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Landslides Triggered by the Mw 7.8 14 November 2016 Kaikoura Earthquake, New Zealand

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Abstract

The M_W 7.8 14 November 2016 Kaikoura earthquake generated more than 10000 landslides over a total area of about 10000 km^2, with the majority concentrated in a smaller area of about 3600 km^2. The largest landslide triggered by the earthquake had an approximate volume of 20 (±2) M m^3, with a runout distance of about 2.7 km, forming a dam on the Hapuku River. In this paper, we present version 1.0 of the landslide inventory we have created for this event. We use the inventory presented in this paper to identify and discuss some of the controls on the spatial distribution of landslides triggered by the Kaikoura earthquake. Our main findings are (1) the number of medium to large landslides (source area ≥10000 m^2) triggered by the Kaikoura earthquake is smaller than for similar sized landslides triggered by similar magnitude earthquakes in New Zealand; (2) seven of the largest eight landslides (from 5 to 20 x 10^6 m^3) occurred on faults that ruptured to the surface during the earthquake; (3) the average landslide density within 200 m of a mapped surface fault rupture is three times that at a distance of 2500 m or more from a mapped surface fault rupture ; (4) the “distance to fault” predictor variable, when used as a proxy for ground-motion intensity, and when combined with slope angle, geology and elevation variables, has more power in predicting landslide probability than the modelled peak ground acceleration or peak ground velocity; and (5) for the same slope angles, the coastal slopes have landslide point densities that are an order of magnitude greater than those in similar materials on the inland slopes, but their source areas are significantly smaller.

Introduction

The M_W 7.8 14 November 2016 Kaikoura earthquake in New Zealand occurred at 12:03 am local time (Kaiser et al. 2017). The epicentre was located about 4 km from the rural town of Waiau (population 250) in North Canterbury (Figure 1), with rupture initiation at a shallow depth of

14.1 km (Nicol et al., this issue). Large, shallow earthquakes in mountain chains typically trigger substantial numbers of landslides (Hovius et al., 1997; Parker, 2013; Hancox et al., 2014; 2016; Xu et al., 2016). The Kaikoura earthquake (Dellow et al., 2017) triggered more than 10000 landslides over an area of about 10000 km^2, with the majority being focused in an area of about 3600 km^2 (Figure 1). Fortunately, the area affected by landslides is comparatively remote and sparsely populated such that only a few dwellings were impacted by landslides, and there were no recorded landslide-related fatalities (Stevenson, 2017). However, the landslides dammed rivers, blocked roads and railways, and disrupted agricultural land throughout this region. Landslides along the coast caused substantial damage to both State Highway (SH) 1 and the northern section of the South Island Main Trunk Railway, blocking both in multiple locations (Davies, 2017). At the time of writing, the section of SH1 north of Kaikoura is due to reopen on 15 December 2017, over a year after the earthquake.

The long-term stability of damaged but as yet unfailed slopes is a cause for concern in light of the risk of future strong earthquakes and significant precipitation events. This has been exemplified by debris flows and floods that occurred during rain associated with cyclones Debbie (23 March to 7 April 2017) and Cook (14 April 2017), which caused several of the dams to breach, releasing debris flows and floods that travelled several kilometres downstream. Debris flows were also triggered on the steep coastal cliffs north and south of Kaikoura, leading to the intermittent closures of the reopened portion of SH1 south of Kaikoura.

Both the number of landslides and the area affected are much less than expected based on worldwide observations for an earthquake of this magnitude (Keefer, 2002; Malamud et al. 2004). To investigate the reason for this, we analyse an inventory we are creating of landslides triggered by this earthquake; our analysis relates the spatial distribution and size characteristics of the triggered landslides to geology, topography, strong shaking, and other geologic factors. The objective of this paper is to describe these characteristics of the triggered landslides and quantify their relationship to the various causative factors.

A broad-based investigation of the triggered landslides began immediately following the earthquake. Dellow et al. (2017) provide a preliminary description of the landslides triggered by this earthquake and the immediate response to document them and evaluate related hazards. Jibson et al. (2017) give an overview of landslide types and distribution accompanied by illustrations of the triggered landslides. In this paper, we present version 1.0 of the landslide inventory we have created for this event, which builds on the earlier preliminary inventories presented by Rathje et al. (2016) and Dellow et al. (2017). Refer to the Data and Resources section of this paper for instructions about how to access this dataset. We present these findings as a preliminary account of the potential controls we have observed on the landslide distribution triggered by this event. It is version 1.0, because mapping is ongoing in those areas where the landslide distribution was initially mapped from satellite images. The high-resolution orthorectified aerial photographs that have been used to map much of the distribution were not available in these areas at the time of publication.

Detailed Landslide Inventory from Mapping

Previous studies of worldwide earthquakes have related earthquake magnitude to the number of landslides. For a M_W 7.8 earthquake, the relationship of Malamud et al. (2004) predicts about 25000 landslides; Keefer's (2002) relation predicts about 60000 landslides. Both relations are based solely on magnitude and do not consider other factors such as earthquake depth, distance to fault, topography, rock type, climate, and vegetation that contribute to landslide occurrence.

These estimates based on worldwide earthquakes are two to six times higher than the approximate 10000 landslides mapped thus far from the Kaikoura earthquake.

The Version 1.0 landslide inventory contains 10195 coseismic landslides (Figures 1 and 2). These landslides are inferred to have been triggered by the Kaikoura earthquake and associated aftershocks as no major rain events occurred in the period between the earthquake and the first low-level aerial photograph survey after the earthquake, dated December 2016, used to map the distribution.

To map the distribution, we have primarily used post-earthquake 0.3 m ground resolution orthorectified air photographs, and digital surface models derived from them, alongside digital elevation models from post-earthquake airborne Light Detection and Ranging (lidar) surveys, and other pre- and post- Kaikoura earthquake imagery and lidar data (these data sets are described in Table A1). Landslides were manually digitised directly into a GIS. This was done because the outputs from the automated landslide detection tools we ran generally performed poorly. They: 1) wrongly identified areas of high albedo (in the images) as landslide sources e.g., identified bare farmland; 2) created multiple landslide source regions for individual landslide sources and vice versa where large sources were in fact multiple individual landslides; and 3) required significant time to manually edit. Several authors have shown how landslide mapping can influence an inventory and therefore the results of any analyses of it. For example, Parker et al. (2011) report more than 56000 landslides for the M_W 7.9 2008 Wenchuan earthquake, China, but Xu et al. (2014; 2016) report 196007 mapped landslides and Li et al. (2014) report 57150 landslides. Li et al. (2014) attribute their increase in numbers over Parker et al. (2011) to them separating individual landslides from amalgamated clusters. This change increased the number of mapped landslides but decreased the total volume reported, e.g., see Li et al. (2014). Because of such issues, we did not use the results of any of the automated landslide-detection algorithms. Low-level orthorectified aerial photographs taken in 2015, before the earthquake, were used to identify the many pre-earthquake landslides in the region, to ensure that such landslides were not attributed to the earthquake. We also relied on the geotagged oblique air photos taken from multiple post-earthquake helicopter reconnaissance missions to support and verify mapping in areas of complex terrain. The landslide mapping was carried out by experienced landslide researchers using the scheme outlined by Dellow et al. (2017). Where possible, we have separated the landslide source area from the debris trail to allow more accurate estimates of landslide size. This was done using a combination of aerial images, pre-and post-earthquake ground surface difference models derived from lidar and photogrammetry, and shade models generated from them, which helped to define landslide morphology. We used the scheme of Hungr et al. (2014) to classify the landslides by their mechanism and dominant material type. To date, the smallest mapped landslide source area is about 5 m^2 and the largest about 550,000 m^2. Refinement of the inventory, in particular at the lower end of the size range, is ongoing. The number of mapped landslides (frequency) with source areas of a given size has been binned using source area bin widths that are equal in logarithmic space (Figure 2a). As expected, the areas of the landslide sources generated by this event exhibit characteristic power-law scaling (Figure 2b) (e.g., Hovius et al., 1997; Guzetti et al., 2002; Malamud et al., 2004; Parker et al., 2015), defined by:

$$p(A_L) = \frac{1}{N_{LT}} \cdot \frac{\delta N_L}{\delta A_L} \qquad (1)$$

where $p(A_L)$ is the probability density of a given area within a near complete inventory—defined as the frequency density of landslides of a given source area bin (A_L), divided by the total number of landslides in the inventory—N_{LT} is the total number of landslides in the inventory, and δN_L is the number of landslides with areas between A_L and $A_L + \delta A_L$. For the landslide area bins, we adopted bin widths (δA_L) that increased with increasing landslide source area (A_L), so that bin widths were equal in logarithmic space. The position of the characteristic rollover (Figure 2b), for smaller landslides occurs at a landslide source area of about 50-100 m^2. The frequency-area distributions of most landslide inventories exhibit a rollover at smaller landslide sizes for various reasons, one of which is mapping resolution (Stark and Hovius, 2001).

The power-law scaling exponent (α) of 1.88, fitted to the Kaikoura landslide distribution using the method of Clauset et al. (2009), with $x_{min} = 500$ m^2, falls within the range of previously observed values of landslide inventories (1.4 to 3.4), but it is below the central tendency of 2.3 to 2.5 (Van Den Eeckhaut et al., 2007; Stark and Guzzetti, 2009). Figure 2 shows the landslide frequency and probability density versus area distributions for comparable inventories of landslides triggered by other notable earthquakes in New Zealand. These are: 1) the 1929 M_W 7.8 Murchison earthquake (Hancox et al., 2016), where $N_{LT} = 6104$, $x_{min} = 10000$ m^2 and $\alpha = 2.62$; and 2) the 1968 M_W 7.1 Inangahua earthquake (Hancox et al., 2014), where $N_{LT} = 1199$, $x_{min} =$ 10000 m^2 and $\alpha = 2.71$.

Our results suggest that the number of large landslides >10000 m^2 generated by this earthquake are less than those generated by the similar magnitude M_W 7.8 1929 Murchison earthquake in New Zealand, but are instead more comparable to those triggered by the smaller magnitude M_W 7.1 1968 Inangahua, New Zealand earthquake. (Figure 2a). Nevertheless, the lower α-value suggests that a higher number of larger landslides were triggered than would typically be expected given the number of smaller landslides. Such comparisons, however, do not consider differences in the physiographic setting, which could affect the numbers of landslides generated.

Although a more detailed comparison of the landslides from these different earthquakes is warranted, it is currently outside the scope of this paper.

The Geology and Topography of the Study Area

The region in which most of the landslides occurred can be subdivided into four main geological units (Figure 1b and Table 1). These are described by Rattenbury et al. (2006), and their descriptions are summarised here in order of oldest to youngest: 1) Lower Cretaceous Torlesse (Pahau terrane) “basement” rocks formed primarily of greywacke; 2) Upper Cretaceous and Paleogene limestones, siltstones, conglomerates and minor volcanic rocks; 3) Neogene limestones, sandstones and siltstones; and 4) Quaternary sands, silts and gravels. These materials and their properties tend to control the types of landslides that occurred within them. For example, the greywacke is highly jointed, and most landslides appear to be debris avalanches, controlled by multiple intersecting joint blocks, which limit the volume of such failures. Conversely the Upper Cretaceous and Neogene sandstones and siltstone tend to be massive with highly persistent bedding planes and clay seams, which allow the development of large translational debris slides and slumps. These relationships are explained further in Table 1.

The earthquake mainly affected the northeastern portion of New Zealand’s South Island. This area is dominated by the Kaikoura Ranges, which rise from sea level to a maximum elevation of 2885 m above mean sea level (AMSL) at Mount Tapuae-o-Uenuku. The Kaikoura Ranges are predominantly formed of greywacke and are dissected by several large rivers. The long, straight Clarence River valley separates the Seaward Kaikoura Ranges from the longer and steeper Inland Kaikoura Ranges, including Mount Tapuae-o-Uenuku. Beyond the Inland Kaikoura Range is the valley of the Awatere River, which runs parallel to that of the Clarence River. As these rivers approach the coast, the slopes reduce in gradient, where they are predominantly formed of faulted slivers of Neogene rocks and Quaternary gravel, sand and silt. The township of Kaikoura is the largest town in the area and is located on a rocky peninsula formed of Cretaceous to Neogene sedimentary rocks and Quaternary marine terraces, about 70 km northeast of the earthquake epicentre (Figure 1). The topography south and west of Kaikoura is relatively gentle compared to the Inland and Seaward Kaikoura Ranges. The slopes have mainly been formed by tectonically driven uplift and fluvial incision through the Neogene sandstones and siltstones, which forms the main bedrock unit in the area. The climate across much of the area is temperate and it typically experiences dry, cold winters..

