Probabilistic seismic hazard analysis of New Zealand

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Abstract We construct probabilistic seismic hazard (PSH) maps for New Zealand that are based on the distribution and long-term recurrence behaviour of active faults and the spatial distribution of earthquakes observed in historic time. Slip rate, single-event displacement, and return time data for 154 active faults (including segments of the Hikurangi and Fiordland subduction zones) are combined with observations of the magnitudes and rupture lengths of large New Zealand earthquakes since 1843 and the instrumental record of seismicity since 1964 to predict the future ground motions that will occur across the country. Maps of the peak ground accelerations and 0.5 s response spectral accelerations expected at 10% probability in 50 yr on “rock” show the highest accelerations (≥0.2 g and locally over 0.6 g) in a belt that extends from the southwestern end of the country to the northeastern end, along the faults that accommodate essentially all of the relative plate motion between the Australian and Pacific plates. Accelerations of >0.4 g are predicted for Wellington City, which lies within the belt of high PSH, whereas accelerations of <0.2 g are predicted for Christchurch, and <0.1 g for Auckland and Dunedin Cities, all outside the belt of high PSH. Maps of the peak ground accelerations and 0.5 s spectral accelerations expected at 2% probability in 50 yr show accelerations of 0.7 g for Wellington and <0.2 g for the other cities. The PSH of Wellington is produced by the predicted occurrence of large-to-great earthquakes at distances of <50 km, whereas the PSH of the other cities is generally produced by the predicted occurrence of large earthquakes at distances of 50 km or more. Of all the major urban areas of New Zealand, Wellington is subject to the highest seismic hazard.

Keywords probabilistic seismic hazard; PSH; neotectonics; active faults; earthquakes; New Zealand

INTRODUCTION

In recent years, there have been major advances in the science of probabilistic seismic hazard analysis (PSHA). Recent efforts have been focused toward developing “multidisciplinary” PSH models that combine geologic data (fault length, slip rate, and paleoearthquake data) with historical seismicity data to estimate the future ground motions that will occur at a site or across a gridwork of sites (e.g., Working Group of California Earthquake Probabilities 1995; Frankel et al. 1996). These efforts have been motivated by the observation that historical records of seismicity are generally much shorter than the repeat time of the largest earthquake on a fault. To date, efforts to characterise the PSH of New Zealand at a national scale have been based largely on the distribution of historical seismicity (Smith & Berryman 1986), though several multidisciplinary PSHAs have recently been completed for specific regions of New Zealand (e.g., Berryman et al. 1997).

In this paper we compile geologic data describing the geometry and activity (fault lengths, slip rates, single-event displacements, and return times) of the 154 major active faults of New Zealand, and combine these data with historical seismicity data to develop PSH maps for the country. Our approach is to use the geologic data and historical observations of large earthquakes to estimate the locations, magnitudes, and recurrence rates of future large earthquakes. We then use instrumental seismicity data recorded by the Institute of Geological & Nuclear Sciences (GNS, formerly the Department of Scientific and Industrial Research, or DSIR) over the period 1964–96 to predict the locations, magnitudes, and recurrence rates of moderate-to-large “background” earthquakes in the areas between the mapped faults, thereby addressing the possibility that damaging earthquakes may also be produced by unknown faults. The PSH maps we develop from these data are maps of the peak ground accelerations and 0.5 s response spectral accelerations expected at 10% probability in 50 yr, and at 2% probability in 50 yr, at a gridwork of rock sites. Though these measures of PSH are of importance to engineers and planners, we stipulate that our maps not be used for engineering or planning studies, since the maps are constructed with an unpublished preliminary attenuation relationship for New Zealand, and also incorporate new experimental research on the treatment of uncertainty in estimates of ground motion from attenuation relationships.
NEOTECTONIC FRAMEWORK AND HISTORICAL SEISMICITY

New Zealand straddles the boundary of the Australian and Pacific plates, where relative plate motion is obliquely convergent across the plate boundary at c. 40 mm/yr at the latitude of central New Zealand, and 30 mm/yr in the south (DeMets et al. 1990; Fig. 1). The relative plate motion is expressed in New Zealand by the presence of numerous active faults (Fig. 1, 2), and a high rate of magnitude ($M \geq 4$) earthquakes, including the occurrence of many large earthquakes (and one great earthquake) in historic time (Fig. 3). The historic record of $M \geq 6.5$ earthquakes dates from 1843, which was the time that European settlement began in New Zealand. A southeast-dipping subduction zone lies at the far southwestern end of the country ("Fiordland subduction zone" in Fig. 1), and this is linked to a major northwest-dipping subduction zone in the eastern North Island ("Hikurangi subduction zone" in Fig. 1) by a 1000 km long zone of dextral oblique slip faults ("axial tectonic belt" in Fig. 1). Essentially all of the relative plate motion is accommodated by the faults of the axial tectonic belt in the area between the Fiordland and Hikurangi subduction zones.

The Hikurangi subduction interface dips to the northwest at c. 15° beneath the eastern North Island (e.g., Robinson 1986; Darby & Beanland 1992), and is at a depth of c. 20 km beneath Wellington City. Geodetic modelling indicates that there is a transition from a fully coupled subduction interface in the south to an uncoupled interface in the north (Beavan & Haines 1997; Fig. 1), and abrupt changes in the spatial and depth distribution of seismicity along the subduction interface.
Fig. 2 Active faults used as input for the PSHA from (A) the South Island, and (B) the North Island. The numbers beside each fault correspond to the numbers given on the left-hand column of Appendix 1. The numbers in large print denote subduction zone sources, and the surface projection of each subduction zone source is shown by the polygon closest to each number. The inset shows the geographic areas used to sort the faults in Appendix 1.

have been suggested to mark “tears” or segment boundaries in the subduction zone (Reyners 1983). However, no large or great earthquakes have occurred on the Hikurangi subduction interface in historic time (since 1843), so little is known about the earthquake potential of this feature. The Fiordland subduction zone dips at c. 15° SE immediately offshore from Fiordland, and is steeply dipping beneath Fiordland. The Fiordland subduction interface also shows abrupt changes in seismicity patterns along strike, and the lateral extent of the aftershock zone of a recent large earthquake (the $M_w$ 7.3, August 10 Fiordland earthquake; Van Dissen et al. 1994) shows that rupture lengths less than the length of the entire subduction zone do occur. High rates of $M \geq 4$ earthquakes occur above the Fiordland subduction zone and to a lesser extent above the Hikurangi subduction zone (Fig. 3).

