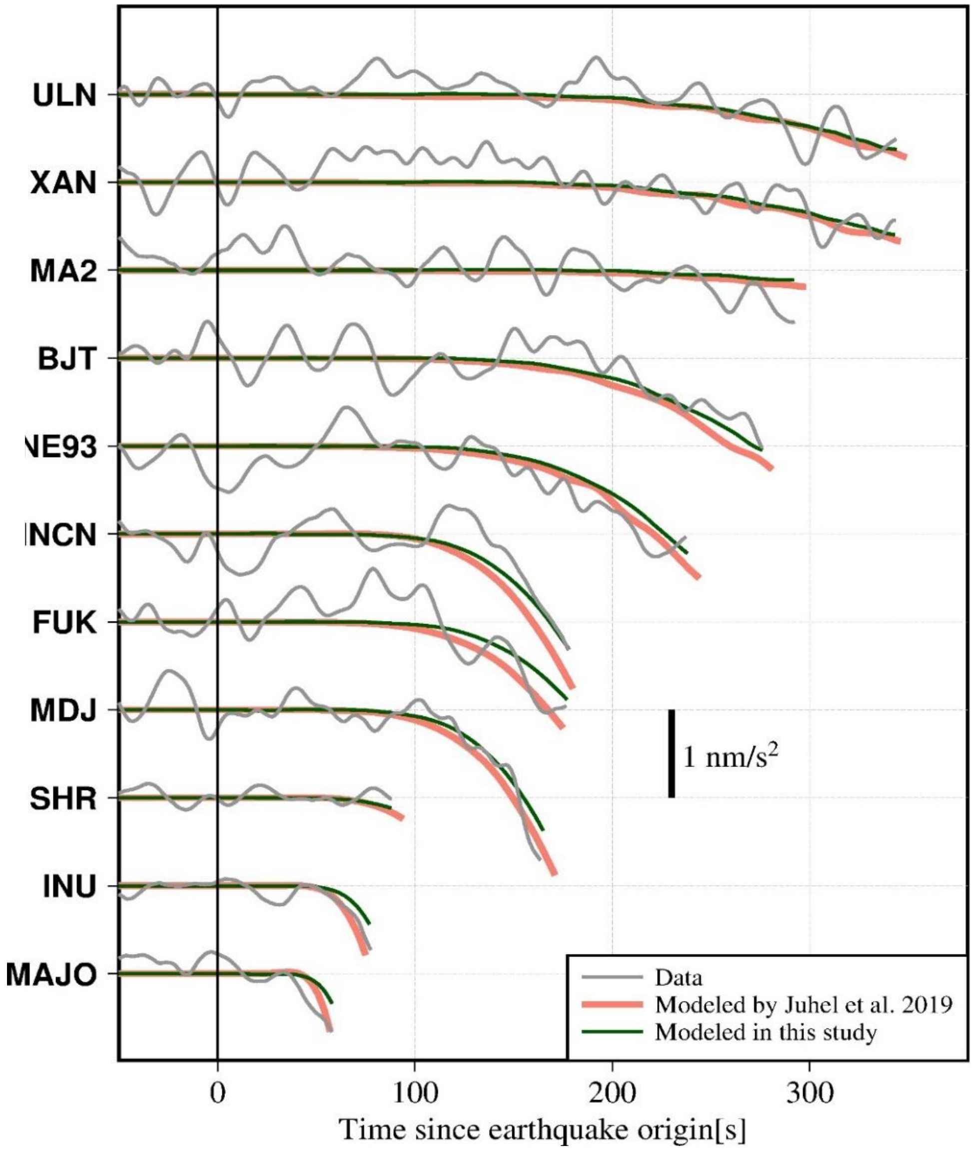
Source model:	Kinematic finite-fault model provided by Wei et al. (2012)
Earth model:	AK135 modified with the local crust structure
QSSP synthetics	
Cutoff frequency:	250 mHz
Cutoff harmonic degree:	2500

