ATTENUATION OF MODIFIED MERCALLI INTENSITY IN NEW ZEALAND EARTHQUAKES

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SUMMARY

The main result of this study is the development of attenuation expressions for Modified Mercalli intensity isoseismals in New Zealand earthquakes, in terms of magnitude M_w and source distance. Attenuation of isoseismal intensities has been modelled in a new 2-dimensional manner, i.e. in terms of three distance measures, i.e. radius a along the strike of the fault rupture, radius b normal to strike, and the mean radius c . Other factors which are included in the modelling are depth, focal mechanism, tectonic type (i.e. crustal, interface or dipping slab), and regional variations. As well as being implied in the source distance, the moderate effect of depth (to increase intensities) is modelled well with a separate linear depth term. Shallow ($h_c \leq 60$ km) Reverse mechanism events were found to give stronger intensities than Strike-Slip events which in turn were stronger than Normal mechanism events, the difference near source between Reverse and Normal events of M_w 7+ events being about 0.4 intensity units. Attenuation for the Central Volcanic Region and Deep events differs markedly from the Main Seismic Region. No magnitude saturation effects were found. Comparisons are made with some earlier New Zealand intensity models, and local and international peak ground acceleration attenuation models.

1.0 INTRODUCTION

The continued high usefulness of the Modified Mercalli (MM) intensity scale in earthquake hazard and risk assessment has prompted the authors to attempt to improve the modelling of attenuation of intensity in earthquakes in New Zealand. Earlier models have been developed by Walley [1], Smith [2-4], Matuschka [5] and Dowrick [6, 7], but all have their weak points, mostly arising from the inadequate quality and quantity of the source and intensity data on which they were based. In the case of the previous models of Dowrick [6, 7], concerns were raised [3, 8] that the model appeared to underestimate the area of intensity MM10 near the source of the two largest earthquakes in his dataset. This will be shown in this paper to be due to the poor MM10 data then available and to natural interevent variability, rather than to the model. The data available for the present study is superior to that available previously, on the following counts:

- Nine isoseismal maps are available for earthquakes which have occurred since the 1987 cut-off for Dowrick's earlier study [6, 7];
- Recent maps have been used of six large early earthquakes: 1855 [9], 1888 [10], 1914 [11, 12], June 1929 [13], 1931 [14] and 5 March 1934 [15];
- In addition the first author has developed (a) MM intensity maps of 12 events [16] dating

from 1922 to 1942 replacing the poor Rossi-Forel intensity conversion maps previously available, e.g. Figure 1(b); (b) 12 completely new maps for events dating from 1901 to 1985 [17], e.g. Figure 1(a); and (c) complete revisions of MM intensity maps for 15 events dating from 1948 - 1990 [18].

- Moment magnitudes are now available for a lot more earthquakes [19-25, 34]. By excluding the use of M_L , the very large scatter in the relationship between M_L and M_w [34] involved in all earlier studies [1-7], was eliminated (M_L differed unsystematically from M_w by up to 0.8 magnitude units).
- Better estimates of source depths, focal mechanisms and tectonic types can now be made as a result of various recent special studies [19-33].

In the earlier work by Dowrick [6, 7] (which was restricted to earthquakes with depths ≤ 65 km), a difference was found between the attenuation of intensity in events of different focal mechanism, but no difference was found with source depth, except that arising from source-to-site distance. A recent study [35] of attenuation of peak ground acceleration (PGA) in New Zealand earthquakes, found that attenuation of PGA depends not only on focal mechanism (for crustal earthquakes) but also on depth and on tectonic type (i.e. on whether the source was in the interface between the crust and the dipping slab or not). These factors are all considered below.

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2.0 EARTHQUAKES STUDIED

The earthquakes used in this study were limited in two ways, i.e.

- Only events for which intensities have been assigned directly in the MM scale were used (i.e. events studied since 1945).
- Only events for which reasonable estimates of moment magnitude M_w were available, as qualified below, were used.

