A DEVELOPMENT IN THE MODELLING OF FAR-FIELD INTENSITIES FOR NEW ZEALAND EARTHQUAKES

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ABSTRACT

With the availability of better estimates of magnitudes for large historical earthquakes, a 1978 study to model the Modified Mercalli intensities in large earthquakes in New Zealand has been revised. While instrumental measures of strong ground motion are obviously valuable, intensities will always be important because most of our large historical events predate the installation of accelerographs, and because intensity does seem to give a measure of the degree of damage.

Crustal earthquakes in the Main Seismic Region have been divided into two classes on the basis of known or apparent focal depth. Events in the volcanic region are treated separately. No revision is attempted for shallow earthquakes in Fiordland or for events at greater than crustal depths anywhere in the country, the data base being too poor. Detailed goodness-of-fit statistics are presented: they compare favourably with those for the 1978 model and also for another model developed recently. The model is for the far field, so is useful for regional estimation of hazard but not for fault-specific studies at short distances. A companion paper addresses near-source modelling.

Intensity is often assumed to be linearly related to the logarithm of ground motion parameters such as displacement or acceleration. This study shows that that assumption cannot hold over a range of distances. It may be possible to formulate a linear relationship at any particular distance, but the parameters of that relationship will change as distance is varied.

1. INTRODUCTION

One of the essential tools in earthquake hazard analysis is a model for the attenuation of strong ground motion with distance from the source. Actual records of strong ground motion in large New Zealand earthquakes are relatively few, so it is essential to be able to estimate the likely ground motion at any given site, from an earthquake for which the location and magnitude are known, in the absence of actual recordings at the site.

The strong motion data available from most of New Zealand's important earthquakes are in the form of intensities. While it is clearly desirable to have instrumental records of strong motion, because of the shortcomings of the intensity scale, the fact remains that most of our large earthquakes occurred before accelerographs were in operation. Many predated any recording instruments at all. Also, intensity does seem to be a good indicator of the degree of damage, often expressed as a damage ratio. It is therefore imperative that as much information as possible be extracted from the intensity database. This study restricts itself to intensity data.

An attenuation relationship was developed in an earlier study [Smith 1978a]. It was used in a hazard analysis [Smith 1978b], and a summary of these two papers was presented as soon as the results were to hand [Smith, 1976]. Because the 1976 paper presented only a summary and not the full detail of the analysis, the study will be referred to herein as the 1978 study. Smith & Berryman [1983,1986] also used the results of the 1978 study to estimate earthquake hazard.

2. NEAR-SOURCE EFFECTS

The term "far-field" in the title is important. As is usual in attenuation studies, there are very few data from sites close to the actual fault rupture. Assumptions must therefore be made about the extrapolation of the derived intensity function back to the source. It is of course at these short distances that the ground motion is of most interest. The extrapolation has been done here with care, but the results should not be used for near-source studies. Studies of hazard near specific faults will involve modelling the fault and its rupture geometry. Such modelling is beyond the scope of the present study. The companion paper [Smith, 1995] addresses this problem, and shows that it is only important at distances less than about two fault lengths. None of the data used in this present study show obvious dependence on fault geometry (with one exception: see Appendix A, event 7). The extrapolation back to zero distance provides a tool for use with the near-source formulation developed in the companion paper.

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3. THE 1978 STUDY

The 1978 study developed an attenuation function which gave, for the first time in New Zealand, a means of calculating the likely intensity at any particular place due to an earthquake elsewhere, given its magnitude and location. The study involved an analysis of isoseismal maps, and suggested that there are regional differences in source mechanism and/or attenuation, so defined three regions for shallow earthquakes (Figure 1). Deep earthquakes were handled separately. The approach was an empirical one, and resulted in an intensity function which involved interpolation from tabulated values rather than an explicit mathematical function. It was observed that the isoseismals were rarely circular, and this ellipticity was modelled and incorporated into the function. The highly elliptical nature of the isoseismals for deep earthquakes was also analysed, including the fact that the centre of maximum ground motion is often displaced south-eastwards from the epicentre, because of the favourable propagation up the inclined slab which is the subducting Pacific Plate. Although deep earthquakes rarely cause intensities greater than MM VI, they were included in the study for the sake of completeness.

The study had some significant limitations. First, the magnitudes of the large earthquakes were not well known. Most of these occurred before magnitudes could be determined from recordings in New Zealand, and some of the values used were very approximate estimates. Dowrick and Smith [1990] have calculated the surface wave magnitudes of many of the large earthquakes since 1990 which were recorded at teleseismic distances, and their results provide a more reliable database for attenuation studies. The results were also presented by Dowrick [1991a]. Anderson, Webb and Jackson [1993] have published a study of large earthquakes in the South Island, which further extends the magnitude database by presenting estimates of the moment magnitude Mw, for 15 large earthquakes since 1964.

Secondly, there were some inadequacies in the extrapolation of the derived intensity function back to short distances. In Region A in particular (see Figure 1) it was observed that the intensity is apparently a linear function of the logarithm of the distance. This fact was used to extrapolate back to zero distance, with an ad hoc limiting of the intensity at very short distances. In Region B, some curvature in the function was observed, so the extrapolation was handled differently. The Region A formula tends to overestimate the intensity at short distances. The present study develops a function which has a similar curvature for the two regions (here assumed to be regions where the predominant focal depths are different), but in Region A that curvature is only evident at distances less than about 20 km, where we have very little data. When constraints are applied to the epicentral intensities, the two sets of curves take very similar form, though with different dependences on distance.

Thirdly, the non-circular nature of the isoseismals was handled very inadequately. There is much variation in ellipticity from one earthquake to the next, but there are some consistencies. The present study has much better statistical justification.

Fourthly, the Central Volcanic Region in the North Island, and the area to the north and west of it, were recognized as anomalous in the 1978 study but little could be done to model the difference from the remainder of Region B. An effort was made to address this using the data from the 1987 Edgecumbe earthquake [Smith, 1990], with only limited success. There are now more data for this region, and it has been possible to define a physically plausible model.

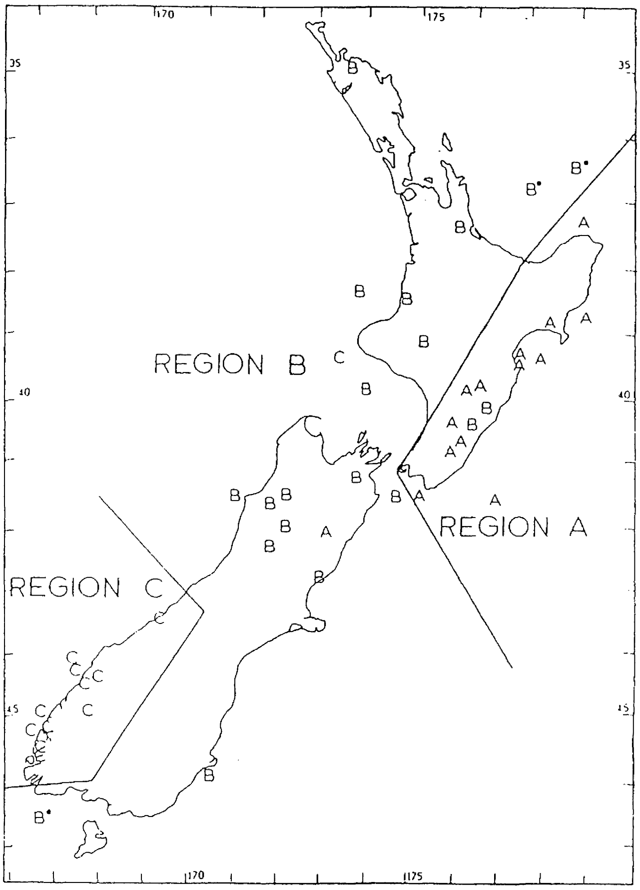

FIGURE 1. Regionalisation proposed in the 1978 study for attenuation from shallow earthquakes (from Smith [1978a]).

4. AVAILABLE DATA

One approach to modelling intensities would be to use the actual intensity spot values. Walley [1976] attempted this, as did Matuschka [1980], but the problem is a very difficult one. The method appears to demand many assumptions, which are not required if one starts from isoseismal maps. Identification of regional differences, and ellipticity of isoseismals, are two aspects which are much more easily handled from isoseismal data, so for the present study (as in the 1978 study) this method has been used. The disadvantage is that one is subject to the personal preferences of the seismologist who drew the maps. In particular, isoseismal lines are defined to identify and enclose areas where a particular intensity is predominant. In practice this is not easy to do in an objective way, and biases can result. Another disadvantage is that isolated observations, inadequate for drawing isoseismals, cannot be used.

Table 1 lists the isoseismal maps which were used in this study, taken from the Seismological Observatory's archive. The list excludes 24 maps for which the epicentres were too poorly constrained, the isoseismals were so poorly controlled by intensity data that they were of little use, or the magnitudes were not known. The data for the 1929 Buller earthquake (Event 7) are taken from a revision by Dowrick [1994].

Earthquakes were classified as follows:

TABLE 1. Isoseismal data used in this study. For each earthquake is listed the date, location, magnitude estimates (value used identified by an asterisk), nominal depth, ellipticity (N50W axis as a percentage of the N40E axis), isoseismal radii (km) for MM IV to X, classification of the event, and references. Classifications are as follows. Upper: Upper crust; Lower: Lower crust; Int: intermediate depth, i.e between 40 and 82 km; Deep: Depth 82 km or more in the Main Seismic Region; Volc: Upper crust event in the Central Volcanic region or the northwest of it; Fiord: Fiordland area; Deep F: Depth 82 km or more in Fiordland. References: 1 - Dowrick & Smith [1990] or Dowrick [1991]; 2 - Anderson, Webb & Jackson [1993]; 3 - Dziewonski et al [1983]; 4 - Dziewonski et al [1984]; 5 - Dziewonski et al [1987]; 6 - Dziewonski et al [1988]; 7 - Dziewonski et al [1991a]; 8 - Dziewonski et al [1991b]; 9 - Darby and Beanland (1992); 10 - This study; 11 - Other events in the Seismological Observatory catalogue.

No.	Date	lat.	long.	ML	Ms	Mw	km	ell	4	5	6	7	8	9	10	Class	Refs
1	1855 Jan 23	41.4	175.0		8.2*					417B	307B	203.5B	141.5B	89.5	45	Upper	9
2	1917 Aug 5	40.8	176.0		6.6*		<45?	77		166.5A	111B	56.5A	31A			Lower	1
3	1921 Jun 28	39.3	176.4		6.4*		80	25		634	407	107.5				Int	1
4	1922 Dec 25	43.0	173.0		6.4*		10	93		180.5B	91A	46.5A				Lower	1
5	1927 Feb 25	38.0	178.0		6.2*		60	106		46.5	40.5					Int	1
6	1929 Mar 9	42.8	171.9		7.1*		<15?	99	341.5A	215A	109A	56.5A	27.5A			Lower	1
7	1929 Jun 16	41.7	172.2		7.8*		20	59		426B	274B	167.5B	97.5B	44		Upper	1
8	1931 Feb 2	39.3	177.0		7.8	7.8*	30	69	494.5A	273.5A	218A	145.5A	78.5A	58	39	Lower	1
9	1931 Feb 13	39.5	177.5		7.3*		<30	74	282A	186A	134A	70A				Lower	1
10	1931 Sep 15	45.0	168.0		5.4*		<60?		145.5	70						Int F	1
11	1931 Sep 21	37.5	178.0		6.1*		80	34	541	407	209.5	52.5				Int	1
12	1932 May 5	39.5	177.5		5.9*		<30	93	148.5A	94.5A	46.5A					Lower	1
13	1932 Jul 20	40.0	174.0		5.1*		60?	120	291	174.5	128					Int	1
14	1932 Sep 15	38.9	177.6		6.9	7.1*	30	80	273.5A	180.5A	125A	55.5A	38A	22		Lower	1
15	1934 Mar 5	40.5	175.5		7.6*		45?	89	387A	260.5A	154A	68.5A				Lower	1
16	1934 Mar 15	38.9	177.2		6.4*		40	85	203.5A	100.5A	52.5A					Lower	1
17	1938 Oct 30	38.5	176.5		5*		150?	43	139.5							Deep	11
18	1938 Dec 15	40.0	177.0		5.6*		<45?	96	209.5B	138B	81.5B	40.5B				Upper	1
19	1938 Dec 16	45.0	167.0		7.0*		60	126	331.5	174.5	67					Int F	1
20	1940 Oct 7	38.5	176.3		7.0*		175	40	255.5	87.5						Deep	10
21	1942 Jun 24	40.9	175.9		7.2*		15	114	366A	232.5A	151.5A	72.5A	33.5A			Lower	1
22	1942 Aug 1	41.0	175.8		7.0	6.8*	43	138	352A	206.5A	116.5A	51A				Lower	1
23	1943 Aug 2	45.8	166.8	6.2	6.6*		50-80	98	180	96	46.5					Int F	1
24	1946 Jun 26	43.2	171.7	6.2*			12	98	238.5A	160B	61A	26A				Lower	11
25	1948 May 22	42.5	173.0	6.4*		12	75	159.5A		104.5A	40.5A	23.5A	13.5A			Lower	11
26	1949 Feb 9	39.7	174.4	6.4*			199	40	270.5	192.5						Deep	11

