Regional kappa (κ) scaling of New Zealand rock GMPEs

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ABSTRACT: New Zealand currently has two main published ground motion prediction equations (GMPEs) to empirically estimate the level of spectral acceleration, $S_A(T)$, due to a given set of earthquake source, path and site parameters. These predictions are designed to fit a dataset of strong motion recordings from around New Zealand. While the data used to create these GMPEs tend to be only from the most active regions of New Zealand, these equations are applied to other regions of low to moderate seismicity, where the data may not be representative of the geological characteristics. This study applies the inverse random vibration theory (IRVT) κ scaling method to adjust the New Zealand GMPEs for regional differences in rock site attenuation. Using recorded small magnitude crustal data, the rock site attenuation parameter, κ_0, is estimated at GeoNet rock sites (www.geonet.org.nz) in several regions around New Zealand. Preliminary results show that κ scaling in Dunedin results in a predicted rock PGA a factor of four greater than the Bradley (2013) prediction for the maximum credible earthquake. For Gisborne, κ scaling may reduce rock PGA by a factor of two, while for Wellington, there is a slight increase in short period motion compared to current predictions. Comparisons with large events from the 2010-2012 Canterbury sequence suggest κ scaling may be beneficial for ground motion prediction.

1 INTRODUCTION

Ground motion prediction models, also known as attenuation relations, are mathematical expressions that relate a specific ground motion intensity parameter (e.g. peak ground acceleration (PGA), or spectral acceleration ($S_A(T)$) to several seismological parameters of an earthquake. The seismological parameters are chosen to quantitatively characterise the earthquake source, the wave propagation path and the local site effects. The ground motion prediction equation (GMPE) is empirically fitted to recorded earthquake data, typically a regional or international database. GMPEs are designed to be a convenient, statistical representation of the median and standard deviation of a ground motion intensity parameter for a given set of earthquake parameters.

To characterise rock site conditions in modern GMPEs, the time-averaged shear-wave velocity in the top 30 metres (V_{S30}) is usually adopted as a predictive parameter in the model. However, V_{S30} cannot fully model the high-frequency spectral content (Laurendeau et al., 2013), which is controlled by damping in the upper crust and parameterised by the 'spectral decay parameter', κ (Anderson and Hough, 1984). κ controls the rate of decay of Fourier amplitudes at high frequencies, modelled as:

$$A(f) = A_0 \exp(-\pi \kappa f) \quad , \quad \text{for } f > f_E \tag{1}$$

where A_0 is a source- and path-dependent Fourier amplitude, f is the frequency and f_E is the frequency above which the decay is approximately linear on a plot of log(A) against f. An example of κ fitted to the high-frequency part of the Fourier amplitude spectrum (FAS) is shown in Figure 1. κ is thought to be a function of epicentral distance (R) and a site variable (S), i.e.

$$\kappa(R,S) = \kappa_0(S) + \tilde{\kappa}(R) \quad , \tag{2}$$

where $\kappa_0(S)$ is the attenuation in the upper few kilometres of the Earth's crust and is unique to every site, and $\tilde{\kappa}(R)$ is the distance-dependence of κ (Anderson, 1991).

Numerous studies have shown that $\kappa_0(S)$, hereafter referred to as κ_0, varies significantly between rock sites and can have a large effect on the severity of high-frequency ground-motion, therefore it is a key parameter in rock site ground motion prediction. Given that κ_0 is not typically included as a site parameter in GMPEs, current models cannot accurately predict short period rock site ground motion. The κ_0 effect will be implicitly included in the GMPE, however the predicted short period motions will not be well constrained, and will be characteristic of an average κ_0 value from the dataset.

As an example, the New Zealand strong motion dataset compiled by Zhao and Gerstenberger (2010) is shown in Figure 2. The dataset of shallow crustal events is primarily made up of events in the most active regions of New Zealand (i.e. in the vicinity of the Alpine fault or on the east coast of the North Island) with very few events in regions of low to moderate seismicity. Therefore, GMPEs regressed on this dataset will model an average κ_0 for sites in the active regions. These regions are often referred to as the 'host' region of a GMPE i.e. the region of the database.