In all, 85 earthquakes from the period 1855 – 1998 inclusive were taken into account, as listed in Table 1. These events range in magnitude from M_w 4.9 to 8.2, with 11 being $M_w \geq 7.0$. The moment magnitudes used have been assigned a class A, B, C or D in descending quality of M_w determination. The 28 class A events had their M_w calculated seismologically [21-25]. Fifty of the events in Table 1 are M_w Class B, which had their moment magnitudes estimated from M_s and centroid depth h_c by Dowrick and Rhoades [34], who called such magnitudes $\hat{M}_w$. For the era pre-1964, $\hat{M}_w$ values were adopted here in preference to M_w determined by body wave inversions [26, 27] because of the high scatter in the latter estimates, the worst example being the body wave inversion M_w for the 1929 Murchison earthquake is M_w 7.0 compared with our [34] $M_s = 7.8$ and $\hat{M}_w = 7.7$.

Of the three M_w Class C events, M_w 8.2 for the 1855 earthquake was determined [19, 20] from dislocation modelling using changes in ground levels of a somewhat sketchy nature. For the other event pre-dating instrumental magnitudes (1888) Cowan [10] estimated its M_s as 7.0-7.3 from its fault rupture length, and $M_w = 7.1$ was adopted, i.e. in the middle of Cowan's M_s range. The four earthquakes with M_w Class D were small events (Nos. 11, 36, 65, 66) for which M_w was assigned approximately, taking into account the number of stations overseas from New Zealand that reported them [34], as neither M_s nor instrumental M_s were available.

As seen in Figure 2, the earthquakes are mainly distributed along the eastern part of the North Island and the western part of the South Island, which are regions close to the subduction zones along the boundary between the Pacific and Australian tectonic plates. Ten events provide data related to attenuation in the Central Volcanic Region (CVR) (Figure 2 and Table 1). Because of the presence of the subducting plates, large magnitude events occur down to considerable depths under New Zealand (Figure 3), as indicated by the deepest event in our data set which has a depth of 300 km and a magnitude $\hat{M}_w$ of 7.3. Of the 85 events in our data set, 12 have depths greater than 50 km (Figure 2), 11 of which occurred in the dipping slab of the subducting Pacific plate while the other event (1988) was in the dipping Australian plate. The remaining 73 shallower events have depths ranging from 3 to 50 km. As in other studies [6, 7, 35] the primary depth term used is the depth to the centroid of the rupture, h_c , rather than the focal depth. The centroid depth estimates have been obtained from various sources, i.e. focal mechanism studies [19-27], or aftershock studies [29, 30], or in their absence

from reliable focal depths, or from considerations of seismogenic depth in the source region [32, 33] and macroseismic information. The macroseismic procedure was undesirable in relation to depth dependency inferences, but was used only for a few smaller events, and only *a priori*.

In the data set there are 51 Crustal events, 4 Interface events, 19 dipping slab events with $h_c \leq 60$ km, and 11 deeper Pacific slab events, denoted C, I, S₆₀ and S_D in Table 1. In addition to these real earthquakes, two M_w 7.0 dummy events (Nos. 86 and 87) were introduced to help control the near source modelling of large shallow Normal faulting events in both the CVR and the Main Seismic Region. Without these two events the largest magnitude Normal events were of M_w 6.5 in both the CVR and the Main Seismic Region Crust.

Information defining source mechanisms is given (Table 1) in two forms, namely rake angle γ , and the three common geological descriptors, normal (N), strike-slip (S) and reverse (R). Rake is the direction of slip in the fault plane, where the angles 0° and 180° represent horizontal movement (pure S), while 90° and 270° correspond to pure R and pure N respectively. The mechanisms N, S, R given in Table 1 are the predominant mechanisms, which are defined such that $\gamma = 315^\circ - 045^\circ$ and $135^\circ - 225^\circ$ is strike-slip; $\gamma = 46^\circ - 134^\circ$ is reverse; and $\gamma = 226^\circ - 314^\circ$ is normal. Of the crustal events there are 11 reverse, 21 strike-slip, 18 normal and one unknown mechanism. Of the shallow ($h_c \leq 60$ km) Slab events, there are two R, two S, 14 N and one unknown mechanism, while there are two R and two S Interface earthquakes.