Table 1 (continued)

No.	Date	lat.	long.	ML	Ms	Mw	km	ell	4	5	6	7	8	9	10	Class	Refs
27	1950 Jan 7	41.1	174.6	5.6*			33	99	187.5B	72.5A						Lower	11
28	1950 Jan 12	41.1	174.4	5.7*			33	103	105A							Lower	11
29	1950 Jan 12	41.2	174.6	5.7*			33	101	93A	30.5A						Lower	11
30	1950 Mar 13	40.6	174.0	5.7*			33		203.5B							Upper	11
31	1950 Jun 17	38.9	175.3	5.9*			185	44	478	116						Deep	11
32	1951 Jan 10	42.8	173.2	5.5*			12	152	128A	56.5A	23.5A					Lower	11
33	1951 Feb 10	40.2	177.0	6.1*			33	113	183A	107.5A	49.5A					Lower	11
34	1951 Mar 28	37.1	176.9	6.4*			370	14	477							Deep	11
35	1951 Apr 23	37.6	177.8	6.2*	5.7		80-125	53	285	209.5	116.5					Deep	1
36	1951 Jun 24	39.5	176.2	6.3*			33	63	243A	134A	45A					Lower	11
37	1951 Oct 3	41.0	172.7	5.5*			12	78	183B	64A						Upper	11
38	1952 Aug 28	40.0	177.0	5.8*			12	90	110.5A	61A						Lower	11
39	1953 Apr 11	40.5	174.8	5.4*			106	67	186							Deep	11
40	1953 Jul 4	38.9	175.7	5.5*			12	108	64	30.5						Volc	11
41	1953 Sep 29	37.6	176.5	7.2*			273	18	785	451	218					Deep	11
42	1954 May 4	44.9	167.6	5.4*			12	104	75.5							Fiord	11
43	1955 Feb 8	39.9	175.8	5.1*			33	96	96B							Upper	10
44	1956 Jan 30	37.1	177.4	5.8*	6.4		10	120	107.5	93						Volc	1,10
45	1956 Dec 28	38.1	178.4	6.3*			12	78	104.5A	70A	25.5A					Lower	10
46	1957 Feb 22	39.2	175.1	5.6*	4.9		5	72	110.5A	67A	22A					Lower	1,10
47	1957 Aug 21	41.0	176.0	5.6*			33	58	103.5A							Lower	10
48	1958 Jan 31	39.9	176.5	6.1*	5.2		33	81	106A	63A	37.5A	13A				Lower	1,10
49	1958 Dec 10	37.1	176.9	6.3*			284	29	610.5							Deep	10
50	1959 May 22	41.2	174.6	5.7*	4.9		86	106	136.5	78.5	36.5					Deep	1,10
51	1960 Feb 3	37.6	178.0	6.0*	5.1		30	74	262	104.5						Int	1,10
52	1960 Feb 21	42.3	173.1	5.8*	5.4		50		116.5	52.5	22					Int	1,10
53	1960 May 24	44.2	167.7	6.3	6.5*		12		424.5	102.5	29					Fiord	1,10
54	1961 Feb 3	37.7	176.0	5.9*			308	21	148.5							Deep	10
55	1961 May 14	40.4	176.2	5.3*			33	94	88.5A							Lower	10
56	1961 Jul 4	44.4	168.1	5.1*			12	145	52.5							Fiord	10
57	1961 Jul 26	37.7	176.8	5.8*			204	32	84.5							Deep	10
58	1962 Jan 23	38.6	174.7	5.5*	4.5		12		189B	48A						Upper	1,10
59	1962 May 10	41.7	171.4	5.6*	5.9		12	75	218B	136.5B	90B	55.5B				Upper	1,10

Table 1 (continued)

No.	Date	lat.	long.	ML	Ms	Mw	km	ell	4	5	6	7	8	9	10	Class	Refs
60	1962 Jul 29	41.4	173.5	5.7*			109	56	253							Deep	10
61	1962 Oct 15	43.5	169.8	5.7*	5.4		12	132	157A	67A	32A					Lower	1,10
62	1963 Apr 12	38.7	176.8	5.5*	5.7		20	82	131A	63.5A	35B					Lower	1,10
63	1963 Jul 14	39.5	174.7	5.5*			166	58	212.5	122						Deep	10
64	1963 Nov 16	35.0	173.5	3.5*			10		24	2.5						Volc	10
65	1963 Dec 22	34.9	173.7	4.9*	4.4		10	77		30	18.5	11.5				Volc	1,10
66	1964 Mar 8	44.3	167.9	5.8	5.8	5.9*	12	93	238.5	180.5	32					Fiord	1,2
67	1965 Apr 11	42.7	174.1	6.1	5.8	6.1*	12	149	135.5A	53A						Lower	2
68	1965 May 20	45.1	167.5	5.6*			105	122	184.5	64						Deep F	11
69	1965 Jun 15	37.9	177.5	5.8*	5.3		65	86	192A	63.5A						Lower	1
70	1966 Mar 4	38.8	178.1	6.0*	5.8		20	158	99A	59.5A	37A	14.5A				Lower	1
71	1966 Apr 23	41.6	174.4	5.8	5.6	5.8*	20	109	248.5B	116.5B	52.5B					Upper	1,2
72	1967 Jan 16	40.3	175.5	4.9*			12	96	114.5B	27B						Upper	11
73	1967 Mar 24	40.7	176.5	5.3*			12	59	98.5A							Lower	11
74	1967 Sep 21	40.9	175.4	4.7*			64	77	95							Int	10
75	1967 Dec 20	40.0	175.1	4.8*			12	82	85.5B	41.5B						Upper	11
76	1968 May 23	41.8	172.0	6.7	7.4	7.1*	12	61	509B	326B	168B	112B	64B	26	10	Upper	1,2
77	1968 Sep 25	46.5	166.7	5.9	6.2	6.3*	10		378B	(331.5)	93B					Upper	1,2
78	1968 Nov 1	41.6	175.1	5.4*	5.0		20-30	126	184.5B	61A	32.5B					Lower	1
79	1969 Jan 2	45.1	167.7	5.8*			139	159	157	70						Deep F	11
80	1970 Jun 12	45.1	167.8	5.2*			124		74							Deep F	11
81	1970 Jul 27	37.9	177.5	5.7*			138		148.5							Deep	11
82	1971 Aug 13	42.1	172.1	5.8	5.5	5.7*	5-12	31	305.5B	160B						Upper	1,2
83	1971 Oct 31	39.0	176.4	5.9*			141	32	244.5	163						Deep	11
84	1972 Jan 8	37.6	175.7	5.3*	5.0		5-12	109	98	51	20.5	7				Volc	1
85	1973 Jan 5	39.1	175.2	7.0*	6.3		160	39	599	263	134					Deep	1
86	1974 Apr 9	46.0	170.5	4.9*			12	107	93.5B	32.5B	13.5B					Upper	11
87	1974 Sep 20	44.4	168.0	5.5*	5.3		12-40	85	177.5	102						Fiord	1
88	1974 Sep 20	43.2	172.5	5.1*			12	83	71.5A							Lower	11
89	1974 Nov 5	39.5	173.5	6.0*	5.4		12-20		247B	129.5B	38A					Upper	1
90	1975 Jan 4	41.1	175.1	5.9*	4.6		80	56	200.5	123.5	27.5					Int	1
91	1975 Feb 11	35.9	174.8	4.4*			12		42.5							Volc	11

Table 1 (continued)

No.	Date	lat.	long.	ML	Ms	Mw	km	ell	4	5	6	7	8	9	10	Class	Refs
92	1975 Jun 10	40.3	176.1	5.8*	5.3		33-50	71		87.5A	26A					Lower	1
93	1976 May 4	44.7	167.4	6.5	6.4	6.5*	12-30		442	136.5	64					Fiord	1,2
94	1976 Dec 3	40.3	173.5	5.6*			214		154							Deep	11
95	1976 Dec 5	38.1	175.5	5.1*			5	83	76	46	37	19.5	5			Volc	1
96	1977 Jan 18	41.8	174.6	6.0	5.9	6.1*	36	131	308.5B	102A						Lower	1,5
97	1977 May 31	37.9	176.8	5.4*			12	101	42	30	18	10				Volc	11
98	1979 Oct 12	46.7	165.7	6.5	7.2	7.3*	12		415A	240A						Lower	1,2
99	1982 Sep 2	39.7	176.9	5.4		5.5*	41	90	183B	57.5A						Lower	3,11
100	1983 Dec 14	38.4	176.3	5.1*	4.6		5	77	30.5	17	10	6.5				Volc	1
101	1984 Mar 8	38.2	177.4	6.4		6.7*	75			174						Int	4,11
102	1987 Mar 2	37.9	176.8	6.1	6.6	6.5*	8	99	156	116.5	42.5	35.5	22.5	10		Volc	1,6
103	1988 Jun 3	45.1	167.2	5.7	6.7	6.7*	57		439	222	106					Int F	1,2
104	1989 May 31	45.3	166.9	6.1	6.2	6.4*	24			270	52					Fiord	2
105	1990 Feb 10	42.2	172.6	5.8		5.7*	13		193B		50B		21B			Upper	7
106	1990 Feb 19	40.5	176.4	5.9		6.2*	34	77		62A	42.5A					Lower	7
107	1990 May 13	40.4	176.5	6.2		6.4*	30	100			51.5A	31.5A	15.5A			Lower	8

(a) Shallow earthquakes (i.e. depths apparently less than 40 km), excluding those in the Volcanic region and in Fiordland. These were further separated into upper and lower crustal classes, as described below. See Section 6.

(b) Earthquakes in the Central Volcanic region. This region was considered by Eiby [1971] to be part of the Main Seismic Region but it is separated here because of known high attenuation. See Section 7.

(c) Deep earthquakes in the Main Seismic Region. For focal depths of 82 km or more, isoseismals become highly elliptical and are offset laterally from the epicentre. (The value of 82 km is the depth at which the offset is zero, from the regression of offset distance against focal depth.)

(d) Intermediate depth earthquakes in the Main Seismic Region. Several earthquakes appear to have been located at depths between 40 and 82 km. The procedure for these has been to interpolate between the intensity functions for 40 km and 82 km events.

(e) Fiordland. In the 1978 study the earthquakes in this region were identified as causing very low intensities.

(f) Intermediate and Deep earthquakes in Fiordland were handled separately.