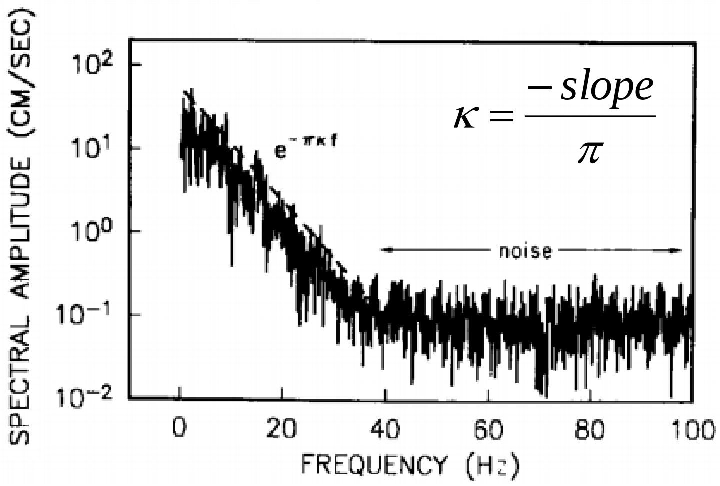

Figure 1. Example FAS of acceleration, with equation (1) fitted to the high-frequency slope (from Anderson and Hough, 1984)

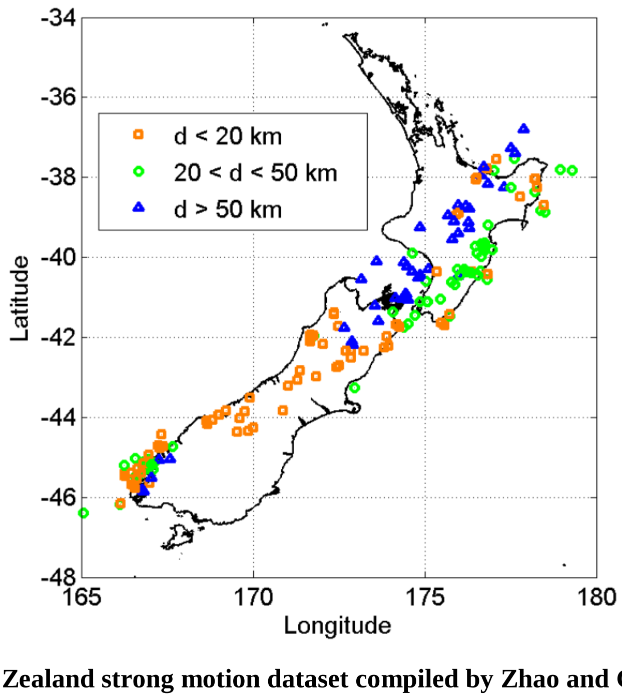

Figure 2, 1966-2010 New Zealand strong motion dataset compiled by Zhao and Gerstenberger (2010) (Figure from Bradley (2010))

For the purposes of probabilistic seismic hazard assessment (PSHA), it is often necessary to use a GMPE to predict ground motion in a region that is outside the GMPE's host region e.g. in low to moderate seismicity regions such as Otago and Christchurch. In these cases, κ_0 scaling factors may be required to correct for the difference in average κ_0 between host and 'target' regions. Campbell (2003) first suggested that scaling factors could be applied to GMPEs, based on regional differences between a number of seismological parameters (including κ_0). In this method, the scaling factors are calculated as the ratio of point-source stochastic simulations between the host and target regions. While the Campbell (2003) method has been widely used, calculating the scaling factors is an onerous task that requires detailed knowledge of many seismological parameters in both regions. Another shortfall of the method is that the scaling factors are calculated in the Fourier domain, but applied in the response spectral domain. Response spectra tend to smooth the effects of κ_0 at high frequencies, and therefore scale differently to Fourier spectra. This means adjustment factors from the Campbell (2003) method may not be applicable at high frequencies.

