# THE IMPORTANCE OF SURFACE WAVES IN STRONG GROUND-MOTION

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ABSTRACT

Surface waves, when present in strong ground-motion, may contribute more to the Modified Mercalli intensity and to damage than does the S wave, even if the S wave contains higher accelerations. This gives a qualitative explanation for the observed regional variation in the felt effects of earthquakes in New Zealand, by assuming a regional variation in the predominant focal depth and hence in the amplitudes of the Love and Rayleigh waves that are excited.

## 1. FELT INTENSITIES IN NEW ZEALAND

An empirical formula for calculating the expected Modified Mercalli intensity at any particular site, given the magnitude and location of the earthquake, has been developed by Smith [1]. Three regions A, B and C are identified (Figure 1), to which different formulae for the decay of intensity with epicentral distance are applied. The aim of the present paper is to provide physical justification for the division into three source regions, and to examine the engineering implications.

The intensity function for region A is shown in Figure 2. The abscissa is the epicentral distance in kilometres, modified to take into account the generally elliptical nature of isoseismals in New Zealand. The geophysical problem is that these intensities are less than observed in region B, and greater than in region C, for earthquakes of equal magnitude. It is the ground motion, as recorded by a Wood-Anderson seismograph, from which the magnitude is calculated. What properties of the three regions, to which the magnitude is insensitive, cause the widely differing felt patterns?

## 2. SURFACE WAVES: A SOLUTION TO THE PROBLEM

Separate microearthquake studies in region A (M. E. Reyners, R. Robinson, pers. Comm.) suggest that most of the earthquake activity there is at depths greater than 20 km, whereas activity in region B seems to be much shallow (e.g. Robinson and Arabasz [2], I Calhaem, pers. comm.). The geometry does have an effect on the ground motion, in that the deeper earthquake is further from the observer so the intensity is less, but the fact that earthquakes in region C, in which the intensities are the least, appear to be predominantly in the upper crust suggests that this is not the whole answer. This is further confirmed by the occurrance of intensities that were much higher than expected from an earthquake in the Waikato area [3], and which could not be explained by a simple argument involving a very shallow source.

The excitation of short-period surface waves by shallow earthquakes offers a much better solution. P and S waves are body waves in that they propagate radially from the focus of the earthquake. The ratio of their velocities is a function of Poisson's ratio for the medium. Surface waves, on the other hand, propagate only in the vicinity of the free surface. The depth of penetration decreases with the period of the signal. They are in fact a resonance phenomenon: the phase velocity is a function of the period and the crustal structure. The velocity is less than that of the S wave, but the dependence on period implies that the propagation is dispersive. Long-period energy travels faster than short-period energy. The particle motion is horizontal for Love waves, transverse to the direction of propagation. Rayleigh waves have elliptical particle motion, retrograde at the surface, that is in the vertical plane containing the direction of propagation.

The range of frequencies causing the effects described by the Modified Mercalli scale is not accurately known, but it certainly includes 2 Hz, adopted here as a representative value. Rayleigh waves of this frequency propagate in only the top few kilometres of the earth's crust, as Figure 3 shows. For P and S velocities of 5.5 and 3.3 km $s^{-1}$ respectively, the displacement drops to less than one per cent out of its peak value by a depth of 5 km. The addition of thin layers of low velocity and density at the surface makes little change to the basic shape of the eigenfunction. The propagation of Love waves is similar except that they require a variation of velocity with depth, and so cannot propagate in a uniform halfspace.

A consequence of this propagation within only the top few kilometres is that these short-period surface waves are excited by earthquakes at focal depths of 12 km or more only because strain is relieved throughout a volume that extends to the critical depth range (0-5 km). If an earthquake were a point source, deeper than 5 km, it would not excite short period surface waves. The amplitude of the surface waves will depend on the amount of strain that is relieved in the top few kilometres. This means that the intensities from an

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earthquake in the lower crust will be less than from a shallower shock, because less strain is relieved near the surface.

The excitation of short-period surface waves will have little effect on the magnitude because the low velocity material of the uppermost layers of the crust attenuates the motion very strongly. Magnitude can be determined only from those seismographs which are so far away that they do not overload, and the surface waves responsible for the high intensities near the epicentre are likely to have been so attenuated as to be insignificant at that distance.