Synthetic PEGS time series at selected stations



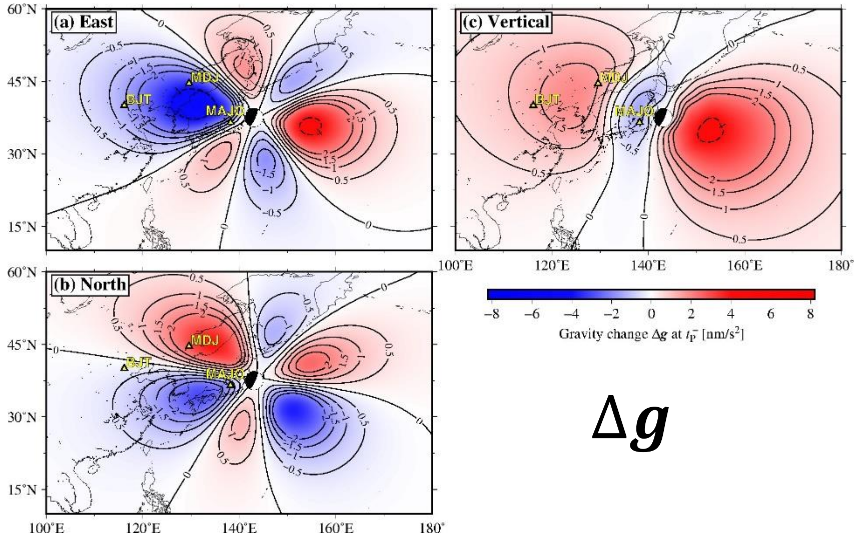
Synthetics vs. data

Vallee et al. (2017)

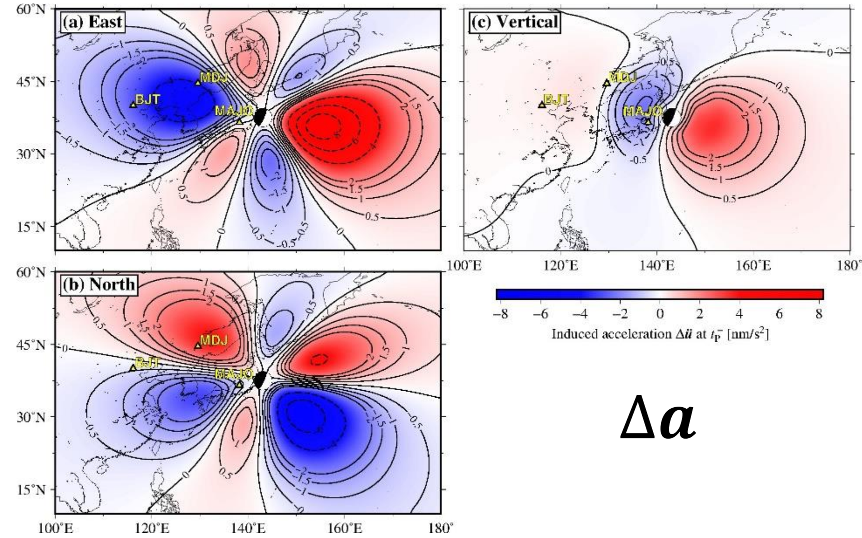


Zhang et al. (2020)

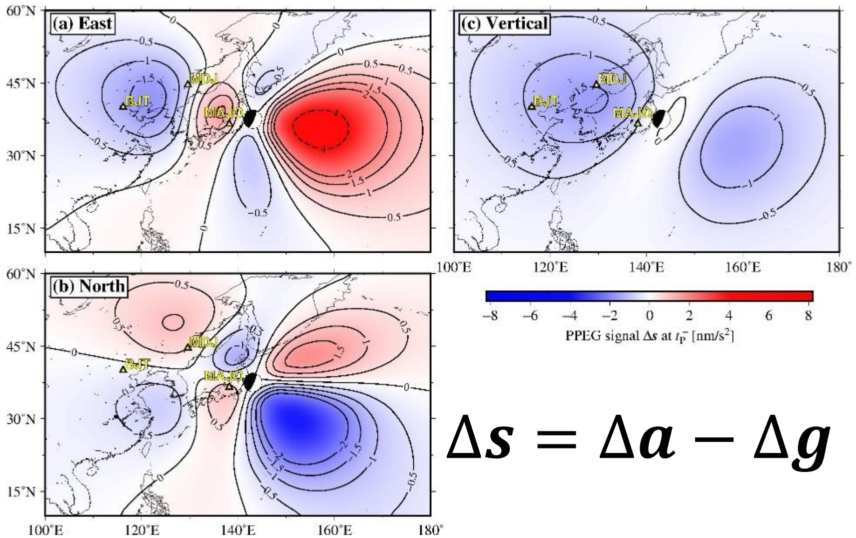
Spatial distribution of peak PEGS signals



