

A PROCEDURE FOR MODELLING NEAR-FIELD EARTHQUAKE INTENSITIES

W.D. Smith¹

ABSTRACT

Close to the rupture surfaces of large earthquakes, the pattern of intensity is expected, on physical grounds, to reflect the fault geometry. But there are usually not enough observational data to constrain isoseismals at such short distances. In order to obtain a plausible model to sustain hazard estimation exercises, a simple procedure for evaluating intensities is presented. It involves using a point-source attenuation function to calculate the contribution to the ground motion due to an element of the source, then integrating along the fault trace or, for dipping faults, over the entire rupture surface. In the far field, the intensity so derived is exactly equivalent to that obtained if the entire rupture is represented by a point source.

1. INTRODUCTION

There is a practical need to estimate likely intensities near the fault rupture of large earthquakes. Insurance studies and hazard planning exercises, for instance, demand such estimates. But a limiting factor in the development of an attenuation formula for strong ground motion is the lack of data at short distances. It is precisely at these distances that the ground motion is greatest, and where estimation of the likely effects is therefore most important.

The companion paper [Smith 1995, hereinafter referred to as Paper 1], examines the available intensity data for New Zealand earthquakes, and draws the conclusion that there are inadequate data to constrain an attenuation model in the near field, i.e. at distances short enough that the fault geometry is important. The attenuation function developed in that paper therefore considers the earthquake as a point source. Several classes of event are recognised: upper crust, lower crust, volcanic region, Fiordland region, and deep earthquakes in the Main Seismic Region and Fiordland. Intermediate depth events are also treated. The data for events deeper than about 40 km were extremely scattered, with depths and magnitudes very unreliable. For this reason, no attempt at a substantial revision of an earlier study [Smith, 1978] is made for these events. But for crustal events, an attenuation function is developed which recognises not only the depth and regional classes but also regional variation of attenuation, expressed as the fact that there is some consistency in the observed ellipticity of isoseismals.

None of the available isoseismal data appear to be constrained by fault geometries. That is, there are no examples of extremely elongated isoseismals, oriented along fault traces. The New Zealand intensity data available at short distances are so sparse that they do not normally permit the drawing of a

near-field isoseismal. The measured ellipticity of isoseismals, in any given earthquake, is remarkably constant from the highest intensities to the lowest (although it is often impossible to draw complete isoseismals at low intensity, because of the lack of observations out to sea). This ellipticity of isoseismals is therefore assumed to be evidence of regional propagation effects and not source geometry.

The present study therefore assumes that as a practical measure the far-field attenuation function may be extrapolated back to the epicentre, as a representation of the ground motion that would result from a point source. This extrapolation was done in Paper 1 by constraining the procedure of fitting the isoseismal data by such near-field point intensity observations as were available. But there were not enough data to provide a reliable near-field formula.

A procedure is then developed for integrating over a distributed source, using the point-source function. In the far-field, where fault geometry is clearly unimportant, this procedure results in an intensity which is exactly that predicted by the point-source function.

2. FORMULATION

A number of authors have proposed ways to compute intensity (or other measures of ground motion) near faults. Boore and Joyner [1982] review various measures of distance that are commonly used in regressions. They tend to favour the shortest distance to the fault. The distance to the nearest part of the rupture surface is another measure that has been proposed. These techniques are mostly not based on any consideration of the effects of the whole rupture at the observation point, but just represent a simple way of obtaining a plausible result.

It is not physically realistic to model a large earthquake as a group of small earthquakes, distributed along the fault. The problem here is that small earthquakes are of short duration and

¹ *Institute of Geological & Nuclear Sciences, Wellington (Fellow)*

high frequency, so even when a group are combined they do not adequately represent the long-period energy and the long duration of a large earthquake.

Ohno et al [1993] present a formulation in terms of an Effective Hypocentral Distance (EHD). This parameter is evaluated in terms of the earthquake source (geometry and moment) and a model of the S-wave attenuation with distance. They find this formulation useful for predicting spectra and maximum accelerations. For such an approximate parameter as intensity, however, it seems that such detailed wave modelling is not warranted, and a simpler approach is justified. The following development is not specific to any geometry, source or attenuation function.