This explains the difference between regions A and B, in terms of the characteristic depth of earthquakes in each region, but the problem of Fiordland (region C) remains. Earthquakes there are predominantly in the upper crust, but intensities are very low. Gibowicz and Hatherton[(4)] classify Fiordland as one of the areas where the earthquakes are of very small source dimension. The rock is very competent, so only a small volume is required to store the necessary strain. An explanation for the very low intensities in Fiordland could therefore be that, although the foci are shallow (12 km or so), the volume from which strain is relieved may still not reach the critical depth range (0-5 km). Felt effects would thus be largely due to body waves, and be consequently much less than for earthquakes with more extended sources. There will also be a spectral effect in that the small source size implies a paucity of low frequencies in the signal, and this is likely to be reflected in reduced intensities. Gibowicz and Hatherton have found other regional variations in earthquake source mechanism, but these do not appear to have any marked expression in the intensities. The Fiordland problem can hardly be regarded as solved. It is of interest to note, however, that the Opunaki earthquake[(5)] had isoseismals similar to earthquakes in Fiordland. The source dimension was also very small.

The increase in ground motion, due to the presence of surface waves, can be estimated from the formulae for regions B and C[(1)]. At a distance of 50 km from an earthquake of magnitude 7, the intensity in region B is 2.5 units greater than in region C. The usual assumption that acceleration and velocity double with each intensity step thus implies that the presence of surface waves in region B increases the ground motion more than fivefold. This assumes that body wave radiation is identical in the two areas, which is most unlikely in view of the different source mechanisms, so the factor of five should be regarded as an upper limit. The increase from region A to region B for the same magnitude and distance is just less than twofold.

It is clear then that an adequate formula for intensity as a function of magnitude and distance should contain body-wave and surface-wave terms, with their different attenuation rates. Howell and Schultz[(6)] suggest such a dual formula, although they make no mention of surface waves. Formulae couched in terms of the attenuation of body waves only are clearly inadequate.

## 3. THE EPICENTRAL REGION

At very short epicentral distances the situation is complicated by near-field effects. The elastic wave field there cannot be separated into P and S waves because of the presence of terms with higher inverse powers of distance which are insignificant elsewhere. Nor can surface waves be distinguished close to the epicentre, because they require a distance of the order of the focal depth to build up in amplitude. The near field is an area about which little information is available. Unfortunately, it is also the region of most interest to engineers. The intensity formula[(1)] extrapolates intensities back to short distances, with plausible assumptions about the source dimensions of large earthquakes. Calculated intensities are therefore much less reliable close to the epicentre than at distances beyond about 30 km, and inferences about the character of the motion very difficult to make.

## 4. IMPLICATIONS FOR EARTHQUAKE ENGINEERING

The argument that intensities depend strongly on surface waves is supported by the observation that the peak acceleration, commonly reported to be contained in the S wave (e.g. Newmark and Rosenblueth[(7)]) is not a good measure of the extent of damage. This has been reported many times in the form of a lack of correlation between intensity and peak acceleration (e.g. Trifunac and Brady[(8)]). And, intensity, despite its limitations of being discrete and descriptive, is nevertheless a scale that was devised to categorize damage.

Consider energy of frequency much greater than 2 Hz. Rayleigh waves will be attenuated very strongly indeed, propagating as they do in the very top layers, but S waves travel in the basement rock until they reach the site in question, so are attenuated less severely. If surface waves are indeed responsible for much of the damage, it follows that while the high frequency energy will affect the acceleration significantly it will not contribute greatly to the damage. The duration of shaking is also important (e.g. Trifunac and Westermo[(9)]). It affects the intensity but not the peak acceleration, and is increased by the presence of surface waves, which travel more slowly than the S wave and are dispersive.

An important conclusion, therefore, is that strong ground motion can be expected to contain, perhaps even be dominated by, the transverse horizontal motion of the Love wave and the retrograde elliptical motion of the Rayleigh wave. But the situation will, of course, be complicated by local geological and topographical irregularities.

A recent text on earthquake engineering (10) does acknowledge that surface waves may be important, but it is interesting that the relevant chapter was written by a seismologist (J. N. Brune). The importance of surface waves appears not to have been fully appreciated by the engineering community.

## ACKNOWLEDGEMENTS

I am grateful for the help of

Drs. R. Robinson and R. D. Adams and Mr. G. A. Eiby, both in preparing the manuscript and in clarifying the concepts presented in it.

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Paper received 3 November, 1977.