Recently, an alternative technique to scale GMPEs for κ_0 has been introduced, which applies the scaling factors in the Fourier domain (Al Atik *et al*., 2014). This method uses inverse random vibration theory (IRVT, Gasparini and Vanmarcke, 1976; Rathje et al., 2005) to derive a FAS from a GMPE-generated response spectrum, then measures 'host' κ_0 values by fitting the high-frequency slope with the Anderson and Hough (1984) model in equation (1). The resulting κ value, known as κ_{IRVT}, represents the 'native' κ of the dataset from which the GMPE was derived. If κ_{IRVT} is measured for short distance scenarios (e.g. less than 20 km), the effect of the seismic quality factor Q can be considered negligible and the measured κ is approximately $\kappa_{0,IRVT}$. The FAS is scaled to a 'target' κ_0 value, then the target response spectrum is calculated using random vibration theory (RVT). An advantage of this method is that only the target κ_0 values are required to calculate the κ scaling factors, rather than a full seismological model. To summarise the method:

1. Generate rock site response spectra for given short-distance earthquake scenarios.
2. Compute response-spectra compatible FAS using IRVT (Rathje *et al*., 2005).
3. Fit the high frequency slope of the FAS with $exp(-\pi \cdot \kappa_{IRVT} \cdot f \cdot)$ for the short-distance earthquake scenarios, then average over all κ_{IRVT} to obtain κ_{host}.
4. Generate a response spectrum for a design earthquake scenario, then convert to FAS using IRVT. Apply κ scaling to the FAS for a design earthquake scenario, by multiplying the host FAS by $exp(-\pi \cdot f \cdot (\kappa_{target} - \kappa_{host}))$.
5. Convert scaled FAS to a response spectrum using RVT.
6. Calculate κ scaling factors by dividing the κ-scaled response spectrum by the initial GMPE response spectrum.

New Zealand currently has two GMPEs in regular use (McVerry *et al*., 2006; Bradley, 2013). The McVerry *et al*. (2006) crustal and subduction models are based on a 1966-1995 New Zealand strong motion database supplemented with foreign, near-source PGA data, while the Bradley (2013) crustal model is derived from an international strong motion database, with some modifications to fit a New Zealand 1966-2010 dataset. While these two GMPEs are deemed to be applicable throughout New Zealand, there are several regions around New Zealand where there are few or no shallow strong motion recordings in the McVerry *et al*. (2006) or Bradley (2013) datasets. These regions can have geology with different attenuating properties to other parts of New Zealand where strong motion data is available, and hence current prediction models may not accurately predict future earthquakes in these regions. The purpose of this study is to apply host-to-target κ_0 scaling to New Zealand GMPEs to obtain more robust predictions for these regions.

2 CALCULATION OF κ_0

2.1 Target region $\kappa_{0,AS}$

The preferred method of obtaining κ_0 for the target region is via directly fitting the high-frequency slope of the acceleration spectra of recorded earthquake data ($\kappa_{0,AS}$). This method is defined by Ktenidou *et al*. (2014) as being part of a family of methods for estimating κ that are suitable for host-to-target adjustments of GMPEs. Many regions around New Zealand have a GeoNet (www.geonet.org.nz) seismic recording instrument located at a surface rock site, from which $\kappa_{0,AS}$ can be directly measured. Therefore, the 'target' region in this study is a GeoNet rock site near an urban area. Target sites identified in this study are Dunedin (GeoNet station OPZ), Wellington (WEL), Gisborne (GKBS) and two sites in the Port Hills south of Christchurch (GODS, MTPS). $\kappa_{0,AS}$ is calculated using the adapted procedures of Ktenidou *et al*. (2013) and Van Houtte *et al*. (2014), with the results shown in Figure 3. A summary is included in Table 1. For the two Christchurch sites, the regional slope was fixed to be the same for both stations.