Controls on the Spatial Distribution of Landslides

The landslide distribution does not represent a homogenous mass of landslides clustered around the earthquake epicentre. Instead, the mapped distribution shows a long, generally linear pattern, with many landslides occurring on either side of the faults that ruptured to the ground surface (Figure 1), to the northeast of the earthquake epicentre. Many smaller landslides concentrate along the coast and in discrete clusters on either side of the faults that ruptured. Many of the larger landslides occurred on faults with surface ruptures that passed through their source areas (Figure 3). Interestingly, the larger landslides, whilst also occurring in clusters along the faults, do not appear to occur at the same locations as the clusters of smaller landslides. A comparison of the mapped distribution with the bedrock geology shows that landslide occurrence is a function of lithological variations across the area, and field observations suggest that such variations control the nature and type of landslides triggered by the earthquake (Figure 1b, Table 1). For example, the landslide point density in the massive, but weaker Neogene sandstones and siltstones is 5.5 landslides km^{-2}, compared to 2.5 landslides km^{-2} in the stronger but closely jointed greywacke (Table 1).

The dynamic response of a slope during an earthquake is not controlled solely by lithology but comprises a complex interaction between seismic waves and the hillslope (e.g., Ashford et al., 1997; Sepulveda et al., 2005; 2011; Massey et al., 2016; Rai et al., 2016). We have used our mapped landslide distribution to explore the relationships between the occurrence of a landslide and the variables that may control its occurrence (Table 2), which we have broadly grouped into: 1) predominantly landslide forcing variables representing the intensity of the event-specific seismic ground motions and their proxies, for the Kaikoura earthquake; and 2) predominantly landslide susceptibility variables that capture the strength of the hillslope materials at a regional scale and the static shear stresses at the slope scale.

We used logistic regression (e.g., Von Ruette et al., 2011; Parker et al., 2015) to investigate the influence that the variables listed in Table 2 have on the spatial distribution of coseismic landslides attributed to the Kaikoura earthquake. The method models the influence of multiple predictor variables on a categorical response variable Y (with possible values 0 or 1) using:

$$P_{LS}(Y=1)=\frac{1}{1+\exp(-(b_0+b_1x_1+b_2x_2+b_3x_3 \ldots b_nx_n))} \tag{2}$$

where logistic regression is used to estimate the coefficients $(b, b_n \ldots)$ for predicting the probability (P_{LS}) that Y = 1, given the values of one or more predictor variables $(x, x_n \ldots)$. The condition Y = 1 corresponds to the occurrence of a landslide within a sample grid cell. The regression coefficients are estimated using a maximum likelihood criterion.

To undertake logistic regression, we have defined a sample grid at 32 m resolution, based upon an 8 m ground resolution digital elevation model, resampled from the 2012 version of the Land Information New Zealand (LINZ) digital elevation model for New Zealand. The 32 m grid mesh is much less than the typical hillslope lengths in the region, which can vary from 100 to >>1000

m. For this assessment, we have used only landslide source areas and not the debris trails. Landslides with areas less than 50 m^2 were removed from our data set to eliminate sample bias, because landslides smaller than this have not been systematically mapped and may be underrepresented in the inventory. Thus we have assumed that Y= 1 for any given sample grid cell in which its centroid falls within a landslide source area, even if the grid cell is not fully occupied by a landslide source.

The predictor variables used in this assessment were chosen based on variables previously found to influence landslide occurrence (listed in Table 2 and shown in Figures 1b, 4, 5 and 6a to d). To represent the landslide forcing variables, we adopted: 1) peak ground velocity models (PGV); 2) peak ground acceleration models (PGA); 3) the proximity of a landslide to a coseismic fault rupture; 4) permanent coseismic "fault" displacement derived from InSAR and GPS measurements (Hamling et al., 2017) (Figure 5); and 5) local slope relief (LSR). We adopted variables 1) and 2) as proxies for ground shaking, and variables 3) and 4) as less direct proxies for ground shaking. The permanent coseismic displacement variable also serves as a proxy for other susceptibility factors such as rock mass damage and steeper and higher relief. This is because displacement can lead to rock mass deformation and displacement in a vertical sense (uplift) is usually associated with reverse fault hanging walls, which in the Kaikoura region are where the steeper slopes are located. The proximity to a fault is inherently included in the estimation of PGV and PGA; however, we included it separately to examine the influence of local ground deformation and other near-field effects that might not be fully taken into account in the ground-motion models. To include this in the model, we determined the horizontal distance of each sample grid cell to the surface projection of the nearest fault that ruptured to the surface. Note that these faults are different than the locations of the simplified faults used in the

Hamling et al. (2017) fault model, and its variations, which was used in the ground motion modelling of PGA and PGV. It should also be noted that the proximity to fault variable does not account for faults that did not rupture to the surface, but which also contribute to the shaking intensity. The location of those faults that ruptured to the surface during the earthquake were taken from the GNS Science Active Faults database (Langridge et al., 2016; Stirling et al., 2017; Litchfield et al., this issue). Local slope relief (LSR) was defined as the maximum height difference within a fixed 80 m radius of the centroid of a given grid cell. It represents a proxy for slopes that could amplify ground shaking due to their "larger-scale relief" (larger than just a sample grid cell-size), where larger values of LSR represent the steeper and higher slopes of the region, which can amplify ground shaking more than lower-in-height and less steep slopes (Ashford et al. 1997; Massey et al., 2016; Rai et al., 2016; see Table 2 for details). We also used slope aspect to investigate directivity effects caused by the earthquake-rupture sequence on landslide occurrence, refer to Table 2 for details.

To estimate the PGV and PGA variables, we have used three different ground motion models, as follows: 1) $PGV_{BRADLEY}$ from Bradley et al. (2017); 2) PGV_{LF}, which is low-frequency (long period) PGV calculated up to 0.33 Hz, and derived using the method described by Holden et al. (2017); 3) PGA_{SM} and PGV_{SM} from ShakeMap NZ (listed in the Data and Resources section of this paper), developed by the USGS (Wald et al., 1999; Worden et al., 2012), and calibrated for New Zealand by Horspool et al. (2015) (Figure 4c and d). The first two models incorporate directivity and basin amplification effects using 3D velocity models and account for along-strike variations in fault slip, whereas the third does not directly account for any of those effects except where they are captured by felt reports or seismic data. All three ground motions models are based on the fault-source model of Hamling et al. (2017). All models use the strong motion data for the earthquake recorded by the GeoNet strong motion stations located within the area affected (Figures 1 and 4). However, there were only four stations within the 3600 km^2 main area that was affected by landslides, about one station for every 900 km^2, and 13 in the wider area affected (10000 km^2), about one station for every 800 km^2. The minimum, maximum and mean distance between these stations was 6.5, 51.3 and 23.6 km, respectively, indicating a sparse coverage of stations for the main area affected by landslides.

We used landslide susceptibility variables of: 1) elevation; 2) slope curvature; and 3) geology. Curvature was used as a proxy to represent potential slope-scale patterns of topographic amplification that tend to occur at breaks in slope (Ashford et al., 1997; Rai et al. 2016) and localised slope morphology that could represent pre-earthquake landslide scarps and therefore potentially unstable slopes, thus representing both a susceptibility and earthquake forcing variable. Curvature is scale dependent and will vary as a result of both the size of the landslide and the slope. For this paper, curvature was calculated using ArcGIS and taken from the curvature of the surface on a cell-by-cell basis, as fitted through that cell and its eight surrounding neighbours. This appeared to best capture the more significant breaks in slope relative to the scale of the morphology of the slopes along the coast and inland, but not the higher peaks of the Kaikoura Ranges. Further work is needed to investigate the scale dependency of slope curvature and its effects on landslide occurrence. Slope gradient and elevation of each 32 m sample grid cell were measured by taking the mean values from the n=16, 8 x 8 m grid cells that fell within it. Table 2 details how these variables were calculated. We used a categorical variable to represent the main geological units present in the area (Table 1 and summarised in Table 2), adopting four categories.

Model fitting was done manually using the Statistica software (Statistica, 2017). For a predictor variable to be included in the model, it must have a logical and statistically significant influence on P_{LS}. We used a significance level (p-value) of $p < 0.05$ (using the Wald statistic) as the threshold for inclusion in the model. During model fitting, multiple variable combinations were iteratively tested. To ensure that the predictor variables included in the model do not exhibit multicollinearity, we used a variance inflation factor matrix (VIF), given as:

$$VIF = \frac{1}{1-R^2} \tag{3}$$

where R^2 is the linear coefficient of determination of the relationship between any pair of predictor variables. Pairs with VIF >10, indicating a high level of multicollinearity, are avoided in our models (Kutner et al., 2004; Parker et al., 2015), (Table A2). The final models represent those variables that produced the best fit whilst meeting the significance level and multicollinearity criteria.

Results

We independently derived two models—one adopting PGA_{SM} and one adopting PGV_{SM} as the ground motion parameter—to hindcast the probability of a landslide occurring in each grid cell. Landslide probability (P_{LS}) is given by the following equation for PGA:

$$P_{LS} = \frac{1}{1+\exp\left(-\left(\begin{matrix} c_{Intercept}+c_{PGA_{SM}} \cdot PGA_{SM}+c_{FaultDist} \cdot FaultDist+c_{Elev_{MEAN}} \cdot Elev_{MEAN} \\ +c_{Slope_{MEAN}} \cdot Slope_{MEAN}+c_{LSR} \cdot LSR+c_{Geology_X} \end{matrix}\right)\right)} \tag{4}$$

where the regression coefficients are denoted by c. The regression coefficients and goodness of fit statistics are shown in Tables 3 and 4 for models adopting PGA_{SM} and PGV_{SM} as the ground motion predictor variables. We found that the best combination of predictor variables used to estimate landslide probability were regional ground motion models (PGA_{SM} or PGV_{SM}), distance to the surface expression of a fault that ruptured, slope gradient, elevation, local slope relief, and geology. All the other variables tested during model fitting were found to be less effective predictors than those included in the models, or they failed either the statistical significance test ($p < 0.05$) or the variance inflation factor test.

Figures 6E and 6F show the spatial distribution of P_{LS} calculated using the two regression models (Tables 3 and 4). The only difference between the combinations of variables used in each model is the ground motion parameter (PGA_{SM} and PGV_{SM}). The results show that there is little statistical (Tables 3 and 4) or spatial (Figures 6e and f) difference between the model outputs of P_{LS}. There was also little difference in the modelled P_{LS}, when substituting other ground motion variables ($PGV_{BRADLEY}$, PGV_{LF} and PGV_{MEAN}) independently in the model, whilst keeping the other variables fixed. To further investigate the explanatory power of the other variables on P_{LS}, we have adopted a model that uses PGV_{SM} (Table 4), because the model results have a marginally higher coefficient of determination—pseudo R^2 adopting Nagelkerke's R^2 method—than those when the other PGV or PGA ground motion variables were adopted. Although the pseudo R^2 of this model is relatively low, it is comparable to other similar studies on landslide data sets from New Zealand (Parker et al., 2015).

Although the predictive power of the model on P_{LS} is low, it has no apparent biases in any part of its range. Figure 7 presents a comparison of observed versus predicted P_{LS}. This relationship was calculated by accumulating (adding) the predicted P_{LS} values for each sample grid cell from smallest to largest, along with the corresponding observed Y value for the same grid cell. For the PGA_{SM} and PGV_{SM} models, the observed and predicted probabilities display a good fit to the line of equality. This shows that the modelled probabilities are broadly consistent with the data. The low pseudo R^2 of the model indicates that there are many landslides in cells with low values of P_{LS}. For example, the model adopting PGV_{SM} has about 43100 cells that are classified as being landslides (Y= 1) where the modelled landslide probability is ≤10%. However, there are over 3.4M cells where the modelled probability of a landslide occurring is less than 10%, resulting in a landslide pixel density of about 0.005 landslides per cell. Conversely, there are only 228 pixels where the modelled landslide probability is >50%, of which 26 are classified as being landslides, resulting in a landslide pixel density of 0.11.

