

Ground motion observations from the 14 November 2016 Mw7.8 Kaikoura, New Zealand earthquake and insights from broadband simulations

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1 Abstract

This paper provide insights into the observed ground motion records from the 14 November 2016 M_w 7.8 Kaikoura earthquake. Specific detail is given to examining observations in the near-source and those in the urban Wellington / Lower Hutt regions through a combination of time series and response spectra to understand salient source, path and site effects. Several ground motions exceeding 1.0g horizontal are recorded, as well as up to 2.7g in the vertical direction at one location. The complexity associated with multi-fault rupture is clearly evident in several strong motion stations in the near-source region. Significant site effect case histories from sedimentary and reclaimed soil deposits and topographic effects are observed in the Wellington and Lower Hutt regions. To provide

further insights, predictions from preliminary broadband ground motion simulations are examined to more directly interpret the observations.

2 Introduction

On 14 November 2016 at 12:02 AM local time, the M_w 7.8 'Kaikoura' earthquake occurred along the east coast of the upper South Island, New Zealand (USGS, 2016; GeoNet, 2016a). The earthquake initiated in the Waiau Plains in North Canterbury, and involved multiple fault segments as the rupture generally propagated northward over 150km to Cape Campbell in Marlborough. It was the largest recorded earthquake in New Zealand since the 1855 M_w 8.2 – 8.3 Wairarapa earthquake (GeoNet, 2016b).

Given the geographical proximity of the source in a largely rural region of New Zealand, the impacts in the near-source region were largely geological and geotechnical in nature, namely surface rupture (Litchfield et al., 2016), landslides (Petley, 2016), and localized ground failure, resulting in damage to coastal transportation infrastructure (NZTA, 2016) and the formation of landslide dams. Significant damage to State Highway 1 (SH1) both North and South of Kaikoura and the inland Kaikoura road (SH70) resulted from landslides and slope instability caused significant disruption. SH70 was opened to unrestricted traffic volumes on 19 Dec 2016 after several weeks open to restricted traffic (NZTA, 2016); SH1 South of Kaikoura was opened to restricted traffic on 21 December (Fairfax Media, 2016c), and SH1 North of Kaikoura is expected to require up to one year to open (NZTA, 2016). Relatively significant damage also occurred to numerous structures in the capital city of Wellington (located approximately 60km to the North of the causative faults). One structure had a partial collapse of precast floor units and is the subject of an on-going government enquiry (New Zealand Government, 2016), while another three structures are undergoing rapid demolition as a result of earthquake damage (Wellington City Council, 2016; Fairfax Media, 2016a,b). As of 7 December 2016, 11% of Wellington office space has been closed due to damage (Wellington City Council, 2016).

This paper is intended to provide insights into the observed ground motion records from the Kaikoura earthquake as a means to both understand the earthquake-induced

demands imposed on the natural and built environment, as well as underlying inferences of the spatial and temporal nature of the causative rupture. Salient features of the observed strong motions are discussed, as well as compared with preliminary broadband ground motion simulations. In the subsequent sections we first summarize the tectonic setting and inferred rupture, followed by a discussion of the observed ground motions in various contexts, and then finally inferences based on broadband ground motion simulations.

3 Tectonic setting and inferred causative rupture

The Marlborough Fault Zone (Figure 1) is one of the most active crustal regions of New Zealand with many mapped active faults, in particular the 'major' Wairau, Awatere, Clarence, and Hope faults, as well as several other 'connecting' faults (see Stirling et al. (2012) and references therein). The W-Phase moment tensor solution from the USGS is available at: <http://earthquake.usgs.gov/earthquakes/eventpage/us1000778i#moment-tensor>). Observations and multiple geophysical and geodetic methods highlight the source complexity of this earthquake with numerous fault segments rupturing.

We draw on three principal datasets in this paper: (1) geologic mapping of surface fault rupture observations¹ and measured displacements from Litchfield et al. (2014); (2) a preliminary source inversion model based on InSAR, GPS and surface uplift measurements (Hamling, 2016); and (3) a heuristic interpretation of the temporal and spatial evolution of recorded ground motions. Figure 1 illustrates the rupture segment geometries of the causative faults that we adopted for ground motion observation and simulation interpretations. For reference, the surface trace of fault segments that are considered in the InSAR+GPS inversion of Hamling (2016), significant aftershocks, and mapped active faults are also shown.

Litchfield et al. (2016) have so far identified ten fault segments (The Humps, Hundeele, Conway-Charwell, Upper Kowai, Fidget, Jordan Thrust, Papatea, Kekerengu, and Needles Faults) that have evidence of surface fault rupture. Nine of these faults are located onshore with displacements obtained principally from direct field measurement of identi-

¹The first author also examined a significant portion of surface ruptures in the field.

fiable features; while rupture of the Needless Fault, located offshore at the northern end of the ruptured faults, has been identified from seabed uplift (NIWA, 2016). Measured fault displacements range up to 9-11m horizontal (Kekerengu Fault) and 6-8m vertical (Papatea Fault) on the northern fault segments (Litchfield et al., 2016), while the southernmost rupturing faults (The Humps, and Hundalee Faults) exhibited maximum surface rupture displacements in the range of 1-5m. Most notably, essentially no surface rupture of the Hope Fault (the major fault identified *a priori* in the region (Stirling et al., 2012)) has been mapped to date (Litchfield et al., 2016).

The surface trace of the InSAR+GPS inversion of Hamling (2016) shown in Figure 1 also highlights the source complexity seen from surface rupture (it is noted that an earlier version of Litchfield et al. (2016) was used in that inversion). In addition to the faults identified as surface rupturing, this model also infers slip on the Hope Fault and the sub-parallel Clarence fault to the north. However, it does not model rupture on the Papatea Fault, which is considered as a pop-up structure [Hamling, pers. comm.]. Our source model utilized for ground motion modelling in Figure 1 started with the model of Hamling (2016), and adopted the following modifications: (1) added the Papatea fault as implied from surface rupture; (2) ignored the small slip inferred on the Clarence, Grassmere, northern end of the Hope, and southern end of the Hundalee faults; (3) considered a maximum down-dip width of 19km for (significant) slip, which for fault dips ranging from 55-80 degrees equates to maximum depths of approximately 15.5-18.5km; and (4) Modified the adopted dip in the inversion for the Hope fault as it was considered excessively low relative to the *a priori* value in the national hazard model (Stirling et al., 2012) for this well-studied fault. The principal reason for these modifications was to reduce the total fault area in the 'starting' source model, which had an area significantly higher than M_w -Area scaling relations would suggest is reasonable, and because this current source inversion moment is significantly higher than the seismologically-determined $M_w = 7.8$. Table S1, available in the electronic supplement to this article, provides the key parameters of each fault segment² depicted in Figure 1, and further details of the apportioned seismic

²Multiple fault segments, each of constant strike, are used for the Humps Fault Zone and Hundalee Fault

moment and temporal initiation for each rupture segment are discussed in a subsequent section on ground motion simulation.

Clearly, significant corroborating evidence is needed to truly understand the source complexities of this event, but, as subsequent results indicate, we believe that the interpretation presented above is broadly consistent with observational data and insightful for the purpose of understanding ground motion observations.

4 Ground motion observations

4.1 Ground motion record processing

Volume 1 ground motion records were obtained from GeoNet (www.geonet.org.nz/). A total of 224 records were obtained for this event, the majority of which were recorded on CUSP-3, Kinematics Etna or Basalt digital strong motion instruments. All ground motions were baseline corrected via removal of the mean and linear trend (if any) in recorded acceleration. The baseline-corrected records were then processed with a low-pass causal four-order Butterworth filter with cutoff frequency, $f_c=50$ -100 Hz (depending on the sampling rate), and a high-pass filter with $f_c=0.05$ Hz. All figures in this paper make use of absolute timing such that $t = 0$ represents the hypocenter rupture time. Subsequently, velocity and displacement signals were obtained. Because of the era of these instruments, all processed records following the above process resulted in physically-realistic Fourier spectra amplitudes and integrated displacement histories. As a result, the processed ground motions can be considered to provide reliable estimates of peak ground accelerations (PGA) and response spectral ordinates over the range $T = 0.01 - 10$ seconds (Douglas and Boore, 2011), which are typically of engineering interest. It should be noted that the above procedure does not lend itself to the computation of residual displacements (from permanent fault displacement or ground failure), which may be non-zero for near-source locations and is left for future study.

4.2 A regional view

Figure 2 illustrates the location of the GeoNet ground motion station recordings relative to the causative fault, which range in source-to-site distances³ of approximately R_{rup} =0-216 km. Figure S1, available in the electronic supplement to this article, illustrates the distribution of recorded geometric mean⁴ peak ground acceleration, PGA . In total, 47, 16, and 5 ground motions were observed with $PGA > 0.1g$, $0.2g$, and $0.5g$, respectively.

Figure 2 illustrates the spatial distribution of the GeoNet strong motion stations, and representative velocity time series in three orthogonal components. As would be expected, the highest velocities are evident in the near-source region. However, the nature of the waveforms vary significantly at similar R_{rup} values depending on the back-azimuth from the site to the site - those stations in proximity to the southern end of the rupture (North Canterbury; stations WTMC, WIGC, CULC, HSES) exhibit high-amplitude short-duration ground motions, while those at the north end of the rupture (northern Marlborough and Wellington; KEKS, WDFS, BWRS, FKPS) have substantially longer duration and also well-defined multiple wave packets indicating 'delayed' rupture initiation of several segments. Central South Island stations, such as Molesworth (MOLS) and Wairau Valley (WVFS), also exhibit two clear velocity wave packets.