Source parameter data in Table 1 incorporate what is considered to be the best quality information available, derived preferably from special studies, as referenced in Table 1.

3.0 INTENSITY AND DISTANCE DATA

In the previous study by the first author [6, 7], the attenuation of intensity was modelled solely in terms of the mean distance (radius) of the isoseismals from the centroid of the earthquake source. Here we have developed and modified that approach to take account of the shape of each isoseismal in relation to the geometry of the fault rupture surface in a new manner. Because of the importance of modelling the intensities in large surface rupturing earthquakes, the geometry of the isoseismals is measured in relation to fault strike. As shown in Figure 4 the dimensions of the isoseismals are defined as:

- a = horizontal radius along the fault strike;
- b = horizontal radius normal to strike;
- c = mean horizontal radius;
- r_a, r_b, r_c = source(-to-isoseismal) distances corresponding to a, b, c , such that for example

$$r_a = (a^2 + h_r^2)^{1/2} \quad (1)$$

where h_r = depth to top of fault rupture.

Table 1: Source parameters of earthquakes considered in this study
(see explanatory notes at foot of table)

No	Date	Time	°S	°E	h	h _c	h _t	M _w	M _s	T	FM	Rn	θ	β	γ	Refs
1	1855 Jan 23	0932	41.40	175.00	sha	19	0	8.2C	-	C	S	M	45	90	180	9, 19, 20
2	1888 Aug 31	1645	42.60	172.40	sha	8	0	7.1C	7.1	C	S	M	80	90	180	10
3	1901 Nov 15	2015	42.70	173.30	12	10	1	6.78B	6.87	C	R?	M	50	70	U	17, 34
4	1911 Oct 5	0736	39.50	177.00	25?	25	23	5.56B	5.32	S ₆₀	N	M	40	U	U	17, 34
5	1912 May 26	0635	38.00	175.00	cru	10	7	5.65B	5.52	C	N	N	0	45	U	17, 34
6	1914 Nov 22	0814	37.50	176.50	300	300	300	7.29B	6.46	S _D	U	D	40	U	U	11, 34
7	1922 Jul 4	0515	38.60	176.10	v.sh	3	0	4.93B	4.39	C	N	V	35	45	U	17, 34
8	1922 Jul 14	0356	38.60	176.10	v.sh	3	0	5.12B	4.73	C	N	V	35	45	U	17, 34
9	1922 Sep 5	0233	38.60	176.10	v.sh	3	0	5.37B	5.14	C	N	V	35	45	U	17, 34
10	1922 Dec 25	0333	43.00	173.00	10?	20	15	6.79B	6.42	C	S	M	72	70	170	16, 34, 26
11	1926 Nov 11	2251	37.70	175.50	sha	5	3	5.0D	-	C	N	N	340	45	U	17, 34
12	1929 Mar 9	1050	42.80	171.90	15?	11	0	6.95B	7.05	C	S	M	65	90	178	16, 34, 26
13	1929 May 7	1614	40.00	175.50	cru	10	8	5.45B	5.22	C	S	M	32	U	U	17, 34
14	1929 Jun 16	2247	41.70	172.20	15?	9	0	7.72B	7.79	C	R	M	5	45	69	13, 34
15	1931 Feb 2	2246	39.30	177.00	30	17	5	7.79B	7.83	C	R	M	40	45	U	14, 34