Figure 8 shows the predictor variables in rank order of significance, which we determined by sequentially removing each of the predictor variables that contribute least to the fit of the models. In each model, and in order of importance, the slope angle, distance to fault, elevation and geology variables contribute most to the fit of the models, followed by PGV_{SM} (or PGA_{SM} when substituted for PGV_{SM} in the model) and local slope relief. Notably, distance to the surface expression of a fault that ruptured has more explanatory power in the regression model than PGA_{SM} or PGV_{SM} ground motion models or any of the other modelled PGV variables when tested independently within the regression model. This variable may be capturing: 1) additional ground motion parameters such as high-frequency ground motions that are not captured by the current PGA and PGV models, but which will sharply decay with distance from a fault; 2) the complex nature of the multi-fault rupture, and the multi-frequency ground motion intensity—and not just the higher frequency ground motions—better than the current PGA and PGV models even though it doesn't take into account the shaking contributed by those faults that did not rupture to the surface; and 3) the influence of rock mass damage on the rock mass strength, where rock masses closer to faults are likely to be more jointed or "damaged" and have lower mass strengths than those less damaged rock masses, formed in similar materials, located farther away from faults. In addition, preexisting persistent discontinuities such as faults and permanent tectonic surface deformation along some of them could have been important in triggering several of the large landslides located directly on or close to faults that ruptured to the surface.

The relatively low pseudo R^2 value of the model might be taken to suggest that variables not considered in the presented models might be important for predicting P_{LS}. For example, only four main geological units have been adopted even though there are significant differences in rock type and their associated physical properties within these four broad groups. Such differences in their properties have not yet been determined in sufficient detail to be included in the models. Also, the northwestern part of the main area affected by landslides contains a cluster of many small landslides (west-southwest of Ward, Figure 1), situated in areas of Neogene mudstone. This area does not “stand out” in the models as having a high P_{LS}, and ground shaking (Figure 4) was relatively low in this area during the Kaikoura earthquake with no nearby faults rupturing, which suggests that some other variable may be needed to explain this distribution. This area was affected by the M_W 6.5 16 August 2013 Lake Grassmere earthquake, and to a lesser extent by the M_W 6.6 21 July 2013 Cook Strait earthquake (Figure 1) (Van Dissen et al. 2013). The Lake Grassmere earthquake generated landslides in this area, and it induced slope cracking. Thus, the Lake Grassmere earthquake may have preconditioned the slopes in the area to fail in the subsequent Kaikoura earthquake (as described by Parker et al., 2015). Alternatively, the mismatch between modelled and observed landsliding could be due to the high amplitude of the shaking in this area from the large amount of slip on the Kekerengu fault (Litchfield et al., this issue), which may not be captured by the current ground motion models.

Discussion

Kaikoura earthquake landslide numbers and their size

Our results suggest that the number of large landslides >10000 m^2 generated by this earthquake is fewer than the number generated by the similar magnitude M_W 7.8 1929 Murchison earthquake in New Zealand and similar to the number triggered by the smaller magnitude M_W 7.1 1968 Inangahua earthquake. One reason for this might be that the area affected by strong shaking and landslides is topographically constrained. Specifically, several of the faults that ruptured to the surface extended off shore, leading to the triggering of many submarine landslides (Mountjoy et al., this issue), which are not taken into account in the terrestrial landslide distribution examined in this paper. Another reason for this difference may be that the Kaikoura earthquake involved the rupture of more than 20 faults that broke to the land surface over a fault-zone length of more than 100 km, suggesting that the earthquake comprised numerous “sub-events” (Kaiser et al., 2017; Stirling et al., 2017) of lower magnitude (Hamling et al., 2017). For each fault that ruptured, an equivalent magnitude can be calculated based on fault dimensions and estimated total slip (estimated either from geodetic and/or seismic data inversion). Hamling et al. (2017) estimated that the cumulative moment from the faults that ruptured south of Kaikoura equates to a M_W 7.5 earthquake. Even though the cumulative moment from the northern faults is larger than from the southern faults, the moment from some of the individual smaller faults that ruptured to the north of Kaikoura was equivalent to a M_W 7.1 earthquake (Hamling et al. 2017). This would conceptually result in the shaking energy being distributed over a larger area but at a smaller amplitude and, possibly, duration. Large landslides are possibly more sensitive to shaking in the range of frequencies that often control the ground motion PGV. If the moment release was distributed across many faults, the shaking duration and frequency content would reflect rupture from many smaller faults rather than a single large fault. Thus, the shaking would not have the same intensity as would be produced by a single fault rupturing with M_W 7.8. With this in mind, it will be important to update the landslide regression models as improved ground motion modelling for the earthquake becomes available.

Possible controls on the size of the landslides triggered by the earthquake:

The strength of the dominant rock type in the area that was strongly shaken was mainly greywacke, and it accounts for 60% of the rocks in the main area affected by landslides. The coseismic landslide distribution in greywacke is dominated by many small landslides with few very large ones. Non-earthquake induced landslides in such materials have in the past been limited in size as greywacke tends to be highly jointed, favouring smaller failures (Hancox et al., 2015). Previous work on fracture spacing in Torlesse Schist of the Southern Alps, suggests that its properties are highly influential in geomorphic response (Hales and Roering, 2009). Additional work is required to assess the landslide distribution source areas and volumes with regards to the main geological materials in which they occurred, and the role of rock mass conditioning of landslide source areas/volumes. Such an assessment is outside the scope of this current paper.

Another contributing factor might be that high-frequency energy radiation during the Kaikoura earthquake is inferred by some researchers as smaller than during other landslide-triggering crustal earthquakes of a similar magnitude. While the energy magnitude (Me) of the Kaikoura and M_W 7.8 2008 Wenchuan, China earthquakes derived from low-frequency (0.5 – 70 s) waveforms (IRIS DMC, 2013a) are similar (Me = 7.93 and 8.06, respectively), there is a significant difference in the energy magnitudes estimated from higher frequency (0.5 – 2 s) waveforms (Me = 7.59 and 8.05, respectively) (IRIS DMC, 2013b).

The role of distance from the surface fault rupture

To explore the relationship between landslide occurrence and proximity to a surface fault rupture, we have plotted the landslide point and area densities as a function of the distance from the surface expression of the nearest fault that ruptured (Figure 9). We did this by creating successive buffer zones around the mapped fault traces that ruptured to the surface (Litchfield et al., this issue; Nicol et al., this issue). We then computed the number and total area of landslide source areas within each successive 200 m buffer to a distance of 3000 m on either side of the mapped fault trace as well as those landslide source areas through which faults pass. The density of landsliding in areas outside the fault buffers was also calculated for comparison. The results show that the landslide densities (both point and area) within 200 m of a fault are as much as three times greater than densities outside the 3000 m buffers. The results also show a general decrease in landslide density with increasing distance from a fault. At a distance of about 2500 to 3000 m, the background landslide density (termed “rest of area” in Figure 9) is reached. This finding may be the result of: a) high-frequency shaking, which declines rapidly with distance from a fault, being an important control on the density of landslides triggered by the Kaikoura earthquake; b) the rock masses close to faults being weaker because of damage from previous fault rupture events; and c) slopes nearer faults often exhibit greater relief and are steeper than those farther away, which is the case for those slopes in the Kaikoura region. Others have reported similar findings, for example, Scheingrosset al. (2013) hypothesized that earthflows tend to cluster near the creeping San Andreas Fault because of a fault-induced zone of reduced bulk-rock strength that increases hillslope susceptibility to failure. Meunier et al. (2007) also suggested that near-field (near-fault), high-frequency shaking is likely to have been an important control on the density of landslides triggered by earthquakes.

Only 44 of the mapped landslide source areas are directly intersected by faults that ruptured to the surface, but this number includes seven of the eight largest landslides triggered by the Kaikoura earthquake. This would suggest that the initiation of these large landslides might have been due to a combination of preexisting discontinuities such as faults and rock mass damage, dynamic strong shaking and permanent tectonic displacement of the fault as it ruptured to the surface within the source area.

Earthquake ground motion frequency, slope amplification and landsliding

As noted above, our logistic regression analysis indicates that PGA (or PGV when substituted for PGA in the model) from the ShakeMap NZ models performs best, but overall the PGA (or PGV) variable has low explanatory power on predicting landslide occurrence. Distance to fault, which may capture additional ground motion parameters, has a much higher explanatory power.

Generally, the shaking nearer the source contains a lot more high-frequency energy than farther away (e.g., Davies, 2015), suggesting that ground motion frequency may play a key role in determining slope response. Therefore, slopes that are near faults that rupture are more likely to experience such high-frequency ground motions. If the fundamental frequency of the slope is similar to the dominant frequency of the ground motion, amplification of shaking may also occur (Geli et al., 1988).

Ashford et al. (1997) showed that the fundamental frequency (f) of a slope behind the crest can be estimated using the following equation:

$$f = \frac{4 \times H}{V_S} \qquad (5a)$$

and a slope/topographic frequency:

$$f = \frac{5 \times H}{V_S} \quad (5b)$$

where H is the slope height (or relief) and V_S is the shear wave velocity of the material forming the slope. More recently, Rai et al. (2016) have developed a model to predict the effects of topography on earthquake ground motions, adopting the relative relief of a slope (like the LSR used in this paper). For slopes in the main area affected by landslides, the mean and modal values for slope relief are 135 m and 85 m for coastal slopes, and for inland slopes they are 588 m and 103 m, respectively. The mean V_S30 of the rock forming the coastal and inland slopes is estimated by Perrin et al. (2015) as 1000 m/s. Equation (5a) yields fundamental frequencies of the coastal slopes ranging from 1.9 to 2.9 Hz, and of the inland slopes from 0.4 to 2.4 Hz, for the mean and modal slope relief, respectively. Such fundamental frequencies are relatively high, suggesting that the combination of high-frequency shaking at close proximity to the faults, and amplification of shaking caused by the slopes responding to such high-frequency shaking, may explain why so many landslides occurred on slopes adjacent to faults. It should be noted that it is not just the fundamental frequency of the hillslope that matters, which will scale with slope morphology and relief, but also the fundamental frequency of the potential failure mass, which is likely to be shallower, and therefore have a higher fundamental frequency than the overall slope. However, such a difference may only be distinguishable from the rest of the slope if there is some preexisting plane or damage resulting in a contrast of density/shear wave velocity between the potential failure mass and the slope (e.g., Massey et al., 2016).

Landslide slope angle and elevation

We have explored the higher density of landslides on the coastal slopes by attributing the centroid of each landslide source area with its mean slope angle and elevation. We split the landslide distribution into coastal and non-coastal slopes—where costal slopes are defined as those that extend from the sea to the first main inland ridge line, an approximate strip about 1 km wide—and calculated the area of coastal/non-coastal slope within each slope angle bin (Figure 10). The results show that coastal slopes consistently have more landslides for a given slope angle than corresponding inland slopes, but that the mean size of the landslide sources on the inland slopes is larger than those on coastal slopes. Variations in slope angle and geology cannot explain this difference because the proportion of inland slopes in the steeper slope angle bins (Figure 10b, inset) is larger than the proportion of slopes on the coast within the same corresponding slope angle bins. The coastal slopes are primarily formed from greywacke, which is also the dominant rock type forming the slopes inland. A possible explanation for these smaller landslides on the coastal slopes is that their size has been limited by the topography, as the coastal slopes have a lower relief (i.e., elevations less than 500 m AMSL) compared to the higher relief slopes inland.

The results of the logistic regression model show that landslide probability increases with decreasing elevation and coastal slopes are at lower elevations. This finding could be due to the coastal-slope geometry and materials (and contrasting materials caused by coastal weathering processes and products), and their effects on amplifying the ground shaking. Studies of similar coastal slopes (Massey et al., 2016), albeit in different materials, have shown that amplification—between the peak acceleration of the free field earthquake motion and the average peak acceleration of the slope—of shaking between the base and crest of a slope could be up to 2.5 times, with a mean of 1.6 times, higher at the crest than the base of the slope. Such values are comparable to the amplification factors reported by Ashford et al. (1997). It is also possible that the predominantly greywacke coastal slopes are more weathered than their inland counterparts. Such hypotheses are likely to form the basis of future research on the landslides generated by the Kaikoura earthquake.