2.2 Host region $\kappa_{0,IRVT}$

$\kappa_{0,IRVT}$ has been previously calculated for the two New Zealand GMPEs, and compared with measured $\kappa_{0,AS}$ estimates in Christchurch (Van Houtte *et al*., 2014). Stable $\kappa_{0,IRVT}$ estimates were not attainable using the McVerry *et al*. (2006) GMPE, due to the limited high-frequency range of the model. $\kappa_{0,IRVT}$ for the Bradley (2013) model is relatively independent of magnitude, distance (up to R = 100 km, except at small distances where GMPEs are not well constrained, see Figure 4a) and V_{S30} for rock sites (i.e. V_{S30} > 800 m/s, see Figure 4b). For lower V_{S30} values, κ_{IRVT} increases slightly, however for this study the effect is considered minimal, as we are primarily focussed on rock sites. Given that the effect of κ is effectively decoupled from the Bradley (2013) GMPE, this study assumes that the host $\kappa_{0,IRVT}$ = 0.034 s.

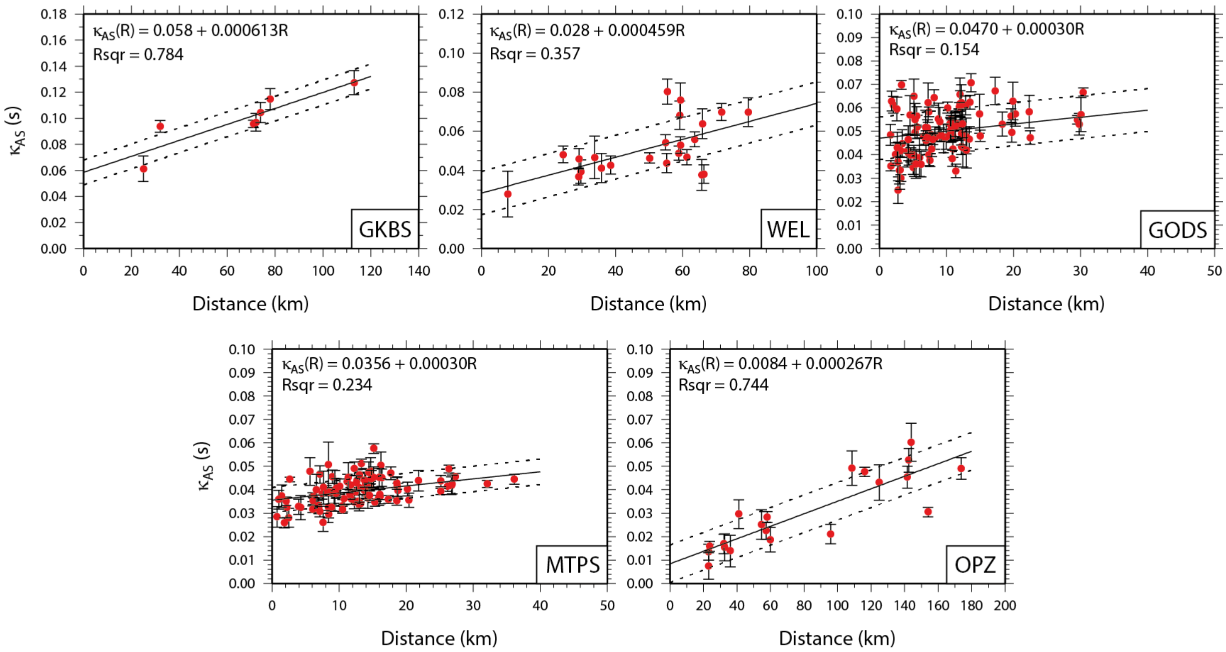

Figure 3. Distance-dependence of κ_{AS} for (a) Gisborne, (b) Wellington, (c) Christchurch Godley Drive, (d) Christchurch Mt Pleasant and (e) Dunedin. Error bars on the data points represent the scatter in κ_{AS} due to orientation of the two horizontal components.