Δg



Δa



$$\Delta s = \Delta a - \Delta g$$

Observable with seismic instruments

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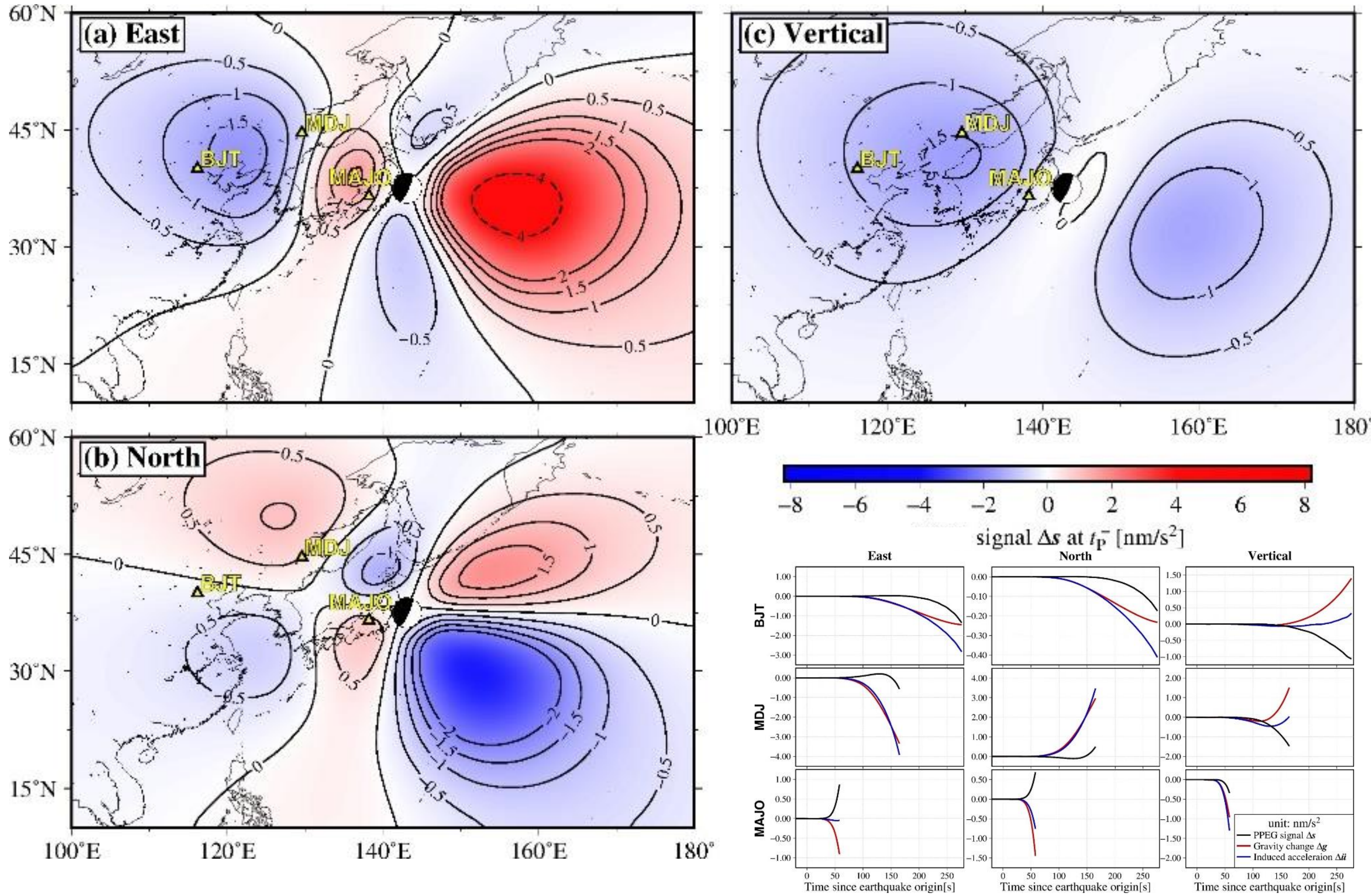


(a) East
(b) North
(c) Vertical
BJT
MDJ
MAJO
60°N
45°N
30°N
15°N
100°E
120°E
140°E
160°E
180°
−8 −6 −4 −2 0 2 4 6 8
signal Δs at t_P^- [nm/s²]
East
North
Vertical
Time since earthquake origin[s]
unit: nm/s²
PPEG signal Δs
Gravity change Δg
Induced acceleraion Δü