Conclusions

Our main findings are: (1) the number of large landslides (with source areas ≥10000 m^2) triggered by the Kaikoura earthquake is fewer than the number of similar sized landslides triggered by other similar magnitude earthquakes in New Zealand; (2) the largest landslides (with source volumes from 5 to 20 M m^3) occurred either on or within 2500 m of the more than 20 mapped faults that ruptured to the surface; (3) the landslide density within 200 m of a mapped surface fault rupture is as much as three times higher than those densities farther than 2500 m from a ruptured fault; (4) for the same slope angles, coastal slopes have landslide point densities that are an order of magnitude greater than those in similar materials on the inland slopes, but their source areas are significantly smaller, possibly indicating that these slopes locally amplified ground shaking, and (5) the "distance to fault" predictor variable, when used as a proxy for ground motion intensity, has more explanatory power in predicting landslide probability than the modelled PGA or PGV variables adopted in the logistic regression modelling, even though this variable does not account for faults that did not rupture to the surface, but which also contribute to the shaking intensity. This relationship might be because the distance to fault variable captures: (a) the high-frequency ground motions and their attenuation with distance from a fault better than the current PGA and PGV models; (b) the complexity of the multi-fault rupture, and therefore the multi-frequency ground motion intensity, better that the current PGA and PGV models; and (c) the more damaged nature of the rock masses close to the faults, where they tend to be more sheared and weakened. The strong explanatory power of the "distance to fault" predictor variable could also reflect the apparent structural control of some of the largest landslides that occur on or near faults.

Data and Resources

A recent update on information relating to submarine landslides triggered by the Kaikoura earthquake was given in the AGU Landslide Blog. 2017. Last accessed October 2017. http://blogs.agu.org/landslideblog/2017/02/27/niwa-1/

The ShakeMap NZ map of peak ground accelerations for the Kaikoura earthquake was published online on the GeoNet website. 2016. Last accessed October 2017. http://www.geonet.org.nz/news/fiBlIE2uNq2qGmmiOg42m

The software package used to carry out the logistic regression is called Statistica. 2017. Last accessed October 2017. http://www.statsoft.com/Products/STATISTICA-Features

The version 1.0 landslide dataset used in this paper can be downloaded from the GNS Science landslide database https://data.gns.cri.nz/landslides/ or the https://www.designsafe-ci.org/ website.

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Table 1. Lithology and landslide types adopted in this paper

Lithology	Proportion of main area affected (%)	Landslide point / area density* (N/km^2 / %)	Dominant landslide types
Quaternary sands, silts and gravels. These typically form river terrace deposits in the region.	19	1.8 / 0.3	Debris avalanches and flows that tend to be relatively small, and their source areas are mainly located in the terrace sands and gravels on top of the steeper coastal slopes. Many other landslides occurred within the shallow regolith, which covers many slopes in the area that were affected by strong ground shaking. These include shallow, translational slides in soil with displacements of a few centimetres to several metres.
Neogene limestones, sandstones and siltstones. These are typically massive, but weak rocks with unconfined compressive strengths (UCS) of typically <2MPa (Read and Miller, 1990).	9	5.5 / 0.9	Relatively shallow debris avalanches and flows that source from the more weathered rocks, or relatively deep-seated slides and slumps, where movement is thought to occur either along bedding or other persistent structural discontinuities, e.g., fault planes, thin clay seams, or through the rock mass. Substantial numbers of pre-Kaikoura earthquake, large landslides were mapped in these materials of which many reactivated (a few centimetres to metres) during the earthquake, forming translational and rotational slides/slumps.
Upper Cretaceous to Paleogene rocks including limestones, sandstones, siltstones and minor volcanic rocks. These are typically massive, but weak rocks with UCS of typically <2MPa.	12	4.6 / 0.5	Rockfalls and debris avalanches in areas of steeper terrain, with some slides and slumps (termed coherent after Keefer 2013) in areas of less steep topography, and their location might be controlled by the presence of thin clay seams or small-scale changes in lithology. Several relict landslides are present in these materials, and there were numerous small rockfalls and debris avalanches from their over-steep head scarps.
Lower Cretaceous Torlesse (Pahau terrane) "basement" rocks predominantly sandstones and argillite (greywacke). These are relatively strong rocks with UCS 10-20 MPa, but they are closely jointed.	60	2.5 / 0.6	Rockfalls (of individual boulders) to debris and rock avalanches. Given the highly discontinuous nature of the rock mass, most landslides are controlled by multiple intersecting joint blocks, hence a potential limitation on the volume of such failures. However, the Kaikoura earthquake triggered several very large and structurally controlled rock avalanches, the Hapuku landslide being the largest mapped landslide.

* Landslide point densities were calculated by dividing the number of landslide sources within a given geological unit by the area of ground within that given unit, within the main area affected by landslides (Figure 1b). Landslide area densities were calculated by dividing the total area of all landslide sources within a given geological unit by the area of ground within that unit, within the main area affected by landslides.

Table 2. Predictor variables evaluated in the logistic regression model, their ID codes, descriptions and units.

Variable type	Variable ID	Description	Units
Susceptibility	Geology 1	Quaternary sands, silts and gravels. These materials typically form terrace deposits on the top of the steep coastal cliffs as well as inland slopes adjacent to the main rivers of the area. Many of these terraces have been incised by rivers.	N/A
	Geology 2	Neogene limestones, sandstones and siltstones, which are typically weak. They occur along sections of the coast north of Kaikoura.	N/A
	Geology 3	Upper Cretaceous to Paleogene rocks including limestones, sandstones, siltstones and minor volcanic rocks. These are typically weak (like the Neogene limestones and sandstones), and easily erodible and they can contain thin clay seams, which are volcanic in origin. They are typically exposed in narrow strips overlying the greywacke basement rocks.	N/A
	Geology 4	Lower Cretaceous Torlesse (Pahau terrane) “basement” rocks are predominantly sandstones and argillite, also known as greywacke. The greywacke rocks are typically moderately to well bedded and tend to be closely jointed. They form many of the coastal slopes as well as the steeper inland Kaikoura mountain ranges.	N/A
	$\text{Slope}_{\text{MEAN}}$	Local hillslope gradient taken from the 8 m resolution digital elevation model generated by GNS Science, adopting the mean value of all of the 8 m by 8 m cells that fall within each cell of the sample 32 m by 32 m grid. This variable is a proxy for the static shear stresses in the slope.	Deg (°)
	$\text{Elev}_{\text{MEAN}}$	Local hillslope elevation taken from the 8 m resolution digital elevation model generated by GNS Science, adopting the mean value of all of the 8 m by 8 m cells that fall within the each cell of the sample 32 m by 32 m grid. This variable represents the observation that topography can limit the size of the landslides. For example, slopes that are higher in elevation tend to have larger surface areas, and can therefore generate larger landslides than slopes at lower elevations, which tend to have smaller surface areas.	mAMSL
	$\text{Curv}_{\text{PROFILE}}$	Profile curvature generated using ArcGIS, taken from the curvature of the surface on a cell-by-cell basis, as fitted through that cell and its eight surrounding	One hundredth

		neighbours. using the 8 m resolution digital elevation model generated by GNS Science. A negative value indicates the surface is upwardly convex at that cell. A positive profile indicates the surface is upwardly concave at that cell. A value of 0 indicates the surface is flat. This variable is a proxy for slope "sharpness" that represents topographic amplification effects, as amplification of shaking has been recorded at sharp breaks in slope (e.g., Massey et al., 2016; Janku, 2017).	(1/100) of a z-unit
	Asp_{MEAN}	The aspect for each sample grid-cell was calculated using ArcGIS using the 8 m resolution digital elevation model generated by GNS Science, adopting the mean of all of the 8 m by 8 m cells that fall within each cell of the 32 m by 32 m sample grid.	Deg (°)
Earthquake forcing	FaultDist	The distance from the centroid of each of the 32 m by 32 m sample grid cells to the nearest fault that ruptured using the mapped surface expression, taken from the GNS Science Active Faults database (Langridge et al., 2016), which includes those faults that ruptured during the Kaikoura earthquake.	Meters
	PGA_{SM}	Grid of the mean peak ground acceleration (PGA) derived from ShakeMap NZ (GeoNet, 2016), developed by the U.S. Geological Survey (Wald, 1999; Worden, 2012), and calibrated for New Zealand by Horspool et al. (2015). Grid resolution is 1000 m by 1000 m. The PGA values were attributed to the sample grid cell, by taking the PGV value at its centroid.	g
	PGV_{SM}	Grid of the mean peak ground velocity (PGV) derived from ShakeMap NZ. Grid resolution is 1000 m by 1000 m. The PGV values were attributed to the sample grid cell, by taking the PGV value at its centroid.	m/s
	PGV_{LF}	Low-frequency (long period) PGV calculated from waveforms up to 0.33 Hz, using the method described by Holden et al. (2017). Grid resolution is 500 m by 500 m. The PGV values were attributed to the sample grid cell, by taking the PGV value at its centroid.	m/s
	$\text{PGV}_{\text{BRADLEY}}$	Grid of PGV derived from modelling carried out by Bradley et al. (2017). Grid resolution is 990 m by 990 m. The PGV values were attributed to the sample grid cell, by taking the PGV value at its centroid.	m/s
	PGV_{MEAN}	Mean PGV calculated for each of the sample grid cells by sampling the PGV value from each of the three PGV models model at the centroid of each sample grid, and taking the mean of the three values.	m/s

	Disp_{V}	The vertical permanent tectonic displacement caused by the earthquake was taken from the 100 m by 100 m resolution three-dimensional displacement field derived from satellite radar and GPS data (Hamling et al., 2017). This variable is a proxy for ground shaking intensity because areas of increasing permanent tectonic displacement should correlate with increased dynamic ground shaking and inertial loading on the soil and rock masses forming the slopes, leading to an increase in landsliding.	Meters
	Disp_{H}	The horizontal permanent tectonic displacement was calculated for each sample grid cell as the vector of the maximum x and y displacement fields taken from Hamling et al. (2017), 100 m by 100 m resolution three-dimensional displacement field. As Disp_{V}, this variable is a proxy for ground-shaking intensity.	Meters
	LSR	Local slope relief calculated using focal statistics in ArcGIS. It represents the local height (and angle) of the sample grid cell. It is calculated as the difference in elevation between the lowest in elevation 8 m by 8 m grid cell, within an 80 m (ten (10) 8 m cells) radius from the centroid of the given sample grid cell, and the mean elevation of that grid cell ($\text{Elev}_{\text{MEAN}}$). This variable represents a proxy for slopes that could amplify ground shaking due to their "larger-scale steepness" (larger than just a sample grid-cell size), where larger values of LSR represent the steeper and higher slopes of the region, which can amplify ground shaking more than lower in height and less steep slopes (Ashford et al. 1997; Massey et al., 2016).	Meters (m)

Table 3. Logistic regression output coefficients and model fit statistics. Input ground-motion variable PGA_{SM}. Binomial logistic regression – modelled probability that Landslide = 1.

Variable	Coefficient (c)	Standard error	95% confidence interval	
			Lower bound	Upper bound
Intercept	-8.2531	0.0471	-8.3454	-8.1608
PGA_{SM}	0.0278	0.0005	0.0268	0.0288
FaultDist	-0.0002	0.000003	-0.0002	-0.0002
$Elev_{MEAN}$	-0.0014	0.00002	-0.0014	-0.0013
$Slope_{MEAN}$	0.0816	0.0012	0.0793	0.0840
LSR	0.0158	0.0006	0.0146	0.0169
Geology 1	0.5813	0.0196	0.5429	0.6197
Geology 2	0.1963	0.0186	0.1599	0.2327
Geology 3	-0.1466	0.0104	-0.3117	-0.2434
Geology 4	-0.6866	0.0084	-0.7031	-0.6700

Number of observations: 3,481,858. Likelihood ratio X^2: 3.41×10^4. All variables have p values of less than 1×10^{-8}. Pseudo R^2: 0.141

Table 4. Logistic regression output coefficients and model fit statistics. input ground-motion variable $\mathrm{PGV}_{\mathrm{SM}}$. Binomial logistic regression – modelled probability that Landslide = 1.