The subsequent sections discuss specific aspects of observed ground motions in three sub-regions of particular interest.

4.3 The near-fault region

Figure 3 illustrates four strong motion stations that are located in the immediate region of the causative faults. Station WTMC is located immediately near the inferred epicenter in Waiau; KIKS in Kaikoura, approximately halfway along the north-south extent of the rupturing faults; and KEKS and WDFS are located in Kekerengu and Ward, respectively, at the northern end of the rupturing faults.

Because of its location near the epicenter (see Figure 1), the ground motion observed

³The source-to-site distance is the closest distance of the site to any of the causative fault segments

⁴Geometric mean metrics are used throughout unless noted.

at WTMC (Figure 3a) indicates a short strong motion duration relative to the other three near-fault stations. The ground motion accelerations at WTMC exceed 1.0g in both horizontal directions, and notably the acceleration in the vertical direction reaches 2.7g (with a 100Hz high-cut filter). It is worth noting that the vertical accelerations exhibit strong asymmetry toward higher amplitudes in the positive direction (e.g. two exceedances of +2.0g, while negative accelerations are limited to a little over -1.0g). This phenomena has been documented in several ground motions from past earthquakes such as the 2011 Christchurch, New Zealand (Bradley and Cubrinovski, 2011; Fry et al., 2011), and the 2008 Iwate-Miyagi, Japan (Aoi et al., 2008; Yamada et al., 2009; Tobita et al., 2010) earthquakes. As demonstrated by Tobita et al. (2010) and Jeong and Bradley (2016), such asymmetry results from the varying near surface soil shear strength during induced compression and dilation. Furthermore, independent evidence of the extreme vertical accelerations in the vicinity of the WTMC site were seen⁵ in the form of bearing pads significantly moving (and, in one case, coming out from) between bridge abutment and girders at two bridges on Inland Road (SH70), within 2km from the WTMC station.

The KIKS station recorded relatively small levels of ground motion acceleration, given its location on a peninsula near the north-south mid-point of the rupturing fault segments, with horizontal and vertical peak accelerations of approximately 0.22g and 0.27g, respectively. The horizontal peak velocities are also a similarly small $PGV=41-45\text{cm/s}$, as compared to the other three stations in Figure 3 which generally have $PGV > 80\text{cm/s}$ (velocity time series at all four stations are shown in Figure 2). Relative to the source representation shown in Figure 1, this implies that either the offshore rupture of the Hundalee Fault (which projects to the surface in Kaikoura) had relatively small slip amplitudes at its northern end, or largely did not rupture offshore. Surface rupture was observed on this fault at the coastline (Litchfield et al., 2016), yet offshore seismic reflection is yet to confirm or deny its offshore extent (and InSAR+GPS-based inversions (Hamling, 2016) naturally provide limited resolution for such offshore inferences).

Figure 3c and 3d illustrate the recorded ground motion accelerations at Kekerengu

⁵by the first author and others during field reconnaissance.

(KEKS) and Ward (WDFS) stations, which are located near the northern end of the rupturing faults, and are themselves located 16km apart. Both of these stations clearly illustrate acceleration records which are dominated by two predominant wave packets, one near $t = 60s$, and the other near $t = 80s$. Because of the fact that the KEKS station is within 2.5km of the mapped surface rupture of the Kekerengu fault (Litchfield et al., 2016) (with surface rupture displacements of 5-10m horizontal, and 1-2m vertical in the vicinity) it is reasonable to assume that the peak accelerations in the KEKS record correspond to the through-going rupture past this location. In contrast to KEKS, at the WDFS station the peak accelerations occur near $t = 80s$, which based on the source model in Figure 1 is inferred as the result of rupture of the Needles fault. Hamling (2016) inferred some (small) slip on the Grassmere fault located onshore from the Needles fault (and thus closer to the WDFS station). While not included in the modelled rupture representation in Figure 1, as no surface major rupture was evident on this fault from (Litchfield et al., 2016), the large accelerations at WDFS may imply that such rupture occurred (since inspections of the site do not indicate any pronounced site effects).

The time delay between the predominant wave packets, and consequent peak accelerations, at the KEKS and WDFS stations (approximately $\Delta t = 16s$) despite only a 16km separation distance also imply some rupture delay from termination of rupture on the northern end of the Kekerengu fault and initiation of rupture at the southern end of the Needles fault. In a similar vein, the through-going rupture seen at KEKS at approximately $t = 60s$ implies significant delays in the rupture of the Kekerengu fault (and Jordan Thrust; see Litchfield et al. (2016)) from earlier ruptures of the southern-most fault segments. While we are not aware of any sufficiently comprehensive kinematic inversions to specifically solve for such initiation times, we utilize the above observations to approximate segment initiation times in subsequently discussed kinematic rupture models and ground motion simulations.

4.4 The epicentral region

In addition to the observations at the WTMC station near the rupture hypocenter discussed in the previous section, several other salient observations were obtained in the epicentral region in North Canterbury (see locations in Figure 2) as depicted in the form of acceleration time series in Figure 4. The strong motion station at Waiau Gorge, WIGC, is located approximately 6km south from the inferred western end of the rupture in The Humps fault zone. Like the WTMC record, the shaking at WIGC is punctuated by a relatively short duration of high amplitude shaking with horizontal and vertical peak accelerations in the range of $PGA=0.5-0.8g$. Interestingly, the Culverden station, CULC, is located only a further 6km south of WIGC and exhibits significantly lower horizontal accelerations of 0.23-0.29g, indicating either some site effect at the WIGC station (although both sites are identified by GeoNet as residing on shallow soil) or significant distance attenuation.

The Hanmer Springs (HSES) and Cheviot (CECS) stations in Figure 4 are located one basin either side of the Waiau Basin area, at distances of 20-30km from the WTMC station. It can be seen that the ground motion acceleration records at these two stations are similar, with peak accelerations on the order of 0.25g horizontal and 0.10/0.16g vertical. Both stations also exhibit recordings with clearly identifiable multiple wave packets associated with time delays in the rupture of subsequent fault segments.

4.5 Urban Wellington

Beyond the immediate near-fault region, the earthquake caused ground motions with the most societal impact in the lower North Island region, in particular, the Wellington region. This impact was the result of the northward rupture propagation causing moderate levels of ground motion intensity in the region, particularly at longer vibration periods, combined with sufficient depths of sedimentary deposits. As a result, these 'long-period long-duration' ground motions were damaging to multi-storey office buildings of comparable vibration periods, and port structures on deep reclaimed deposits. As an example of the ground motions recorded in the region, Figure 5 illustrates the acceleration time series

recorded at a typical rock (POTS) and soil (TEPS) station. Similar to the WDFS station (Figure 3d), it can be seen that the ground motions at both stations are dominated by two wave packets, inferred as a result of rupture along the Jordan Thrust/Kekerengu Faults and Needles Fault, respectively. Given its closer location to Wellington, the rupture of the Needles fault clearly dominates the strong motions in Wellington at both stations. The mono-chromatic nature of the shaking at the TEPS soil site is clearly evident.

Figure 6a illustrates the specific locations of eight strong motion stations in the central Wellington region relative to the predominant basin edge defined by the Wellington Fault (Van Dissen and Berryman, 1996), and the regions of either quarry or hydraulic reclamation (Semmens et al., 2011). There are several reference rock stations in the Wellington/Hutt Valley region, and the (geometric mean) response spectra at five such stations are illustrated in Figure 6b. Stations POTS and PTOS are annotated, in particular, because they are considered as reference stations in subsequent figures; however, it can be seen that all five stations have very similar response spectral amplitudes across the full period range depicted. It is noted that all five stations are located on the hanging wall immediately to the west of the Wellington Fault trace. Although not apparent at all five rock stations, it can be seen that at the POTS and PTOS stations there are relative spectral peaks at approximately $T = 1.5s$ and $T = 5 - 6s$ (or conversely, spectral valleys at periods in between these, as the sites are located at the base of topography). Because of the large source to site distance (approximately 60km to the northern extent of the modelled rupture of the Needles Fault), it can be seen that even the rock ground motions have relatively weak high frequency ground motion amplitudes (a comparison to the 500yr return period rock site design spectra NZS1170.5 (2004) is provided in Figure 6b). As a result, the ground motions were not damaging to any short period structures in the region.

As noted in Figure 6a, there is significant commercial, industrial, and social activity in reclaimed regions of central Wellington. Figure 6c illustrates the response spectra at the three stations on reclaimed land relative to that at the POTS reference rock station. The FKPS and TEPS stations are located relatively close to each other (350m apart; 2km from

POTS), while PIPS is located closer to the POTS rock reference station (1.0km). Despite the proximity of the FKPS and TEPS stations, it can be seen that the response spectra are quite different for vibration periods less than $T = 2.0s$. This can be explained by the notably different depths of sedimentary soils between these two sites, which are inferred to be approximately 30m at FKPS and 100m at TEPS (Semmens et al., 2011). Based on sedimentary soil shear wave velocities near the surface of approximately $V_s=200m/s$, increasing gradually with depth, the fundamental period of a soil column of depth H subject to shear deformation gives vibration periods which are very similar to the spectral peaks of 0.4s and 1.4s seen at FKPS and TEPS, respectively.

The PIPS station, located on hydraulic fill near the northern end of Wellington City, is seen to have substantially higher spectral amplitudes than the other two reclaimed sites for vibration periods between $T=0.6-6.0s$. Although direct constraints are poor, Semmens et al. (2011) infer that the soil depth at the location of PIPS is on the order of 250-300m. The combination of deep sediments, hydraulic fill (i.e. harbour muds/sands as opposed to gravelly material in other reclaimed areas), and the location immediately beside the Wellington Fault resulting in long period basin-edge effects (Kawase, 1996), are the inferred causes of the observed long period amplification.