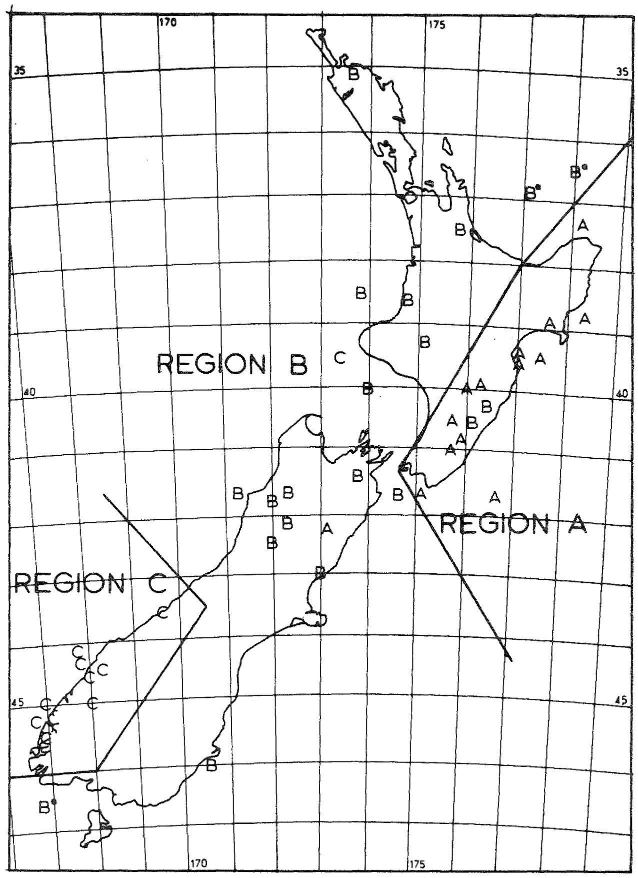


FIGURE 1: DEFINITION OF REGIONS A, B AND C FROM EARTHQUAKE ISOSEISMAL PATTERNS (FROM SMITH, 1976).

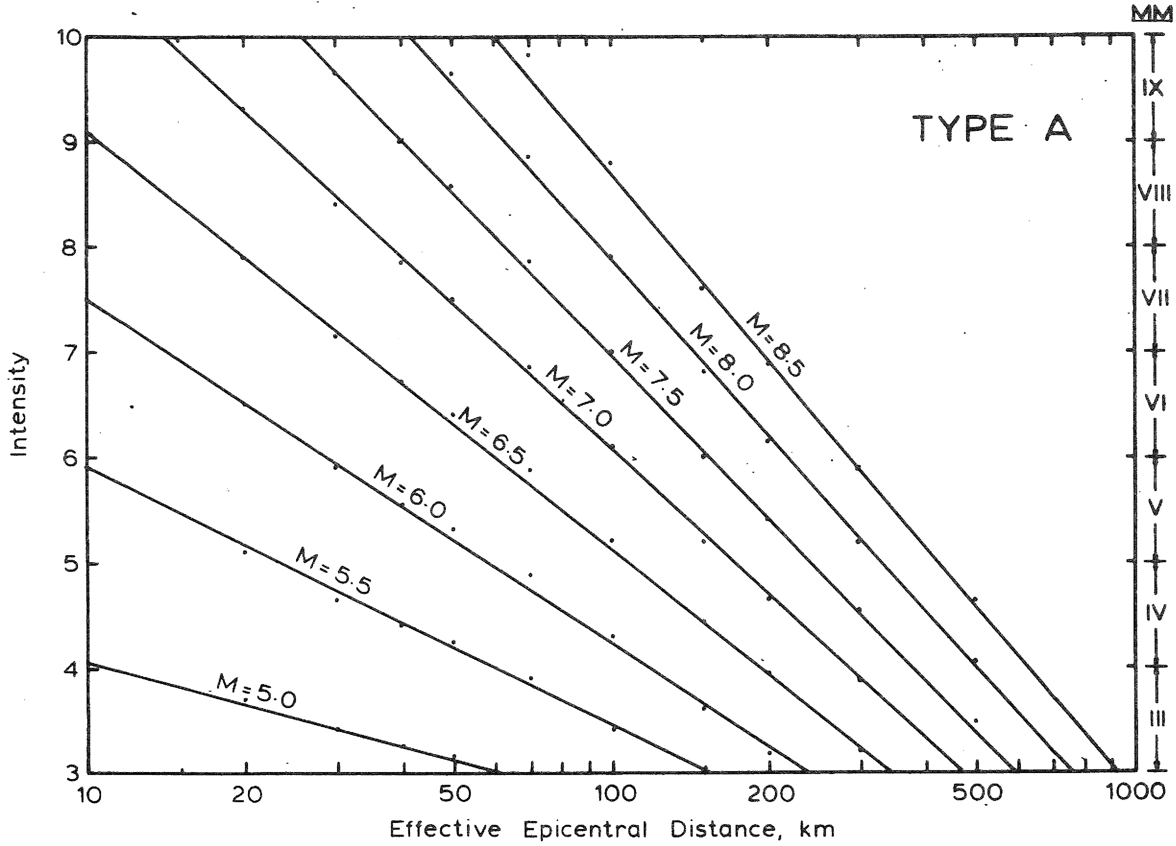


FIGURE 2: INTENSITY AS A FUNCTION OF MAGNITUDE AND EPICENTRAL DISTANCE, REGION A (FROM SMITH, 1976)