Variable	Coefficient (c)	Standard error	95% confidence interval	
			Lower bound	Upper bound
Intercept	-8.5968	0.0494	-8.6937	-8.4999
$\mathrm{PGV}_{\mathrm{MEAN}}$	0.0294	0.0005	0.0284	0.0303
FaultDist	-0.0002	0.000003	-0.0002	-0.0002
$\mathrm{Elev}_{\mathrm{MEAN}}$	-0.0013	0.00002	-0.0013	-0.0012
$\mathrm{Slope}_{\mathrm{MEAN}}$	0.0835	0.0012	0.0812	0.0858
LSR	0.0158	0.0006	0.0147	0.0170
Geology 1	0.1537	0.0214	0.1117	0.1957
Geology 2	0.3005	0.0188	0.2637	0.3372
Geology 3	-0.0978	0.0174	-0.1320	-0.0636
Geology 4	-0.3563	0.0137	-0.3831	-0.3295

Number of observations: 3,481,858. Likelihood ratio X^2: 3.49×10^4. All variables have p values of less than 1×10^{-8}. Pseudo R^2: 0.144

List of Figure Captions

Figure 1. Inset map shows the area of New Zealand affected by coseismic landslides triggered by the M_W 7.8 2016 Kaikoura earthquake. a) Shows the mapped 10195 coseismic landslide source areas and their size (area) triggered by the earthquake, superimposed on the 8 m by 8 m digital elevation model for New Zealand, classified by elevation in meters above sea level. b) The landslide source area distribution overlain on the main geological units. c) Landslide source area distribution shown on the 8 m ground resolution digital elevation model for New Zealand.

Figure 2. a) The number of landslides (frequency) with source areas within each source area bin. Landslide source-area bin widths are equal in logarithmic space for all data sets. b) Landslide probability density plotted against landslide area (for the landslide source areas only), for landslides generated by the Kaikoura earthquake, the M_W 7.1 1968 Inangahua, New Zealand earthquake (Hancox et al., 2014) and the M_W 7.8 1929 Murchison, New Zealand earthquake (Hancox et al., 2016). For Figure 2b the power-law fitting statistics are: 1) M_W 7.8 2016 Kaikoura earthquake, where $N_{LT} = 10195$, $x_{min} = 500$ m^2 and $\alpha = 1.88$; 2) M_W 7.8 1929 Murchison earthquake (Hancox et al., 2016), where $N_{LT} = 6104$, $x_{min} = 10000$ m^2 and $\alpha = 2.62$; and 3) the M_W 7.1 1968 Inangahua earthquake (Hancox et al., 2014), where $N_{LT} = 1199$, $x_{min} = 10000$ m^2 and $\alpha = 2.71$.

Figure 3. a) Hapuku rock avalanche in Lower Cretaceous basement rocks – this is the largest of the mapped landslides with an estimated volume of about 20 (±2) M m^3. In this case, the slide surface appears to correspond to multiple persistent discontinuities such as old and recent fault planes. Several faults that ruptured to the surface pass through the source area of the landslide. The debris left the source and blocked the Hapuku River creating a dam about 100 m

high. Multiple lobes of debris of different clast size can be mapped in the deposit, indicating multiple pulses of debris deposition. The dam subsequently overtopped and the downstream face was partially eroded (due to headward erosion initiated by seepage through the dam) following Cyclone Cook in April 2017. The debris left in the source is still unstable and several debris flows have occurred, which have eroded the debris down to bedrock in places. The debris forming the dam continues to erode as water from the impounded lake flows over the crest and down the outflow channel. b) Seafront rock slide/slump in Paleogene limestone – This is the largest mapped landslide in these materials with an approximate volume of 18 (±2) M m^3. This slide surface is assumed to be deep seated (>100 m below the surface), with the field observations and cross sections suggesting a semi-rotational failure through the rock mass. Much of the debris has remained intact, and so the slide/slump would be classified as coherent (Keefer, 2013). The displaced mass is still creeping and several debris flows have occurred off the toe of the intact displaced debris and also the head scarp. The Papatea fault (Hamling et al., 2017) ruptured through the source area suggesting that surface rupture of this fault caused the landslide to initiate. The vertical displacement of this fault measured approximately 0.5 km away from the landslide is about 6 m. We are not sure whether the landslide initiated either from permanent coseismic displacement of the ground or dynamic displacement caused by shaking, or some combination of both. c) Leader River rock slide/slump in Neogene mudstone – The largest mapped landslide in these materials is the Leader River landslide with an approximate volume of 8 (±1) M m^3. This rock slide/slump is predominantly within Neogene mudstone (including sandstone and siltstone), and the slide surface is assumed to be deep seated (about 80 m below the surface) with the displacement vectors suggesting a translational failure (with some rotation at the head scarp), possibly along bedding, which is inclined at about 20° to 25° out of the slope (measured near the toe of the debris) and has the same dip direction as the vectors of landslide displacement. A faulted contact between the Lower Cretaceous greywacke and Neogene mudstone is also present in the landslide head scarp. Although there is no field-evidence to suggest this contact ruptured, it is possible that a fault also ruptured through the source area of this landslide (Nicol et al., this issue), but more investigation is needed to determine whether this is the case or not. All photos D. Townsend.

Figure 4. Peak ground velocities (PGV) and peak ground accelerations (PGA) from: a) Bradley et al. (2017) ($PGV_{BRADLEY}$), calculated up to frequencies of >10 Hz, grid resolution 1000 m; b) PGV_{LF} calculated using the method by Holden et al. (2017) up to a frequency of 0.33Hz, grid resolution 500 m; and c) PGV_{SM} from Shake Map NZ (median estimates), calculated up to frequencies of 50 Hz, grid resolution 1000 m. d) PGA_{SM} from Shake Map NZ (median estimates), calculated up to frequencies of 50 Hz, grid resolution 1000 m. The Kaikoura earthquake landslide distribution (shown as grey polygons, N=10195 landslides) are overlain on all the maps.

Figure 5. Permanent ground displacement: a) horizontal; and b) Vertical, and the inferred fault model taken from InSAR and GPS measurements relating to the Kaikoura earthquake presented by Hamling et al. (2017), grid resolution of 100 m by 100 m, overlain by the Kaikoura earthquake landslide distribution (shown as grey polygons, N=10195 landslides).

Figure 6. Maps a) to d) showing the distributions of the main susceptibility predictor variables used in the logistic regression model. a) Elevation ($Elev_{MEAN}$); b) Slope ($Slope_{MEAN}$); c) Distance to fault (FaultDist); d) Local slope relief (LSR). Maps e) and f) show the estimated landslide probabilities (P_{LS}) from the logistic regression model: e) adopting the PGA_{SM} variable as the

input ground motion; f) adopting the PGV_{SM} variable as the input ground motion. The faults that ruptured to the surface during the earthquake are shown as red lines.

Figure 7. Consistency of the logistic regression model probabilities with the data, adopting the variables listed in Table 3 and ground motion parameter PGV_{SM}. The graph shows a comparison of observed and predicted landslide probabilities, calculated by accumulating (adding) the predicted landslide probability (P_{LS}) values for each sample grid cell from smallest to largest, along with the corresponding observed Y value for the same grid cell.

Figure 8. Logistic regression model performance adopting the variables listed in Table 4 and ground motion parameter PGV_{SM}. The graph shows the relative contributions of predictor variable to the fit of the overall model. The sequence of model variables and the resulting pseudo R^2 values are shown in rank order of their significance, which we determined by sequentially removing each of the predictor variables contributing least to the fit of the model.

Figure 9. Landslide point and area density (N=10195 landslides) within each 200 m distance from fault buffer. Landslide density is calculated by taking the centroid of each landslide source area that falls within each 200 m distance buffer from the mapped surface expression of the faults that ruptured during the earthquake. The number (N) of landslide points within each distance from fault bin range is then divided by the area of slope (km^2) within each bin. The landslide area density is also shown, which is calculated in the same way as the landslide point density; however, the area of each landslide source (km^2) within each distance from fault bin is summed and divided by the total area of ground within each 200 m bin.

Figure 10. Landslide source areas (N=10195 landslides) normalised relative to the largest mapped landslide (area in km^2) and their associated elevation and slope angle taken from the 8 m

by 8 m New Zealand digital elevation model. The slope angle and elevation values attributed to each landslide source area were sampled from the digital elevation grid by calculating the mean values within each source area polygon. a) Landslides on coastal slopes only; and b) landslides on non-coastal slopes. c) Area of slope within a given slope angle bin as a proportion of the total area of coastal and non-coastal slopes. d) Landslide point density for each slope angle bin adopting 10-degree bins. Landslide density is calculated by taking the number of landslide sources that have mean slope angles that are within each 10-degree slope-angle bin range. The number (N) of landslides within each slope-angle bin range is then divided by the area of slope (km^2) within each bin. The point densities are calculated for coastal and non-coastal slopes and landslides.

Appendices

The information contained in the appendices comprise the following: 1) A summary of the imagery and topographic data used to map Version 1.0 of the landslide distribution presented in the paper; and 2) The variance inflation factor matrix (VIF), which was used as a method to ensure that the predictor variables included in the logistic regression model did not exhibit multicollinearity.

Table A1. Summary of data used to compile the landslide inventory.

	Item	Data	Type	Date (NZST)	Source	Ground resolution (m)	Public availability	Notes
Pre Kaikoura earthquake data	1	Kaikoura District aerial photographs	Orthorectified mosaics Individual tiled tiffs (provided by Council) converted to one mosaic by GNS Science.	2014-2015	Environment Canterbury (ECAN), (captured by Aerial Surveys)	0.3	Yes	
	2	Marlborough District aerial photographs	Orthorectified mosaic Individual tiled tiffs (provided by Council) converted to one mosaic by GNS Science.	2011-2012	Marlborough District Council (MDC), (captured by Aerial Surveys)	0.4	Yes	
		Marlborough District aerial photographs	Orthorectified mosaic Individual tiled tiff format files	2015-2016	MDC, captured by AAM Group Ltd.	0.2	Early 2018	
	3	Kaikoura Digital Surface Model (DSM), generated from the photographs taken for 1 and 2.	ESRI Grid file	2014-2015	ECAN, captured by Aerial Surveys Ltd.	1.0	Early 2018	
	4		ESRI Grid file already provided	2014-2015	ECAN, captured by Aerial Surveys Ltd.	10.0	Early 2018	

	5	Airborne lidar	Point clouds converted to Digital Elevation Models (DEMs) and Hillshades by GNS Science	2012	Captured by AAM Group Ltd.	1.0	Yes	Only the coastal strip from Ward through to Cheviot
Post Kaikoura earthquake data	5	WorldView-2 satellite imagery	Multispectral bands supplied raw. Orthorectified as an Imagine file and converted to mosaics by GNS Science	22 November 2016	Captured by Digital Globe	2.4	Yes	EAGLE technology processed the same raw images and provided to all of government.
	6	WorldView-3 satellite imagery		15 November 2016	Captured by Digital Globe	1.4	Yes	
	7	GeoEye satellite imagery		15 November 2016	Captured by Digital Globe	2.0	Yes	
	8	Aerial photographs	RGB stereo-tiff files with image coordinates, processed to individual orthorectified images and DSMs by GNS Science. Aerial Surveys to provide complete processed data set	December 2016	Captured by Aerial surveys Ltd. commissioned by LINZ at the request of GNS Science and other New Zealand agencies	0.3	Early 2018	Area covered is the main area affected by landslides and the total area affected by landslides.
	9	Airborne LIDAR	Point clouds converted to DEM and Hillshades by GNS	November to December 2016	Captured by AAM Group Ltd. commissioned by LINZ at the request of GNS Science and other New Zealand agencies	1.0	On request	Only the coastal strip, main faults and Goose Bay provided to date. Additional areas (dam sites) to be provided later.
	10	Terrestrial LIDAR of landslides and landslide dams on the rivers called – Hapuku, Ote	Point clouds, orthorectified images,	November and December 2016	Captured by GNS Science	Variable	Yes	Multiple surveys of each dam. Several of the dams failed following Cyclone Debbie and Cook, and

		Makura, Linton, Conway, Towy, Stanton and Leader		March, April, May and September 2017				surveys of these dams were carried out post failure.

Table A2. Variance inflation factor matrix (VIF) for the variables included in the logistic regression models. VIF values greater than 10 indicate a high level of multicollinearity (Kutner et al., 2004) and are avoided in our models.