The axial tectonic belt is a zone of dextral transpression, most dramatically illustrated by the central to south Westland section of the Alpine Fault (Fig. 1), where dextral slip rates of 15–35 mm/yr, and uplift rates of 17 mm/yr, are observed (Berrymen & Beanland 1988; Berrymen et al. 1992; Sutherland & Norris 1995). The Alpine Fault accommodates virtually all of the relative plate motion in the central South Island, whereas several faults with slip rates $>1$ mm/yr accommodate the plate motion in central New Zealand (Fig. 1). Many $M \geq 6.5$ earthquakes have occurred within the axial tectonic belt in historic time (Fig. 3; Table 1), including the two largest earthquakes to occur in historic time (the $M_w$ 8.1–8.2, 1855 Wairarapa earthquake, and $M_w$ 7.8, 1931 Napier earthquake). However, only a few of these earthquakes have produced clear evidence of surface ruptures, despite their large magnitudes. The reasons for the lack of surface ruptures are: (1) some of the earthquakes are blind thrust events (e.g., the Napier earthquake); and (2) the combination of sparse historical records and rapid modification of surface ruptures in the humid New Zealand climate may have prevented the identification of some surface ruptures. The Alpine Fault has not produced any large or great earthquakes in historic time, and is presently characterised by low rates of small-to-moderate earthquakes ($M < 6.5$; Fig. 3). Geologic data provide evidence for the occurrence of great earthquakes on the Alpine Fault with return times of hundreds of years (references in Appendix 1).

The Taupo Volcanic Zone (Fig. 1) is a zone of active crustal extension that has developed in response to the southward migration of back arc spreading from the Havre Trough (Fig. 1) into the continental margin of New Zealand in the last million years (Cole & Lewis 1981). Normal faults with slip rates of up to 3 mm/yr accommodate crustal extension within the Taupo Volcanic Zone. Several moderate-sized earthquakes have produced surface ruptures in the Taupo Volcanic Zone in historic time, the most recent being the $M_w$ 6.3, 1987 March 2 Edgecumbe earthquake, which produced a normal slip rupture along the Edgecumbe Fault (139 in Fig. 2B). High rates of small earthquakes also characterise the Taupo Volcanic Zone (Fig. 3).

Faults located away from the axial tectonic belt and Taupo Volcanic Zone tend to have slip rates that are about an order of magnitude less than the faults in those areas. Reverse
faults with slip rates of 0.1–1 mm/yr characterise the style of faulting in central Otago and south Canterbury (Fig. 1), and similar slip rates characterise the reverse faults in north Westland and Nelson (Fig. 1). The reverse faults have developed in response to the oblique compression across the plate boundary. The $M_w$ 7.6, 1929 Buller, and $M_s$ 7.4, 1968 Inangahua earthquakes occurred on reverse faults in the Nelson–north Westland area, and high rates of $M \geq 4$ earthquakes are observed near the epicentres of these earthquakes (Fig. 3). The western North Island is a broad zone of relatively stable crust, disrupted only by normal faults in the northeast and southwest (Fig. 1). Several $M \geq 6.5$ earthquakes have occurred within the western North Island area in historic time, all in the southwest. Finally, the Canterbury–Chathams platform is an area of stable continental crust that stretches well east of the map boundary in Fig. 1. Very few earthquakes have occurred on the Canterbury–Chathams platform in historic time (Fig. 3).

METHOD AND ANALYSIS

The steps taken to perform our PSHA are: (1) to use geologic data and the historical earthquake record to define the locations of earthquake sources and the likely magnitudes and frequencies of earthquakes that may be produced by each source; and (2) to estimate the ground motions that the sources will produce at a gridwork of sites that covers the entire country. The computation of ground motions in (2) is achieved with a seismic hazard code that is based on a code developed for a PSHA of southern California by Wesnosky (1986). The code has been updated for our PSHA of New Zealand, in that faults are treated as 3-D sources (i.e.,
“planar” or “dipping” sources), background seismicity is now included as input to the PSHA, and new ground motion attenuation relationships for New Zealand (Abrahamson pers. comm.) are incorporated into the code.

Earthquake sources

We show the 154 fault sources used in our PSHA in Fig. 2, and list them in Appendix 1. The fault data are obtained largely from published sources, but we have also been able to use some unpublished data and interim reports from the Institute of Geological & Nuclear Sciences (reference 13, Appendix 1). The fault traces shown on Fig. 2 are simplified from the original mapped fault traces, the purpose being to reduce the number of calculations in the PSHA. In Fig. 2 and Appendix 1, we sometimes divide a given fault into more than one source if: (1) geological data and/or the rupture length of a historic earthquake provide evidence for a fault having separate rupture segments (e.g., the Hope Fault is divided into three sources); or (2) a fault has wide (≥25 km) steps in the fault trace (e.g., the Kerepehi Fault is divided into sources 140 and 141 in Fig. 2B). Data bearing on the geometry (e.g., fault dip) and activity (slip rates, single-event displacements, and return times) of the fault sources are also listed in Appendix 1. Our method of estimating the likely maximum magnitude ($M_{\text{max}}$) and return time of earthquakes produced by each fault source in Fig. 2 varies according to the quantity and quality of available data for each fault. Where possible, the magnitudes of large historical earthquakes (usually well constrained from instrumental records or from MM intensity data) and lengths of the associated surface ruptures are used to define the $M_{\text{max}}$ and length of particular fault sources. If historical observations are unavailable for a fault source, then the next most preferable method of defining $M_{\text{max}}$ is to use published estimates of single-event displacements and fault area, and the equations for seismic moment and moment magnitude:

$$M_o = \mu AD$$

and,

$$\log M_o = 16.1 + 1.5 M_{\text{max}}$$

in which $M_o$ is the seismic moment of $M_{\text{max}}$, $\mu$ is the rigidity modulus, $A$ is the fault area, and $D$ is the single-event displacement (Eq. 1 is from Aki & Richards 1980, and Eq. 2 is from Hanks & Kanamori 1979). To calculate fault area we assume an average depth of 15 km to the base of the seismogenic layer for all of the faults. Lastly, if single-event displacement data are unavailable, then an empirical regression of Wells & Coppersmith (1994) is used to estimate $M_{\text{max}}$ from fault rupture area. The return time ($T$) assigned to $M_{\text{max}}$ is either the published estimate from geological investigations, the return time calculated with the equation:

$$T = D/S$$

if a published return time estimate is unavailable ($D$ is single-event displacement and $S$ is the fault slip rate), or the return time calculated with the equation of Wesnousky (1986):

$$T = M_o / M_{\text{rate}}$$

if single-event displacement data are unavailable ($M_{\text{rate}}$ is the rate of seismic moment release on the fault, equal to $\mu / AS$, in which $\mu$ is the rigidity modulus, $3 \times 10^{11}$ dyn cm$^{-2}$, $A$ is fault area, and $S$ is fault slip rate). Where possible, we use the preferred values of $D$, $S$, and $T$ in Appendix 1 in Eq. 1–4, and otherwise use values that are the means of the minimum and maximum values given in the appendix. We also use the mean of minimum and maximum values of $M_{\text{max}}$ given in Appendix 1 in the equations.

We define the geometry of five earthquake sources along the Hikurangi subduction interface from the results of studies that have used the spatial and depth distributions of seismicity to map the 3-D geometry of the coupled subduction interface, and the likely positions of “tears” or segment boundaries in the subduction zone (references listed in Appendix 1 for sources 85–89). We set the top and base of the sources at 10 and 40 km, respectively, and assume that each source dips 15° NW (geometry based on cross-section shown in Darby & Beanland 1992). The surface projection of each source is shown by the large polygons numbered 85–89 in Fig. 2. We then use an empirical regression of Wells & Coppersmith (1994) to estimate the $M_{\text{max}}$ for each source from the fault area. Geodetic modelling indicates that there is a transition from a coupled Hikurangi subduction zone in the southern North Island to an uncoupled subduction zone in the north (Beavan & Haines 1997). We therefore assume that our northernmost subduction source (85 in Fig. 2) is uncoupled and has zero earthquake potential. Our basis for defining the recurrence rate of $M_{\text{max}}$ events on the remaining Hikurangi subduction zone sources (sources 86–89) is to assume that the recurrence rate is proportional to the component of relative plate motion not taken up by the faults in the hanging wall of the subduction zone. To explain, we sum the slip rates of all faults in the hanging wall of a particular subduction source, and subtract this total slip rate from the relative plate motion rate of 40 mm/yr (DeMets et al. 1990) to get the slip.
rate for the subduction source. The slip rate is then used to calculate the return time of $M_{\text{max}}$ earthquakes with Eq. 2 and 4. Since the faults of the axial tectonic belt accommodate all of the relative plate motion in the northern South Island, zero earthquake potential is assigned to sources 88 and 89 (Fig. 2B). However, only about one-half of the relative plate motion rate is taken up by the faults of the tectonic belt in the southern North Island, and as a consequence we calculate return times of 464 and 354 yr for $M_{\text{max}}$ events on sources 86 and 87, respectively (Fig. 2B; Appendix 1). Our return time estimates for great earthquakes on the Hikurangi subduction zone are shorter than the return time estimates of Berryman (1994) and Berryman et al. (1997), since our estimates are based on the assumption that all of the subduction zone slip occurs seismically (supported by the recent work of Beavan & Haines 1997), whereas the Berryman (1994) and Berryman et al. (1997) estimates are based on the assumption that half of the subduction zone slip rate occurs aseismically.

The geometry of the coupled portion of the Fiordland subduction zone is defined from the 3-D geometry of the aftershock zone of the $M_{\text{W}}$ 7.0, 1993 August 10 Fiordland earthquake (Van Dissen et al. 1994). We divide the subduction zone into four sources in total (sources 53–56 on Fig. 2). Source 53 is defined to extend from the southern end of the Alpine Fault to the vertically dipping aftershock zone of the $M_{\text{f}}$ 6.1, 1988 June 3 earthquake (interpreted to represent a tear in the subduction zone; Reynolds et al. 1991); source 54 is defined to extend from the 1988 aftershock zone to the northern end of the 1993 aftershock zone; source 55 is simply defined from the extent of the 1993 aftershock zone ($M_{\text{max}}$ set at 7.0); and source 56 is defined to extend from the 1994 aftershock zone to the southwestern tip of the South Island, where there is a northwest-trending gap in the shallow crustal seismicity (Fig. 3). Estimation of the $M_{\text{max}}$ for each source (other than source 55) is achieved by the same method used for the Hikurangi subduction zone sources, and the return time of $M_{\text{max}}$ events on each source is estimated by using the relative plate motion rate (30 mm/yr; DeMets et al. 1990) with Eq. 2 and 4.