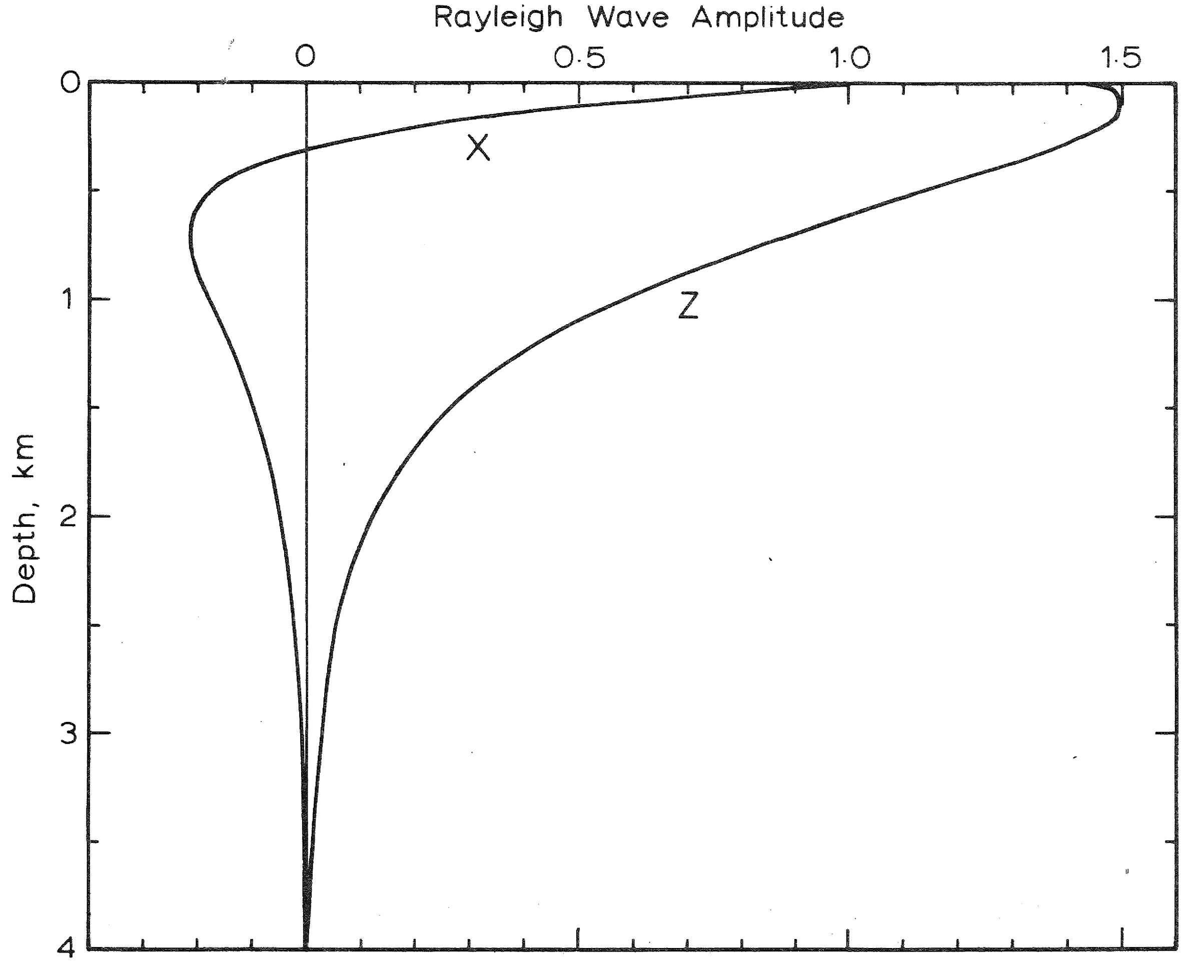


FIGURE 3: HORIZONTAL (x) AND VERTICAL (z) DISPLACEMENTS FOR THE FUNDAMENTAL RAYLEIGH WAVE IN A UNIFORM HALF-SPACE WHERE P AND S VELOCITIES ARE 5.5 AND 3.3 km.s-1 RESPECTIVELY, AT A FREQUENCY OF 2 Hz. THE TWO COMPONENTS ARE 90$^{o}$ OUT OF PHASE.