In Table 1, magnitudes M_L, M_s and M_w are listed separately, where values are available. Some explanation of the three measures of earthquake size is in order. Richter's original local magnitude M_L is computed from the amplitude on a short-period Wood-Anderson seismograph. It is thus a measure of ground displacement above the resonant frequency of the seismograph (1.0 second), and therefore does not represent adequately the long period content of earthquakes larger than about 6.0. The surface wave magnitude M_s is determined from the amplitude of surface waves with a period of 20 seconds, and is a better measure of size for large earthquakes but it has the same shortcoming of not representing well the largest events, though at a higher threshold than M_L. Hanks and Kanamori [1979] devised the moment magnitude scale M_w, which is more linear with respect to energy release over the whole range of earthquake sizes. Ideally, if M_w could be evaluated for all events, it would be the one to use in attenuation studies, but it is determined by modelling the earthquake source, and this is easily done only for large earthquakes and requires broadband recordings. Nor can M_s be measured for earthquakes much less than about 5.5, because they do not generate enough surface waves. We are in fact forced to use a combination of all three.

The value used in the analysis is marked with an asterisk in Table 1, chosen by the following rule: M_w is used wherever it is available; if it is not available, and there are values for both M_L and M_s, M_L is used for events up to M_L 6.0 and M_s for larger events. M_s is however not used for events with depths greater than 40 km, because less surface waves are generated by these events.

5. FORM OF THE ATTENUATION FUNCTION

It is common to assume an algebraic expression for the attenuation of intensity or instrumental ground motion parameters with distance from the earthquake. Dowrick [1991b, 1992] chose one of the form

$$I = a + bM + cr + d \log r \qquad (1)$$

where a,b,c and d are constants, and r is the distance in kilometres as determined from the sum of squares of the mean radius to a given isoseismal line and an "effective depth".

In the 1978 study, an empirical expression was developed, with interpolation necessary for determination of intensity at specific distances and magnitudes. It was based on several observations from the data:

(i) At any given distance, there is a linear relationship between intensity and magnitude. This is encouraging, because it suggests that it is feasible to express a descriptive parameter such as intensity in a numerical formula. However,

(ii) The coefficients of the linear relationship between intensity and magnitude change with distance. This means that Equation (1) does not model the intensities accurately, because it predicts a linear relationship with the same slope at all distances. An empirical approach, producing tabulated results for interpolation, has therefore been retained.

(iii) Crustal earthquakes (i.e. depths less than 40 km) seemed to fall into at least two classes. It was recognized by Hayes [1936] that the attenuation patterns for the 1929 Buller and 1931 Hawke's Bay earthquakes were very different, even at distances of 200 km or so. In the 1978 study, this difference was interpreted as a regional one, and Regions A and B were defined (Figure 1). There were a few exceptions, but most shallow earthquakes in the Main Seismic Region could be classified as being of Type A or B on the basis of their locations.

In another paper [Smith, 1977] it was suggested that the difference might be one of focal depth: that the predominant focal depth in Region A is greater than in Region B, and therefore the generation of short-period surface waves will be greater in Region B, with accompanying higher intensities.

Dowrick [1991b, 1992] chose to account for a range of focal depths by the use of his "effective focal depth" device, but this is inadequate because when radial distance is computed, the distinction between focal depths of, say, 10 km and 30 km is insignificant at epicentral distances greater than 100 km. The data clearly show very significant differences at those distances, as first noticed by Hayes. Some of the differences which are here attributed to focal depths were ascribed by Dowrick to different focal mechanisms. There is however a problem with Dowrick's approach in that (i) the depths are not well known for most of our large earthquakes, and (ii) nor are the mechanisms. The approach adopted here is to ignore any difference which might be expected from variation in focal mechanism. Events are classified into ranges of focal depths, following the suggestion of Smith [1977] that surface wave excitation (which depends strongly on focal depth) is important for strong ground motion. See Section 6 for details.

Resolution of the dependence on mechanism (small in Dowrick's results) awaits the acquisition of more reliable data. It may be that this classification into depth classes is also a classification into mechanism classes, i.e. that there is a relationship between focal depth and mechanism. However, the two functions derived in this study show a greater difference than do Dowrick's two functions, classified by mechanism. It is also important to note that the two earthquakes of magnitude 7.8 in the dataset (events 7 and 8, Table 1) both had compressive mechanisms, but they had very different intensity patterns. It is suggested below that the difference is in their focal depths.

There remains another possibility: that there may indeed be regional differences in attenuation, as was proposed in the 1978 study. In that case, the depth classification can be retained as a computational device, although the term would be a misnomer. See the comment on the North Canterbury events in Section 6.

(iv) Isoseismals are in general not circular. In the 1978 study, the isoseismals were approximated by ellipses, with one axis parallel to the strike of the country (about N40E) and the ellipticity expressed as the ratio of the two axes. Dowrick chose to make a very detailed measurement of isoseismals, averaging radii at intervals of 22.5 degrees. He also assumed that the area of the isoseismal is the same as that of a circle with the mean radius, which is not quite true although the error is usually small. But if the reason for the ellipticity is anisotropic attenuation, the area of the isoseismal is not a good measure of its size. That is, two earthquakes of the same magnitude, mechanism and focal depth, in areas where the propagation in the north-west direction has different attenuation, will have isoseismals of different area.

For the present study, the semi-axis of the isoseismal in the N40E direction has been retained as the measure of the size, together with the semi-axis in the perpendicular direction, from which a model for ellipticity is developed. In practice the N40E axis is easier to measure because this is the general strike of the country, and isoseismals are usually on land. The fact that, in cases when isoseismals are elliptical, one axis is usually in approximately this direction makes this a plausible procedure. This procedure is significantly different from that of Dowrick, who did not attempt to model the ellipticity, suggesting that a further study might address this problem. Such a study is offered here.

6. CRUSTAL EARTHQUAKES

Table 1 lists all the data as measured from isoseismal maps. The distance parameter given is the N40E semi-axis, in kilometres, for isoseismals MM IV and greater. Some maps also have isoseismals for MM III, but these have been ignored because (i) they are usually poorly constrained, and (ii) they have almost no contribution to seismic hazard.

In preparing these data, some revisions were necessary because the procedure currently being used at the Observatory for computing magnitude M_L has not been used for the whole catalogue. It has been in use since 1977, and has been applied in revisions for the periods 1943-1954 and 1963-1976, but the remaining period 1955-1962 has not yet been revised systematically. For the purposes of this study, therefore, all events in Table 1 from 1955 to 1962 have been revised according to current procedures and their magnitudes M_L recomputed (Reference 10 in Table 1).

If crustal earthquakes are classified into two groups on the basis of their isoseismals, it is found that these two groups are essentially those believed to be in the upper and the lower crust. Instrumentally determined focal depths are often not reliable, but it is recognized, for instance, that the 1929 Buller earthquake was in the upper crust and the 1931 Hawke's Bay event was deeper. For many years it was the practice of the Seismological Observatory to assign Shallow and Normal focal depth classifications to crustal earthquakes, which were later given the nominal values of 12 and 33 km. It is found that in particular parts of the country, either upper or lower crustal events generally predominate. Earthquakes were therefore examined individually, on the basis of their isoseismals and also considering the known tectonics of the local area. Each event was assigned a depth classification, and in only a few cases was that at variance with the instrumentally determined focal depth.

It was necessary to exercise a degree of judgement in making this classification, because of the poor quality of the data. Figure 2a shows the isoseismal semi-axis in the N40E direction (Table 1) as a function of magnitude for all the observations of MM IV. The open circles indicate earthquakes which are believed to be in the upper crust and the closed circles the events which are believed to be in the lower crust. A smooth curve separates these two classes, with only a few exceptions. In Table 1, each MM IV isoseismal semi-axis is classified as A or B according to whether the symbol falls below (A) of the curve in Figure 2a, or above it (B). The same exercise was done for intensities V to VIII; Figure 2b shows the data for MM VI. This classification of individual isoseismals is then used to give an overall classification for each event, which is also shown in Table 1. An earthquake for which the isoseismals are all of type A, for instance, will normally be classified as a lower crustal event. In some cases judgement had to be exercised. The justification for the classification of each event is given in Appendix A. It is this classification which is the basis for the two sets of symbols in Figure 2.

Events 4,6,24,25,32,61,67 and 88 deserve special comment. One of these events (61) was in Westland, the others in North Canterbury. They were all assigned shallow depths in the Observatory catalogue, but these were often poorly controlled. They have isoseismal patterns like others with definitive lower crustal depths. Event 6 (Arthur's Pass, 1929) has been described by Yang [1991] as being on the Kakapo fault, but his data are largely circumstantial, being based only on landslides. Reyners and Cowan [1993] have suggested a tectonic explanation for earthquakes in the lower crust in this part of the country. As mentioned in Section 5, it is possible that events such as these were actually in the upper crust but that there is a true regional attenuation effect. For the purposes of computation, however, a lower crustal classification has been assigned. The data for this event are discussed further below.

Crustal earthquakes were therefore considered as two separate data sets. Before developing attenuation functions for them, however, evidence of epicentral intensities was also examined, and this also demanded judgement. Figure 3 shows the estimates of maximum intensities as a function of magnitude for the two classes. The value plotted is that of the innermost isoseismal, increased by 0.5 units. This arbitrary increase represents middle ground between those events for which the epicentral intensity barely exceeds that of the innermost isoseismal and those for which it almost reaches the next level. In many cases it is apparent that the intensity data lack good observations near the epicentre, or that there were insufficient

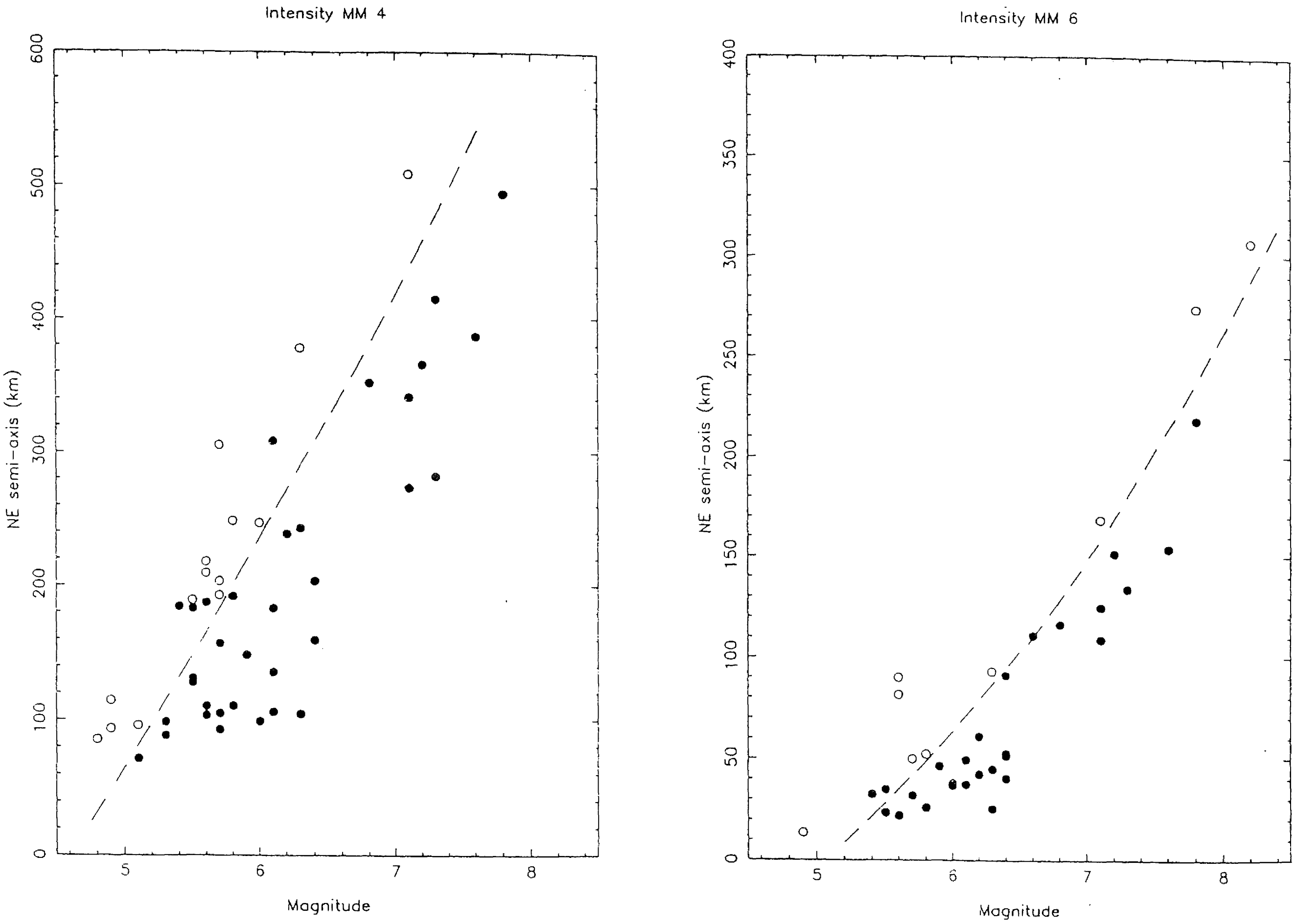

FIGURE 2. Semi-axes of the MM IV isoseismals (Figure 2a, left) and MM VI (Figure 2b, right) as a function of magnitude for crustal earthquakes. Events classified as upper crustal are shown by open circles and lower crustal by solid circles.