(i) Vertical fault

The attenuation function developed in Paper 1 estimates the intensity at any given location, due to a (point source) earthquake elsewhere. Consider an earthquake caused by a fault rupture of length L , and assume that the displacement is uniform along the entire length L . If the earthquake source (magnitude m) were confined to element dl the intensity at point P could be determined by the point-source formula; call this intensity I (see Figure 1).

Note that the attenuation function developed in Paper 1 does not involve a simple dependence on epicentral distance. It includes all the effects of regional attenuation and focal depth. Azimuthal dependence is expressed in the form of elliptical isoseismals. The major axis is oriented N40E in some parts of the country, N50W in others, and elsewhere where the data are too few, isoseismals are circular. For deep earthquakes the ellipticity is extreme and the centre of the isoseismal pattern is offset from the epicentre, apparently because of waveguide propagation up the subduction zone. But the following development does not rely on the particular attenuation function used. One could choose that of Dowrick [1992], for instance,

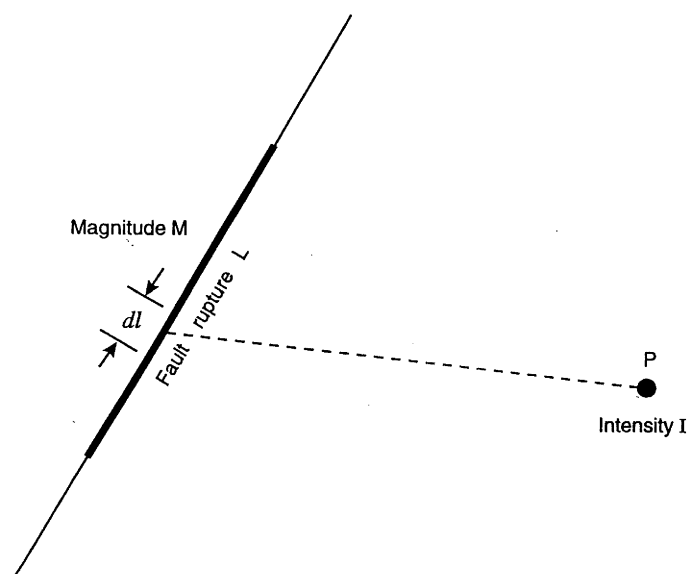


FIGURE 1. Schematic of the proposed integration scheme for intensity, for a fault rupture of length L . An earthquake of magnitude m located at the fault element of length dl would cause intensity I at point P , according to the point-source formula which may be isotropic or incorporate azimuthal variation of attenuation.

or any other. This paper addresses only the implementation of the point-source function at short distances.

Ground motion is often considered to approximately double for each increase of one unit in intensity. Willmore [1979] gives conversions between intensity and displacement, velocity and acceleration which, though valid for different ranges of frequency, all double for each intensity step. Also, wave energy is related to the square of the amplitude. Write 2^{2I} as a measure of ground motion, therefore, and scale this by the length of the fault so that the contribution dx to energy at P , due to the fault element dl is

$$dx = \frac{2^{2I} dl}{L} \quad (1)$$

Now integrate along the entire length L of the fault, and take the logarithm to base 2 to derive the intensity:

$$x = \int dx = \int \frac{2^{2I} dl}{L} \quad (2)$$

$$\hat{I} = 0.5 \log_2 x \quad (3)$$

If P is sufficiently distant from the fault that I is constant for all fault elements dl , this reduces to $\hat{I} = I$. Close to the fault, however, the integral expresses the overall effect at P of the whole fault length, taking account of the fact that some parts of the fault may be much closer to P than others.

Note that in the computation of I , all the effects of regional attenuation, focal depth, etc, are included. The procedure could be applied using simpler models of attenuation than is developed in Paper 1 (see Section 3 below). But the important point is that it does fulfil the requirements that there be a physically plausible basis for the computation, and that it be exactly equivalent to the point-source function in the far field.