In addition to defining the locations, magnitudes, and frequencies of large to great earthquakes on the crustal faults and subduction zones, we also allow for the occurrence of moderate-to-large background earthquakes ($M_{5-7.5}$), both on and away from the major faults. Our reasons for considering background earthquakes in our PSHA are twofold. First, earthquakes of $M < 6.5$ are not generally large enough to rupture to the surface (e.g., Wesnousky 1986), so it is possible that some future earthquakes of these magnitudes will occur on fault sources that have no surface expression and have escaped mapping by geologists. Second, since many of the historic $M \geq 6.5$ earthquakes listed in Table 1 have not been able to be assigned to specific faults, the possibility exists that some future large earthquakes may also occur on faults not listed in Appendix 1. We apply the methodology of Frankel (1995) to characterise the PSH from background earthquakes. Frankel (1995) made the observation that moderate-to-large earthquakes in the western United States generally occur in areas where high rates of small earthquakes are observed, and this observation became the basis for estimating the PSH from background earthquakes for the US Geological Survey PSH model. We therefore use the spatial distribution of seismicity recorded by GNS and DSIR from 1964 to 1996 (Fig. 3) to predict the likely locations and recurrence rates of background earthquakes ($M_{5-7.5}$) at a gridwork of point sources across New Zealand. We restrict our analysis of recorded seismicity to the time period 1964–96 because the GNS seismic network was significantly upgraded and computerised in 1964, which greatly reduced errors in locating earthquakes. Our upper magnitude of $M = 7.5$ is the approximate magnitude of the largest historical earthquakes that have not been able to be assigned to specific faults (e.g., the $M_{\text{W}}$ 7.6, 1934 Pahiatua earthquake; Berryman 1994) due to the sparseness of historical records for these events and rapid degradation of the associated surface ruptures in the humid New Zealand climate. A Gutenberg-Richter distribution, $\log N = A – BM$ ($N$ = number of events > magnitude $M$), and $A$ and $B$ are empirical constants; Ishimoto & Iida 1939; Gutenberg & Richter 1944), is then used to predict the recurrence rates of background earthquakes at each point source. The basis for using a Gutenberg-Richter distribution is that the seismicity of large areas (in this case New Zealand), and the earthquakes that occur along fault zones that are less in size than the $M_{\text{max}}$ of the fault, will generally follow a Gutenberg-Richter distribution (e.g., Stirling et al. 1996). We first remove events less than the $M_4$ catalog completeness level (Stirling et al. 1996) from the catalog, remove events recorded at depths greater than the 15 km assumed for the seismogenic depth (15 km is assumed for the whole country for computational simplicity), and decluster the catalog by the method of Reasenberg (1985). A gridwork with cell dimensions of 0.1° in latitude and longitude is then placed over the map area, and the earthquakes found inside each grid cell are counted to give “$N$ values” for each grid cell. The gridwork of $N$ values is then spatially smoothed with a Gaussian smoothing function, following the methodology of Frankel (1995). For each grid cell, this involves multiplying the $N$ value for the grid cell and all of the neighbouring $N$ values (i.e., the $N$ values that are within a specified distance from the grid cell) by the Gaussian function, summing all of the products, and then dividing by the sum of all of the Gaussian functions. The equation is:

$$N(\text{smoothed}) = \Sigma(N(\text{each site})e^{-d^2/c^2})/\Sigma(e^{-d^2/c^2})$$

in which $c$ is the correlation distance (50 km), and $d$ is the distance from the centre of the grid cell to the centre of each neighbouring grid cell (neighbouring grid cells greater than $3 \times$ the correlation distance from the grid cell are not used in Eq. 5). The Gaussian smoothing preserves the total number of earthquakes in the catalog after every $N$ value in the gridwork has been smoothed with Eq. 5. The 50 km correlation distance is chosen since it produces a spatial distribution of $N$ values that correlates well with the general seismicity patterns across the country (cf. Fig. 3 and 4). Shorter correlation distances tend to produce grainy maps of $N$ values, and longer correlation distances smooth the maps too much by forcing $N$ values into areas of low seismicity rates. The recurrence rates of $M_{5-7.5}$ events at each point source are then calculated from the smoothed $N$ values by dividing each $N$ value by 33 yr (the number of years of catalog recording) to get $N/yr$ for $M_{5-4}$ (cumulative rates), solving for the $A$ value in the Gutenberg-Richter relationship (this time equal to $\log N/yr$ ($M_{5-4}$) = $A – BM$, with $M = 4$ and $b = 1.1$, this being the $b$-value calculated for New Zealand by Stirling et al. 1996, and applied to every point source in the country for computational simplicity), and then using the cumulative rates to calculate the number of events per year.
for each 0.1 increment of magnitude (incremental rates, \( n/\text{yr} \)) for \( M = 5.0 - 7.5 \). For example, if \( n/\text{yr} (M = 5.5) \) is equal to 0.01, and \( n/\text{yr} (M = 5.4) \) is equal to 0.03, then \( n/\text{yr} (M = 5.4) \) is equal to 0.02.

A common test of a PSH model is to compare the rate of earthquakes predicted from the source model to the historical record of earthquakes (e.g., WGCEP 1995). In Fig. 5A we show the total cumulative number of events per year greater than or equal to magnitude \( M \) predicted from our source model by solid circles, and the cumulative number of events per year observed historically by open squares. The historical rates are calculated by combining a catalog of \( M \geq 6.5 \) earthquakes that have occurred within the map area since 1843 (Table 1) and events of \( M = 6.4 \) that have been recorded by GNS and DSIR since 1964. The cumulative rates predicted with our source model at each magnitude increment (solid circles) are generally similar to the rates observed historically (squares), except at about \( M = 7.5 \) where the observed rates are more than a factor of three higher than the predicted rates. Our source model therefore predicts earthquake rates that are similar to the earthquake rates that have occurred in New Zealand in the last 154 yr, except at about \( M = 7.5 \). A possible reason for the discrepancy at about \( M = 7.5 \) is that the historical record has “captured” earthquakes that have return times much longer than the 154 yr historical record. An example is the 1968 Inangahua earthquake, which is interpreted from geologic data to have a return time of 4400 yr (Appendix 1).