On the potential use of PEGS

1. Most PEGS signals increase/decrease monotonically before the P wave arrival. For a Mw ~ 9.0 earthquake, their maximum is in the order of **a few tenths of μGal**, which is very small, but significantly over the noise level at several quiet stations.
2. It can be best detected at epicentral distances between 500-3000 km, but difficult without information of the P wave onset, implying **a major limitation for earthquake early warning**.
3. Far-field peak PEGS signals are proportional to the total seismic moment, providing **strong constraints on the earthquake magnitude** and therefore **useful for tsunami early warning** under certain ideal conditions.
4. The radiation pattern depends not only on the fastest P wave, but also on many other seismic wave phases (pP, sP, …), which can provide **complementary constrains on the source mechanism**.
5. …

How does the peak Δs signal depend on the moment magnitude Mw and rupture duration T?

Two point source tests

Dip-slip

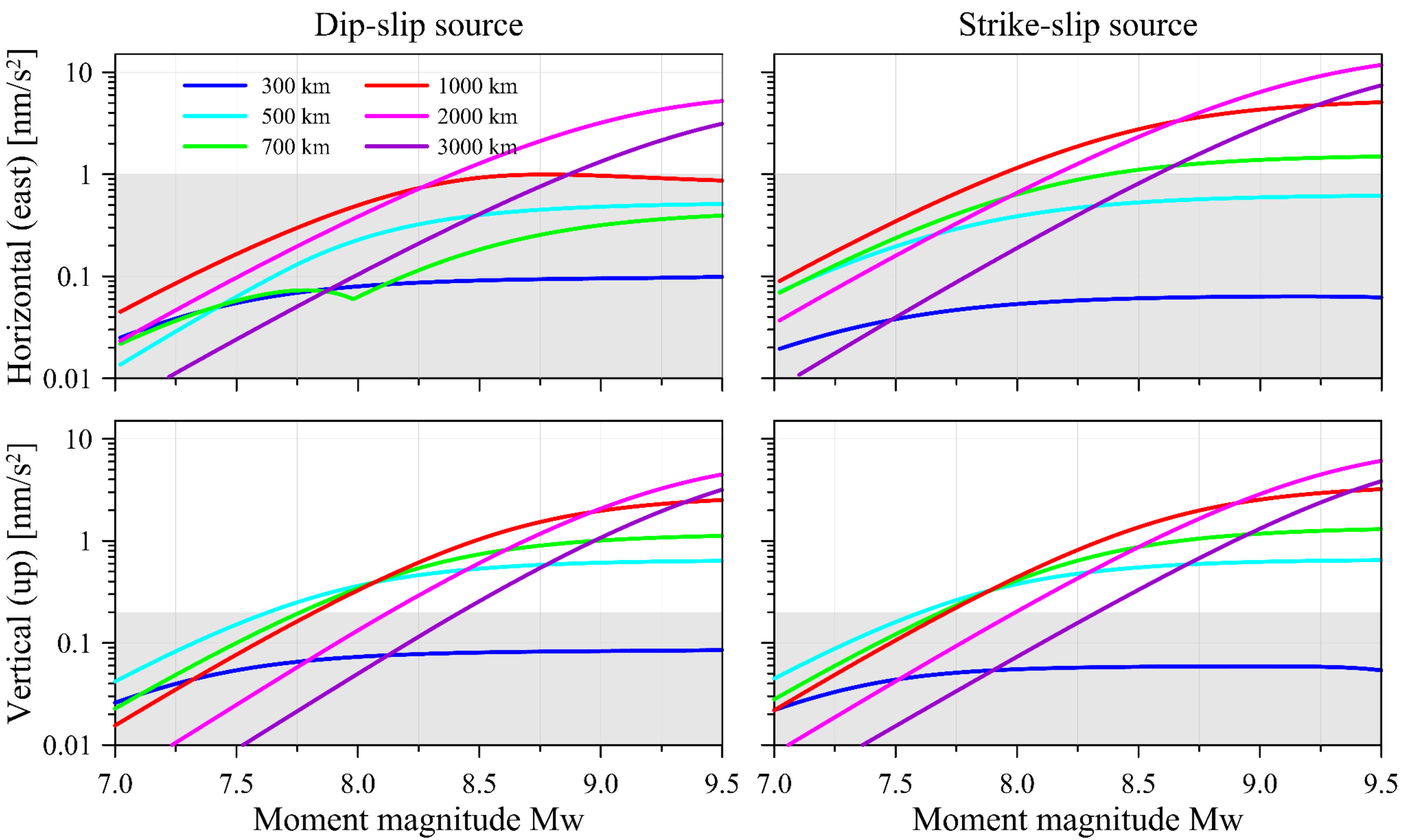
strike/dip/rake = $0^o/20^o/90^o$, receiver azimuth 90^o