Variables	LSR	$\text{Slope}_{\text{MEAN}}$	$\text{Elev}_{\text{MEAN}}$	FaultDist	PGV_{MEAN}	$\text{PGV}_{\text{BRADLEY}}$	PGV_{LF}	PGV_{SM}	PGA_{SM}
LSR	-	4.54	1.60	1.00	1.01	1.03	1.03	1.13	1.00
$\text{Slope}_{\text{MEAN}}$		-	1.48	1.00	1.00	1.03	1.03	1.16	1.00
$\text{Elev}_{\text{MEAN}}$			-	1.04	1.00	1.00	1.02	1.11	1.02
FaultDist				-	1.10	1.04	1.02	1.30	1.60
PGV_{MEAN}					-	8.89	3.04	1.09	1.16
$\text{PGV}_{\text{BRADLEY}}$						-	1.96	1.00	1.05
PGV_{LF}							-	1.00	1.03
PGV_{SM}								-	1.62
PGA_{SM}									-

Figure 1

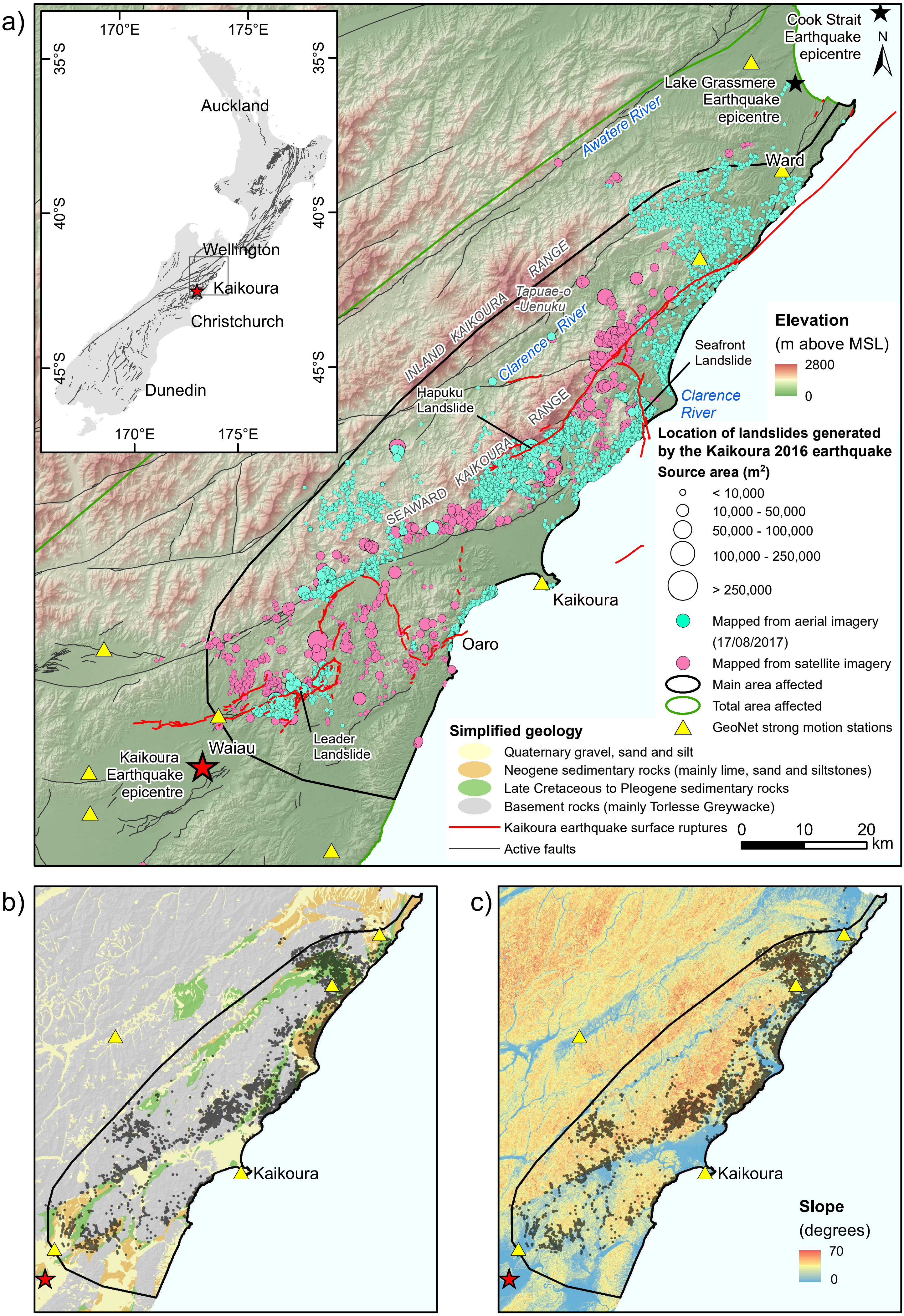

a)
170°E
175°E
35°S
40°S
45°S
Auckland
Wellington
Kaikoura
Christchurch
Dunedin
Cook Strait Earthquake epicentre
N
Lake Grassmere Earthquake epicentre
Awatere River
Ward
INLAND KAIKOURA RANGE
Tapuae-o -Uenuku
Clarence River
Seafront Landslide
Hapuku Landslide
SEAWARD KAIKOURA RANGE
Clarence River
Kaikoura
Oaro
Leader Landslide
Waiau
Kaikoura Earthquake epicentre
Elevation
(m above MSL)
2800
0
Location of landslides generated by the Kaikoura 2016 earthquake
Source area (m^2)
< 10,000
10,000 - 50,000
50,000 - 100,000
100,000 - 250,000
> 250,000
Mapped from aerial imagery (17/08/2017)
Mapped from satellite imagery
Main area affected
Total area affected
GeoNet strong motion stations
Simplified geology
Quaternary gravel, sand and silt
Neogene sedimentary rocks (mainly lime, sand and siltstones)
Late Cretaceous to Pleogene sedimentary rocks
Basement rocks (mainly Torlesse Greywacke)
Kaikoura earthquake surface ruptures
Active faults
0
10
20
km
b)
Kaikoura
c)
Kaikoura
Slope
(degrees)
70
0

Figure 2

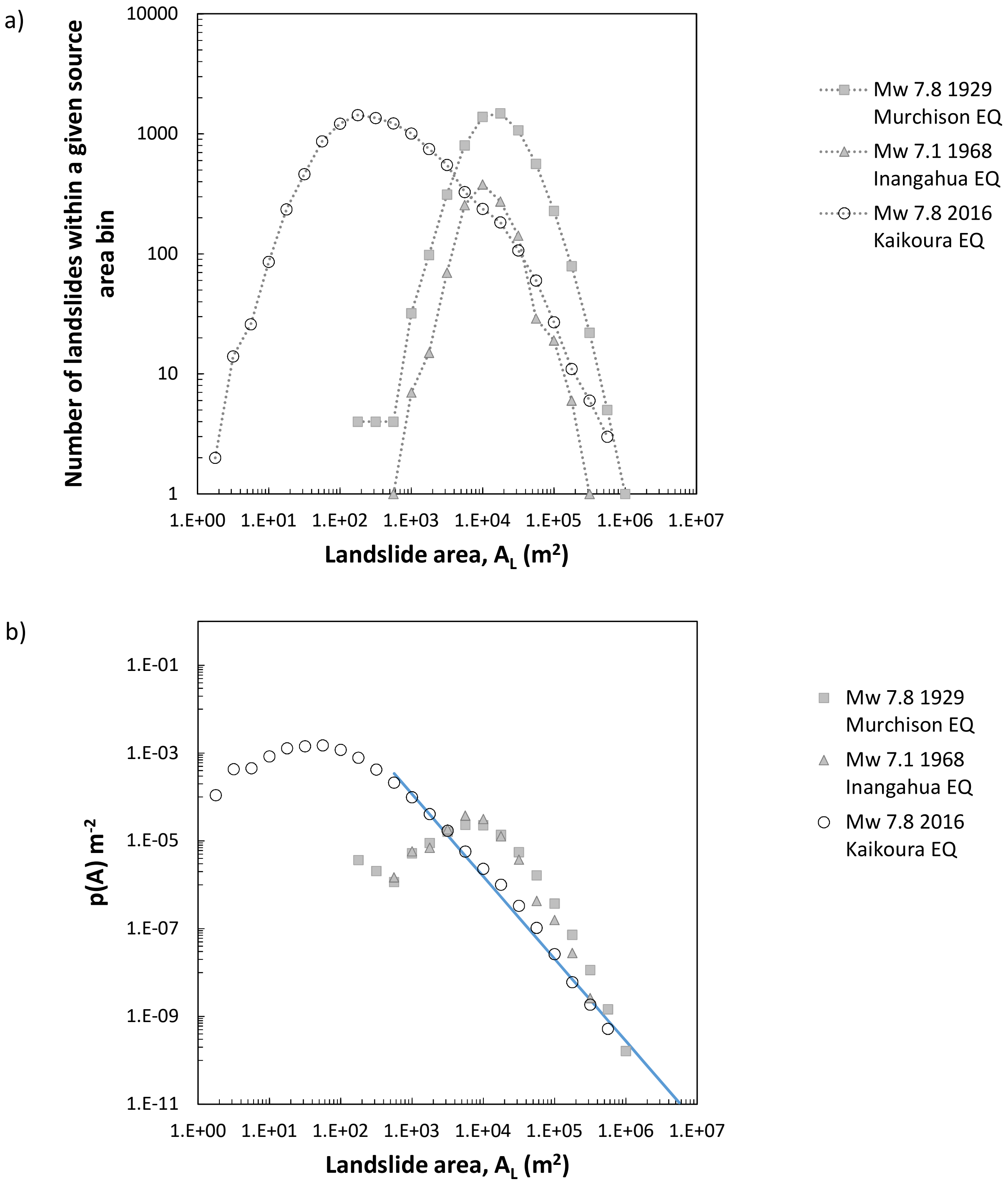

a)
Number of landslides within a given source area bin
10000
1000
100
10
1
1.E+00
1.E+01
1.E+02
1.E+03
1.E+04
1.E+05
1.E+06
1.E+07
Landslide area, A_L (m^2)
Mw 7.8 1929 Murchison EQ
Mw 7.1 1968 Inangahua EQ
Mw 7.8 2016 Kaikoura EQ
b)
p(A) m^{-2}
1.E-01
1.E-03
1.E-05
1.E-07
1.E-09
1.E-11
1.E+00
1.E+01
1.E+02
1.E+03
1.E+04
1.E+05
1.E+06
1.E+07
Landslide area, A_L (m^2)
Mw 7.8 1929 Murchison EQ
Mw 7.1 1968 Inangahua EQ
Mw 7.8 2016 Kaikoura EQ