of these to justify an isoseismal. In some cases the epicentres were offshore. The observations plotted as open circles have therefore been ignored. This simply reflects the difficulty of obtaining closely spaced observations near the epicentre: they are often not available. In the case of Event No. 7 (the 1929 Buller earthquake, magnitude 7.8), Dowrick's [1994] revision of the isoseismal map has an isoseismal for MM9, and he comments that it is reasonable to assume that the intensity reached MM X close to the fault, but there were no buildings in the area on which to base an intensity assessment. So for this earthquake the epicentral intensity is plotted as 10.0. This is an important earthquake and warrants this special attention.

Regression lines fitted to the solid symbols in Figures 3a and b are

$$\text{Fig 3a:} \quad I_0 = 1.65M - 2.63 \qquad (2)$$
$$\text{Fig 3b:} \quad I_0 = 1.20M + 1.00$$

These have been used as zero distance constraints in the fitting procedure. This approach is crude but practical, because the epicentral intensities presented by Equations 2 are plausible and fit the available data.

Figure 4 shows the resulting intensity functions for magnitudes 5 to 8, for upper crustal (solid lines) and lower crustal events (broken lines). As a practical procedure, intensities at short distances have been limited by Equations (2), producing an "epicentral tableau" which is shown by the straight line segments in Figure 4. The actual output for short distances from the fitting procedure is shown by dotted lines. The fact that this truncation was necessary further justifies the rejection of low intensity data (open circles) in Figure 3, because the extrapolation of the far-field data back to short distance produced epicentral intensities which exceed even the regression from the data represented by solid circles in Figure 3. Including the lower intensity data in the regression for Equations 2 would have resulted in even more severe truncation of the extrapolated curves.

Beyond the epicentral tableau, the intensity for any specific magnitude and distance is obtained by inverse interpolation from the values presented in Table 2. A suitable interpolation procedure is that proposed by Wiggins [1976]. For magnitudes less than 5.0 and greater than 8.0, it is necessary to extrapolate, on the basis that at any particular distance the intensity is a linear function of magnitude. This was an empirical observation in the 1978 study and is borne out by Figure 4 despite the fact that it was not assumed in the fitting process.

The fitting has been done by a neural network procedure, [Robinson, 1991]. This technique makes no ad hoc assumption about the form of the function, but "learns" as each new datum is added. The data supplied were the magnitudes and the N40E semi-axes for intensities MM IV to X, as in Table 1, together with the epicentral constraints in the form of the magnitudes for intensity values of 4 to 10, from Equations 2.

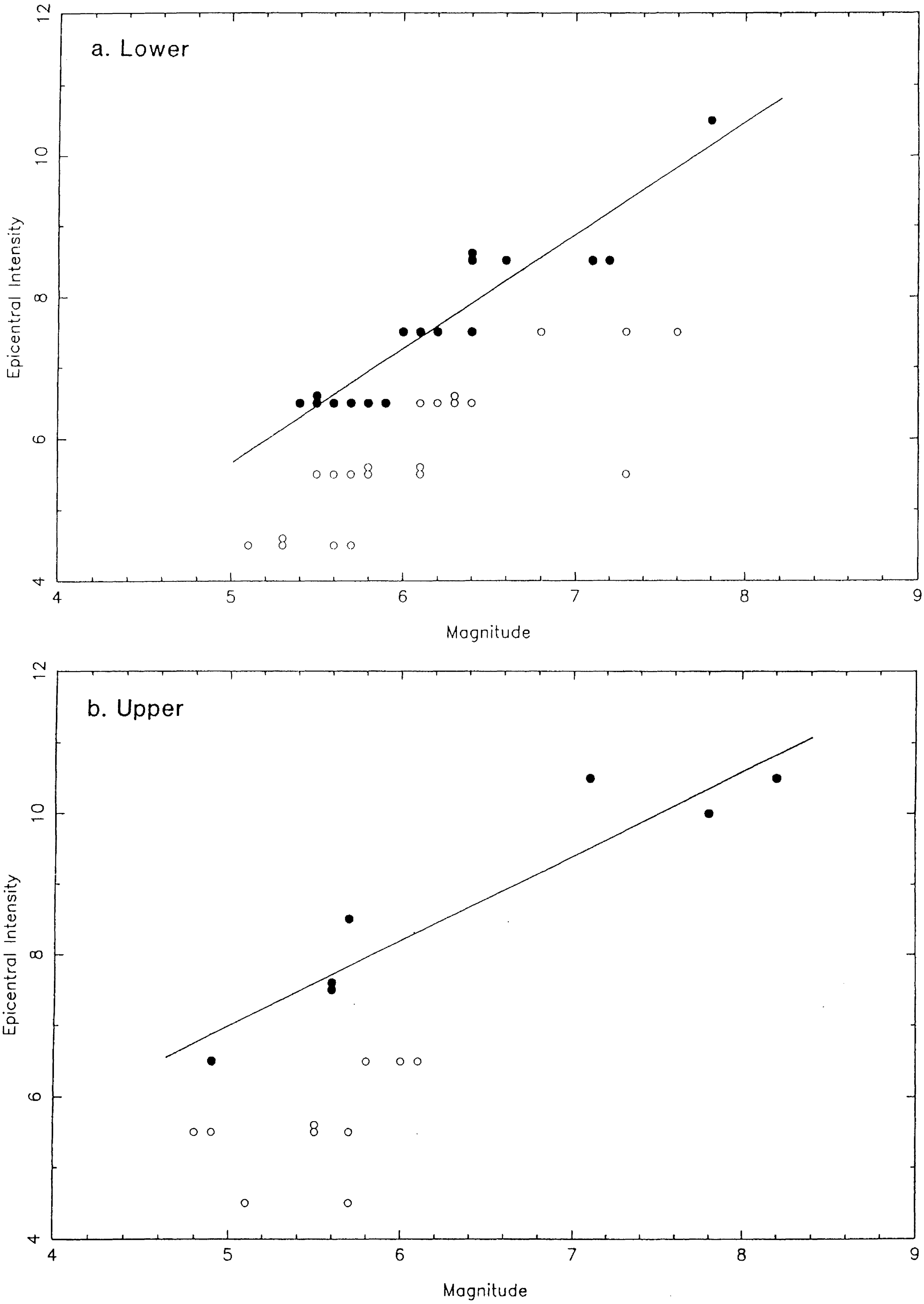

FIGURE 3. Intensities of the innermost isoseismal, as a function of magnitude, for lower crustal events (Figure 3a, top) and upper crustal events (Figure 3b, bottom). See text for explanation.

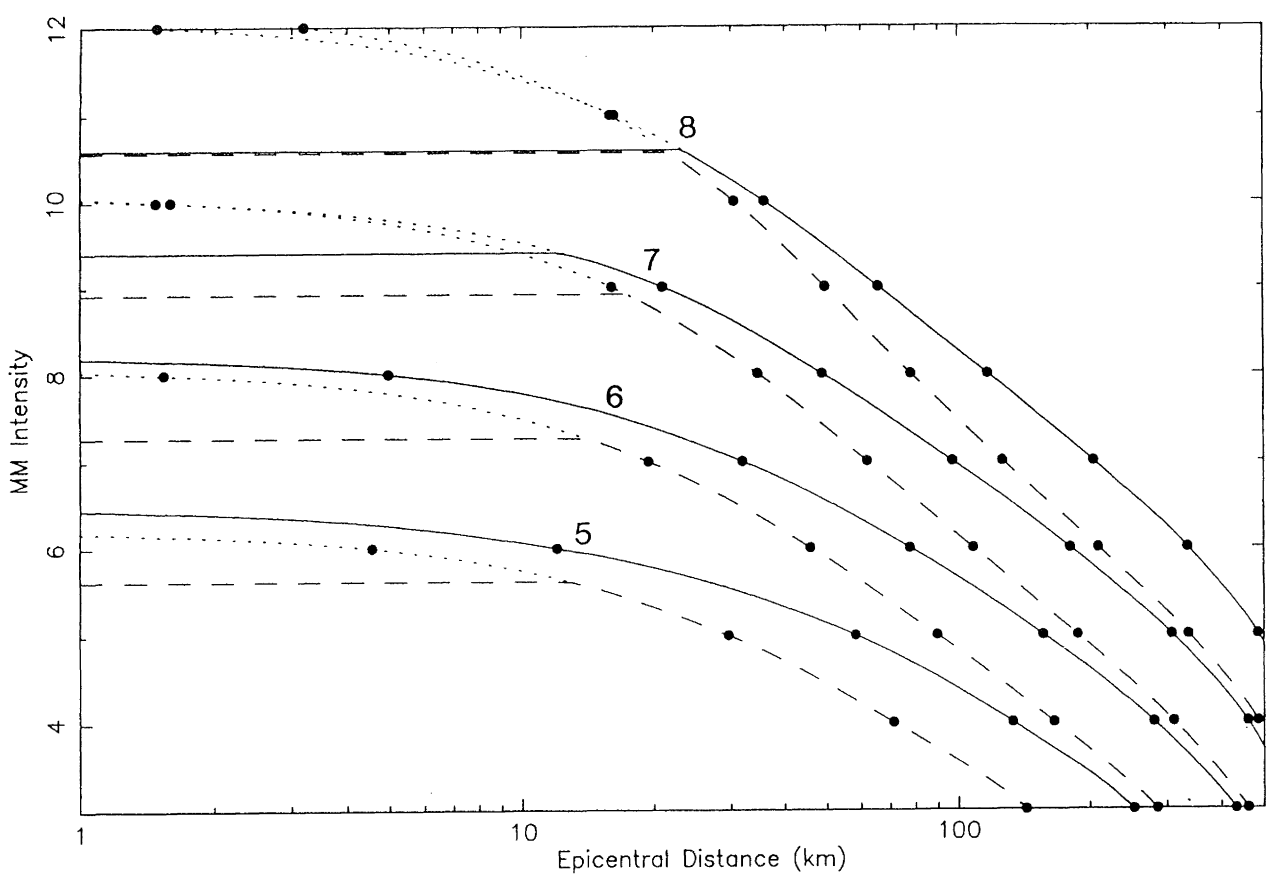

FIGURE 4. Intensity model for upper crustal earthquakes (solid lines) and lower crustal earthquakes (broken lines). The dotted lines represent the extrapolations back to zero distance, truncated by Equations 2. Point values given in Table 2 are also shown.

TABLE 2. Isoseismal semi-axes (km) in the N40E direction, for intensities MM 4 to 12 and magnitudes 5 to 8. Where negative values are given, these are to provide for smooth second-difference interpolation back to zero epicentral distance. Intensities at short distances, obtained from these tables by inverse interpolation, should replaced by Equation 2 if they exceed the values given in that equation.