Strike-slip

strike/dip/rake = $45^o/90^o/90^o$, receiver azimuth 90^o

Mw-T scaling law

$$T = T_o 10^{Mw/2}$$



Zhang et al. (2020)

The first magnitude/duration inversion

Cost function:

$$R = \sum_{i=1}^{N} \frac{1}{\sigma_i^2} \int_0^{t_i^P} [d_i(t) - s_i(t)]^2 \mathrm{d}t$$

$d_i(t)$: data

$s_i(t)$: model

σ_i^2: standard pre-seismic noise variance

Unknowns:

Moment magnitude: Mw

Rupture duration: T

Approach 1:

Assume Mw-T scaling law

$T = T_o 10^{Mw/2}$

2D grid search Mw and T_o

Approach 2:

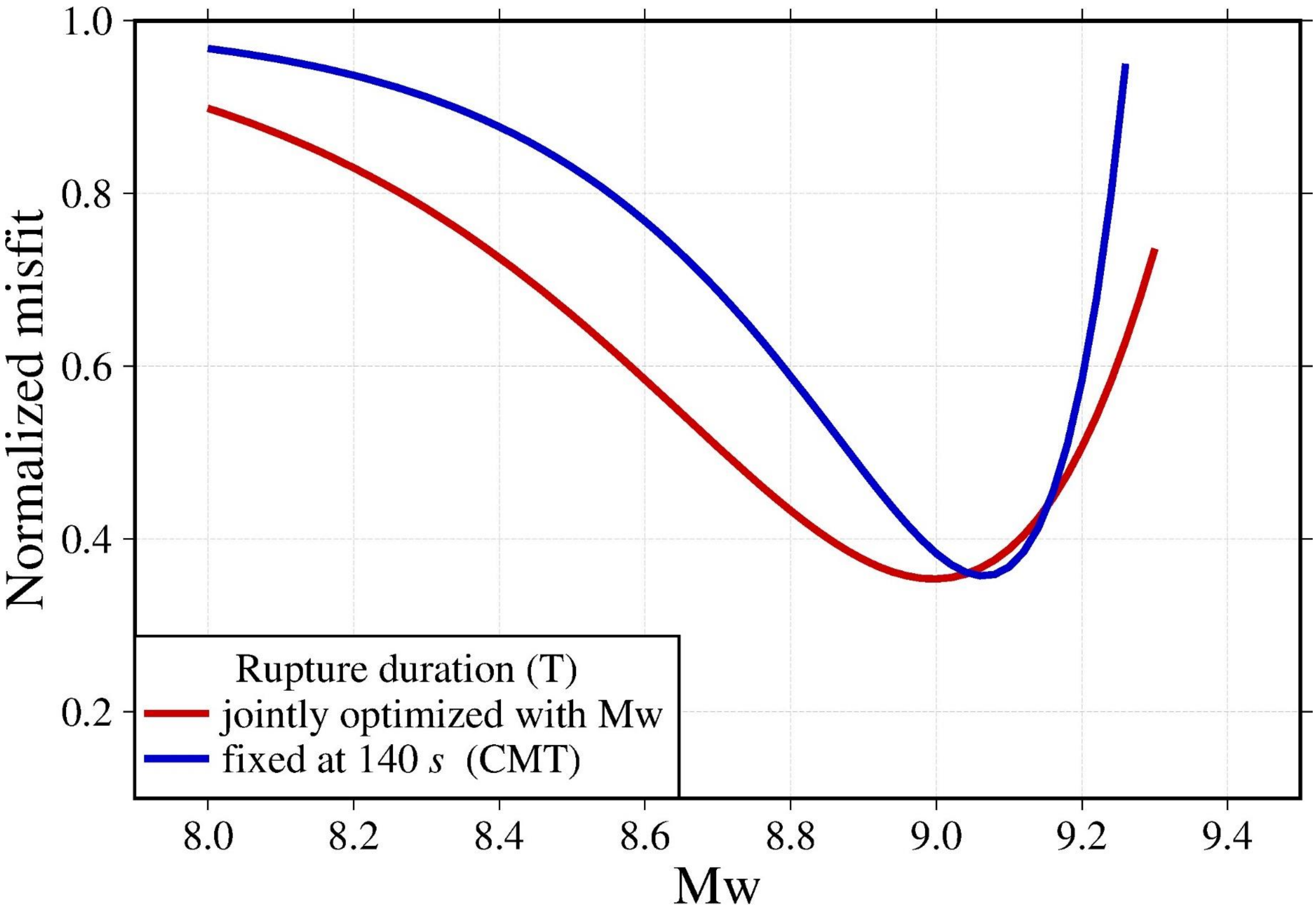
Fix

T = 140 s (CMT)

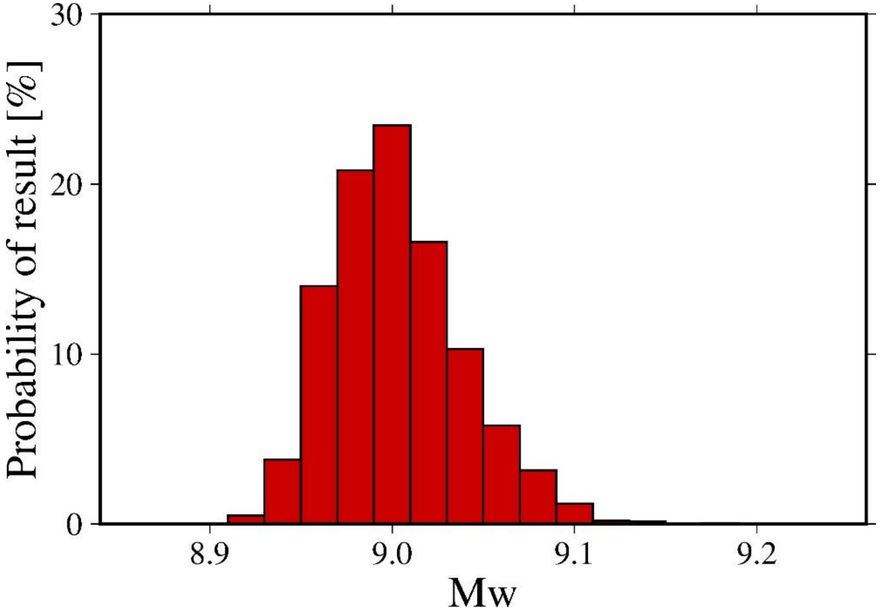
1D grid search for Mw

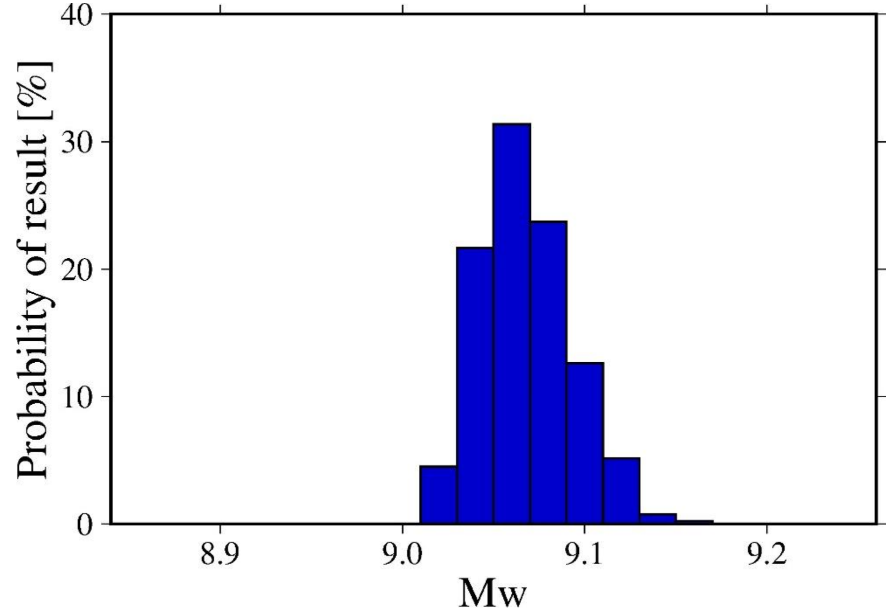
Confidence interval through Bootstrap tests