Figure 6d illustrates three stations (WEMS, TFSS, VUWS) in the central and northern portions of Wellington City that are located on alluvial deposits, with the rock and hydraulic fill stations of POTS and PIPS for comparative reference. It can be seen that all three alluvial sites exhibit amplification for $T < 4.0s$ relative to the POTS reference station. At long periods ($T = 2 - 6s$) the amplitudes at the TFSS site are slightly larger than the other two alluvial sites, consistent with its (poorly constrained) greater sediment depth on the order of 100m (Semmens et al., 2011), compared to depths on the order of 50m for the other two sites. At shorter periods ($T = 0.2 - 1.0s$) the VUWS site does exhibit amplitudes up to 50% larger than the other two sites.

Figure 6e illustrates the observed spectral amplitudes at several other miscellaneous stations in the Wellington suburbs, with the POTS and VUWS stations providing references for rock and alluvial soil conditions in central Wellington. WNKS station is located

in Karori, an elevated basin suburb located to the west of central Wellington. It can be seen that this site exhibits a strong spectral peak at short periods ($T = 0.4s$) with no amplification relative to the POTS rock reference station for $T > 1.0s$. This site response is indicative of shallow sediments over stiff bedrock. WEL station is located just above central Wellington on a relatively small hill near its crest, with what can be seen as consistently smaller spectral amplitudes than the VUWS alluvial soil reference, but larger than the POTS rock reference. Finally, WNKS station is located at Wellington Airport, on raised beach deposits that were lifted above sea level as a result of the 1855 Wairarapa earthquake (GeoNet, 2016b). Similar to the prior comments, it can be seen that this site also exhibits notably smaller amplifications than the VUWS alluvial reference site.

4.6 Lower Hutt / Petone

The Hutt Valley is located directly to the north of Central Wellington and is another urbanized area. Figure 7a illustrates six strong motion stations that span the mouth of the Hutt Valley at its southern end, and provide a nice illustration of site effects in an alluvial valley that, on face value, is notably simpler than the site conditions in central Wellington. The western edge of the valley is bounded by the Wellington Fault, with sediment depths up to 300m, based on Boon et al. (2011). Progressing across to the east of the valley along the transect of stations, the basin depth gradually decreases. Figure 7b illustrates that, as seen in central Wellington, the four soil sites exhibit amplifications over the period range $T = 0.4 - 4.0s$ relative to the PTOS rock reference station⁶. It can also be seen that over the period range $T = 1.5 - 5.0s$ there is a greater amplification for the two stations (PVCS, PGMS) located in the deeper deposits to the west than those shallower stations to the east (LHUS, LRSS). The LIRS station on the eastern boundary of the valley can be seen to exhibit mild amplification at short periods ($T = 0.2s$) from its shallow soil depth overlying rock.

⁶The higher amplitudes at periods shorter than $T = 0.1s$ are not reflective of any amplification, but a result of the fact that the peak oscillator response is being driven by longer period ground motions, the theoretical details of which are discussed in detail by Bora et al. (2016)

5 Insights from broadband ground motion simulations

The previous sections provided a discussion of ground motion observations in an effort to implicitly understand seismic source, path, and site effects associated with this event. In this section we explore these prior inferences further by supplementing them directly with numerical results from preliminary hybrid broadband ground motion simulations.

5.1 Simulation methodology and inputs

Ground motion simulation was undertaken using the Graves and Pitarka (2015) methodology, a 'hybrid' approach in which the low frequency (LF) waveforms are comprehensively computed by solving the elastodynamic equation in the 3D crustal model domain (using finite differences), while the high frequency (HF) waveforms utilise a phenomenological simplified physics approach.

In the adopted methodology, the seismic source is prescribed in the form of a kinematic rupture description. The specific kinematic rupture generator is described by Graves and Pitarka (2015), and we use default parameters of that model unless described below. We extend the methodology presented in that paper by generating kinematic ruptures on the multiple fault segments to produce the rupture shown in Figure 1. The rupture description for each fault segment requires the specification of fault geometry (location, strike, dip, length, width), moment magnitude, M_w (which implies average slip for a given fault area), average rake angle, hypocenter location and hypocenter rupture initiation time (even though there is only one real hypocenter, in the rupture generation a hypocenter is specified for each fault segment). The high frequency source description utilized a stress parameter of $\Delta\sigma = 5MPa$, anelastic attenuation of $\kappa = 0.045$, and generic 1D crustal model which have been used for simulations of the 2010-2011 Canterbury earthquakes (Bradley et al., 2015) (the stress parameter value is also the same as that recommended for California (Graves and Pitarka, 2010, 2015)). Additional modelling specifics for the seismic source are provided in a subsequent section.

A New Zealand seismic velocity model (NZVM) (Lee et al., 2016; Thomson et al., 2016) was utilized to describe the P- and S-wave velocity, and density of the crust over the simulation domain. This model contains a 3D travel-time tomography-based crustal model from Eberhart-Phillips et al. (2010), a high-resolution representation of sedimentary deposits in the Canterbury region (Lee et al., 2016) (which is not present over the wider region) related to research on the 2010-2011 Canterbury earthquakes, and near-surface estimates of shear wave velocity at strong motion stations based on geology, topography, and direct measurements (Foster et al., 2016). Ground motion simulations utilizing this NZVM have been validated against observations for events in the 2010-2011 Canterbury earthquake sequence (Razafindrakoto et al., 2016, 2015), as well as small magnitude events in the Canterbury foothills (Nazer et al., 2016).

Near-surface site response was modelled following the V_{s30} -based empirical approach discussed in Graves and Pitarka (2010) based on the empirical amplification model of Campbell and Bozorgnia (2014). V_{s30} values of the stations are provided in Table S2, available in the electronic supplement to this article.

The ground motion simulations were performed within a computational domain of 640km x 300km x 80km, with a finite difference grid spacing of 0.2km. In the Canterbury sedimentary basin a minimum shear wave velocity of $V_s=500\text{m/s}$ was considered (yielding maximum frequencies of the low frequency waveforms of 0.5Hz), while outside the Canterbury basin the minimum shear wave velocity is generally greater than $V_s=1000\text{m/s}$ (giving maximum frequencies of 1.0Hz). A time increment of $\Delta t=0.005\text{s}$ was utilised, resulting in 50,000 time steps to produce simulated ground motions for a target duration of 250s. The ground motion simulations were performed on the NeSI 'Fitzroy' Power6 cluster (<https://www.nesi.org.nz/services/high-performance-computing/platforms>). The LF finite difference calculation, using 512 compute cores, required approximately 9 hours of wall clock compute time, with a subsequent 5 hours for the HF calculation and post-processing. Additional computational workflow details are discussed in Bae et al. (2016).

5.2 Source model assumptions

As alluded to earlier, Figure 1 illustrates the kinematic rupture model used in the subsequently presented ground motion simulations, with the standard rupture format (SRF) parameters for each fault segment provided in Table S1, available in the electronic supplement to this article. The arguments for the adopted source geometry were provided in Section 3, so the discussion here focuses on the specified seismic moment and rupture initiation time on each fault segment.

Hamling (2016) provides InSAR+GPS-derived slip estimates, but as already noted, such estimates result in a seismic moment release that is notably above $M_w 7.8$. In addition, it is now well recognised that the generation of strong ground motion is often not directly coincident with inferred location of large static slip. As a result, we draw relative inferences from the InSAR+GPS-derived slip across all fault segments, but determine absolute slip amplitudes (via specification of fault segment M_w values) based on consideration of empirical M_w -Area scaling relations. The total fault area in the model presented in Figure 1 is 5510km², which using the inter-plate strike-slip correlation of Leonard (2010) yields a median $M_w = 7.74$ - i.e. consistent with the seismologically-derived value of $M_w 7.8$. The application of the Leonard scaling relation to each fault segment directly does not work for a multi-fault rupture (as it implies that average slip would be independent of fault area and hence magnitude), so we therefore use it in a relative sense to apportion the total seismic moment, M_o , across the different segments shown. Specific details of the moment magnitude apportioned to each segment are provided in Table S1, available in the electronic supplement to this article. It can be seen from the distribution of slip amplitudes in Figure 1 that the largest average slip amplitudes occur on the northern-most fault segments (i.e. Jordan Thrust, Kekerengu, Papatea, and Needles Faults), consistent with both magnitudes of surface rupture displacements, and also implied slip amplitudes from InSAR+GPS inversion (Hamling, 2016).

The rupture initiation time on each fault segment was based on a combination of two factors: (1) the time to kinematic triggering, when the rupture of an immediately adjacent fault reaches the (approximate) location of the segment initiation point; and (2)

any imposed rupture time delay. As noted in Section 3, the observed ground motions do indicate several significant rupture initiation delays are present, and therefore we attempt to account for this in a heuristic manner (based on the present limited information). Given that the considered fault segments are all approximately connected, we performed a simulation with a previous, but similar, version of fault geometry without consideration of any time delays (i.e. all segments initiated rupture via kinematic triggering). The main discrepancy we observed when comparing observed and simulated ground motion waveforms, was the substantially later high-amplitude wave packet arrivals in observed ground motions at stations such as KEKS and WDFS (see Figure 3 and related text). The simplest adjustment we could identify was to delay the initiation of rupture at the southern end of the Jordan Thrust (connected to the Hope Fault, see Litchfield et al. (2016)). A delay of almost $\Delta t = 20s$ was necessary for the Jordan Thrust and subsequently Kekerengu Faults to rupture with an approximately correct arrival time of the large amplitude motion at KEKS station. In addition, a several second delay was also required between the Kekerengu and Needles faults to resemble the nature of the ground motion observed at the WDFS station (and also in the lower North Island). Assumed rupture initiation times for each of the considered segments are provided in Table S1, available in the electronic supplement to this article.