Upper crust

Mag.	MM 4	5	6	7	8	9	10	11	12
5	129.9	56.4	10.0	(-12.2)	(-30.1)				
6	282.1	155.5	77.3	32.2	4.7	(-14.0)	(-27.7)		
7	477.7	314.8	182.8	99.6	51.8	23.0	3.5	(-11.0)	(-22.2)
8	645.7	510.7	348.6	211.6	123.2	72.5	42.1	21.7	6.4

Lower crust

Mag.	MM 4	5	6	7	8	9	10	11	12
5	71.8	31.0	4.0	(-16.1)	(-32.4)				
6	161.9	89.1	46.4	18.8	(-1.6)	(-18.1)			
7	309.6	184.6	107.4	62.8	34.6	14.1	(-2.5)	(-16.5)	
8	498.5	338.9	208.7	126.9	80.2	51.4	31.0	14.3	0.1

It was decided to fit the distances rather than the intensities. The procedure of drawing an isoseismal is one of defining the region within which a particular intensity is experienced. The independent parameter is thus the intensity, for which a distance is estimated. A modelling procedure should therefore minimise the residuals in distance, rather than in intensity. Note the distinction between isoseismal data and instrumental measures of ground motion. Measures such as velocity and acceleration are made at known distances, so the situation is reversed: distance is the independent parameter, and ground motion residuals should be minimised in the fitting procedure. Joyner and Boore [1988] did exactly this. But because isoseismal data are measures of distance rather than of intensity, they have been fitted in the opposite way. Goodness-of-fit statistics are given in Table 4 (see Section 10).

Some salient features of the short distance modelling are:

(i) Intensities are slightly lower for lower crustal events than for the upper crustal class, but this effect is very small at large magnitudes. This convergence for large events is to be expected on the grounds that, for very extended sources in this depth range, intensity at the epicentre will not be strongly dependent on depth to the focus.
(ii) The spatial extent of the epicentral tableau decreases with magnitude. This is also a very plausible result, because earthquake source volume decreases with magnitude.
(iii) The epicentral tableau is larger for lower crustal events than for the upper crustal class. This is a natural consequence of focal depth.
(iv) From trials with variations of the parameters in Equations 2 it is apparent that the uncertainty in epicentral intensity is at least 0.5 units, especially for large magnitudes.

It is interesting to compare the isoseismals of Event 6 with the two attenuation functions developed (Figure 4). Figure 5 shows this comparison: solid lines are the upper crustal model and broken lines the lower crustal model. The five isoseismal semi-axes can each be used to estimate the magnitude: the mean of these five estimates is 6.4 (std. error 0.1) when compared with the upper crustal model and 7.0 (std. error 0.2) for the lower crustal model (recognizing of course that this event was included in the data set used to derive the curves for the lower crustal model). The actual magnitude was 7.1, i.e. the lower crustal classification is justified for this event.

7. VOLCANIC REGION

Smith & Berryman [1986] defined a seismicity model for New Zealand. It comprises 15 regions, within each of which seismicity is considered to be spatially uniform. This regionalisation has recently been modified slightly [Berryman, pers. comm.], and is presented in Figure 6.

Events 40,44,84,97,100 and 102 are in the Central Volcanic Region (Region C of Figure 6). Events 64, 65 and 91 are in Region A. Events 58 and 95 are in Region B, although event 95 is very close close to the boundary of Region C. These events are all considered together in this analysis, with the exception of event 58, and the combined region referred to as the volcanic region, for brevity. For event 58 the felt effects appear to be more widespread than is typical for the volcanic region. The classification for events in Regions A and B of Figure 6 was justified later (See Section 11).

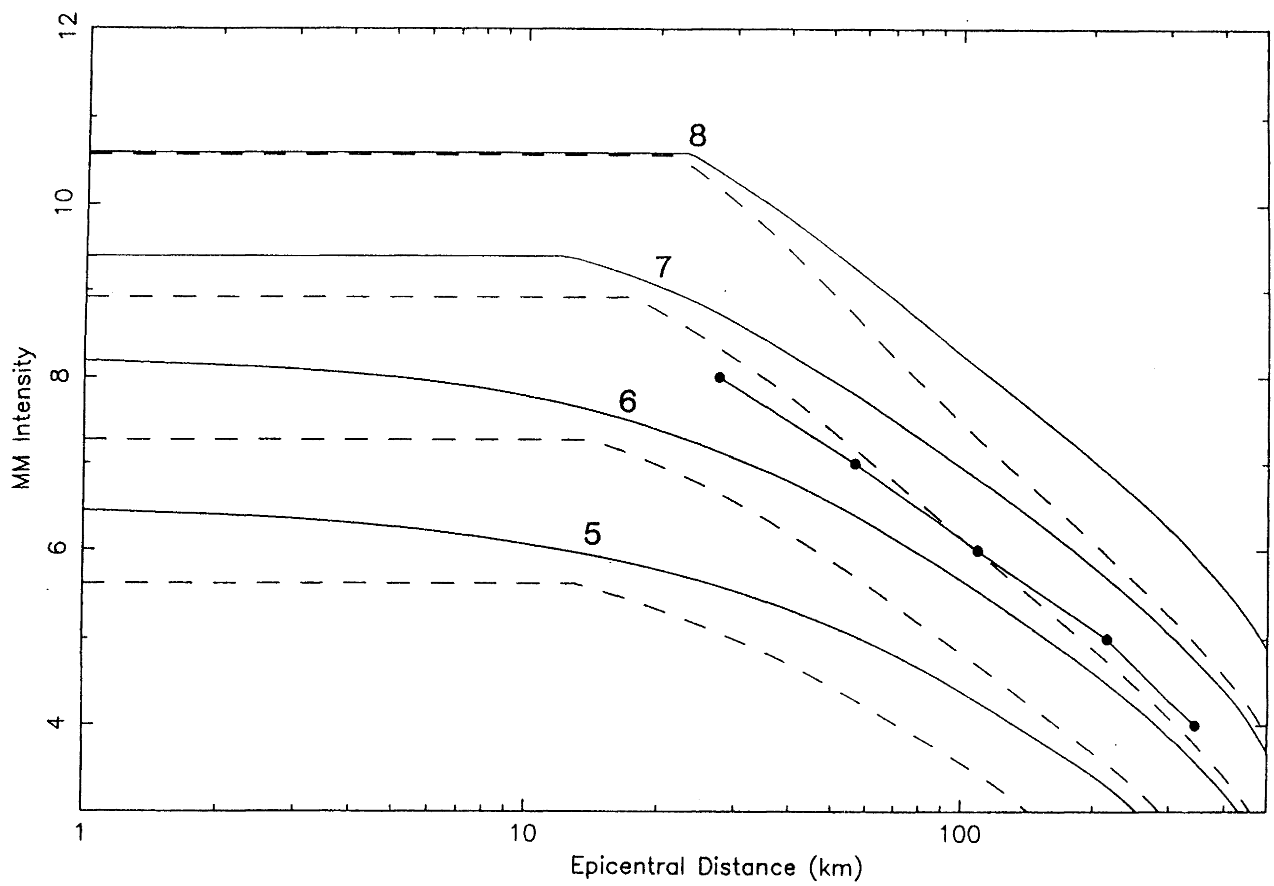

FIGURE 5. Isoseismal semi-axes (N40E) for Event 6 (magnitude 7.1), with the upper crustal model (solid lines) and the lower crustal model (broken lines).

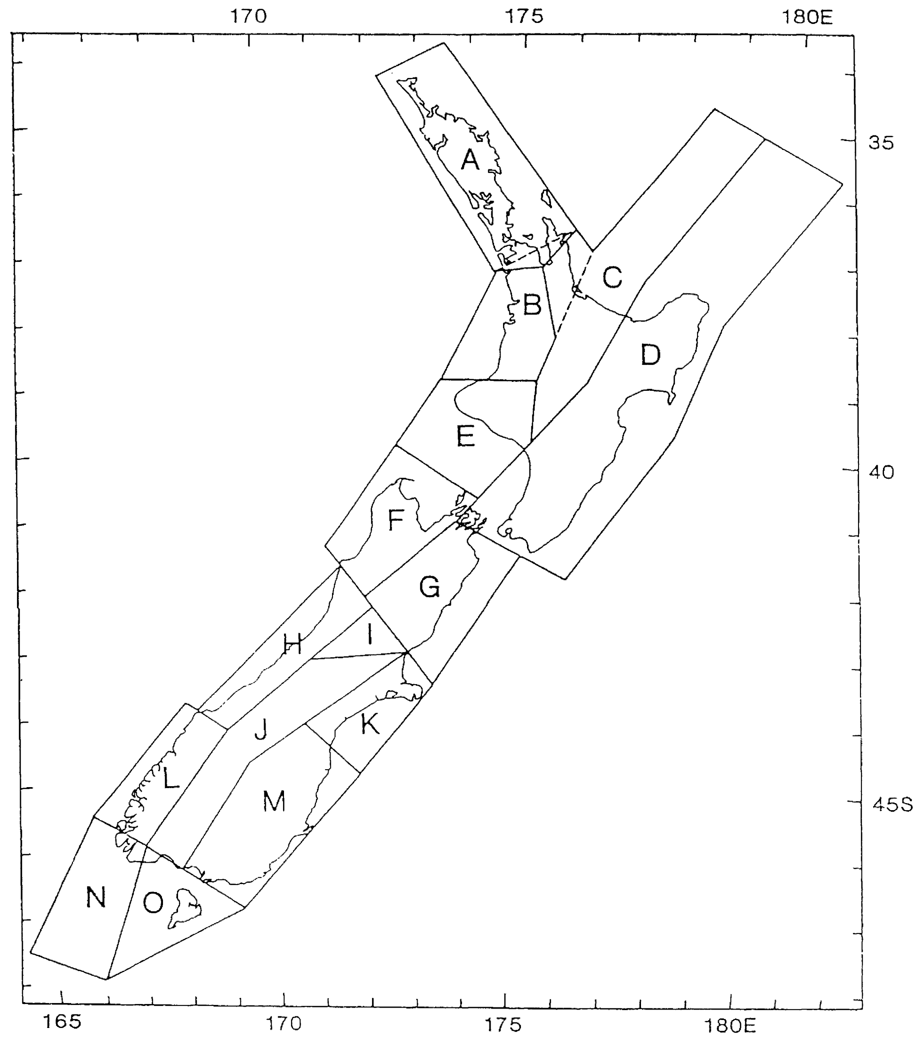

FIGURE 6. Regionalisation of shallow seismicity (Berryman, pers. comm.). The broken lines indicate the boundaries that were used in the 1986 study.

Focal depths are known to be very shallow (e.g. Eiby [1977]; Anderson and Webb [1989]). Attenuation is high, at least in the Central Volcanic Region itself [Haines, 1981]. Figure 7a shows the intensities observed in these earthquakes, compared with the upper crustal model developed in Section 6. There is considerable scatter, but most events show rather more attenuation with distance than is accounted for in the model. A different function is warranted.

The data are inadequate for a neural network solution as in Section 6. The only practical solution is to fit a simple function, and a suitable one is the form chosen by Dowrick [1991b, 1992], expressed in Equation 1. Its use is justified here because there is very little depth variation throughout the volcanic region. The "effective depth" is regarded simply as a device to fit the data, required because of the logarithm term in Equation 1. Boore et al [1993] comment that it is "a fictitious depth determined by the regression". A range of values was tried, and this parameter was set to 15 km, although there is very little control as it affects only the few data at very short epicentral distances. Note that in Dowrick's use of the "effective depth" device he chose to identify it with the centroid of the actual rupture surface.

Equation 1 could be considered to have some physical justification, in that the terms in c and d can be related to anelastic attenuation and geometric spreading respectively, and intensity considered to be proportional to the logarithm of the displacement. Parameter d should be -2.0 for head waves, -1.0 for body waves, -0.5 for surface waves [after Brekhovskikh, 1960].