The 2011 Tohoku earthquake (11 best BB stations)



Confidence level via Bootstrap tests





Approach 1:

Mw = 9.00 ± 0.05

Optimal T_o = 0.00371s

(T = **117s**)

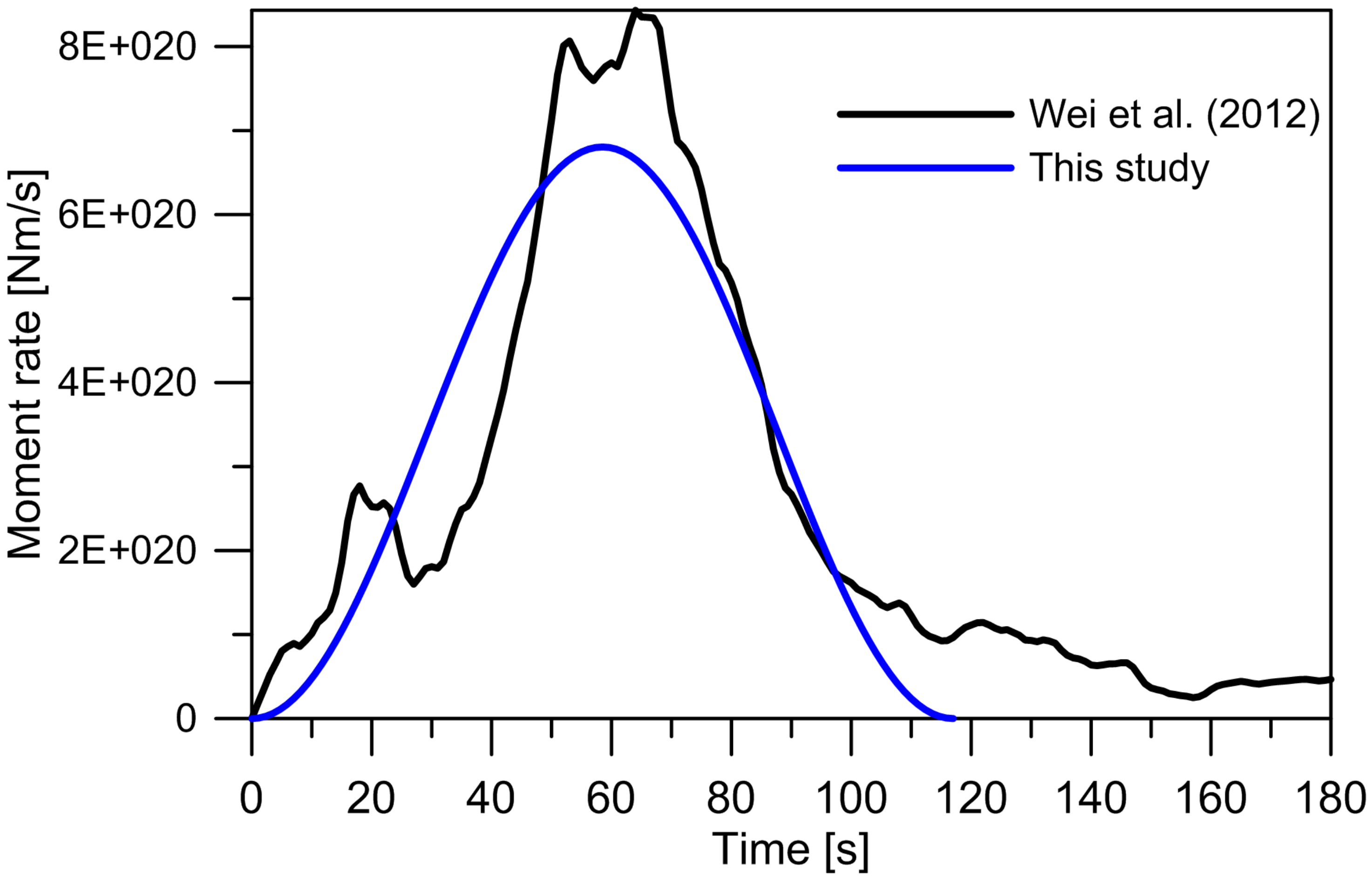
Approach 2:

Mw = 9.06 ± 0.03

Using fixed T adopted

from CMT (**9.08**/**140s**)

