[EE] Evening Poster | S (Solid Earth Sciences) | S-SS Seismology

[S-SS06]CSEP, earthquake forecast testing, and the role of SSE in earthquake occurrence.

convener:Danijel Schorlemmer(GFZ German Research Centre for Geosciences), Naoshi Hirata(Earthquake Research Institute, the University of Tokyo), Matt Gerstenberger(共同), Hiroshi Tsuruoka(Earthquake Research Institute, Tokyo Univ.)

Mon. May 21, 2018 5:15 PM - 6:30 PM Poster Hall (International Exhibition Hall7, Makuhari Messe)

The Collaboratory for the Study of Earthquake Predictability (CSEP) has expanded over the years to many different testing areas hosted at multiple testing centers. One of which is the Japan testing center at the University of Tokyo, operated in collaboration with GFZ Potsdam. Hundreds of earthquake forecast models have been submitted to CSEP and are being tested. New testing metrics were developed and implemented and a lot of progress was made to establish CSEP as an institution that cannot be ignored when issuing earthquake forecasts. Its rigor and independence became the standard in evaluating earthquake forecasts and in reporting on the results.

Although the tests CSEP has conducted have been successful and well-received, they have also shown the limitations of the CSEP approach. What is a sufficient testing period for models? Are time-invarying models really describing the long-term seismic activity? Are long-term models testable at all? Do short-term models provide significant information for the forecasting problem or do they only model aftershock sequences? What other signals should be included in forecasting models to improve them? Do improvements in forecasting models translate into improvements of hazard models? Many aspects of seismic hazard or earthquake forecasting remain inherently untestable if only the model forecasts are tested and not the model ingredients. We propose to create new areas of activity for CSEP, namely targeted experiments that cannot be conducted with the current CSEP software system.

We solicit contributions addressing forecasting models, forecast testing problems, new ideas for CSEP experiments, possibilities of further CSEP developments, ways of expanding CSEP into the hazard and risk domain, and more general views on the forecasting problem. This is aimed at fostering the discussion in the community about further goals of earthquake forecasting experiments.

[SSS06-P03]Estimating centroid location and source dimension of the Te Araroa earthquake (Mw 7.1), New Zealand by analyzing direct and reflected tsunamis

*Tatsuya Kubota[1], Tatsuhiko Saito[1], Yoshihiro Ito[2], Ryota Hino[3], Syuichi Suzuki[3], Yoshihiro Kaneko[4], Laura Wallace[4,5] (1.National Research Institute for Earth Science and Disaster Resilience, 2.Disaster Prevention Research Institute, Kyoto University, 3.Graduate School of Science, Tohoku University, 4.GNS Science, 5.Institute for Geophysics, The University at Texas at Austin)

Keywords:The 2016 Te Araroa earthquake, New Zealand, Tsunami, Coastal reflected wave, Centroid moment tensor

On September 1, 2016 (UTC), the Te Araroa earthquake occurred ~ 80 km northeast of the coast of the North Island, New Zealand (NZ) (Mw 7.1, GCMT). This earthquake had a normal-faulting mechanism and occurred in the subducting plate in the Hikurangi subduction zone (Warren-Smith et al., 2018). When this event occurred, ocean bottom pressure gauges (OBPs) were installed at ~150 km south of the source area, and first-arrival of tsunami with amplitude of −2 cm and the coastal reflected tsunami with amplitude of +1.5 cm were clearly observed. Centroid moment tensor (CMT) solution of this earthquake was estimated by GCMT, USGS, and the regional seismic network (GeoNet) of GNS Science, NZ. Although their depths, strikes, dips and rakes, and seismic moments were similar, their horizontal locations were quite different. Both the GCMT and USGS centroids were located ~80 km northeast from the coast, but they were ~20 km apart from each other in NS direction. The GeoNet centroid was located further ~50 km northeast from the GCMT and USGS centroids. The robustness of the centroid location estimation from onshore seismic data depends on the various factors, such as the station coverage, S/N ratio, or the uncertainty of the velocity structure. Since this earthquake occurred far from the coast, the centroid location estimated using the onshore seismic data may not be accurate. On the other hand, the tsunami propagates much slower than the seismic waves and reliable bathymetry data is available for the accurate tsunami propagation modeling. Tsunami also has an advantage in constraining the earthquake source dimension which determines the extent of the tsunami source area, because the tradeoff between the earthquake source dimension and the rupture velocity is much more significant for seismic wave than tsunami. In this study, we estimated the centroid location and source dimension of the Mw7.1 Te Araroa earthquake using the offshore tsunami data.

We estimated the centroid location using grid search approach. Fixing the magnitude, fault geometry and centroid depth (GCMT), we assumed the rectangular planar fault using the scaling law (Wells and Coppersmith, 1994). Then we searched for the centroid location that best reproduces the observed waveforms, based on the VR (variance reduction) between the observed and calculated waveforms. We first used the direct waves and obtained the centroid location at ~15 km north of the GeoNet centroid (~ 130 km northeast from the coast). However, the high-VR area (> 90% of the best solution’s VR) extended ~100 km in the WSW-ENE direction, suggesting the centroid location was not constrained well. We then used tsunami reflected from the coast to calculate VR, and obtained the centroid location at ~10 km northwest of the GCMT centroid (~ 80 km northeast from the coast). The extent of the high-VR area was reduced to ~40 km in the WSW–ENE direction and does not include the GeoNet and USGS centroids. This suggests that the GCMT is the most suitable solution for the centroid location, and the reflected waves contributed for constraining the centroid location.

We then searched for the earthquake source dimension by fixing the seismic moment and fault geometry to the GCMT value and assuming the ratio of source length L to width W such that L/W = 2 (rigidity *μ* = 40 GPa). With the direct waves alone, the best source length was obtained as L = 40 ± 20 km. On the other hand, when using both the direct and reflected waves, the source length was L = 50 ± 15 km. Although the source dimension is not constrained well, the upper limit of the possible source length was almost the same in both analyses, and the modeling result using the coastal reflection suggests that a smaller source dimension (L <~ 30 km) is not plausible. Using the range of source dimensions obtained from the analysis of the coastal reflection, we calculated the stress drop Δ*σ* of ~ 0.5 – 3.0 MPa, which is a typical value for earthquake stress drop (e.g., Kanamori and Anderson, 1978).