(ii) Dipping faults

It is a simple procedure to extend the above formula to a dipping fault, whatever the angle of dip - even a shallowly dipping fault surface such as a subduction zone. The line integral in (2) above becomes a surface integral, over the entire rupture surface of the fault, and it is scaled by fault area instead of fault length. The computation is straightforward, using the point-source function to compute I from each area element on the fault plane.

3. EXAMPLES

Figure 2 is a scenario for the Wellington Fault, with a magnitude 7.5 earthquake and a 75 km rupture. The broken lines are the isoseismals for a point source, and the solid lines those for the fault which is shown, modelled as a vertical fault with integration along the fault trace.

In Figure 2a (intensities estimated using the results of Paper 1) the attenuation function incorporates slight ellipticity with the major axis in the N50W direction. The effect of the fault is (i) to elongate the MM IX isoseismal along the strike of the fault; (ii) make similar but very slight changes to MM VIII and lower isoseismals; (iii) spread out the source so that the MM X isoseismal, which is predicted by the point-source model, actually disappears.

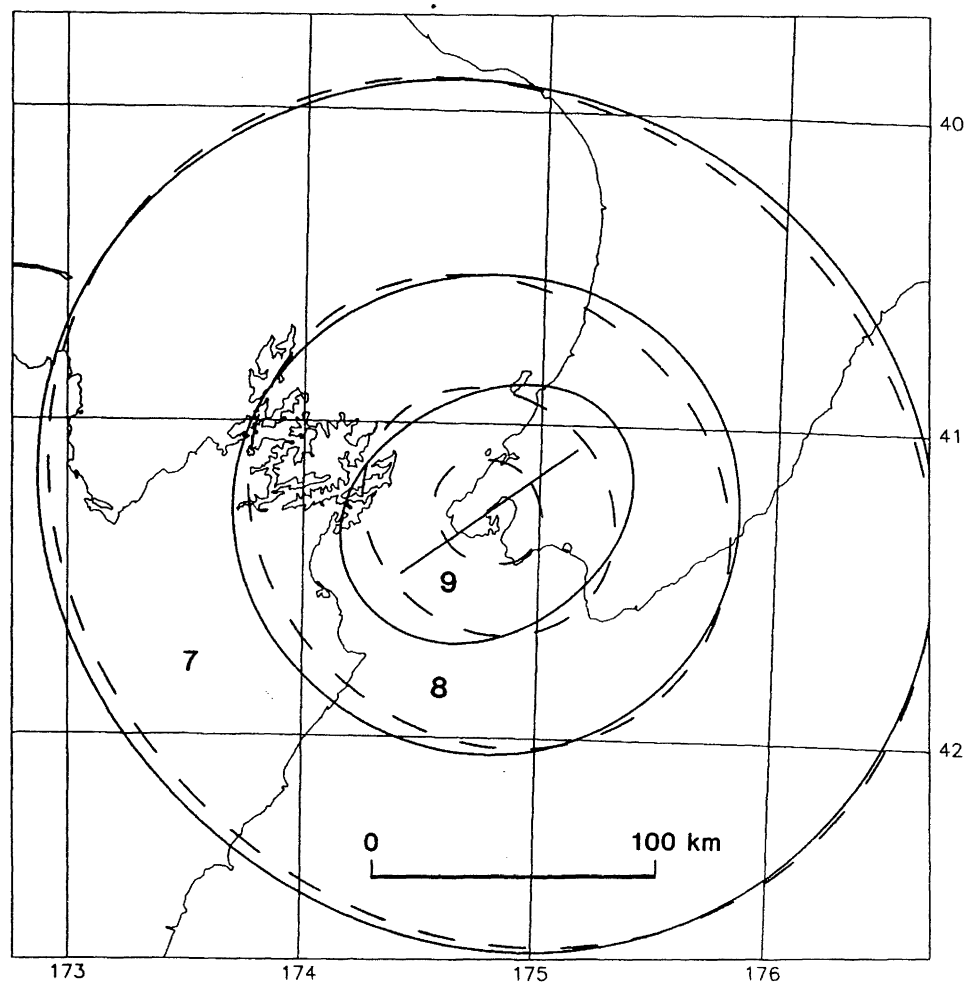


FIGURE 2a. Isoseismals modelled for an earthquake of magnitude 7.5 on the Wellington fault (rupture length 75 km). The dashed lines are the isoseismals predicted by the point-source model for an earthquake of the same magnitude, located at the centre of the fault. The point-source function is that from Paper 1, which incorporates slight ellipticity in this region, with major axis oriented N50W.

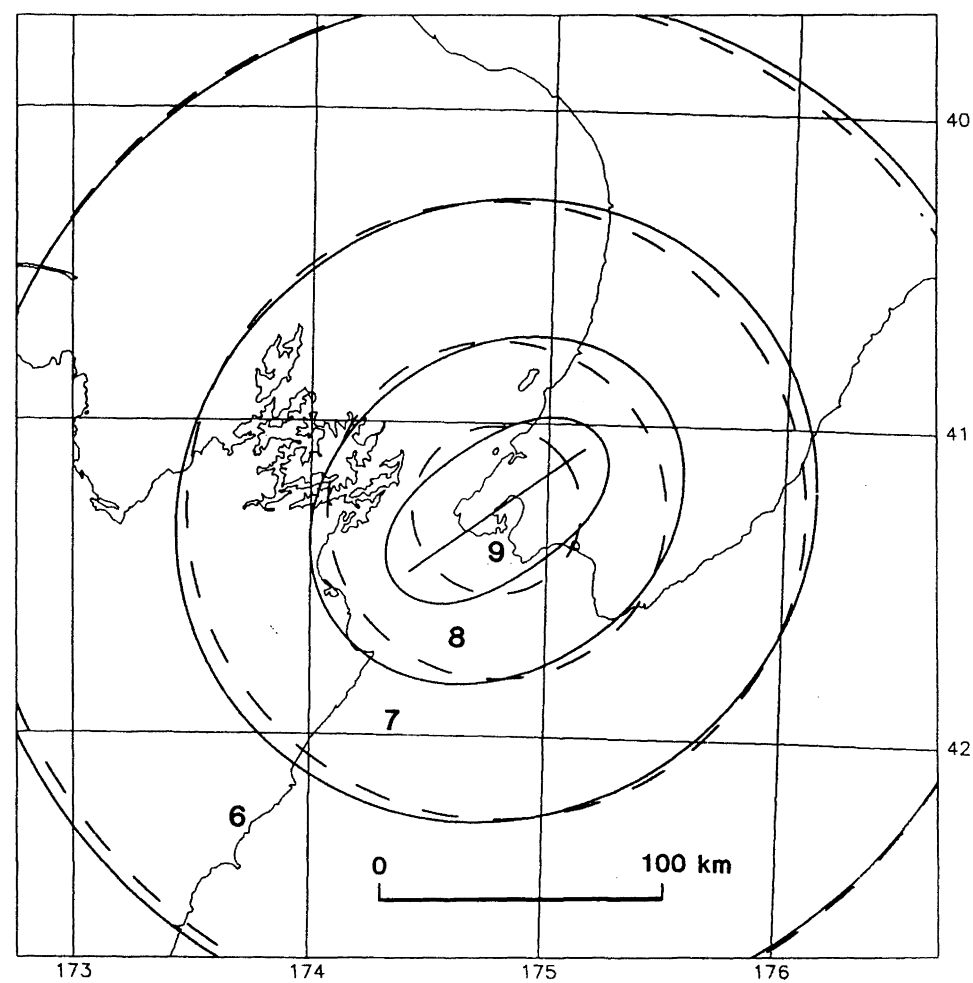


FIGURE 2b. Isoseismals as for Figure 2a, but using Dowrick's formula for a reverse fault with an effective depth of 10 km.