Figure 3

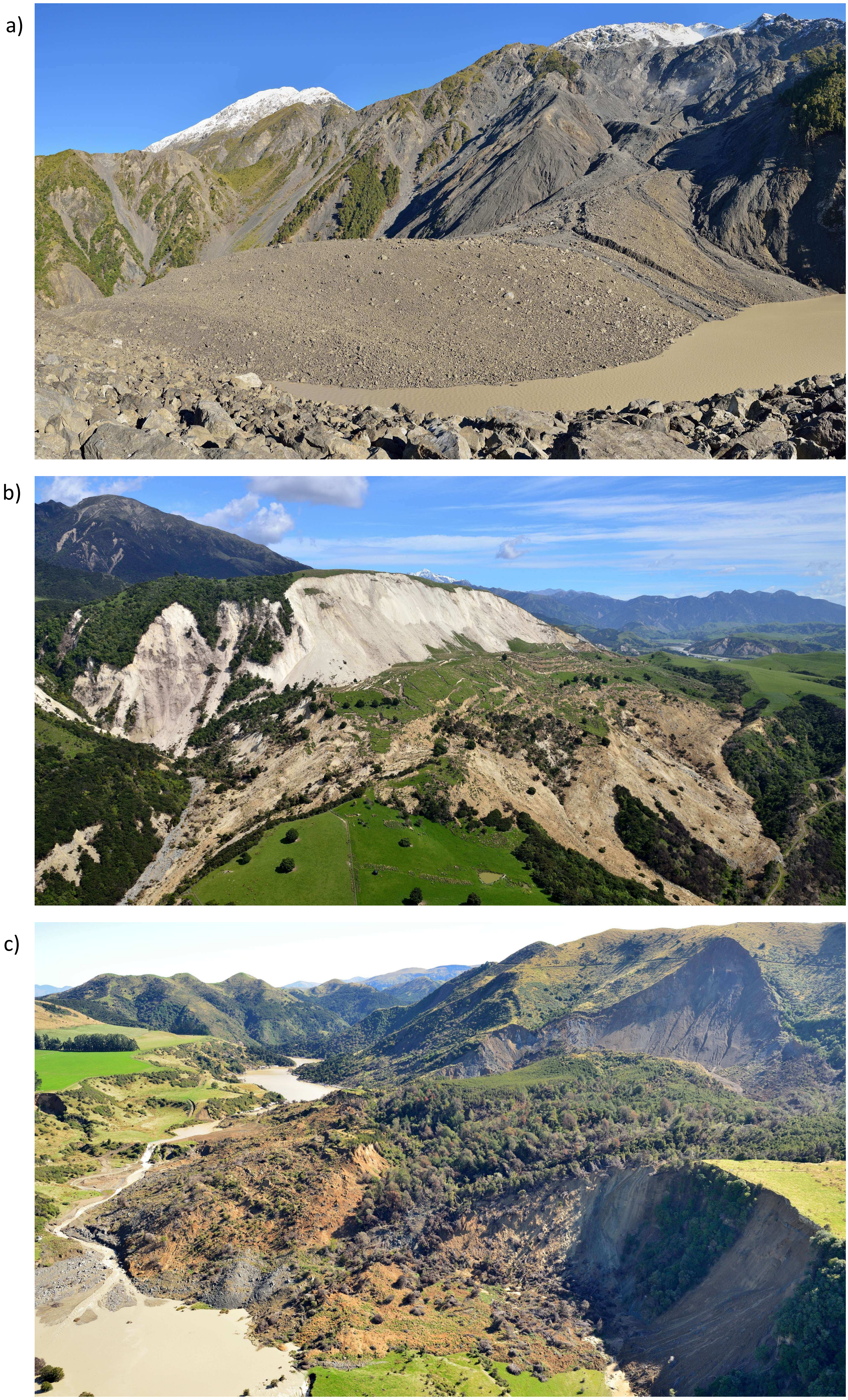


Figure 4

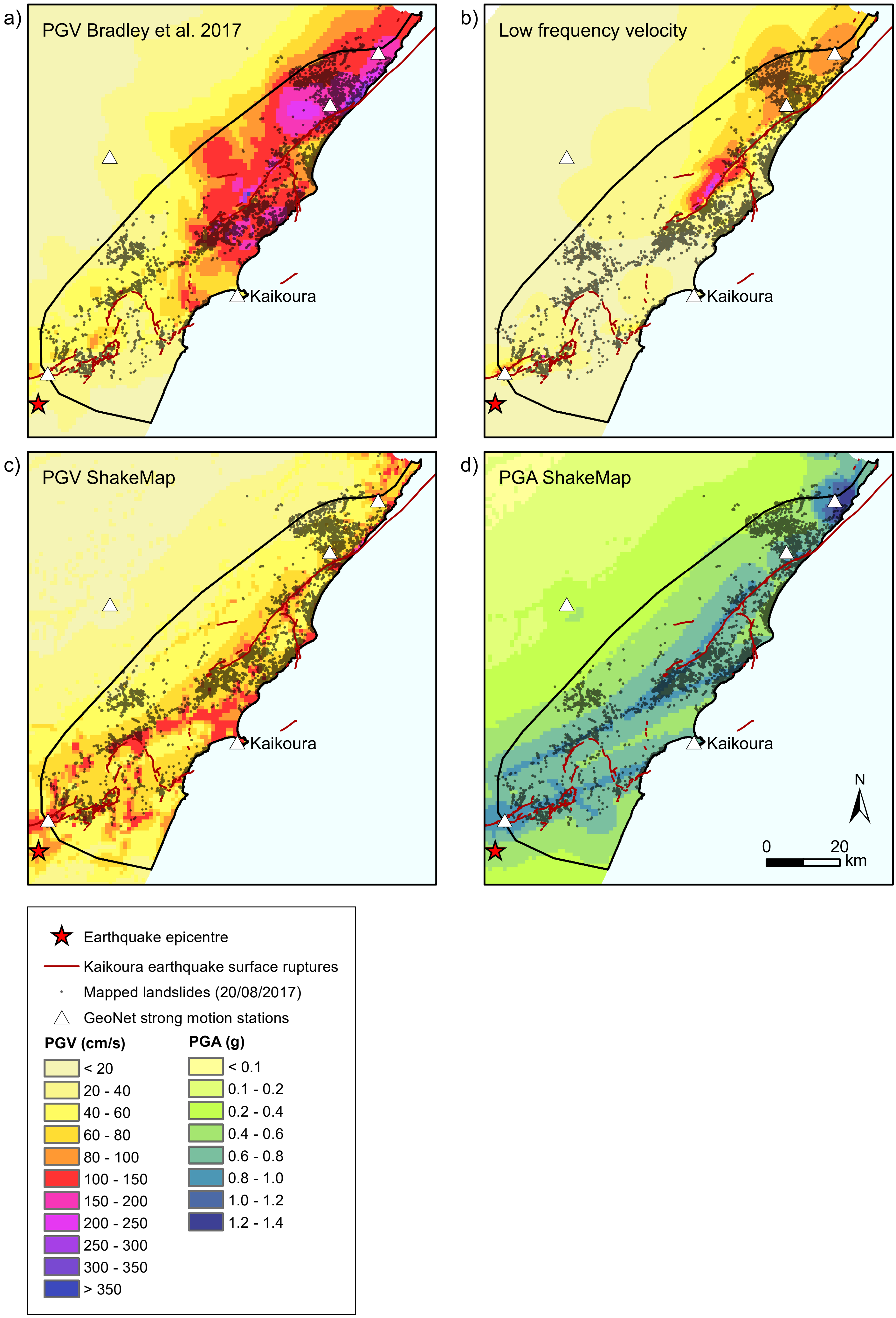

a)
PGV Bradley et al. 2017
Kaikoura
b)
Low frequency velocity
Kaikoura
c)
PGV ShakeMap
Kaikoura
d)
PGA ShakeMap
Kaikoura
N
0
20
km
Earthquake epicentre
Kaikoura earthquake surface ruptures
Mapped landslides (20/08/2017)
GeoNet strong motion stations
PGV (cm/s)
< 20
20 - 40
40 - 60
60 - 80
80 - 100
100 - 150
150 - 200
200 - 250
250 - 300
300 - 350
> 350
PGA (g)
< 0.1
0.1 - 0.2
0.2 - 0.4
0.4 - 0.6
0.6 - 0.8
0.8 - 1.0
1.0 - 1.2
1.2 - 1.4

Figure 5

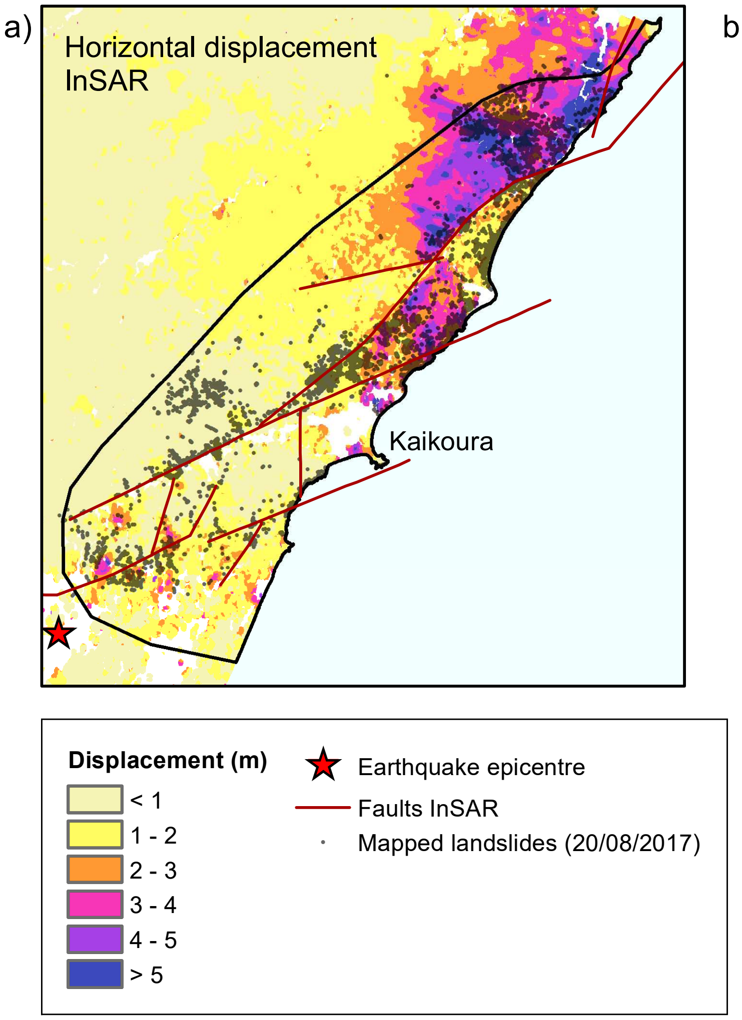

a)
Horizontal displacement
InSAR
Kaikoura
b)
Vertical displacement
InSAR
Kaikoura
N
0
20
km
Displacement (m)
< 1
1 - 2
2 - 3
3 - 4
4 - 5
> 5
Earthquake epicentre
Faults InSAR
Mapped landslides (20/08/2017)