It should be relatively obvious that there are numerous heuristics in the method by which the seismic source has been modelled. We have considered multiple variations among different rupture geometries, moment partitioning, and segment initiation, all in the climate of rapid acquisition of observational data, and corroborating evidence (or otherwise) from third-party analyses using different datasets and/or methodologies. As the results in the subsequent sections demonstrate, several minor tweaks can be made to obtain simulations which align better with observations, but major improvements will require a more systematic understanding of the source, as well as improved representation of the crustal structure in the region, and near-surface site effects representation.

5.3 High-level simulation results

Figure 8 illustrates snapshots of the simulated ground motion wavefield (in the form of ground motion velocity at the surface) for nine time instants during the simulation (A video is available in the electronic supplement to this article and at: <https://youtu.be/ZbI7rgnZ2U8>)⁷ During the first 30 seconds of the simulation it can be that the 'southern' fault segments (The Humps, Hundalee, Hope) rupture in a northerly direction. Approximately at $t = 40s$ the delayed rupture initiation at the southern end of the Jordan Thrust starts, and these 'northern' faults (Jordan, Kekerengu, Papatea, Needles) rupture over the following 40 seconds. Finally, after the rupture itself ceases at approximately $t = 80s$, the wavefield, with pronounced directivity migrating to the north east, approaches the lower North Island.

Figure 9 provides a summary of the ground motion intensities over the simulation domain in the form of the peak ground acceleration, PGV (three-component maximum), and the Modified Mercalli Intensity (MMI) values, based on the MMI-to- PGV correlation of Worden et al. (2012). As implied by the wavefield snapshots in Figure 8, it can be seen that significant directivity occurs to the north east as a result of the fault geometries and rupture sequence (fortunately a significant portion of the $MMI > 8$ region occurs offshore, or in low-population density areas). As a result, the ground motion amplitudes to the south and west of the causative faults were modest in comparison. It can be seen that the Marlborough and Lower North Island was subject to approximately $MMI = 7$ ground motion amplitudes with $PGV = 20 - 40cm/s$.

5.4 Simulation validation

5.4.1 Waveform comparison

Figure 10 provides a comparison of the observed and simulated velocity time series at ten stations of interest because of their locations in the near-fault, epicentral, and Wellington

⁷The waveform anomaly near Lon:174.5° Lat:-43.2° is the result of a discontinuity between the domain-wide model of Eberhart-Phillips et al. (2010), with the offshore portion of the Canterbury Velocity model of Lee et al. (2016). It is present only in the offshore region and the resulting localized spurious wavefield does not have a material effect on the simulated motion onshore

regions, as discussed in relation to Figures 3-5. It can be seen that the simulated ground motions capture the general features of the strong motion arrival time, amplitude, and significant duration. A comparison of observed and simulated three-component waveforms at all 162 stations is given in Figure S3, available in the electronic supplement to this article.

5.4.2 Ground motion intensity with distance

Figure 11 illustrates the ground motion spectral amplitudes for four vibration periods ($T = 0.0, 0.2, 3.0$, and $10.0s$) as a function of source-to-site distance. The observed and simulated ground motion amplitudes for the 162 stations within the simulation domain are shown. The stations are also separately annotated based on their location in either the North or South Island. In addition to the simulation predictions, for reference, the NZ-specific empirical ground motion model of Bradley (2013) is also shown. It can be seen that the simulation provides a generally good comparison with the observed amplitudes. In particular, the distance attenuation in the observations at short periods (i.e. $T = 0.0$ and $0.2s$) is consistently predicted by the simulations, while the empirical model predicts a slower attenuation; conversely at long periods, the empirical model predicts a faster attenuation than exhibited by both the observed and simulated amplitudes. Although, not easily evident because of the large number of data points present, the simulations and observations are also broadly consistent in the higher-than-average amplitudes of North Island ground motions relative to those in the South Island for the same source-to-site distance - because of the aforementioned effect of rupture directivity.

5.4.3 Spectral acceleration bias with period

Figure 12 provides a summary over all vibration periods of the results noted in Figure 11. It can be seen that the broadband simulations have negligible bias for vibration periods up to $T = 2.0s$, with a bias increasing to an average of -0.4 natural log units (negative implying an over-prediction of observations) over $T = 3 - 7s$, and then reducing slightly for $T > 7s$. The empirical model; which does not account for rupture features such

as partitioning of slip across the various rupture segments, or the temporal evolution of rupture, and hence directivity; has a bias of -0.4 to -0.6 units at short vibration periods, and then a bias of up to +0.7 units at $T = 10s$.

6 Discussion and conclusions

This paper examined several salient aspects of the observed ground motions from the 14 November 2016 $M_w 7.8$ Kaikoura earthquake. Particular attention was given to an engineering-focused interpretation of the observations, and the causal inferences related to source, path, and site effects. In order to aid in such causal inferences, comparison was made with hybrid broadband ground motion simulations based on a kinematic rupture informed by surface rupture observations and a preliminary InSAR+GPS inversion, with several modifications based on the authors' experience and use of heuristics.

Overall, the simulations provide a favourable agreement with the observations, and thus their further exploration will evidentially enable a more fundamental understanding of the salient source, path, and site aspects which contributed in this event, and the propensity for repeat occurrences in the future. Some systematic bias was evident in the simulations at long vibration periods. Such bias could be addressed through reasonable modifications to some of the assumptions regarding the moment partitioning and rupture initiation of the various fault segments, as well as the addition/removal of other fault segments whose inclusion remains somewhat controversial (i.e. where surface rupture was not evident, but are inferred to have ruptured from InSAR+GPS data; and vice-versa in the case of the Papatea Fault). In addition, bias in the ground motion simulations at individual stations, or subsets of stations in a specific geographical region, are likely to be reduced through the use of improved crustal velocity and near-surface site modelling. Clearly, progress in these three areas is likely to lead to the most fruitful advancements in understanding.

7 Data and resources

Observed ground motions were obtained from GeoNet (www.geonet.org). Dr. Ian Hamling (GNS Science) provided preliminary unpublished source models. Ground motion simulations were performed on the National eScience Infrastructure (NeSI) 'Fitzroy' Power6 cluster. The New Zealand seismic velocity model (Lee et al., 2016; Thomson et al., 2016) source code is available at: <https://github.com/ucgmsim/Velocity-Model>. The adopted kinematic rupture generator was that of Graves and Pitarka (2015). The adopted QuakeCoRE ground motion simulation workflow is available on GitHub (Bae et al., 2016). Plotting was performed using GMT <http://gmt.soest.hawaii.edu/> and python <https://www.python.org/>.

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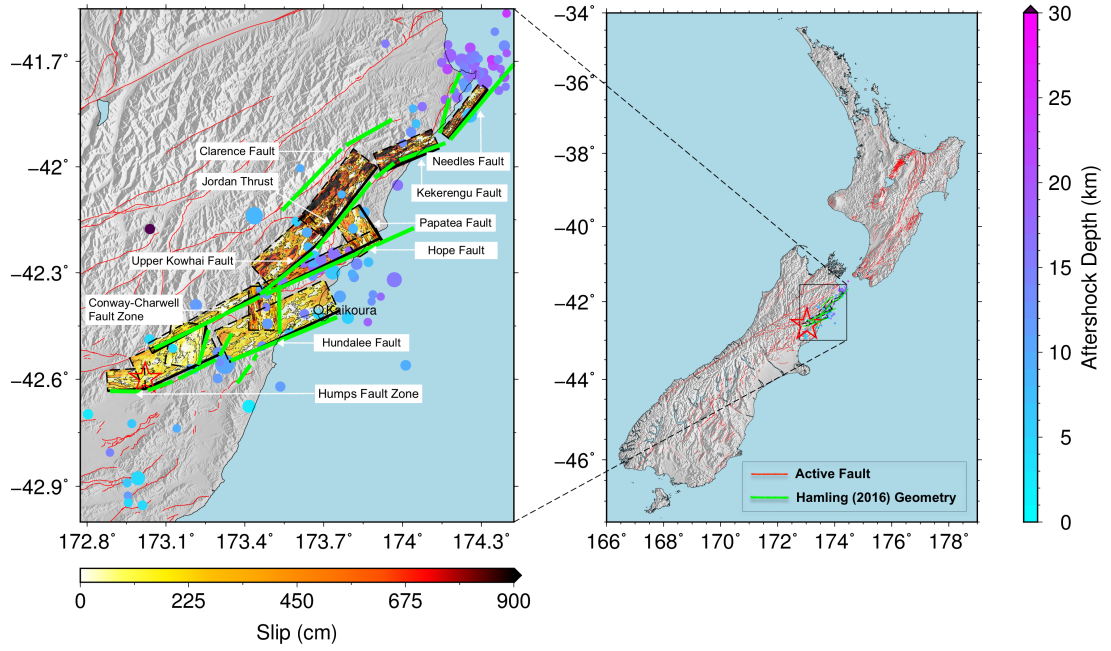


Figure 1: Location of the M_w 7.8 Kaikoura earthquake and causative faults on the east coast of the South Island, New Zealand. Hypocentre is marked with a red star. Surface trace of the source inversion of Hamling (2016) shown in green, and adopted kinematic rupture model faults shown as planes coloured by slip amplitude and contoured by rupture time. $M_w > 4.5$ GeoNet CMT aftershocks till 15 Dec 2016 also illustrated as a function of centroid depth. The inferred rupturing faults are discussed further subsequently.