We can test the statistical significance of the discrepancies observed in Fig. 5A between the predicted and observed rates of \( M = 7.5 \) earthquakes. We use the methodology of Stirling & Wesnousky (1997) to examine the likelihood that a 154 yr “sample” from the predicted rates would yield the rate of seismicity observed in the last 154 yr. We convert the predicted cumulative number of events per year (squares) to the equivalent incremental rates (\( n/\text{yr} = M \)), and then multiply the rate by the sampling time (154 yr) to give the expected number of events (\( n \)) of magnitude \( M \) for that time period. We then assume that each \( n \) is described by a Poisson distribution, and choose a final value at random from a Poisson distribution with a mean equal to \( n \). The final value of \( n \) is then converted back to the cumulative number of events per year. In Fig. 5B we show the cumulative number of events per year produced from 500 repetitions of the above procedure (dots) superimposed on top of the rates observed historically (squares). The range of simulated rates clearly overlaps with the rates observed historically throughout the range of magnitudes. We therefore observe that the historical seismicity rates fall within the rates calculated from our source model for the same period of time and for the entire range of magnitudes.

**Computation of hazard**

We use the locations, sizes, and recurrence rates of earthquakes defined in our source model to estimate the PSH for a gridwork of New Zealand sites with a grid spacing of 0.1° in latitude and longitude. Our measures of PSH are the ground motion levels (peak ground acceleration and spectral acceleration at 0.5 s period) expected to be exceeded with a probability of 10% in the next 50 yr on “rock”. We use the standard methodology of PSHA (Cornell 1968) to construct PSH maps. This is to: (1) calculate the annual frequencies of exceedance for a suite of ground motion levels at each site within the gridwork of sites (i.e., develop a gridwork of “hazard curves”, or a “hazard matrix”) as a function of the magnitude, recurrence rate, and source-to-site distance of earthquakes predicted from the source model; and (2) construct maps that show the maximum ground motion level that is expected to be exceeded with a certain probability (in this case 10%) in a certain time period (in this case 50 yr) at each site. For each site, step (1) is repeated for all sources in the source model, and the results summed to give the annual frequencies of exceedance at the site due to all sources. In step (2) we estimate the maximum ground motion level at each site according to a Poisson model. For a Poisson model, probability \( P \) (in this case \( P = 0.1 \)) is equal to \( 1 - e^{-r} \), in which \( t \) is the time period of interest (50 yr), and \( r \) is the rate. The ground motion level expected to be exceeded at 10% probability in 50 yr is determined by solving for \( r \) with \( t = 50 \) yr and \( P = 0.1 \), and then determining which ground motion level in the hazard curve occurs at a rate that is equal to \( r \). Ground motion levels with rates of 0.0021 or more (return times of up to 475 yr) contribute to this measure of PSH.

In modern PSHA it is common practice in step (1) above to take into account the uncertainty in estimates of ground motion from attenuation relationships in the calculation of PSH. A ground motion calculated with an attenuation
relationship represents the median of a log-normal distribution that is characterised by a standard deviation (here referred to as \( \sigma_T \)). Many early PSH maps were constructed from median estimates of ground motion (e.g., Algirmisssen et al. 1982; Wesnousky 1986), whereas modern PSH maps are constructed by using the median estimates of ground motion with \( \sigma_T \) to calculate the probability of exceedance for a suite of ground motion levels (e.g., Frankel et al. 1996). The former approach produces ground motion estimates that are considerably lower than the estimates produced with the latter approach (cf. Wesnousky 1986 and Frankel et al. 1996). The reason for the discrepancy is that values of \( \sigma_T \) tend to be large (usually about 0.5 in natural log units of ground motion), which is a consequence of the attenuation relationships being constructed from strong motion datasets recorded in a great diversity of tectonic and geomorphic settings, and then being applied to individual sites. However, recent studies of earthquakes such as the Northridge earthquake have shown that the values of \( \sigma_T \) for attenuation relationships appear to correctly predict the large differences in ground motions that are observed at sites equidistant to a single earthquake, the differences attributed to basin effects and rupture directivity (Frankel pers. comm.).

Anderson & Brune (in press) have taken an alternative approach to the treatment of uncertainty in estimates of ground motion from attenuation relationships. They have suggested that the “ergodic” assumption implicit in modern PSHA, that the total range of ground motions represented by \( \sigma_T \) will eventually be observed at a site (i.e., all uncertainty is “aleatory” or random uncertainty), may be wrong, since some of the uncertainty may be non-ergodic (i.e., “epistemic” or model uncertainty). Anderson & Brune (in press) have developed a methodology for the separate treatment of aleatory and epistemic uncertainties in a PSHA. They assume that the aleatory and epistemic uncertainties (\( \sigma_A \) and \( \sigma_E \)) are related to the total standard deviation for an attenuation relationship \( \sigma_T \) by:

\[
\sigma_T^2 = \sigma_A^2 + \sigma_E^2
\]

Their methodology involves using an attenuation relationship to estimate the median ground motion at a site as a function of magnitude and source-to-site distance, and then defining a series of ground motion values that encompass the range of possibilities defined by \( \sigma_E \). Each of these ground motion values is then used independently with \( \sigma_A \) to calculate the probability of exceedance for a suite of ground motion levels (i.e., \( \sigma_A \) is used where \( \sigma_T \) is normally used in modern PSHA), and the weighted average of the probabilities of exceedance for each ground motion level is then used in the PSHA. In our study we develop PSH maps based on the standard approach to modern PSHA (e.g., Frankel et al. 1996), maps based on the early approach to PSHA (e.g., Algirmisssen et al. 1982; Wesnousky 1986), and experimental PSH maps based on the approach of Anderson & Brune (in press). We use an unpublished attenuation relationship developed from New Zealand strong motion data by Abrahamson (pers. comm.) that has a form similar to that of Abrahamson & Silva (1997). Abrahamson’s attenuation relationship allows for the separate treatment of earthquake sources according to slip type and tectonic setting, and we use the attenuation relationship appropriately for each of the sources shown in Appendix 1. However, for computational simplicity, we assume that all background earthquake sources produce oblique slip earthquakes. We do not show