It is instructive to set d to -2.0, and fit parameters a,b,c. Figure 7b shows the result, an extremely poor fit from which it is concluded that: (i) -2.0 is an inappropriate value for parameter d, because much more additional attenuation is required, but if it is modelled by an anelastic term it is not modelled well; (ii) the above physical argument for setting d to -2.0 does not hold, i.e. intensity is not proportional to the logarithm of displacement; (iii) the other possibilities -1.0 and -0.5 can be discounted on the same grounds. In the same way that a linear relationship between intensity and magnitude was found in the 1978 study (See Section 5 above) there may be a linear relationship between intensity and the logarithm of displacement at any particular distance, but the parameters change with distance. This result is not too surprising, however, because of the nature of intensity: it is likely to be dependent on the predominant frequency content of the ground motion, which will vary with distance, and no doubt on the particular wave types which are dominant at any given distance.

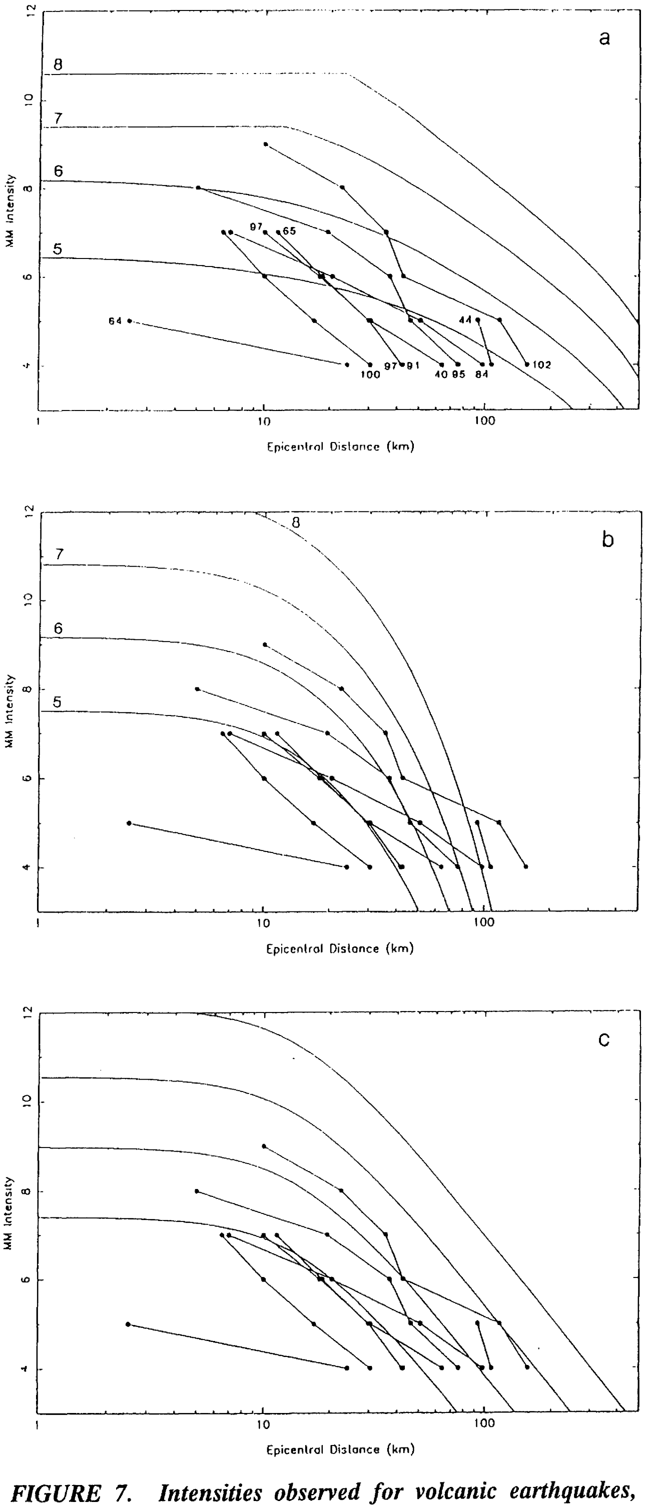

FIGURE 7. Intensities observed for volcanic earthquakes, with (a) the upper crustal model which clearly does not have enough attenuation, (b) a model given by Equation (1) with parameter d set to -2.0 and a,b,c fitted, and (c) the adopted model for the Volcanic region (Equation 3).

An attempt was made to fit the four parameters of Equation 1, but c and d were found to be too highly correlated, and the solution would not converge. However the optimum appeared to have a very small value of c, much less than its standard error. As a practical solution, parameter c was therefore set to zero, and a solution obtained for a,b and d. The values obtained for these three parameters were 6.901 +/- 1.082, 1.567 +/- 0.191, -6.220 +/- 0.627 respectively, with effective depth 15 km,

$$\text{i.e.} \quad I = 6.901 + 1.567M - 6.220 \log(r) \qquad (3)$$

where r is the radial distance from the point of interest to an effective focus at depth 15 km. This function is shown in Figure 7c plotted against epicentral distance.

This fitting was done by optimising the ratio of the observed and predicted distances, i.e. minimising the difference from unity of the ratio of the observed isoseismal distance to the predicted distance. This device gives extra weight to the inner isoseismals, because while a residual of a few kilometres is of no consequence at a distance of 100 km or more, it is significant close to the epicentre.

This procedure differs from that employed for the neural network data for the upper and lower crustal classes of events (Section 6) where a solution was found by simple minimisation of the distance residuals. Using the distance ratio in the neural network procedure would have been preferable, but it could not be implemented there easily.

Equation (3) should be regarded simply as an empirical model. Given the above discussion about attenuation and geometric spreading, no physical significance should be attached to the parameter values. Goodness-of-fit statistics are given in Table 4 (see Section 10).

8. ELLIPTICITY OF ISOSEISMALS

The analysis thus far has examined only the semi-axes of the isoseismals in the N40E direction. The perpendicular axis was also measured, and an average value of the ratio obtained for each event. This ratio (N50W axis divided by N40E axis) is given in Table 1 and shown in Figure 8, plotted as a percentage at the epicentre. Three regions can be defined, within each of which there are sufficient data to extract a mean value of the axis ratio. These mean values, with standard deviations, are shown in Table 3.

Earthquakes outside these regions have been ignored; there are insufficient of these to define a model. Some are out of character with such other data as are available from nearby and may be subject to considerable uncertainties. In particular, the epicentre for the 1966 Gisborne earthquake (event 70, ellipticity 1.58) is offshore, but there was considerable doubt about this location [e.g. Hamilton, 1969]. It was very poorly constrained, and could in fact have been onshore. This would imply an ellipticity closer to or less than 1.0, in conformity with other earthquakes in that region. No doubt many of the events within the three regions are also subject to such uncertainties: this finds expression in the standard deviations.

Outside these three regions, the axis ratio is assumed to be 1.0 (circular isoseismals), for lack of evidence to the contrary. In Region E2 the mean axis ratio is greater than 1.0, and in the other two it is less than 1.0, so it is appropriate to set the model value to 1.0 on all the boundaries of the three regions. Within these regions, the model axis ratio has been set to the designated values given in Table 3, at the locations given. Where the mean is less than unity (Regions E1 and E3) this designated value is the mean for that region minus one standard deviation. In Region E2, where the mean is greater than unity, the mean plus one standard deviation is used. At all other points the axis ratio is determined by linear interpolation between the central point and the boundary. This provides for a continuous model of the axis ratio, though not a smooth one, representing the data in

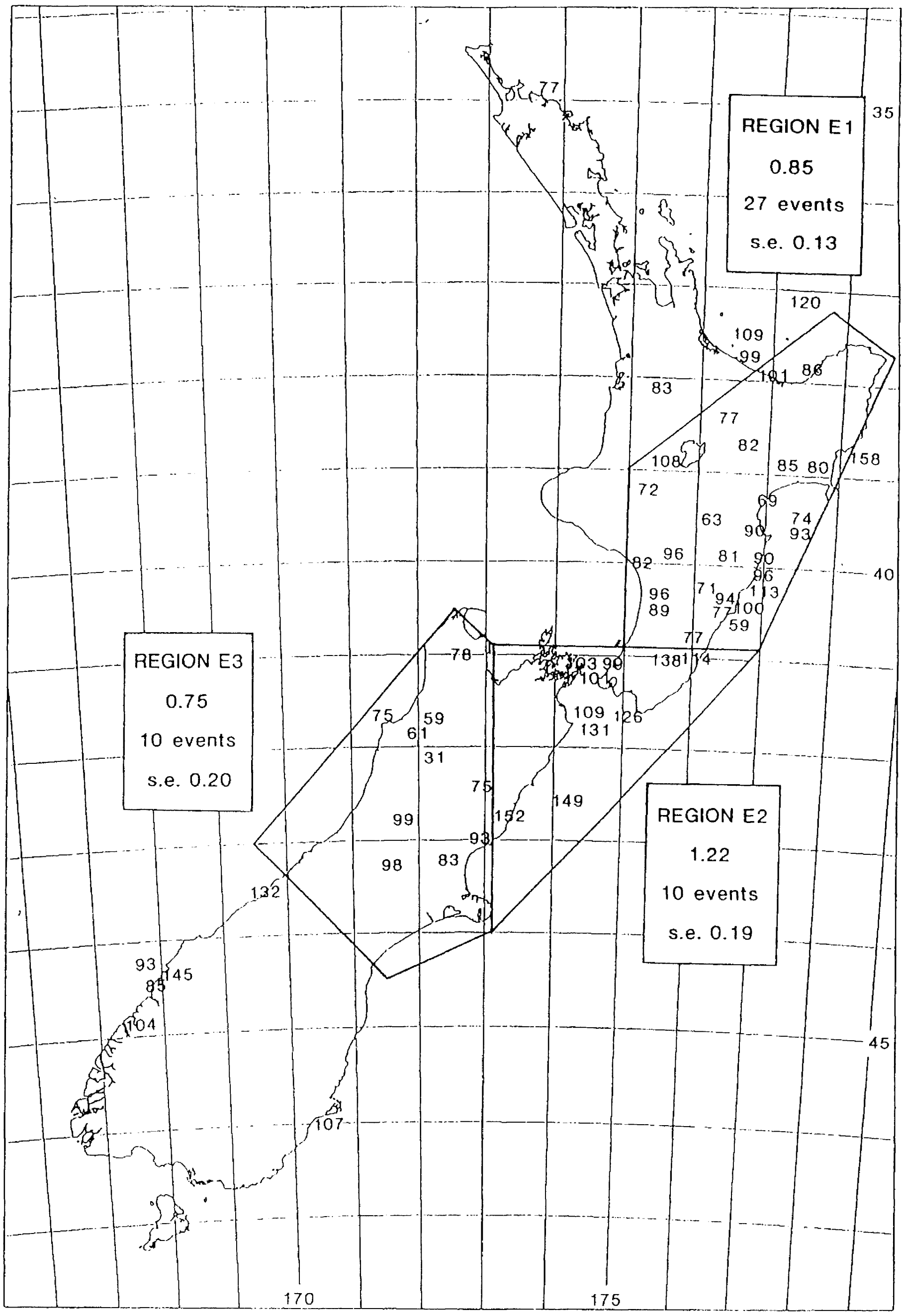

FIGURE 8. Axis ratios for crustal earthquakes (N50W axis divided by the N40E axis, expressed as a percentage), defining three regions where values differ consistently from 100%.

TABLE 3. Model for the axis ratio of isoseismals (see Figure 8).

	Region	mean	std. dev.	central pt.	value
I	Eastern N.I.	0.84	0.16	39.5, 176.5	0.68
II	Wellington-Marlborough	1.21	0.19	42.0, 174.7	1.40
III	Nelson/Canterbury	0.74	0.21	42.6, 171.9	0.53

each region by as much as one standard deviation beyond the mean and equal to 1.0 on the boundaries of the regions. It varies from 0.53 to 1.40.