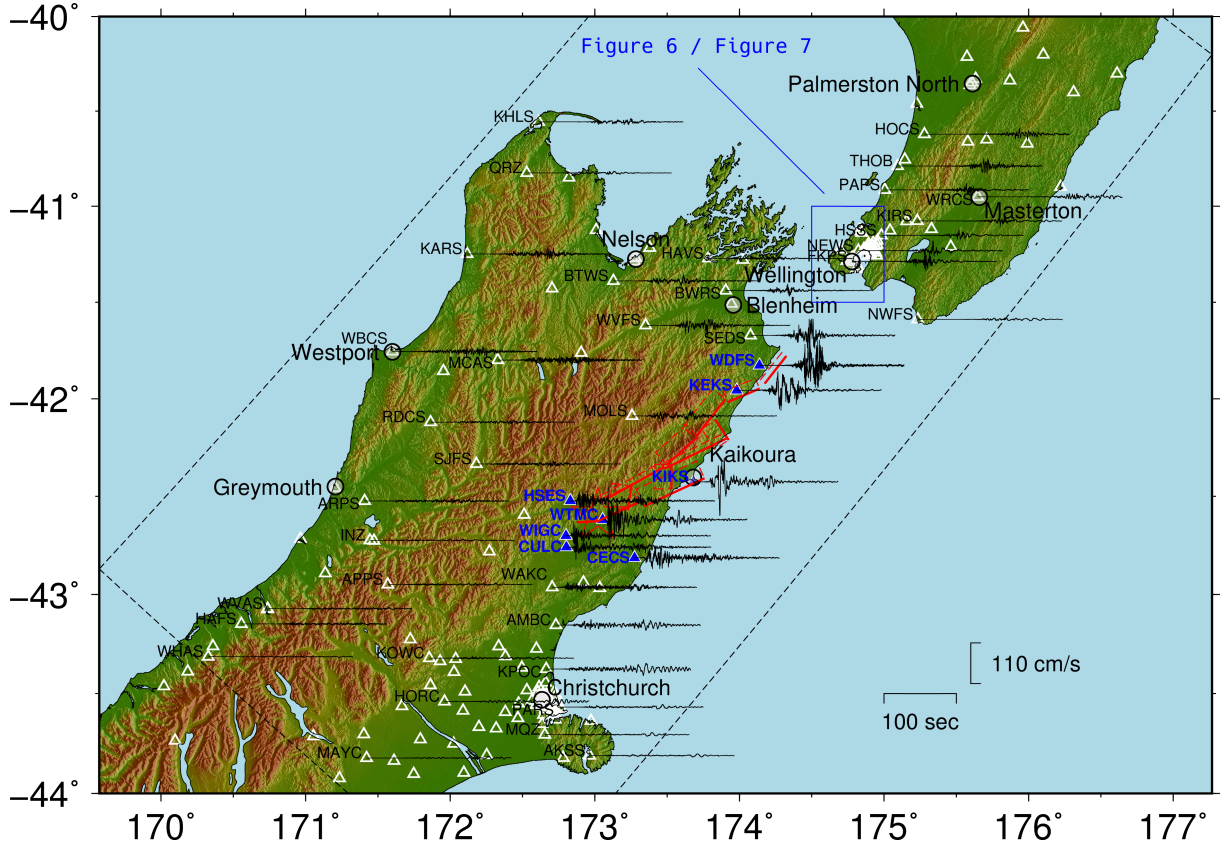


Figure 2: Spatial illustration of the variation in observed ground motion velocities at selected strong motion stations in the east-west (090) component. Stations which are referenced in Figures 3 and 4 are distinguished with different color, as is the geographic location of stations discussed in Figures 6 and 7. Figure S2, available in the electronic supplement to this article, presents all three ground motion components.

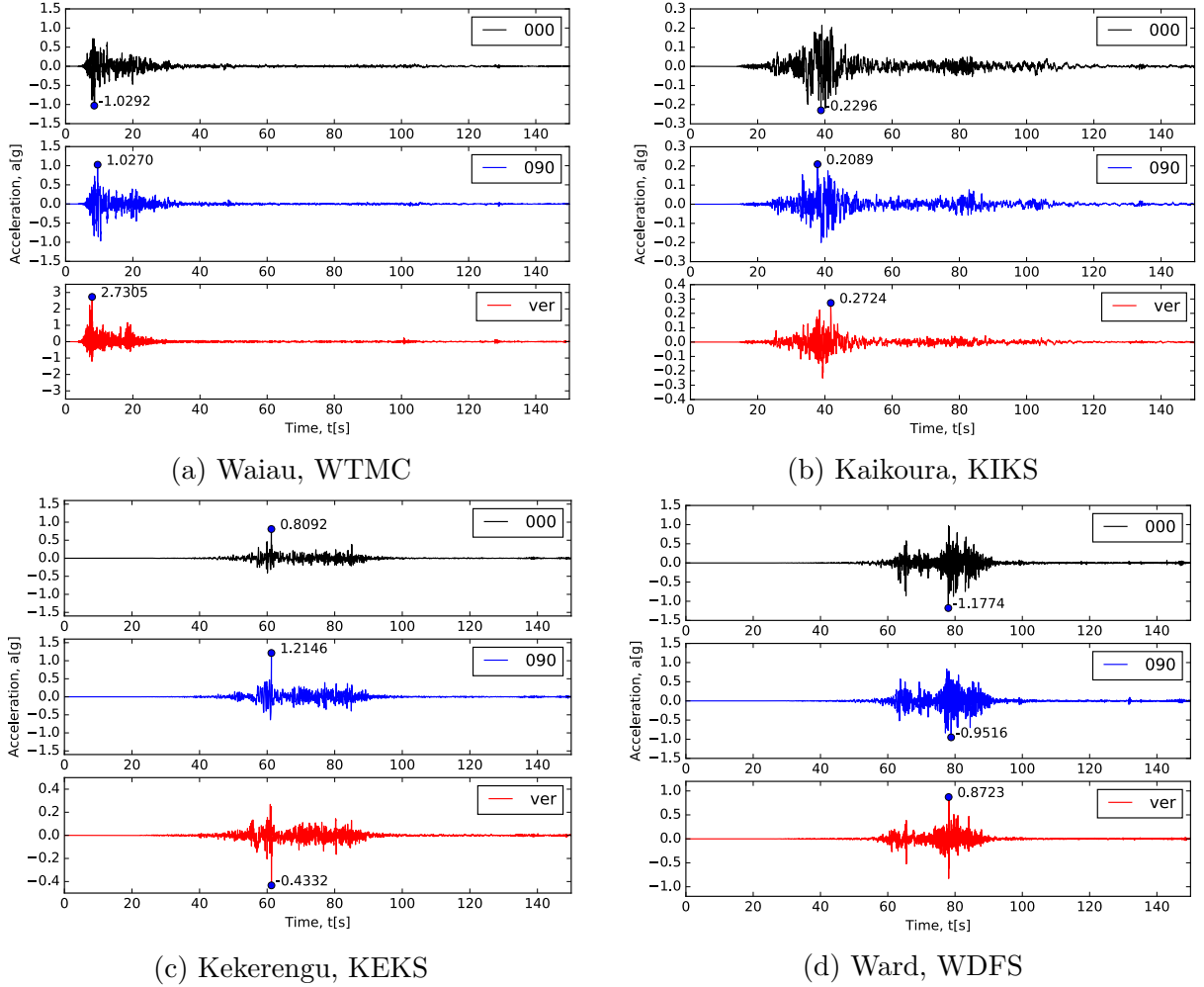


Figure 3: Recorded ground motion accelerations in the immediate vicinity of the rupturing faults (locations noted in Figure 2). 000, 090, and ver represent north-south, east-west and vertical components, respectively. Maximum accelerations in each component are explicitly noted. Different vertical axis scales are used for each station, and between horizontal and vertical components for clarity.