Fig. 5 A. The cumulative number of events per year versus magnitude observed historically in New Zealand (squares) and predicted from our source model (solid circles). We also show the contribution to the predicted rates from the fault and subduction sources (dash-dot line) and background seismicity sources (dotted line). B. The cumulative number of events per year versus magnitude observed historically in New Zealand (squares), and predicted from 500 random catalogs drawn from the total predicted rates (solid circles in A). Many of the dots form vertical lines at magnitudes c. <6.5, and many simulations produce 0 earthquakes at magnitudes c. >6.5.
We show maps of peak ground acceleration expected at 10% probability in 50 yr on rock (Site Category A or “rock and stiff soil” of Standards New Zealand 1992) in Fig. 7. Model 1 (Fig. 7A) is based on median estimates of peak ground acceleration from Abrahamson's attenuation relationship (i.e., the approach used by Algernissen et al. 1982 and Wesnousky 1986). Model 2 (Fig. 7B) is based on the standard approach of modern PSHA, which is to fully incorporate Abrahamson’s $\sigma_T$ (0.5858) into the aleatory component of uncertainty in the PSHA calculations ($\sigma_A = \sigma_T = 0.5858$ and $\sigma_E = 0$). Model 3 (Fig. 7C) is based on the methodology of Anderson & Brune (in press) described above, in which $\sigma_T$ (0.5858) is partitioned into $\sigma_A$ and $\sigma_E$ according to the partitioning scheme derived from Eq. 6 and the precarious rock data in Fig. 6. Maps constructed from models 1 and 2 (Fig. 7A, B) show differences in peak ground acceleration of up to 0.2g, and the map constructed from model 3 (Fig. 7C) shows peak accelerations intermediate between models 1 and 2. All three maps show peak ground accelerations of >0.2g, and locally >0.6g in a belt that extends from the Fiordland subduction zone along the axial tectonic belt to the Hikurangi subduction zone, and also along the Taupo Volcanic Zone. The highest PSHA therefore tends to occur along the faults that accommodate most of the relative plate motion. Of the major cities in the country (Fig. 1), Wellington is the only one that lies within this belt of high PSHA. From model 3 (Fig. 7C), peak ground accelerations of >0.4g are predicted to occur in Wellington at 10% probability in 50 yr, and there are differences of ±0.1g between model 3 and models 1 and 2 (Fig. 7D). In contrast, Auckland, Christchurch, and Dunedin are expected to experience shaking of only <0.1g in this time frame. Hazard curves drawn from model 3 for Auckland, Wellington, Christchurch, and Dunedin (Fig. 8) show that the annual frequencies of exceedance for peak ground accelerations of 0.1–1g at Wellington are more than 10× greater than for the other cities.

There are many fault sources that produce high hazard on the PSHA maps in Fig. 7 (peak ground accelerations of ≥2.4g) that have not produced any large earthquakes in historic time. This applies to the Alpine Fault and Hikurangi subduction zone sources in particular, where the largest earthquakes in our source model are predicted to occur (Appendix 1). Return times for large-to-great earthquakes on the Alpine Fault and Hikurangi subduction zone (Appendix 1) are longer than the historical record (154 yr), so a likely reason for the absence of earthquakes is that the historic record is so short to have recorded earthquakes on these sources. Therefore, it should not be surprising if the Alpine Fault and Hikurangi subduction sources produce large-to-great earthquakes in the near future.

The maps in Fig. 7 show considerable differences in estimates of PSHA to the maps produced by Smith & Berryman (1986) from the distribution of historical seismicity. In Fig. 9 we show Smith & Berryman’s map of the MM intensity expected at 10% probability in 50 yr (Fig. 10 in Smith & Berryman 1986), and also show the equivalent levels of peak ground acceleration represented by each MM intensity level, using the conversion table between MM intensity and peak ground acceleration given by Bolt (1993). Comparison of the peak ground accelerations produced from model 3 (Fig. 7C) to the peak accelerations derived from Smith & Berryman’s MM intensity map (Fig. 9) reveal differences of up to 0.5g.
Fig. 7  PSH maps of peak ground acceleration expected at 10% probability in 50 yr on rock. The maps are constructed from: (A) model 1, (B) model 2, and (C) model 3. See the text for an explanation of the three models. The insets in (D) show the differences in peak ground acceleration between the three models in the Wellington region, which range up to about 0.2g. The map area is the same as in Fig. 1 and 3.
between the maps. Our PSH models generally produce higher estimates of PSH than Smith & Berryman’s model along the major faults of the axial tectonic belt and Taupo Volcanic Zone. There are two reasons for the differences between our maps and those of Smith & Berryman (1986). First, we incorporate geologically derived earthquake recurrence rates into our PSH model, whereas Smith & Berryman did not. Many of the faults are characterised by low rates of historical seismicity but show geologic evidence for the occurrence of large earthquakes with return times of hundreds of years (e.g., the Alpine Fault). Second, Smith & Berryman smoothed their earthquake recurrence rates over large source zones, whereas we do not. Smoothing of earthquake rates over large source zones greatly reduces estimates of PSH in areas of high seismicity rates.

We also construct a map from model 3 of the 0.5 s spectral accelerations expected at 10% probability in 50 yr on rock (Fig. 10). Maps of response spectral acceleration at periods of 0.1 s or greater can be more relevant to the siting of tall buildings than maps of peak ground acceleration, since tall structures are generally more sensitive to these longer periods than to peak ground acceleration. The map of 0.5 s response spectral acceleration shows similar patterns of hazard to the maps of peak ground acceleration (Fig. 7C).