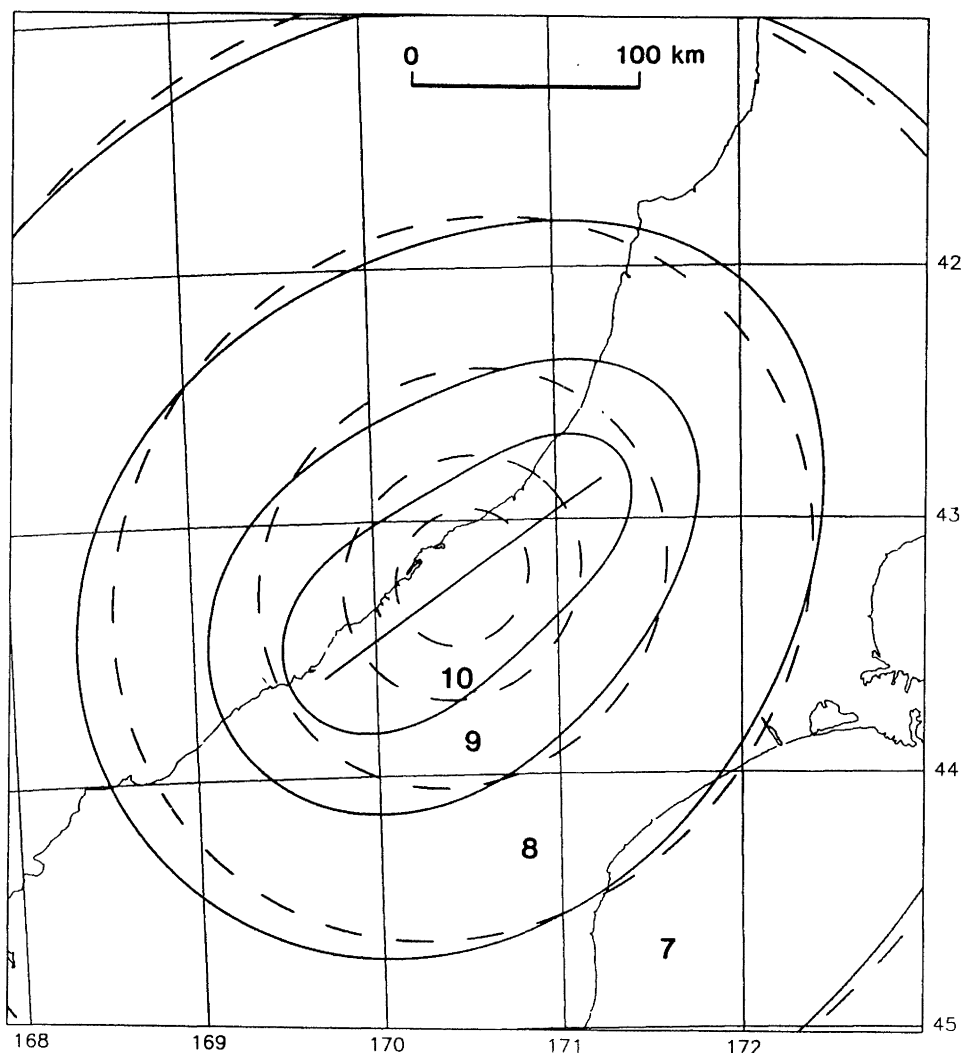


FIGURE 3. *Isoseismals modelled for an earthquake of magnitude 8.5 on the Alpine Fault (rupture length 150 km).*

In Figure 2b the attenuation function used is that of Dowrick [1992], for which the point-source isoseismals are circular. (His reverse fault formula has been used, with an effective depth of 10 km). In the same way as for Figure 2a, the MM X isoseismal predicted by the point-source model disappears, the MM IX isoseismal is elongated along the strike of the fault, and there are lesser effects at lower intensities.

It is clear that in both cases the integration procedure makes no significant difference to the computed intensities at distances beyond about two fault lengths.