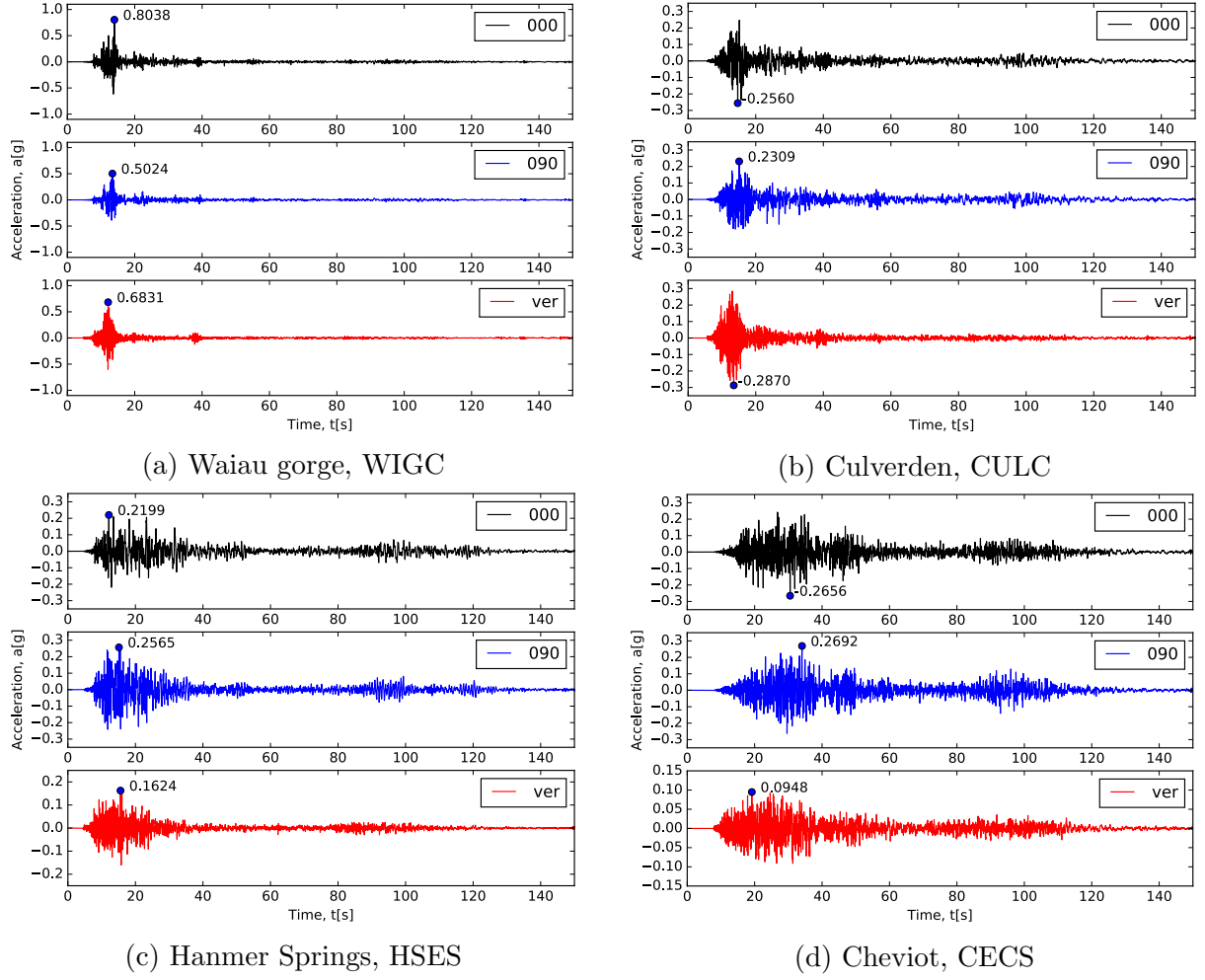
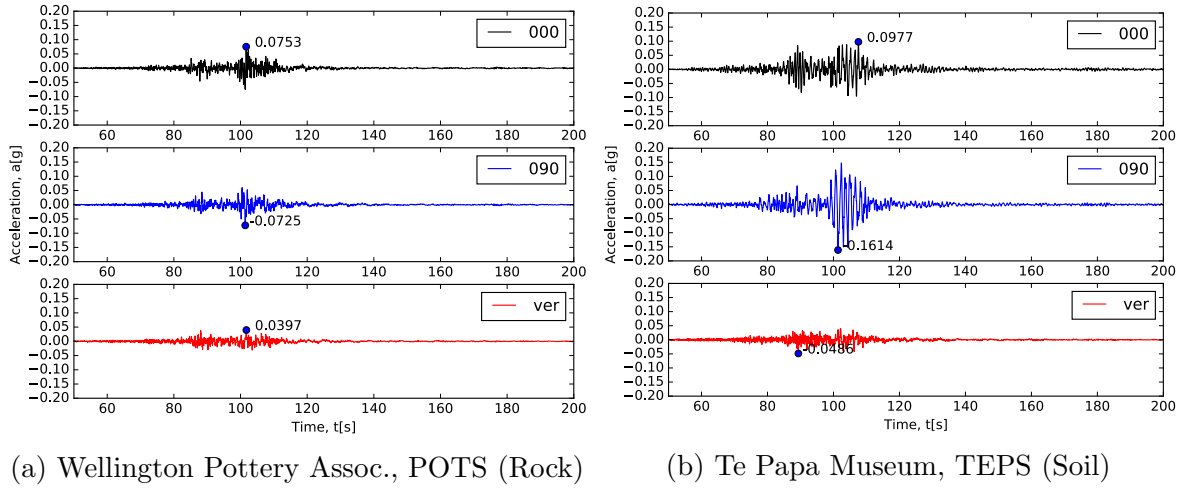


Figure 4: Recorded ground motions in the epicentral region of the initiating rupture on The Humps Fault zone (locations noted in Figure 2). 000, 090, and ver represent north-south, east-west and vertical components, respectively. Maximum accelerations in each component are explicitly noted. Different vertical axis scales are used for each station, and between horizontal and vertical components for clarity.



(a) Wellington Pottery Assoc., POTS (Rock)

(b) Te Papa Museum, TEPS (Soil)

Figure 5: Ground motion accelerations in Wellington (location noted in Figure 2) illustrating the role of shallow and deep site effects. 000, 090, and ver represent north-south, east-west and vertical components, respectively. Maximum accelerations in each component are explicitly noted.

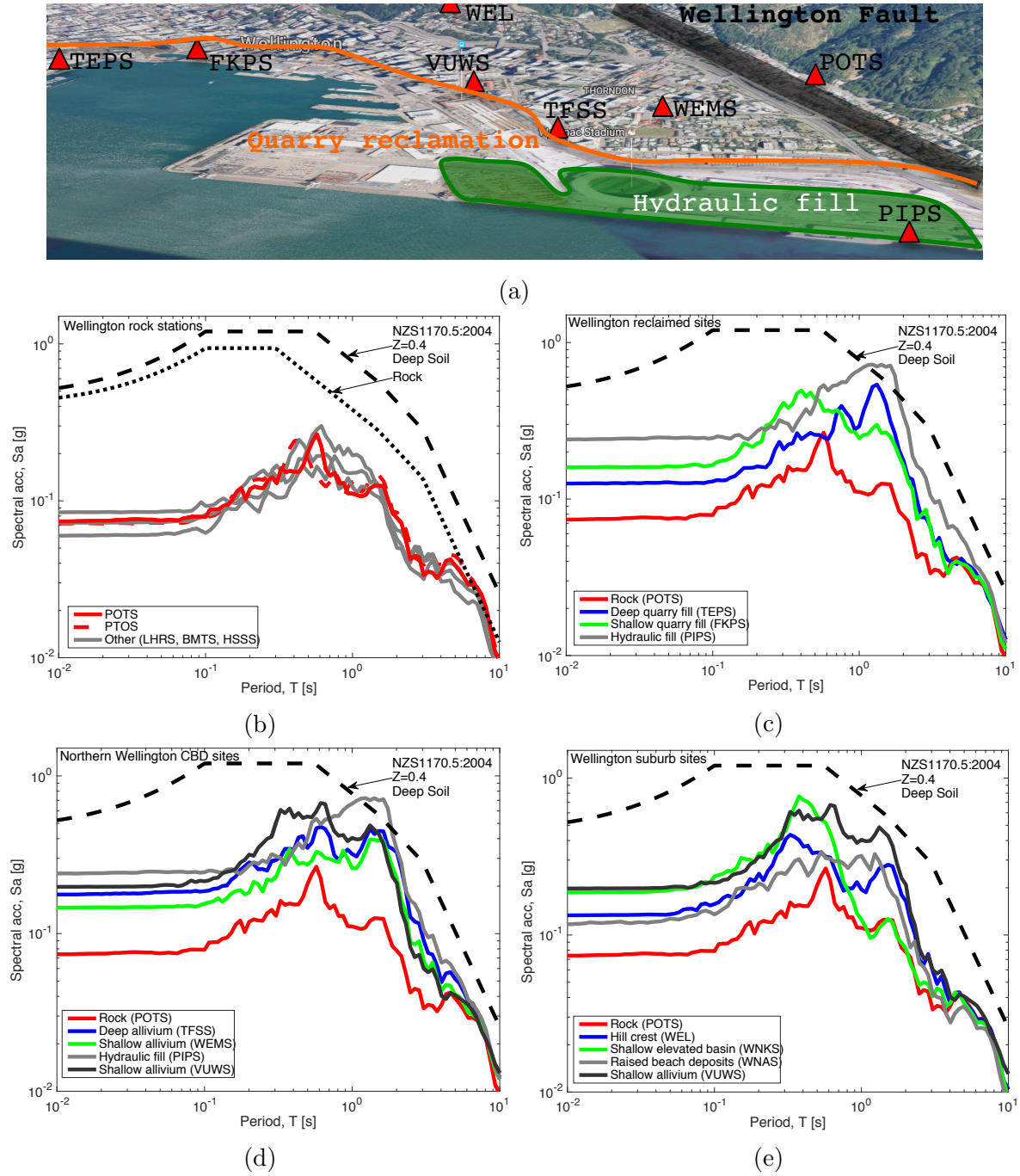


Figure 6: (a) Location of ground motion stations in central Wellington, and observed geometric mean response spectra relative to the design spectra for: (b) rock stations; (c) reclaimed sites; (d) northern central business district (CBD); and (e) suburban sites. Figure 6a notes the location of the Wellington Fault as well as areas of reclamation. The Wellington region is identified in Figure 2. Generalized near surface conditions are described in the figure legends.