The main difference between the peak ground acceleration and spectral acceleration maps (Fig. 7C, 10) is that areas of high peak accelerations that are attributed to background seismicity sources (e.g., offshore Taupo Volcanic Zone) do not show as high spectral accelerations. The spectral accelerations tend to be about 0.1–0.2g lower than the peak accelerations in these areas. This is because the vast majority of background earthquakes are moderate-sized earthquakes, and factored into the Abrahamson attenuation relationships is the common observation in strong motion seismology that moderate earthquakes tend not to produce much long period motion. Again, Wellington is the only major New Zealand city located within a zone of high PSH in Fig. 10. In Wellington, 0.5 s spectral accelerations of &gt;0.4g are predicted to occur at 10% probability in 50 yr. In contrast, tall buildings in Christchurch are expected to experience spectral accelerations of 0.1–0.2g in this time frame, and shaking of &lt;0.1g is expected to occur in Auckland and Dunedin.

Maps of the peak ground accelerations and 0.5 s spectral accelerations expected at 2% probability in 50 yr are also constructed from model 3, and these are shown in Fig. 11A, B, respectively. These maps show much higher PSH than the maps in Fig. 7 and 10 because ground motions with return times of up to 2475 yr contribute to this measure of PSH. Peak ground accelerations and 0.5 s spectral accelerations of &gt;0.1g are predicted for all of the major cities. The peak and spectral accelerations predicted for Wellington are &gt;0.7g. Again, the background earthquake sources make relatively greater contribution to the peak ground accelerations than to the 0.5 s spectral accelerations in Fig. 11A, B. However, the spectral accelerations are slightly higher than the peak accelerations in the distal areas of the map (width of the lightest shaded areas in Fig. 11A, B), especially offshore to the east of central New Zealand. Since these areas are at distances of 100 km or more from the sources in our source model, the PSH of these areas is largely controlled by the great earthquake sources (Alpine and Wairarapa Faults, and Hikurangi subduction zone). The higher PSH in the distal areas of the 0.5 s spectral acceleration map (Fig. 11B) than in the peak acceleration map (Fig. 11A) is therefore a consequence of the 0.5 s spectral accelerations predicted for great earthquakes being larger than the peak accelerations predicted for these same earthquakes at large source-to-site distances.

We also compare the estimates of peak ground acceleration predicted with model 3 at 2% probability in 50 yr (Fig. 11A).
to the peak accelerations predicted with models 1 and 2 (Fig. 11C, D), to examine the effect that the choice of model has on ground motions at low probabilities of exceedance. There are differences of up to 1.0g between the maps produced from models 1 and 2 along the faults of the axial tectonic belt (especially along the Hope Fault), and the peak accelerations predicted from model 3 (Fig. 11A) are approximately intermediate between the maps produced from models 1 and 2. Differences between the maps are generally <0.2g in the areas away from the axial tectonic belt. The three PSH models therefore produce large differences in estimates of PSH at low probabilities of exceedance, especially along the faults of the axial tectonic belt.

A final but important exercise is to understand the dominant magnitudes and source-to-site distances that control the PSH of the major New Zealand cities. The process of deaggregating hazard into the causative magnitudes and distances may lead to the definition of "design earthquakes" for a city, which is information that is often useful for engineering and planning purposes (McGuire 1995). We use a method similar to that of McGuire (1995) to deaggregate the PSH of model 3 for Auckland, Wellington, Christchurch, and Dunedin. In Fig. 12 we show for each city the annual frequency of exceedance for 0.1g (peak ground acceleration in Fig. 12A–D, and 0.5 s response spectral acceleration in Fig. 12E–H) plotted as a function of magnitude and source-to-site distance. In Auckland, peak ground accelerations (Fig. 12A) are predicted to be produced by moderate-sized background earthquakes at close distances (the raised part of the graph at M5–6.5, and at distances <20 km). However, the 0.5 s spectral accelerations predicted for Auckland (Fig. 12E) are also due to large earthquakes on the Kerepahi Fault, which appears on Fig. 12E as a small peak at about M7 and 60–70 km. Both peak accelerations and 0.5 s spectral accelerations in Wellington (Fig. 12B, F) are produced by a large range of earthquake sizes that are all at close distances. The crustal faults of the axial tectonic belt, Hikurangi subduction zone sources, and background seismicity sources all contribute to the PSH of Wellington. There are also sources in the northern South Island and Hawke's Bay (the Hawke's Bay segment of the Hikurangi subduction zone; source 86 on Fig. 2) that contribute to the 0.5 s spectral accelerations at Wellington. They produce the small peaks at approximately M7.5/70 km and M8/200 km in Fig. 12F. Background seismicity is a considerably less important source of 0.5 s spectral accelerations than it is for peak accelerations in Wellington. The peaks at M5–6.5 at distances of <20 km are attributed to background seismicity sources, and these are lower on Fig. 12F than 12B. The PSH of Christchurch (Fig. 12G) is controlled by large-to-great earthquakes on faults in the axial tectonic belt, and to a lesser extent by background earthquakes at close distances. The Porters Pass Fault is the structure that contributes most to the peak ground accelerations at Christchurch, being responsible for the largest peak in Fig. 12C, at M7.3/50 km. The Hope and Alpine Faults also contribute to the peak accelerations at Christchurch, producing small peaks at about M7/100–125 km and M8/125 km in Fig. 12C. However, the Alpine Fault contributes as much to the 0.5 s spectral accelerations at Christchurch as the Porters Pass Fault, being responsible for the peak at about M8/125 km in Fig. 12G. Finally, the PSH of Dunedin is wholly dominated by the predicted occurrence of M6.9 earthquakes on the Akatore Fault (the peak at about M7/50 km on Fig. 12D, H).