This regionalisation is empirical. It is not related to any physical model, because the reason for the ellipticity of isoseismals is not completely clear. It is unlikely to be due to the orientation of faults, which are predominantly along the strike of the country, because this cannot have a significant effect at distances of more than about two fault lengths from the epicentre [Smith, 1995]. In particular, this rules out any explanation for ellipticity for earthquakes of magnitude 6 or less, in terms of fault geometry, because the length of rupture is too short. Ellipticity of isoseismals is more probably related to topography and structure, i.e. attenuation of energy as it crosses mountain ranges. This would explain the predominant orientation of the major axis along the strike of the country in Regions E1 and E3, and one might argue for the influence of the Cook Strait structures in Region E2 having the opposite effect.

9. FIORDLAND REGION, DEEP AND INTERMEDIATE EARTHQUAKES

No revison is suggested for the intensity function derived in the 1978 study for shallow earthquakes in Fiordland, or for that for intermediate or deep earthquakes. Very few data are available, and they are extremely scattered, doubtless due to very poor focal depths and magnitudes for early events. Fiordland events are characterised by very low intensities (e.g. MM VI near the epicentre of event 93, of magnitude 6.5). It is not clear why intensities should be as low as this, although Smith [1977] offered a suggestion in terms of the excitation of short period surface waves. Earthquakes in the Main Seismic Region that are deeper than 82 km are characterised by the centre of the isoseismal pattern being offset to the south-east from the epicentre, and by extremely elliptical isoseismals. (The figure of 82 km was obtained from a regression of offset against focal depth). These effects are apparently due to the favourable propagation of energy up the Benioff zone, and were both modelled in the 1978 study. For earthquakes at depths between 40 and 82 km, the procedure of interpolating the intensities from those computed for those two focal depths has been retained, in the absence of any better algorithm.

10. GOODNESS OF FIT

Despite the fact that the distance data were fitted, rather than the intensities, it is nevertheless important to examine the residuals of both intensity and distance. Table 4 lists the mean residuals in epicentral distance (observed distance minus predicted distance), the mean ratio of the observed distances to the predicted distances, and also the mean intensity residuals at the observed isoseismal distances. These are presented for each of the classes in Section 4 separately, with standard deviations, and finally for all events together.

The fit is poor for intermediate and deep events in the Main Seismic Region, and also for all Fiordland events (witness the large standard deviations for all these and the large bias for intermediate MSR events). However little can be done with the currently available data because (i) the depths are not well enough known and in many cases are obviously much in error; and (ii) the magnitudes are very uncertain. Fortunately, these events have only a small contribution to seismic hazard estimates.

Table 5 shows the residuals obtained using Dowrick's [1991b, 1992] formula, compared with those obtained using the 1978 formula and the new results. Only the 30 events and isoseismals used by Dowrick are included here (MM IV and greater), because Dowrick's measure of isoseismal size is not available for the other events in Table 1. Data for intensity MM X in events 7 and 14 of Table 1, used by Dowrick in his regression but subsequently ascertained by him to be unreliable [Dowrick, pers. comm], have been excluded. Dowrick used a different measure of distance, and classified earthquakes by focal mechanism. Some of his magnitudes were different. The statistics in the first two rows are determined from his data, using his classifications. The remainder use the magnitudes and distances from Table 1.

This comparison is inadequate in that (i) it does not include the modelling of ellipticity of isoseismals, which was not addressed by Dowrick; (ii) few of the events he used are in the upper crustal class; and (iii) the full data set of Table 1 could not be included because Dowrick's measure of isoseismal size is not available. But it is all that can be done at this stage.

TABLE 4. Mean distance residuals (Dres), mean distance ratios (Drat) and intensity residuals (Ires), with standard deviations, for all data and for the separate classes. Ne is the number of events, NI the number of isoseismals, and Nx is the number of isoseismals for which the model does not predict that intensity at all. See text for comment and explanation.

	Dres	s.d.	Drat	s.d.	Ires	s.d.	Ne	NI	Nx
Upper crustal	-4.3	36.6	1.10	0.59	0.00	0.45	17	53	2
Lower crustal	-0.5	33.0	1.00	0.28	-0.06	0.45	39	117	2
Volcanic	1.5	15.9	1.00	0.40	-0.05	0.76	10	33	1
Deep MSR	-23.1	155.4	1.09	0.52	0.00	0.78	18	31	3
Int, MSR	74.5	142.6	2.44	2.86	0.54	1.13	9	22	2
Fiordland	-1.0	69.5	0.98	0.48	0.02	0.51	7	15	1
Deep Fiordland	-19.2	35.1	0.93	0.31	0.17	0.26	3	5	1
Int Fiordland	-46.8	72.2	0.80	0.28	-0.18	0.55	4	11	1
All events	0.2	74.1	1.12	0.93	0.01	0.62	107	287	13

TABLE 5. Goodness of fit statistics as for Table 4, but for only those events used by Dowrick [1991].

	Dres	s.d.	Drat	s.d.	Ires	s.d.	Ne	NI	Nx
DJD 1991	0.9	29.0	1.12	0.50	0.03	0.47	30	116	4
MM IX and X	7.4	9.8	1.67	0.96	0.26	0.41	5	7	0
1978 model	4.8	42.9	1.16	0.62	0.09	0.70	30	117	7
MM IX and X	-8.8	18.2	0.83	0.25	-0.23	0.51	5	7	1
1994 model	0.3	29.0	1.05	0.41	-0.02	0.51	30	117	3
MM IX and X	3.0	10.6	1.11	0.32	0.21	0.47	5	7	1

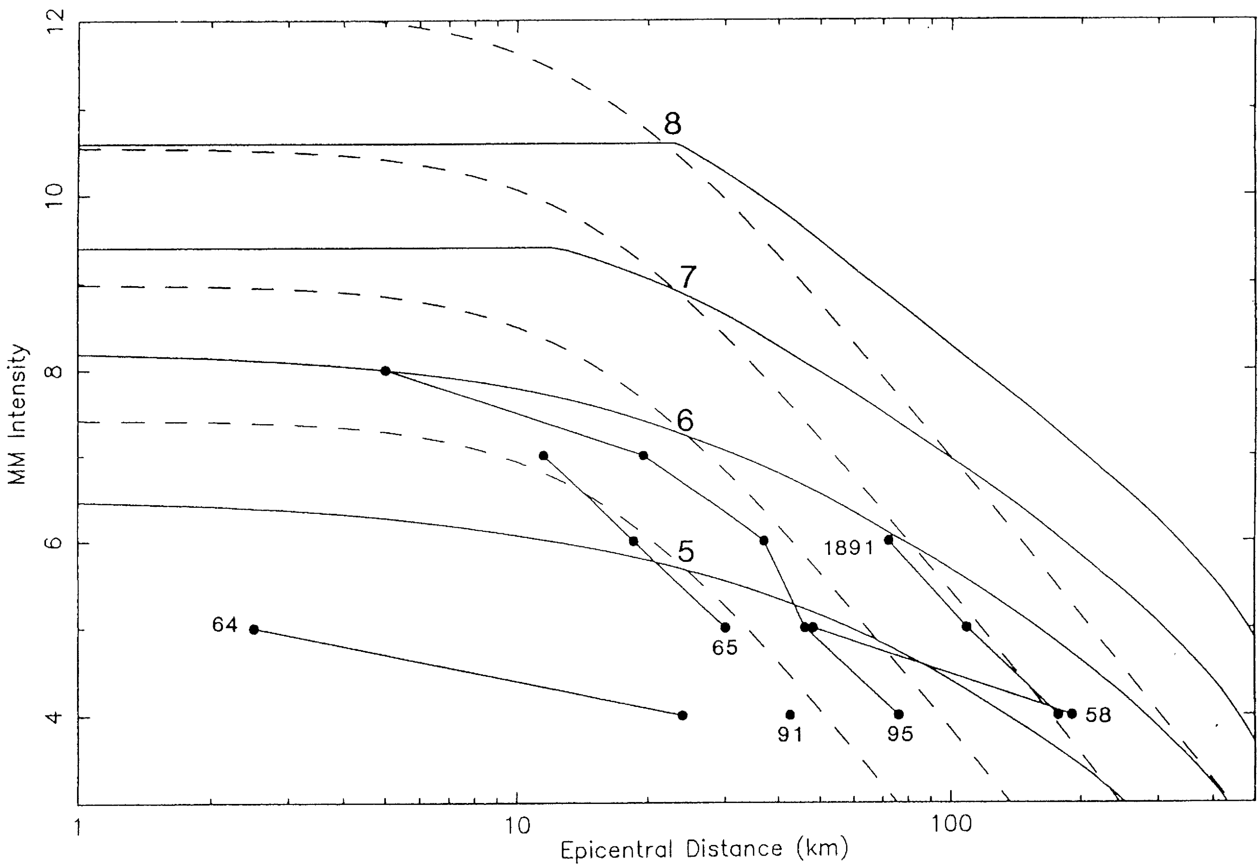

FIGURE 9. Isoseismal semi-axes for events 58,64,65,91,95 and the 1891 Waikato earthquake, with the upper crustal intensity model (solid lines) and the volcanic model (broken lines).

Table 5 includes statistics for MM IX and MM X only. While the 1978 model overestimates these intensities by 0.23 units on average, and the new model underestimates them by 0.21 units, Dowrick's model underestimates them by 0.26 units (parameter Ires). The greatest difference between the models is in the mean distance ratio (parameter Drat), which is 1.67 for MM IX and X, using Dowrick's formula. This is equivalent to underestimating the distance by 40% and the area by 64%. The new model underestimates the MM IX and X distances by 10% and the areas by 19%.

There is a suggestion [Dowrick, pers. comm.] that these high intensity assessments were unreliable, especially in that some were based on conversions from Rossi-Forel figures. There will no doubt be continuous reassessment of intensities and therefore a need for repeated revison of attenuation relationships. The above comparison, however, assesses the goodness of fit of the two models to the data used in deriving them.

11. IMPLEMENTATION IN SEISMIC HAZARD ESTIMATION STUDIES

Smith and Berryman [1983, 1986] used the 1978 attenuation function, together with a seismicity model to estimate return periods of MM intensities VI to IX throughout New Zealand. This study could now be repeated with the new attenuation function and a newly revised seismicity model [Berryman, pers. comm.] But the identification of source regions (Figure 1) must be changed. An appropriate procedure would be to identify the predominant focal depth in each of the regions in Figure 6, and use the upper or lower crustal model accordingly. That is, if the predominant depth is less than about 20 m, the upper crustal model should be used; if greater, the lower crustal model. Where there seems to be a range of focal depths, such as in Region E, Figure 6, the seismicity could be apportioned between the two depth classes. The Fiordland model should be used in Region L.

There remains the question of the extent of the Volcanic region. Should the volcanic model (Equation 3) be used throughout Regions A and B as well as C? Several earthquakes throw light on this problem.

Isoseismal data for events 58,64,65,91 and 95 are shown in Figure 9, together with data for an event in 1891. The solid lines represent the upper crustal model developed above, and the broken lines the volcanic model. Events 64, 65, 91 and 95 were used in the definition of the volcanic model; events 65 and 95 appear to fit it better than they do the upper crustal model. The fit for events 64 and 91 is not conclusive, because of the low magnitude for event 64 (3.5) and the single isoseismal for event 91. Event 58 appears to fit the upper crustal class better, though at a magnitude somewhat less than 5.5. The attenuation rate is much less than for volcanic earthquakes.