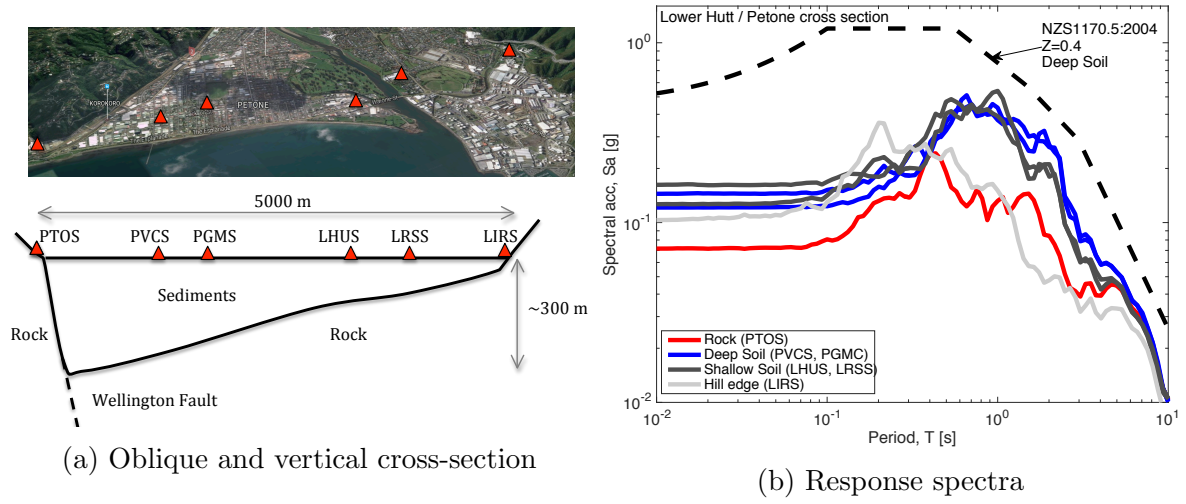


Figure 7: Site response effects in the Lower Hutt / Petone area: (a) Oblique view and projected vertical cross-section of the station locations at the mouth of the valley; and (b) Observed geometric mean response spectra relative to the deep soil design spectrum. The Lower Hutt region is identified in Figure 2. The schematic sedimentary soil depth profile based on Boon et al. (2011)

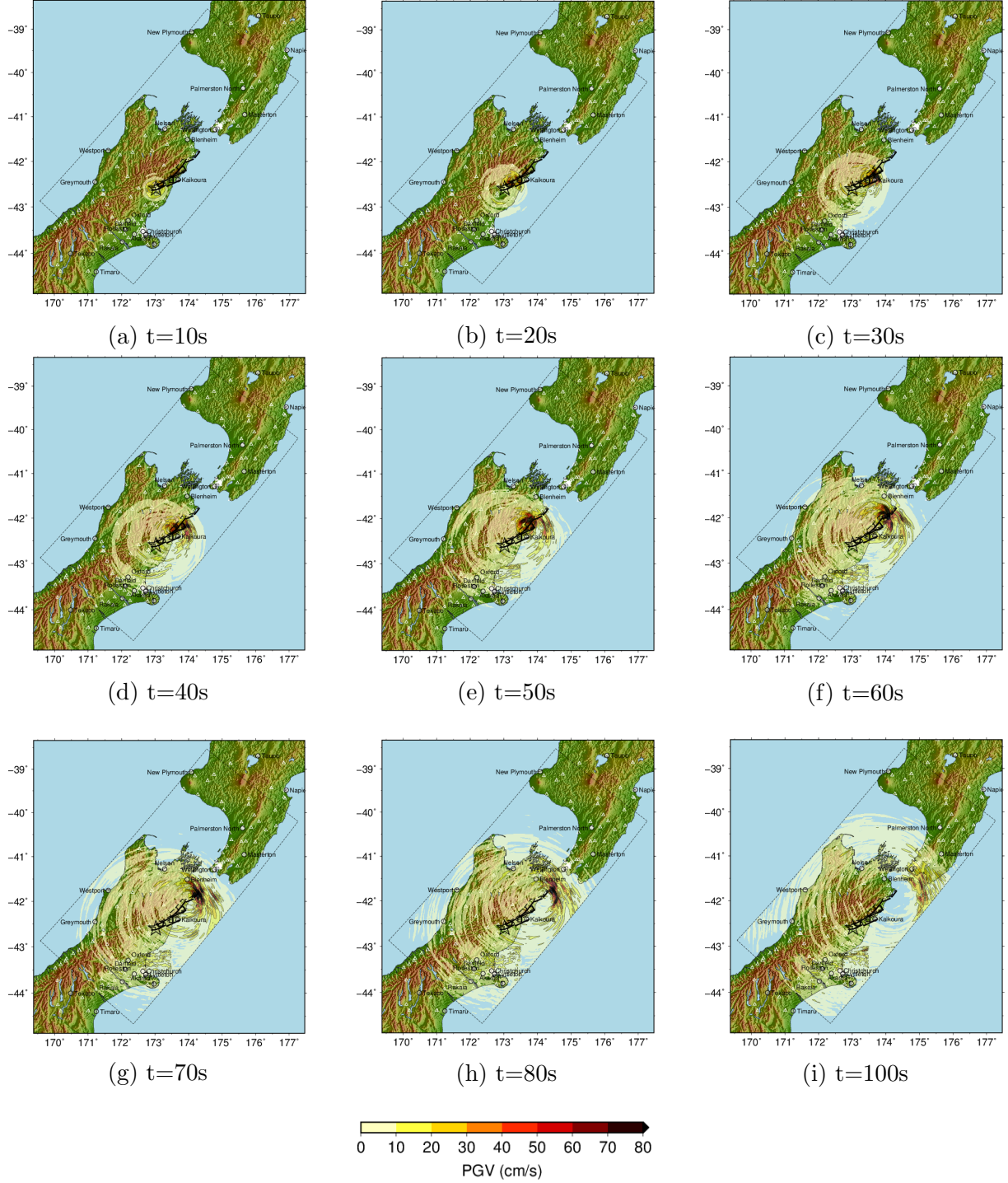


Figure 8: Time snapshots of simulated peak ground velocity (three component maximum). The causative fault segments and simulation domain are also indicated. A video of the simulation is available in the electronic supplement to this article and at: <https://youtu.be/ZbI7rgnZ2U8>.

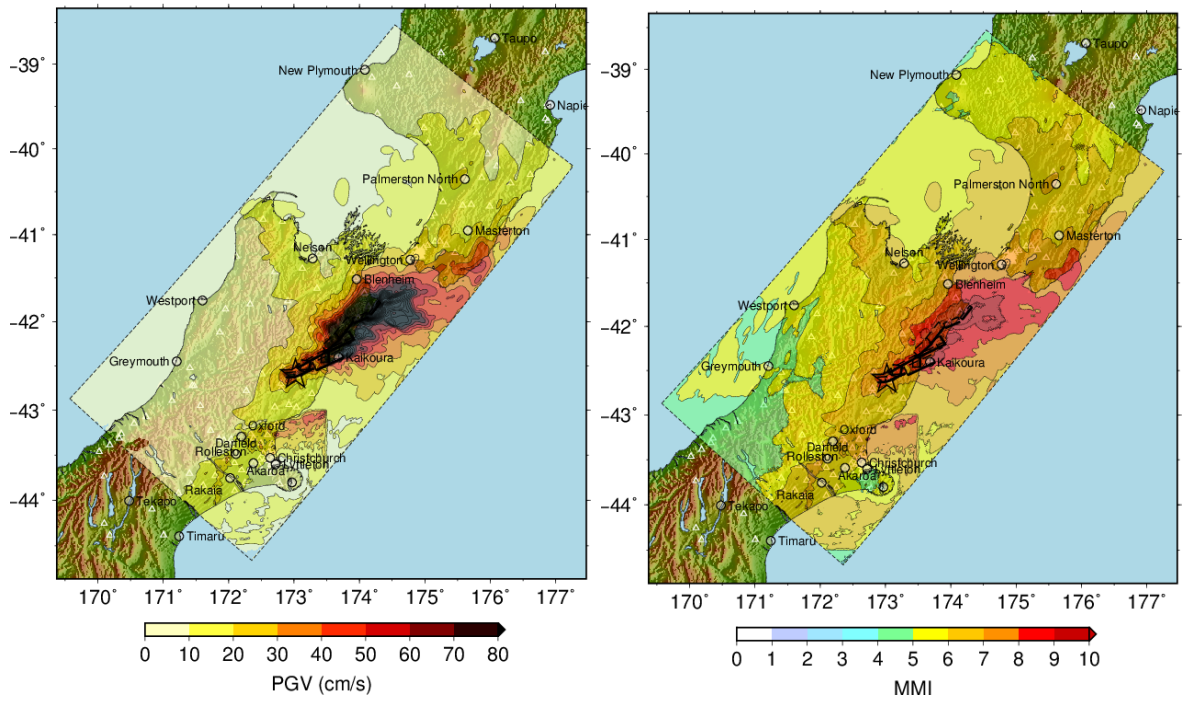


Figure 9: Spatial variation of: (a) peak ground velocity (PGV); and (b) Modified Mercalli Intensity (MMI) from the ground motion simulation (vector maximum of the two horizontal components).