Table 1. $\kappa_{0,AS}$ for five rock sites near urban areas in New Zealand

City	GeoNet station	$\kappa_{0,AS}$ (s)
Gisborne	GKBS	0.058 ± 0.010
Wellington	WEL	0.028 ± 0.011
Christchurch (Godley Head)	GODS	0.0470 ± 0.009
Christchurch (Mt Pleasant)	MTPS	0.0356 ± 0.005
Dunedin (Otago Peninsula)	OPZ	0.0084 ± 0.005

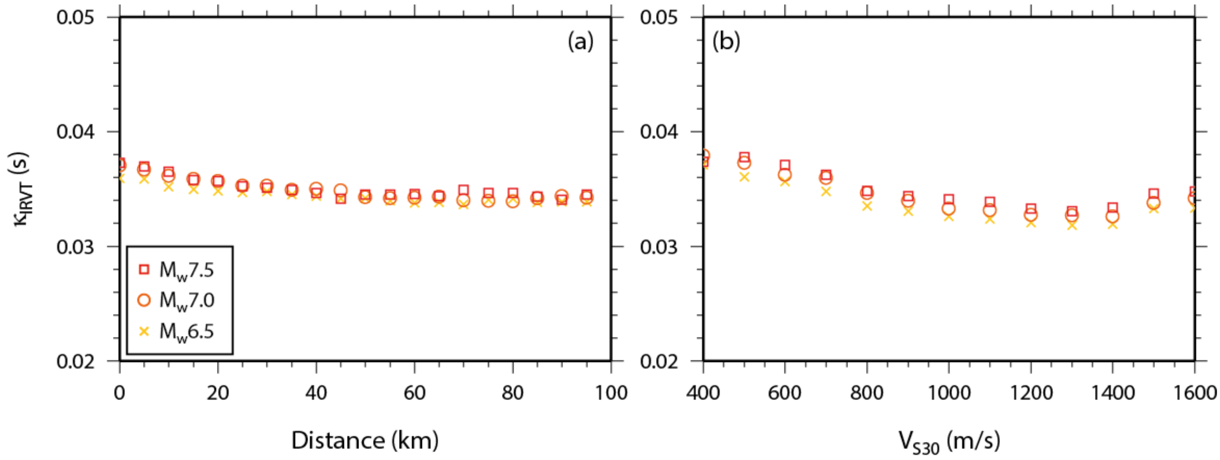

Figure 4. Using the Bradley (2013) GMPE, (a) κ_{IRVT} against distance for V_{S30} = 760 m/s, and (b) κ_{IRVT} against V_{S30} for R = 20 km.

3 PROPOSED SCALING FACTORS

While κ_{IRVT} is independent of distance up to 100 km for the Bradley (2013) model, κ_{AS} typically shows distance-dependence. Therefore, the target κ value is different for near-source and distant earthquake sources, and a κ scaling factor is calculated on an event-by-event basis. Based on 1 in 475 year PGA deaggregations from the New Zealand seismic hazard model (Stirling et al., 2012), Table 2 shows earthquake sources that have large contributions to the hazard for Gisborne, Wellington, Christchurch and Dunedin. The dominant source for Gisborne is unknown to these authors, therefore an example event of M_w6.0 at 10 km distance is considered the design event. κ scaling factors are computed for these sources, using the κ_{IRVT} method outlined in the introduction.

The mean scaling factors computed using this method are shown in Figure 5 for Gisborne, Wellington, two areas of Christchurch, and Dunedin. All adjustment factors tend to one at long periods, as κ is a high-frequency effect. Due to the low κ_{AS} values observed in Dunedin, there is a large increase in short period predictions for the rupture of the Akatore fault, up to a factor of four at T = 0.04 s. There is a slight increase in short period rock motions in Wellington city for a Wellington fault rupture. Christchurch and Gisborne sites are likely to have lower short period motion for events that dominate their PGA hazard.

Table 2. Earthquake sources from the New Zealand seismic hazard model and corresponding target κ.