Seismic waveform inversion (Wie et al. 2012): Mw = 9.06



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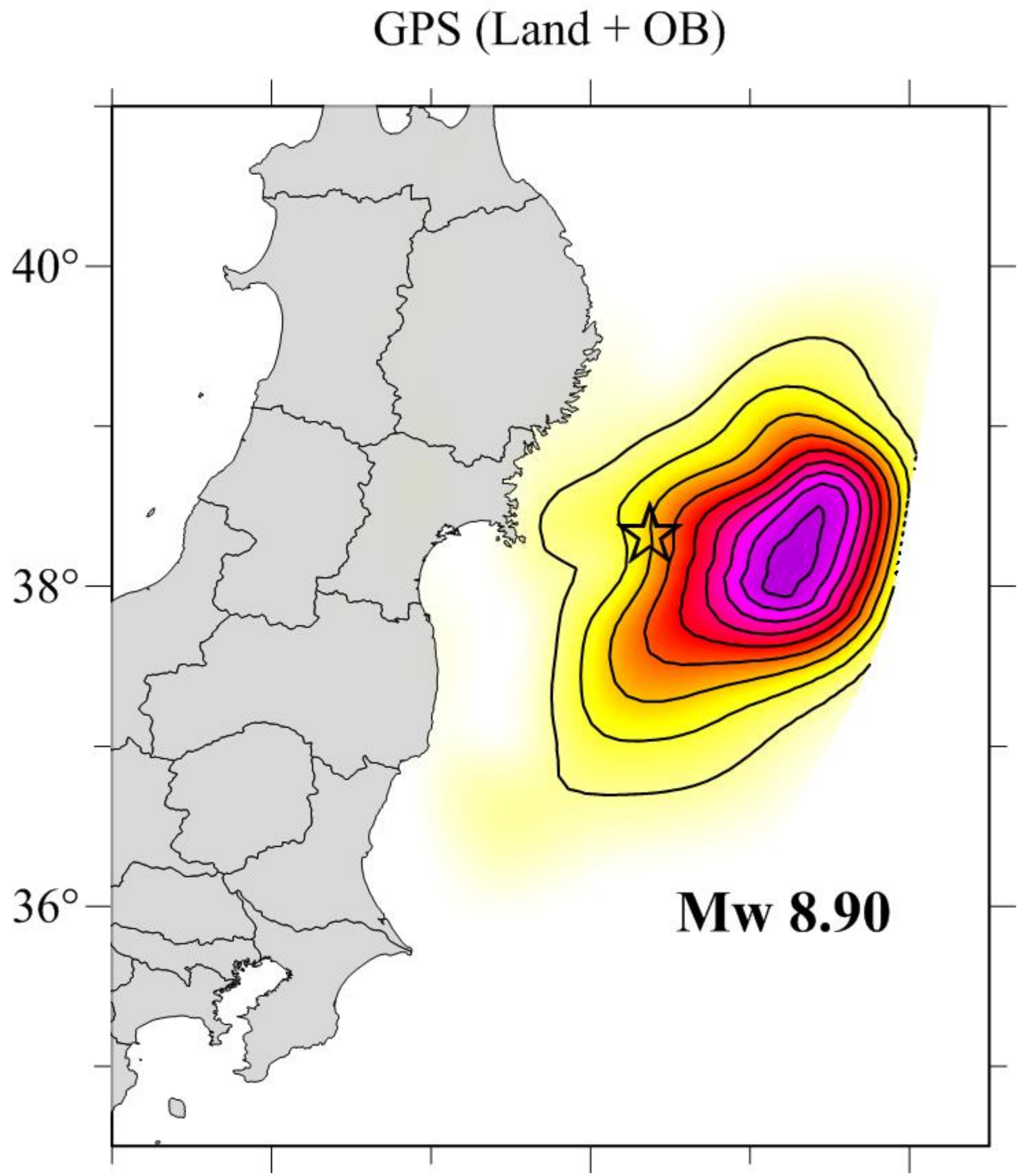
Geodetic inversion (Wang et al. 2013): Mw = 8.90



Conclusions

- ✓ A brand new approach, which is considerably simpler, faster and more accurate than previously used for simulating PEGS signals.
- ✓ A robust estimate of major source parameters of the 2011 Tohoku earthquake using the PEGS data recorded at 11 low-noise broadband stations.

Outlook

- ➢ Further theoretical investigations (it is just the beginning).
- ➢ Development of new measurement systems (e.g., gravity strainmeters).
- ➢ …

Thank you!