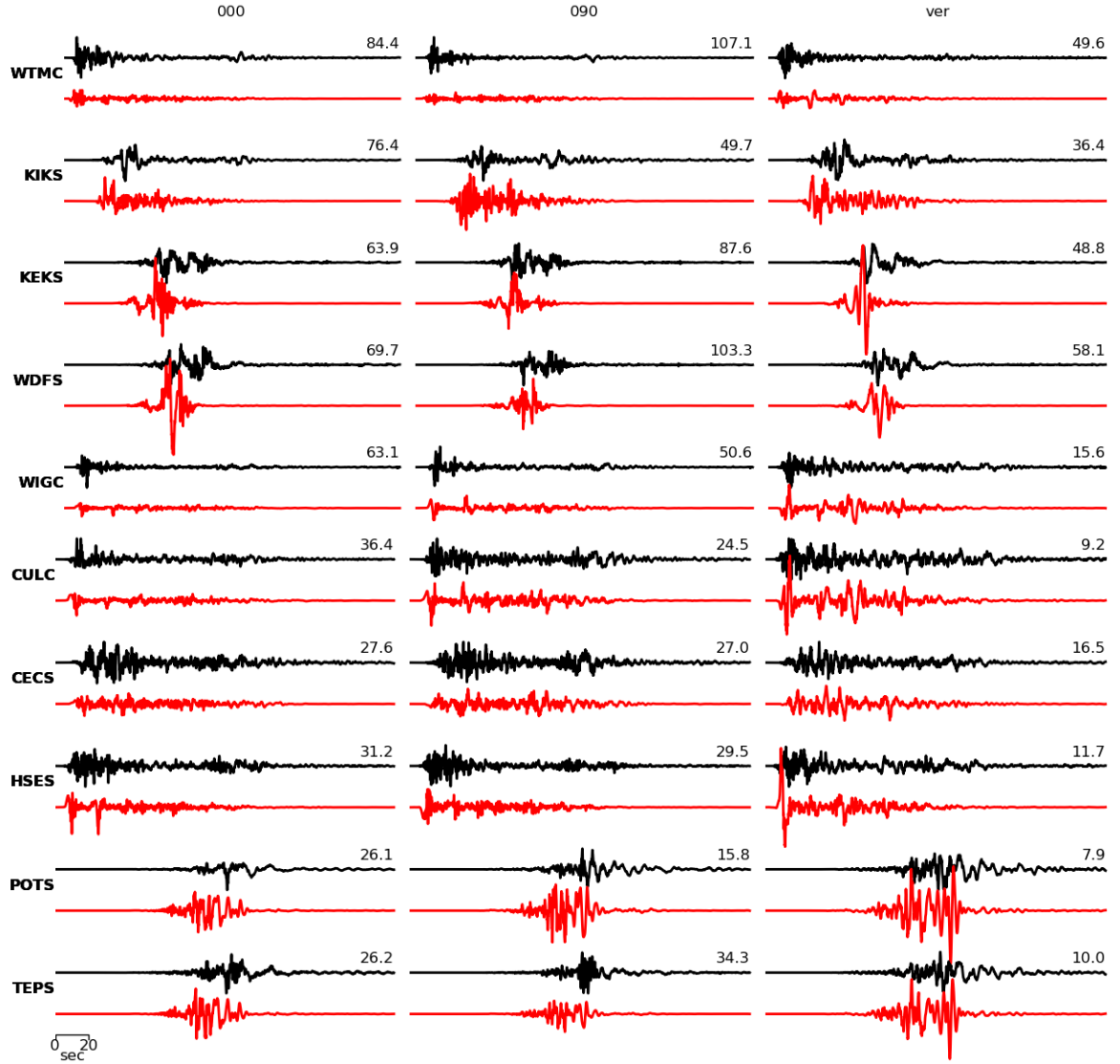


Figure 10: Comparison of observed and simulated velocity time series at strong motion stations of interest in the near-fault region (WTMC, KIKS, KEKS, WDFS), the epicentral region (WIGC, CULC, CECS, HSES) and Wellington (POTS, TEPS). Observed and simulated waveforms are shown in black and red, respectively, with the displayed amplitude normalized to the maximum observed value which is provided at the right of each waveform. The three columns represent the 000 (North-South), 090 (East-West), and vertical components. A comparison for all stations is provided in Figure S3, available in the electronic supplement to this article.

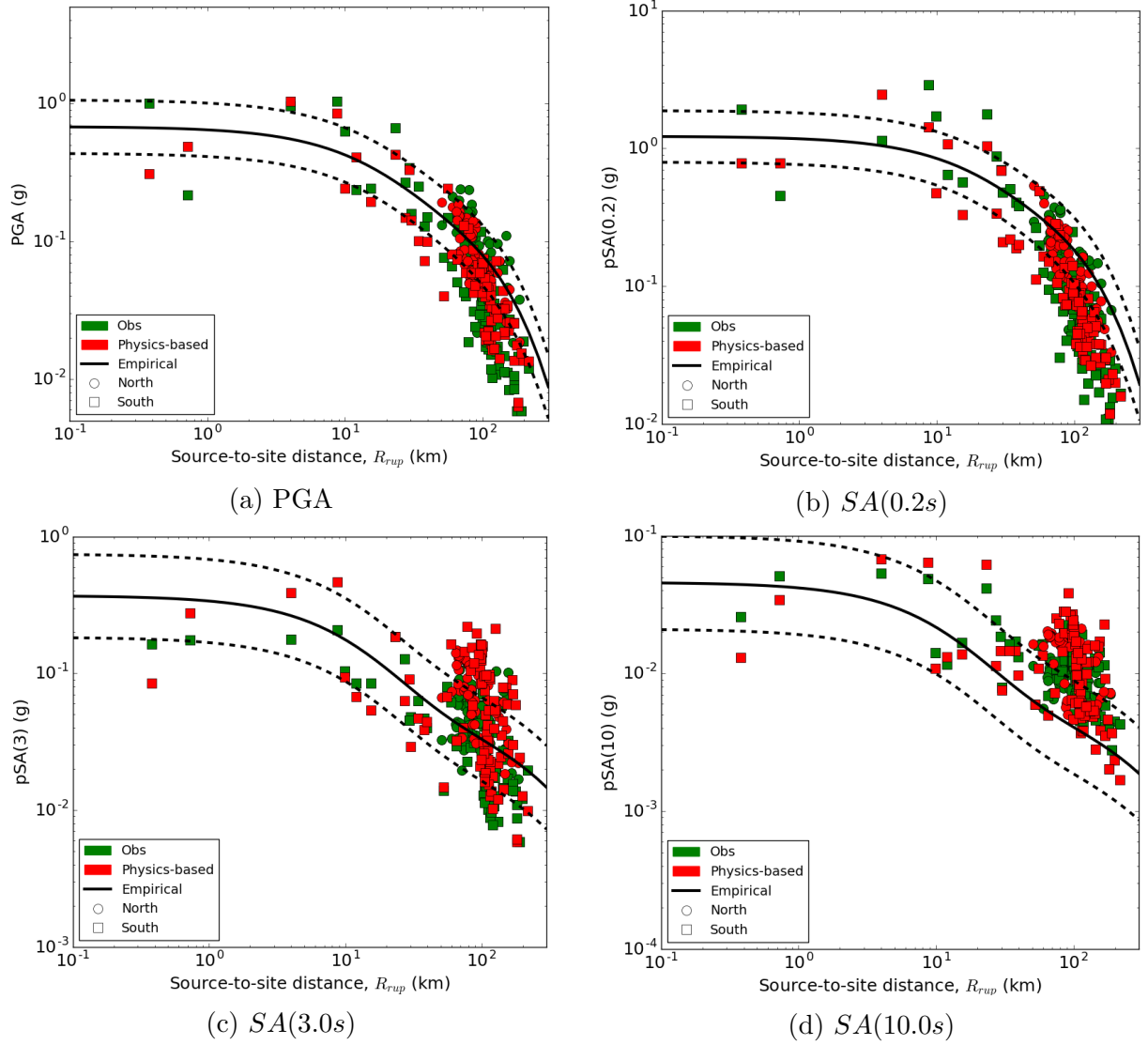


Figure 11: Observed, simulated, and empirically-predicted geometric mean ground motion intensities as a function of source-to-site distance, R_{rup} . Symbol shape indicates location of the station in the North or South Island. The median, and 16th/84th percentiles of the empirical prediction Bradley (2013) are represented by solid and dashed lines, respectively.

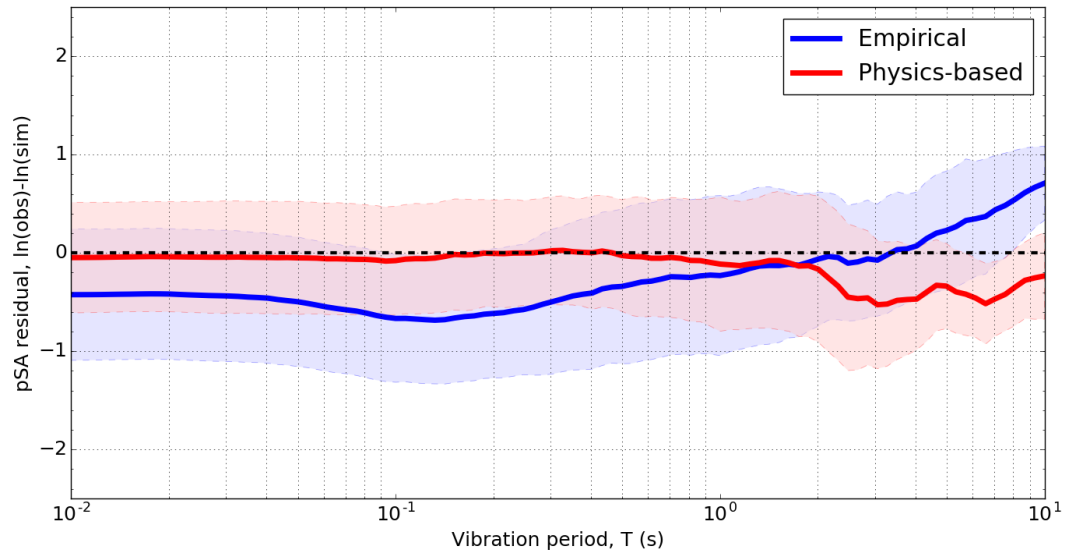


Figure 12: Distribution of residuals between observed and either empirically- or physics-based-predictions of (pseudo) spectral accelerations as a function of vibration period for all stations. The solid line indicates the mean residual, and the shaded region $\pm$ one standard deviation.