City	Earthquake source	Contribution to 1/475 year PGA hazard	Moment magnitude	Distance (km)	Mean target κ (s)
Gisborne	?	?	6.0	10	0.064
Wellington	Wellington fault	20%	7.5	< 1	0.028
Christchurch (Godley Head)	Distributed	15%	5 – 6.8	~ 30	0.056
Christchurch (Mt Pleasant)	Distributed	15%	5 – 6.8	~ 30	0.045
Dunedin	Akatore fault	14%	7.4	13	0.011

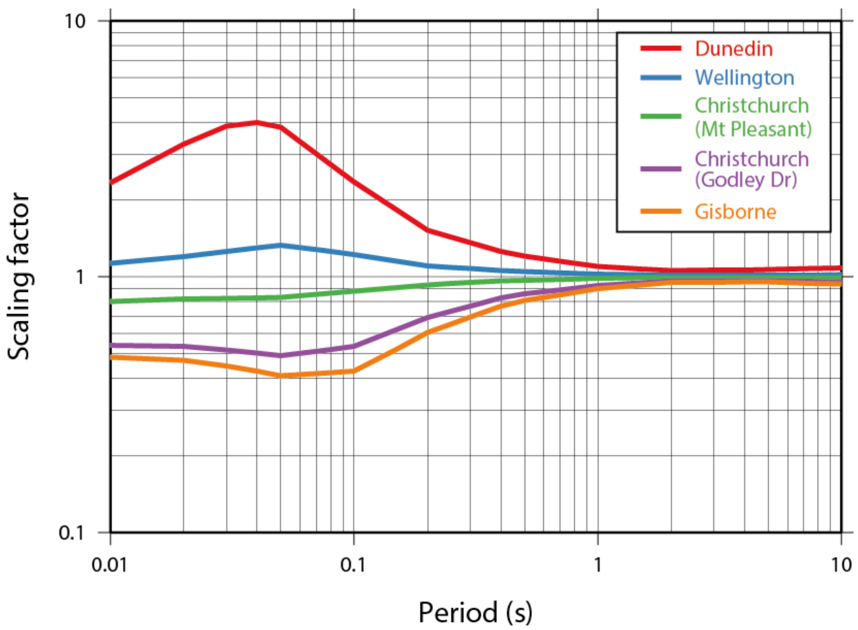

Figure 5. Proposed mean regional κ scaling factors for the Bradley (2013) GMPE for the events that dominate the 1/475 year PGA hazard.

4 DISCUSSION AND CONCLUSIONS

There are several limitations and uncertainties associated with these scaling factors. For the IRVT scaling method to be valid, the crustal amplification function (i.e. V_S amplification) should be flat at high frequencies, to ensure the high-frequency shape of the FAS is controlled only by κ. While this is likely the case for hard rock sites e.g. in Dunedin, it may not be the case for the softer rock sites. For these soft rock sites, the IRVT-generated FAS should be divided by the crustal amplification function before applying a κ correction. While work is currently underway to obtain a crustal amplification function for Dunedin, these factors are unavailable for other regions in New Zealand. As this study assumes that the crustal amplification is flat at high frequencies for all sites, the κ scaling factors presented here should be considered provisional until crustal V_S scaling functions are known.

Also note that in each region, we have elected to scale the Bradley (2013) model to a target κ value, rather than to a zero-distance κ_0 value. The distance-dependence of κ is proportional to the seismic quality factor Q in the frequency band of measurement (i.e. 10-40 Hz). This means that the κ scaling factors proposed here are also adjusting for regional high-frequency Q effects. It is typically preferable to adjust GMPEs for rock site effects (V_S and κ_0) and Q separately, but we choose to include some path adjustment here because no Q scaling functions are currently available. The κ adjustment factors we propose here will be more representative of a deterministic ground-motion scenario, however if future studies adjust the Bradley (2013) model for Q, care will be required to ensure the difference in the high-frequency Q effect is not being double-counted.