Figure 3 shows computed intensities for a magnitude 8.5 earthquake with a 150 km rupture on the Alpine Fault. The broken lines represent the point-source model (Paper 1). The ellipticity model in this part of the country has a major axis in the N40E direction. This ellipticity is accentuated by the fault geometry, but not significantly so for intensities of MM VII and less. The "egg-shaped" nature of the inner isoseismals is a consequence of the ellipticity model: ellipticity is more severe at the north-eastern end of the fault trace than at the south-western end. The MM X isoseismal extends just past the ends of the fault, and the MM XI isoseismal predicted by the point-source function disappears.

Figure 4 shows a magnitude 8.5 earthquake on the subduction zone. The rupture surface is shown - it is 100 km long and dips northwestwards at 20 degrees to the horizontal, with a downdip length of 50 km, beginning at a depth of 20 km. This geometry approximates the modelling of Darby and Beanland [1992]. The integration has been performed over the full rupture surface. In this case it was not possible to include in the Figure the point-source results, for a source at the centre of the rupture, but the changes introduced by the integration procedure are given in Table 1. There is actually no significant change in the intensities in the N50W direction from the centre of the rupture (except for the disappearance of the MM XI isoseismal), but there is some increase in the N40E direction.

4. LIMITATIONS

This formulation relies totally on the extrapolation of far-field intensities back to the source, as an adequate representation of a point source. This paper does not address that issue. Any point-source function could be used. But the extension to distributed sources rests on some further assumptions which limit the reliability of the representation of near-source effects.