DISCUSSION AND CONCLUSIONS

Our PSH maps for New Zealand are developed by combining geologic data describing the long-term recurrence behaviour of the major active faults with observations of the size and location of large historical earthquakes, and seismicity data recorded by GNS and DSIR. As such, they represent the first multidisciplinary PSH maps to be developed for the entire country. The maps show the highest levels of peak ground acceleration and 0.5 s spectral accelerations to occur along the axial tectonic belt, the subduction zones, and the Taupo Volcanic Zone, which together accommodate essentially all of the relative plate motion between the Australian and Pacific plates. While the geologic input to our PSHA is generally restricted to published sources, we have managed to characterise recurrence behaviours for all of the major mapped active faults in New Zealand. Our characterisation of PSH from background earthquakes in essence compensates for any incompleteness in the fault dataset by allowing for the occurrence of moderate-to-large earthquakes in areas between the faults listed in Appendix 1. The inability to assign several large historic earthquakes to specific faults warrants such treatment of background earthquakes in the PSHA.
Fig. 11 PSH map of (A) peak ground acceleration, and (B) 0.5 s spectral acceleration expected at 2% probability in 50 yr on rock. The maps are constructed from model 3. Maps of peak ground acceleration expected at 2% probability in 50 yr on rock are also constructed from model 1 and model 2, and are shown in (C) and (D), respectively. The map area is the same as in Fig. 1 and 3.
Fig. 12 The annual frequency of exceedance for 0.1g (peak ground acceleration in A, B, C and D, and 0.5 s spectral acceleration in E, F, G and H) plotted as a function of magnitude and source-to-site distance for the four main cities of New Zealand. The graphs are constructed from model 3, with a magnitude increment of 0.2 and a distance increment of 20 km. See Fig. 1 for the location of the cities.
We predict that strong earthquake shaking will occur in Wellington at least 10 times more frequently than in Auckland, Christchurch, and Dunedin. Deaggregation of our PSH model 3 at Wellington shows that the PSH is due to $M \approx 7$–8.5 earthquakes at close distances, consistent with the city being close to six faults, two of which are known or assumed to produce great earthquakes (the Wairarapa Fault and Hikurangi subduction zone, respectively). In contrast, Auckland may be shaken only by moderate-sized earthquakes at close distances, but the possibility exists that 0.5 s accelerations of 0.1 g or more will occur when the Kerepehi Fault (source 140) next ruptures. Peak ground accelerations and 0.5 s accelerations at Christchurch and Dunedin will be produced by events of about $M 7$ occurring at distances of 50–100 km. Great earthquakes on the Alpine Fault will also contribute significantly to the 0.5 s spectral accelerations in Christchurch, since the Alpine Fault is only 125 km away from Christchurch.

In the construction of our PSH maps we assume that all of the fault and subduction sources shown in Fig. 2 accommodate relative plate motion by seismic slip, and none (with the exception of source 85) by aseismic slip. This is a valid assumption for the crustal faults, since there are no geologic data that provide evidence for aseismic slip on any crustal fault in New Zealand. Aseismic slip is readily observed along the surface traces of major San Andreas system faults in central California (US Geological Survey 1990), so we would expect it to have been observed along the fastest slipping faults in the axial tectonic belt if it were occurring. However, our assumption that all of the plate boundary slip occurring at the subduction interfaces is seismic is more difficult to confirm, since these features do not break the surface. However, geodetic observations (e.g., Beavan & Haines 1997) provide evidence for the Hikurangi subduction zone being coupled, and large earthquakes have occurred on the Fiordland subduction zone in recent years (e.g., Van Dissen et al. 1994), so our conservative approach to defining the earthquake potential of subduction sources is not inconsistent with data. Further research on the relationship between geodetic strain rates and the location and recurrence rate of large-to-great earthquakes, heat flow modelling to constrain the extent of the seismogenic portion of the subduction interfaces (e.g., Hyndman & Wang 1993), and further analysis of the spatial distribution of subduction zone seismicity are expected to provide additional constraints on the earthquake potential of the Hikurangi and Fiordland subduction zones.

Our fault database should be updated with new data as soon as they become available, and future PSHAs would also benefit from the development of new equations relating earthquake magnitude to fault rupture length and area for New Zealand earthquakes, because it appears that the equations of Wells & Coppersmith (1994) provide underestimates of many of the magnitudes of New Zealand earthquakes that have been observed historically or derived from single-event displacement data. The best examples here are the magnitudes of historical earthquakes along the Wairarapa and Hope Faults (Appendix 1), which are about 0.5 and 0.2 magnitude units larger, respectively, than the magnitudes estimated from the rupture lengths of those earthquakes with equations of Wells & Coppersmith (1994).

Third, since historical earthquake catalogs are the only datasets that are currently available for testing the predictions of PSH models (Fig. 5), increased effort should be focused towards developing field criteria to test the PSH models. It has been suggested that precariously balanced rocks may provide upper-bound estimates of the peak ground accelerations that have occurred at specific sites in southern California for thousands of years (Brune 1996). Thus, documentation of the distribution of precariously balanced rocks and determination of the upper bounds of ground motions necessary for precariously balanced rocks to topple may place independent constraints on the ground motions predicted by our PSHA. A potential problem with this approach is that many New Zealand lithologies, such as Torlesse greywacke, may be unsuitable for the formation of precariously balanced rocks. However, the importance of verifying the predictions of PSH models, particularly in populated areas, makes it worthwhile to undertake a search for precariously balanced rocks in New Zealand, and to use them to provide constraints on the predictions of our PSHA.

Finally, we stipulate that our PSH maps not be used for engineering or planning studies, since we regard them as experimental. They incorporate an unpublished attenuation relationship, as well as a method for the treatment of uncertainty in attenuation relationships that differs from the standard approach of modern PSHA. Though we have shown that the standard treatment of attenuation uncertainty in modern PSHA overpredicts the ground motions at sites of precariously balanced rocks in southern California (Fig. 6), it correctly predicts the large differences in ground motions that have been observed at sites equidistant to a single earthquake (Frankel pers. comm.). We therefore recommend that the issue of attenuation uncertainty be formally debated before any changes to the standard methodology of modern PSHA are made.

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