It is difficult to validate the proposed κ adjustments with strong motion crustal data, as there are few recorded earthquakes for Gisborne, Wellington or Dunedin. While the Canterbury earthquake sequence (Bannister and Gledhill, 2012) provided a large number of strong motion recordings for NZS1170.5:2004 class D and class E sites, there were no near-source strong motion instruments on rock sites for the September 2010 M_w7.1 Darfield earthquake and February 2011 M_w6.3 Christchurch earthquakes. However, a number of instruments were installed on rock sites after the Christchurch earthquake, recording the June 2011 M_w6.0 Sumner earthquake and events from the December 2011 Pegasus Bay sub-sequence (Ristau *et al.*, 2013). While we are unable to use these events to perform statistically significant validation, a preliminary residual analysis of the κ scaling factors is shown in Figure 6. The M_w6.0 Sumner earthquake was neglected from the analysis, as there are still large uncertainties about the character of the source, and the Bradley (2013) model fits the data poorly both before and after κ scaling. Therefore the only five remaining rock site recordings that fall within the

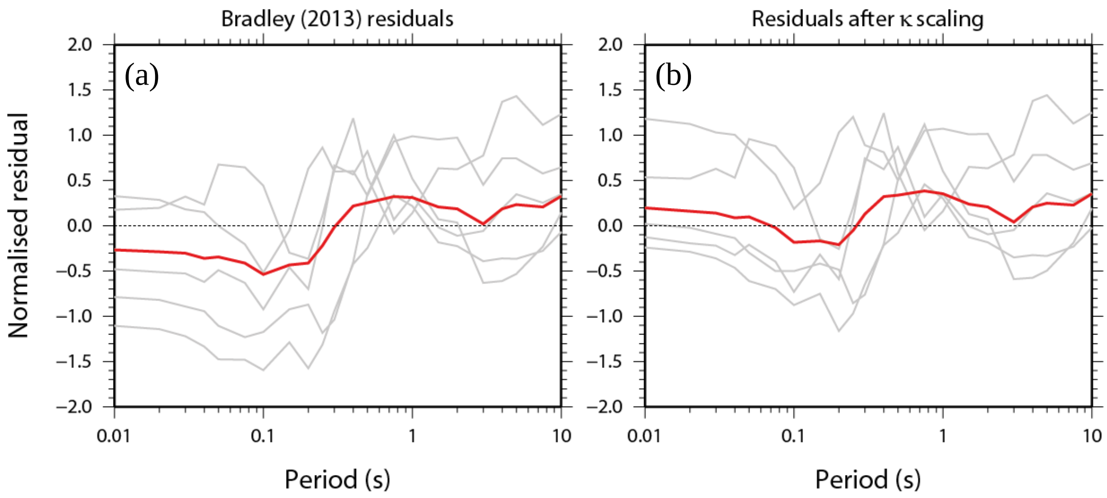

Figure 6. Period-dependent normalised residuals for the Bradley (2013) GMPE (a) before κ scaling and (b) after κ scaling for five events recorded at the MTPS and GODS strong motion stations.

applicable moment magnitude range of the Bradley (2013) model are:

- 21 June 2011, M_w5.2 (recorded by the GODS strong motion station);
- 23 December 2011 12:58pm, M_w5.8 (recorded by GODS & MTPS); and
- 23 December 2011 1:18pm, M_w5.9 (recorded by GODS & MTPS).

The normalised residuals for these events are plotted as grey lines in Figure 6, with the red line representing the mean of the normalised residuals. As these five recordings are at short distances, $\kappa_{0,AS}$ for the GODS and MTPS stations are used as the target κ_0 values for the κ scaling factors, rather than the adjustment factors plotted in Figure 5.

Figure 6a shows the Bradley (2013) model is slightly overpredicting the short period spectral accelerations for these five recordings. The normalised residuals have a peak at ~0.8 s, however this is likely to be due to the large topographic amplification at the GODS station observed in Van Houtte *et al.* (2012). The overprediction at short periods is in agreement with the results of this study, as the host $\kappa_{0,IRVT}$ from the Bradley (2013) model (0.034 s) is lower than target $\kappa_{0,AS}$ for the GODS and MTPS sites (0.037 and 0.052 s respectively). After κ scaling, the trend in the residuals appears to be reduced and predictions improved (shown in Figure 6b). While the residual analysis here is by no means statistically significant, these preliminary results suggest that κ scaling for the rock sites may be beneficial for ground motion prediction.

5 ACKNOWLEDGMENTS

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