Figure 6

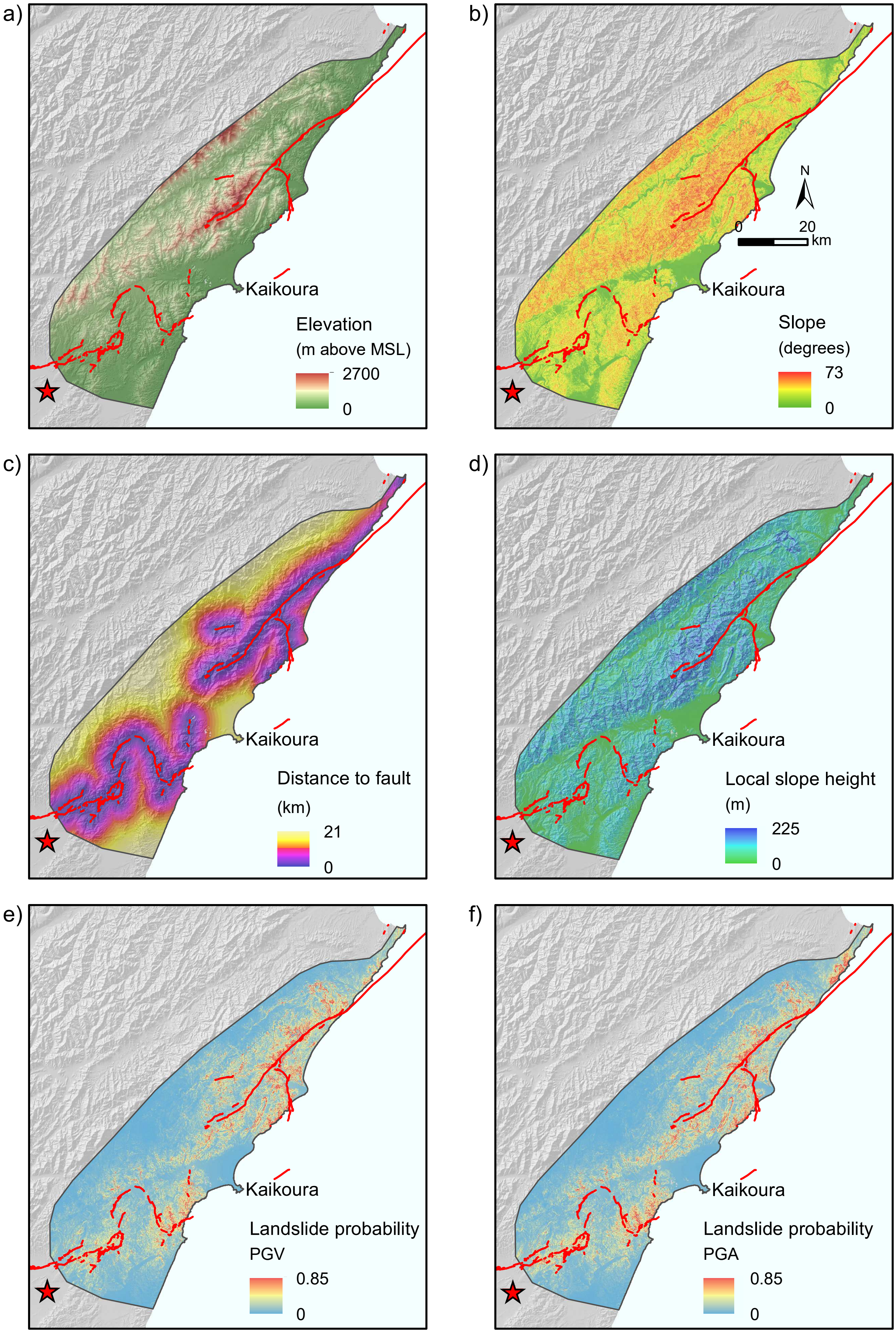
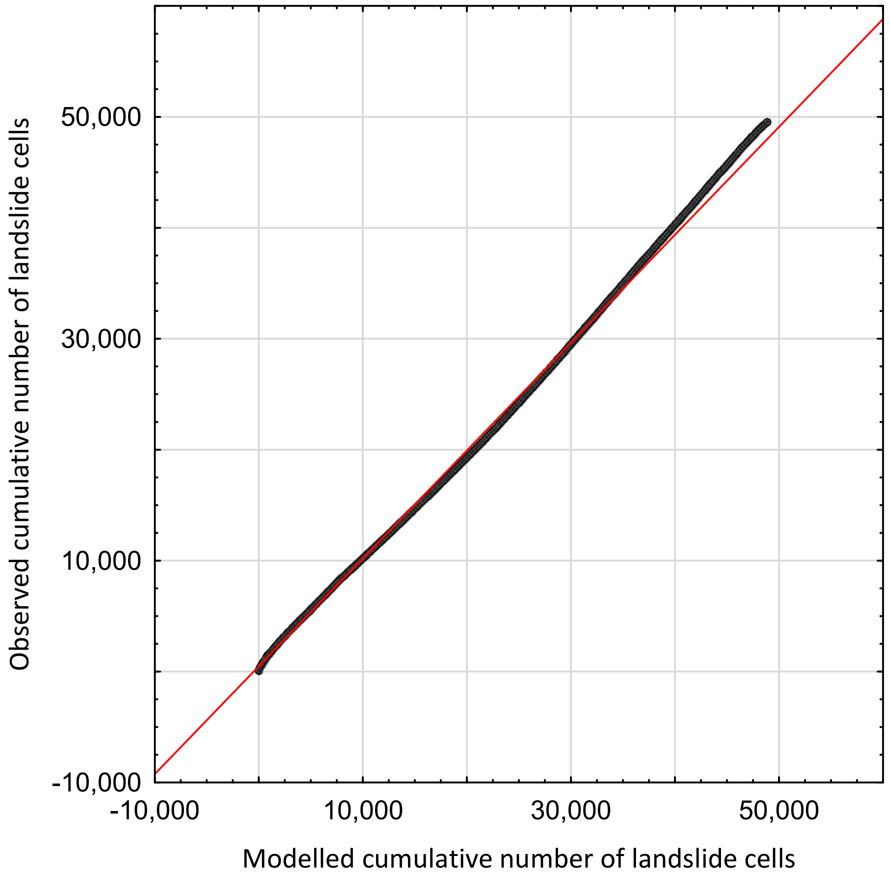
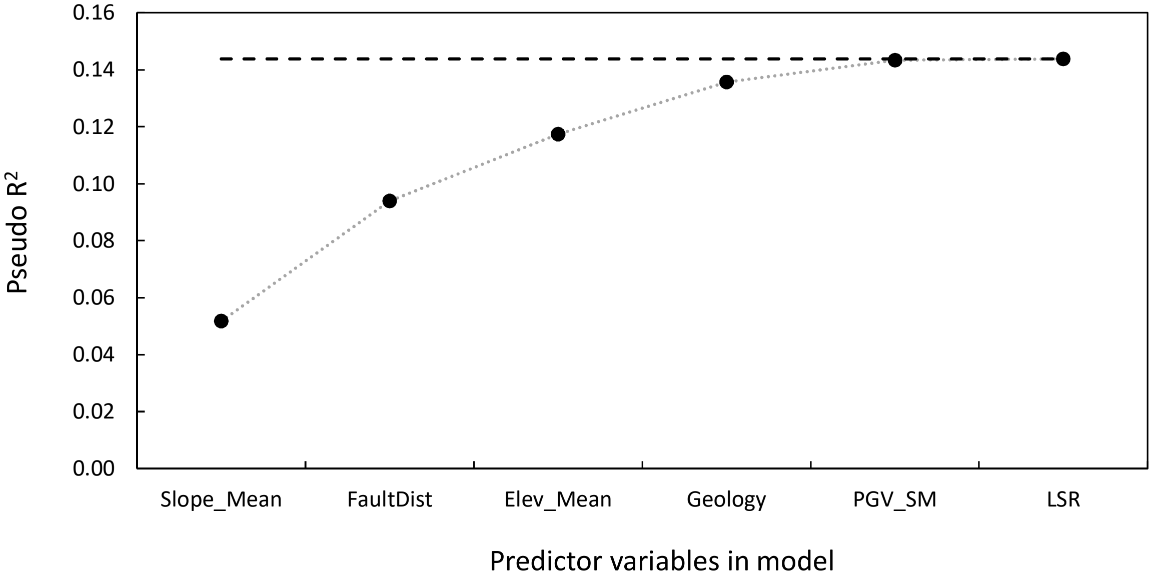

a)
Kaikoura
Elevation
(m above MSL)
2700
0
b)
N
0
20
km
Kaikoura
Slope
(degrees)
73
0
c)
Kaikoura
Distance to fault
(km)
21
0
d)
Kaikoura
Local slope height
(m)
225
0
e)
Kaikoura
Landslide probability
PGV
0.85
0
f)
Kaikoura
Landslide probability
PGA
0.85
0

Figure 7


50,000
30,000
10,000
-10,000
-10,000
10,000
30,000
50,000
Observed cumulative number of landslide cells
Modelled cumulative number of landslide cells


0.16
0.14
0.12
0.10
0.08
0.06
0.04
0.02
0.00
Pseudo R²
Slope_Mean
FaultDist
Elev_Mean
Geology
PGV_SM
LSR
Predictor variables in model

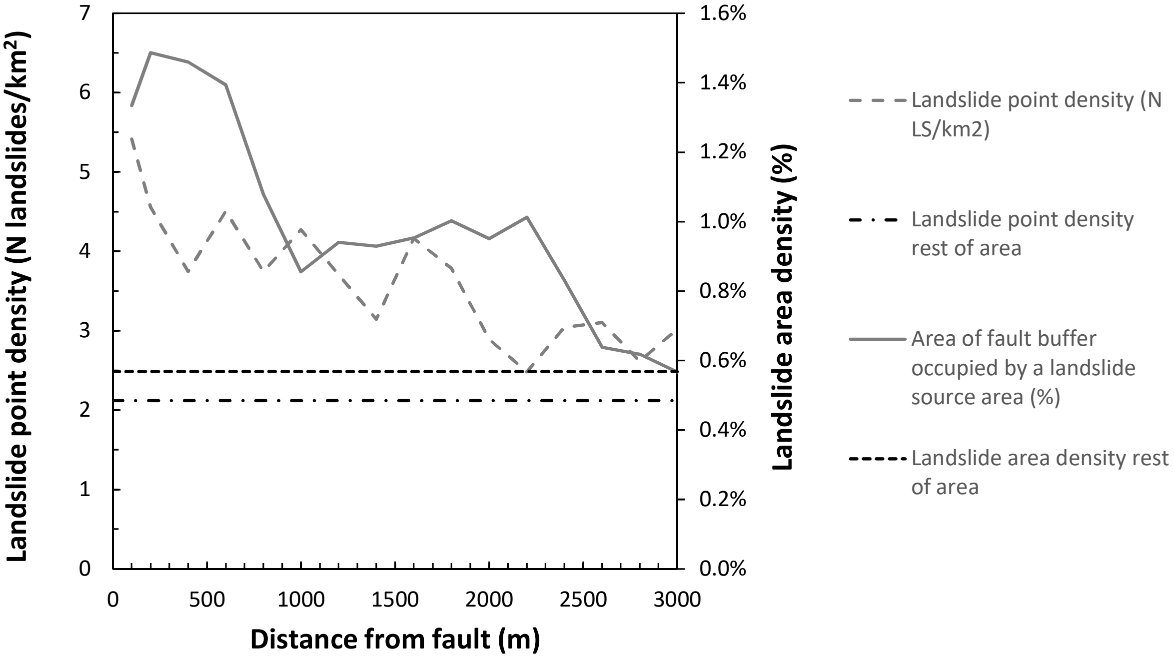

Landslide point density (N landslides/km²)
7
6
5
4
3
2
1
0
Landslide area density (%)
1.6%
1.4%
1.2%
1.0%
0.8%
0.6%
0.4%
0.2%
0.0%
0
500
1000
1500
2000
2500
3000
Distance from fault (m)
Landslide point density (N LS/km2)
Landslide point density rest of area
Area of fault buffer occupied by a landslide source area (%)
Landslide area density rest of area

Figure 10a & 10b

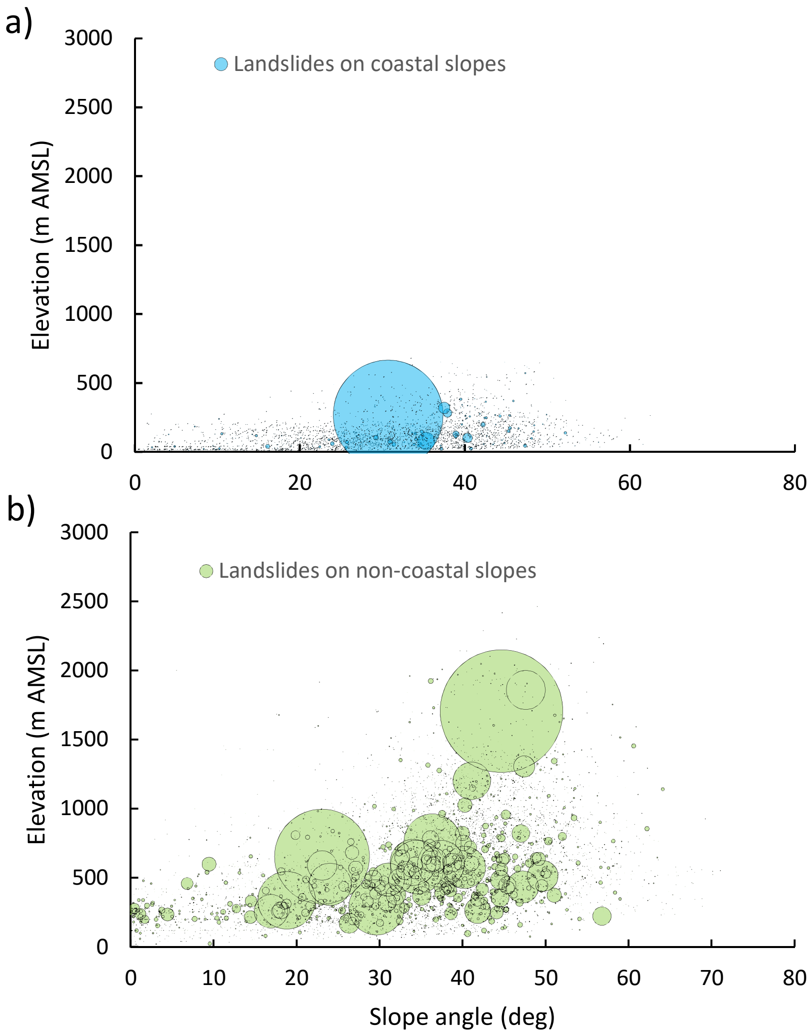
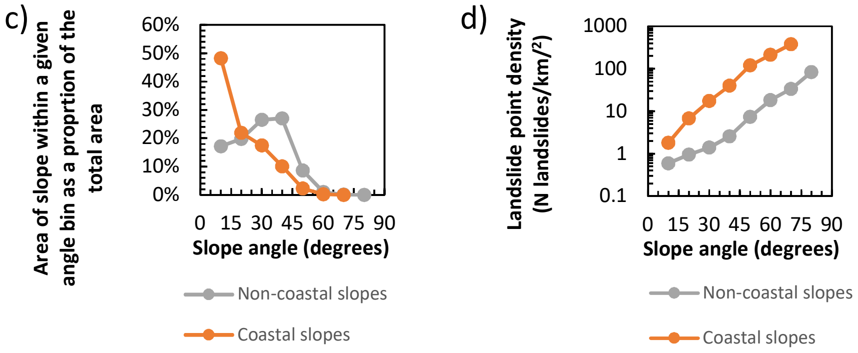


Figure 10c & 10d



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Authors: Chris Massey____

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