In 1891 there was an earthquake apparently centred near 37.4S 174.6E. It caused minor damage throughout the surrounding area. The magnitude is not known, but isoseismals have recently been prepared for this earthquake by G.L. Downes, from newspaper reports and other contemporary accounts. The isoseismal data are included in Figure 9, from which a magnitude of 5.4-5.6 may be inferred if the upper crustal model is used, or 6.5-6.7 for the volcanic model. The upper crustal model is judged to be more appropriate, because an earthquake of magnitude 6.5-6.7 would, according to the volcanic model, cause intensities of MM IX or more at short epicentral distances. The highest intensity reported is about MM VII, which appears to have been close to the epicentre. The conclusion to be drawn is that the attenuation of intensity with distance appears not to be as great here as in the volcanic region.

These data argue for a western boundary of the volcanic region to include event 95 (epicentre 38.1S 175.5E), and for the region to the west of this, including the epicentre of the 1891 event and south to event 58 (38.6S 174.6E) to be modelled with the upper crustal classification. In Region A, however, the volcanic model seems to be more appropriate. Event 65, of magnitude 4.9, fits the volcanic model much better than the upper crustal model (Figure 9). In particular, there is strong attenuation which is represented well by the volcanic model.

The conclusion to be drawn is that Regions A and C of Figure 6 should be modelled using the volcanic model of Equation 3, and Region B with the upper crustal model of Figure 4 (solid lines); but the boundary between Regions B and C should be moved westwards at least as far as 175.5E.

The uncertainty in the intensity estimation could also be included in the estimation of seismic hazard. Table 4 indicates that for the earthquakes in the upper and lower crustal classes, the standard deviations in the intensity residuals is 0.45 units in each case. For shallow Fiordland events it is 0.51 and for volcanic events 0.74 units. These figures could be used to modify the deterministic estimation of hazard. This was not done by Smith and Berryman [1986]. Incorporating the standard error increases the estimate of hazard by a small exponential factor, and it would not be difficult to implement this.

12. CONCLUSIONS

The availability of revised estimates of the magnitudes of large historical earthquakes has enabled a revision of the 1978 study which developed an attenuation function for Modified Mercalli intensities in New Zealand. Rather than use assigned focal depths for crustal earthquakes, which are almost always poorly controlled, it is preferable to classify them as either upper or lower crustal, and to treat these as two separate classes. This classification was done on the basis of (i) nominal focal depths, (ii) isoseismal dimensions and (iii) tectonic considerations. Crustal earthquakes in the volcanic region are treated separately; they admit a simple regression. Analysis of the volcanic earthquakes demonstrates that there is no simple linear relationship between intensity and ground motion parameters such as displacement or acceleration. Such a relationship may hold at a given distance, but it is inappropriate to assume one over a range of distances. It was not possible to revise the 1978 results for shallow earthquakes in Fiordland or for intermediate and deep events. Treatment of these must still be considered unsatisfactory. Dowrick's [1991b, 1992] model addresses the overestimation of hazard which results from the 1978 study, but it underestimates the hazard at intensities of MM IX and MM X. The high intensity data on which both models were based may yet need revision, however. The new formulation would be of use in revising estimates of earthquake hazard.

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APPENDIX A

Classification of Earthquakes

Table 1 classifies events by focal depth and geographic region. The justification for this classification is presented below. It was done by examining individual isoseismals, and plots like that in Figure 2, for intensities MM IV to VIII. Thus each isoseismal was assigned a class A (lower crustal type) or B (upper crustal All the isoseismals for a given event were then compared, taking account of the estimated focal depth and the tectonic setting of the area in which the event occurred, in order to arrive at an overall classification for the event. Classifications A and B correspond with the regionalisation which was used in the 1987 study.

1. This is the largest event in New Zealand in historical times. No instrumental depth is available, but there was extensive surface faulting. Darby and Beanland [1992] assign Mw = 8.2, on the basis of which all isoseismals are Type B. Classification: Upper.

2. A for MM V,VII,VIII, B for VI. Depth is not well constrained, probably less than 45 km. Suggest a lower crustal depth, because that is predominant in that part of the country (East Coast, North Island). Individual intensity data are not shown on the isoseismal map. Suggest poor control for MM VI, VII. Classification: Lower.

3. Depth apparently 80 km. Classification: Intermediate

4. B for V, A for VI,VII. Shallow depth. Classification: Lower. See note below.

5. Depth apparently 60. Classification: Intermediate

6. All isoseismals A. Depth <15? Yang [1992] has suggested that this earthquake was caused by rupture on the Kakapo fault (and was therefore shallow), but his evidence is largely circumstantial because it is based only on landslides. See note below. Classification: Lower.

7. All isoseismals B. Dramatic faulting implies shallow depth. Nominal depth of 20 km is not well constrained. The MM IX isoseismal for this earthquake is the one datum in this study which may be affected by near-source geometry, but because this is the only one it is not considered to have affected the results significantly. Classification: Upper.

8. All isoseismals A. Depth 30. Classification: Lower.

9. All isoseismals A. Depth 30. Classification: Lower.

10. Fiordland. 60 km is common there. Intermediate Fiordland.

11. Depth 80. Classification: Intermediate.

12. All isoseismals A. Depth <30. Classification: Lower.

13. Depth 60?. Classification: Intermediate.

14. All isoseismals A. Depth 30. Classification: Lower.

15. All isoseismals A. Depth 45? Classification: Lower.

16 All isoseismals A. Depth 40. Classification: Lower.

17. Depth 150? Classification: Deep.

18. All isoseismals B. Depth <45? is not well constrained. Suggest that it was an upper crustal event, despite the fact that most of the earthquakes here are lower crustal. There are some exceptions, however. Classification: Upper.

19. Fiordland. Depth 60. Classification: Intermediate Fiordland.

20. Depth 175. Classification: Deep.

21. All isoseismals A. Depth 15 is unusual here, and is not well constrained. (Terry Webb, pers. comm.) Classification: Lower.

22. All isoseismals A. Depth 43. Classification: Lower.

23. Fiordland. Depth 50-80. Classification: Intermediate Fiordland.

24. B for MM V, A for VI, VI, VII. Depth 12 is commonly assigned to earthquakes here but the control is usually very poor. Suggest lower crust. See note below. Classification: Lower.

25. All isoseismals A. Depth 12. See note below. Classification: A as for Event 23.

26. Depth 199. Classification: Deep.

27. B for MM IV, A for MM V. Depth 33, as is common here. Intensity data are extremely sparse, and isoseismals poorly constrained. Classification: Lower.

28. A for MM IV. Depth 33. Lower crustal events common here. Similar location to Event 28. Classification: Lower.

29. A for MM IV,V. Depth 33. Classification: Lower.

30. B for MM IV. Depth 33 is very unusual here, and poorly constrained. Classification: Upper.

31. Depth 185. Classification: Deep.

32. All isoseismals A. Depth 12 is not well constrained. See note below.. Classification: Lower.

33. All isoseismals A. Depth 33. Classification: Lower.

34 Depth 370. Classification: Deep.

35. Depth 125. Classification: Deep.

36. All isoseismals A. Depth 33. Classification: Lower.

37. B for MM IV, A for MM V. Depth 12 which is common here. MM V isoseismal is very poorly constrained. Classification: Upper.

38. A for MM IV, V. Depth 12 is unusual here. Classification: Lower.

39. Depth 106. Classification: Deep.

40. Volcanic region, depth 12. Classification: Volcanic.

41. Depth 273. Classification: Deep.

42. Fiordland. Depth 12. Classification: Fiordland.

43. B for MM IV. Depth 33, but is not well constrained. Classification: Upper.

44. Volcanic region, depth 10. Classification: Volcanic

45. All isoseismals A. Depth 12 is very poorly constrained. Classification: Lower.

46. All isoseismals A. Depth 5, constrained by S-P from close data. Nearly in Volcanic region, where higher epicentral intensities would be expected. This event thus remains an anomaly. Classification: Lower.

47. A for MM IV. Depth 33. Classification: Lower.

48. All isoseismals A. Depth 33. Classification: Lower.

49. Depth 284. Classification: Deep.

50. Depth 86. Classification: Deep.

51. Depth 30 is not well constrained, and Ms much less than ML. Classification: Intermediate.

52. Depth 114. Classification: Intermediate.

53. Fiordland. Depth 12. Classification: Fiordland.

54. Depth 308. Classification: Deep.

55. A for MM IV. Depth 33. Classification: Lower.

56. Fiordland. Depth 12. Classification: Fiordland.

57. Depth 204. Classification: Deep.

58. B for MM IV, A for MM V. Depth 12 is typical for that part of the country. Classification: Upper.

59. All isoseismals B. Depth 12. Classification: Upper.

60. Depth 109. Classification: Deep.

61. All isoseismals A. Depth 12 has very little control. See note below. Classification: Lower.

62. A for IV,V, B for VI. Depth 20 which is likely there. Classification: Lower.

63. Depth 166. Classification: Deep.

64. Northland. Depth 10. Classification: Volcanic

65. Northland. Depth 10. Classification: Volcanic

66. Fiordland. Depth 12. Classification: Fiordland.

67. A for MM IV, V. Depth 12 on the basis of supposedly crustal phases observed. Nearest station over 150 km away. See note below. Classification: Lower.

68. Depth 105. Classification: Deep Fiordland.

69. A for MM IV,V. Depth 33 which is common here. Classification: Lower.

70. All isoseismals A. Depth 33. Classification: Lower.

71. All isoseismals B. Depth 12 is well constrained, although MM VI isoseismal is not. Classification: Upper.

72. B for IV and V. Depth 12. Classification: Upper.

73. A for MM IV is quite well constrained. Depth 12 is unusual here. Classification: Lower.

74. Depth 64. Classification: Intermediate.

75. B for MM IV, V. Depth 12. Classification: Upper.

76. All isoseismals B. Depth 12. Classification: Upper.

77. B for MM IV,VI, also for MM V but this looks excessively large and affected by microzoning. Depth 10 but control is poor. This is one of only two events to the south of Fiordland (other is 97). Classification: Upper. (MM V data not used).

78. B for MM IV,VI A for V. Depth 33 which is typical here. Classification: Lower.

79. Depth 139. Classification: Deep Fiordland.

80. Depth 124. Classification: Deep Fiordland.

81. Depth 138. Classification: Deep.

82. B for MM IV, V. Depth 12. Classification: Upper.

83. Depth 141. Classification: Deep

84. Volcanic region. Depth 5-12. Classification: Volcanic.

85. Depth 160. Classification: Deep.

86. All isoseismals B. Depth 12. Classification: Upper.

87. Fiordland. Depth 12-40. Classification: Fiordland.

88. A for MM IV. Depth 12. See note below. Classification: Lower.

89. B for MM IV,V. A for VI. Depth 12-29 which is typical there. MM VI isoseismal very poorly constrained. Classification: Upper.

90. Depth 80. Classification: Intermediate.

91. Off Northland coast. Classification: Volcanic

92. A for MM V,VI. Depth 33-50. Classification: Lower.

93. Fiordland. Depth 12-30. Classification: Fiordland.

94. Fiordland. Depth 214. Classification: Deep.

95. Volcanic region. Depth 5. Classification: Volcanic.

96. A for MM IV, B for V. Depth 36 which is well constrained. Classification: Lower.

97. Volcanic region. Classification: Volcanic.

98. A for MM IV,V. Depth 12 but control is very poor. This is one of only two events to the south of Fiordland (other is 76). Classification: Lower.

99. B for MM IV, A for MM V. Depth 41 is well constrained and typical there. MM IV isoseismal looks exaggerated. Classification: Lower.

100. Volcanic region. Depth 5. Classification: Volcanic.

101. Depth 75. Classification: Intermediate.

102. Volcanic region. Depth 8. Classification: Volcanic.

103. Fiordland. Depth 57. Classification: Intermediate Fiordland.

104. Fiordland. Depth 24. Classification: Fiordland.

105. All isoseismals B. Depth 13 km is very well constrained, and is common for earthquakes here. MM VI isoseismal is not well controlled (none for MM V,VII). Classification: Upper.

106. A for MM V, VI. Depth 34 km. Classification: Lower.

107. All isoseismals A. Depth 30. Classification: Lower.

For Events 4, 6, 24, 25, 32, 61, 67 and 88 see comment in text.