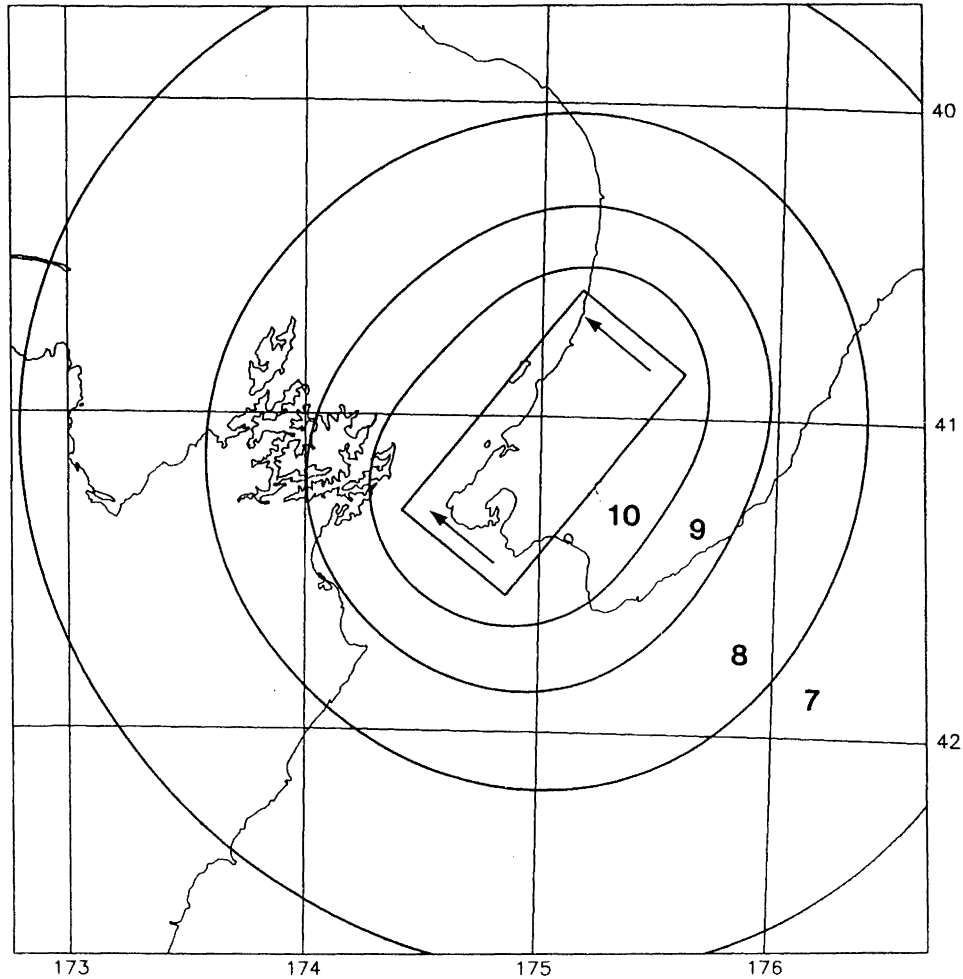


FIGURE 4. *Isoseismals modelled for an earthquake of magnitude 8.5 on the subduction zone. The rupture surface is 100 km long. It begins 20 km deep beneath the trace of the Wairarapa Fault and dips 20 degrees NW for 50 km. The surface is shown as a rectangle, with the arrows indicating the down-dip direction. The comparison of these isoseismals with those computed for a point source is given in Table 1.*

TABLE 1. *Changes to the isoseismal semi-axis along the strike of the rupture, with respect to a point-source function for an event at the centre of the rupture surface, when the integration procedure was implemented to calculate the intensities shown in Figure 4.*

Intensity	N40E semi-axis
MM VIII	8 km increase
MM IX	14 km increase
MM X	20 km increase
MM XI	22 km increase
MM XII	(disappears)

The main one is that all segments of the fault plane generate equal energy to contribute to the intensity at any given location. This is almost certainly not true, especially in the light of work such as that of Liu and Helmberger [1985] and others. Secondly, there are likely to be some rupture propagation effects, which radiate energy preferentially in particular directions. Thirdly, it must be remembered that microzoning problems are not addressed by this formulation: it models only average ground conditions.

Despite these limitations, this formulation does nevertheless represent a significant improvement over either the modelling of an earthquake as a point source, at locations close to the rupturing fault, or using a simple measure such as distance to the fault.

5. CONCLUSIONS

It is not difficult to extend a point-source attenuation function to account for distributed sources. Simple integration along the fault trace, or over the rupture surface, can be used to compute the resultant intensity from the whole source. The result differs from that determined from the point-source function only at distances less than about two fault lengths. At greater distances, the point-source function can be used without significant error.

ACKNOWLEDGMENTS

I am grateful to Graeme McVerry, Bill Stephenson and David Dowrick for their very helpful comments which identified a number of points that needed clarification. Graeme McVerry also drew my attention to the work by Ohno et al. [1993].

REFERENCES

- Boore, D.M. and W.B. Joyner. 1982. *The empirical prediction of ground motion*. Bull. Seism. Soc. Am. 72:S43-S60.
- Darby, D.J. and S. Beanland. 1992. *Possible source models for the 1855 Wairarapa earthquake*. J. Geophys. Res. 97(12):375-389.
- Dowrick D.J. 1992. *Attenuation of Modified Mercalli intensity in New Zealand earthquakes*. Earthquake Engineering and Structural Dynamics, 21:181-196.
- Liu, Hsui-Lin and D.V. Helmberger. 1985. *The 23:19 aftershock of the 15 October 1979 Imperial Valley earthquake: more evidence for an asperity*. Bull. Seism. Soc. Am. 75:689-708.
- Ohno, S., T. Ohta, T. Ikeura, M. Takemura. 1993. *Revision of attenuation formula considering the effect of fault size to evaluate strong motion spectra in the near field*. Report No. 94, Kajima Technical Research Institute, Tokyo.
- Smith, W.D. 1978. *Spatial Distribution of Felt Intensities for New Zealand Earthquakes*. N.Z. J. Geol. Geophys. 21:293-311.
- Smith, W.D. 1995. *A development in the modelling of far-field intensities for New Zealand earthquakes*. Bull. N.Z. Nat. Soc. Earthq. Eng. 28(3):196-217 (This issue).
- Willmore, P.L. (ed.) 1979. *Manual of Seismological Observatory Practice*, World Data Center A for Solid Earth Geophysics, NOAA, U.S. Dept of Commerce.