16	1932 May 5	0823	39.60	176.90	25?	20	17	6.05B	5.87	S ₆₀	N	M	40	U	U	16, 34
17	1932 Sep 15	1355	38.90	177.60	30?	13	3	6.79B	6.87	C	S	M	56	82	167	16, 34, 27
18	1934 Mar 5	1146	40.51	176.29	12	10?	0	7.3C	7.56	C	S	M	40	90	U	16, 34, 27
19	1934 Mar 15	1046	39.31	177.17	10	40	35	6.42B	6.37	S ₆₀	N	M	35	U	U	16, 34
20	1938 Dec 15	0912	40.00	177.00	30?	25	22	5.76B	5.61	S ₆₀	N	M	36	U	U	16, 34
21	1938 Dec 16	1721	45.00	167.00	60	47	40	7.06B	7.05	S ₆₀	R	F	117	88	72	16, 34, 26
22	1942 Jun 24	1116	40.90	175.90	15	12	4?	7.07B	7.16	C	S	M	35	86	170	16, 34, 27
23	1942 Aug 1	1234	41.00	175.80	43	40	35	6.96B	6.97	S ₆₀	N	M	28	67	293	16, 34, 27
24	1943 Feb 17	0215	45.19	167.04	33R	36	32	6.37B	6.33	S ₆₀	S	F	50	52	15	17, 34, 26
25	1943 Aug 2	0046	46.27	166.72	31R	31	27	6.57B	6.58	C	S	F	34	53	6	17, 34, 26
26	1946 Feb 12	0616	39.79	174.85	132	132	132	5.77B	5.14	S _D	U	D	40	U	U	17, 34
27	1946 Jun 26	1234	43.46	171.26	12R	9	5	6.32B	6.37	C	S	M	44	89	180	12, 34, 26
28	1948 May 22	1921	42.48	172.99	12R	4	1	6.36B	6.43	C	S	M	60	90	177	18, 34, 26
29	1951 Jan 10	1915	42.79	173.18	12R	10	7	5.93B	5.89	C	R?	M	30	U	U	18, 34
30	1951 Feb 10	0327	40.21	177.04	33R	20	15	6.21B	6.20	S ₆₀	N	M	35	U	U	12, 34
31	1951 Apr 23	0650	37.53	177.84	80R	80	76	5.99B	5.65	S _D	U	D	40	U	U	12, 34
32	1951 Jun 24	0441	39.46	176.20	33R	25	23	5.51B	5.25	C	S?	M	27	U	U	12, 34
33	1952 Aug 28	1040	39.99	176.96	12R	30	26	6.01B	5.91	S ₆₀	U	M	35	U	U	12, 34
34	1953 Jul 4	0207	38.86	175.68	12R	5	3	5.28B	4.99	C	N	VM	0	45	U	12, 34
35	1956 Jan 30	0843	37.10	177.42	12R	6	0	6.34B	6.37	C	N	V	35	65	U	12, 34
36	1956 Mar 2	2243	38.90	175.80	12R	5	3	5.0D	-	C	N	V	U	45	U	18, 34
37	1957 Feb 22	0030	39.19	175.14	5R	15	13	5.25B	4.88	C	N	M	35	45	U	18, 34
38	1957 Aug 11	0512	39.05	175.97	99	99	96	5.52B	4.90	S _D	U	D	40	U	U	17, 34
39	1958 Jan 31	0632	39.85	176.58	12R	35	33	5.46B	5.12	S ₆₀	N	M	36	U	U	18, 34
40	1959 May 22	0657	41.07	174.30	33R	40	40	5.29B	4.83	C	U	M	40	30	U	12, 34
41	1960 Feb 3	0221	37.61	178.04	144	144	144	5.79B	5.11	S _D	U	D	40	U	U	18, 34
42	1960 May 24	1446	44.17	167.73	50R	9	4	6.46B	6.53	C	N	F	265	67	260	18, 34
43	1962 Jan 23	0649	38.58	174.80	12R	15	13	5.01B	4.47	C	N	N	0	45	U	12, 34
44	1962 May 10	0027	41.67	171.44	12R	8	5	5.91B	5.88	C	R	M	55	40	88	12, 34, 26
45	1962 Oct 15	2336	43.54	169.77	12R	20	18	5.57B	5.36	C	S	M	55	U	U	12, 34
46	1963 Apr 12	0841	38.71	176.76	12R	20	17	5.82B	5.71	C	S	M	15	U	U	12, 34
47	1963 Dec 22	1335	35.10	173.50	12R	6	4	4.93B	4.36	C	N	N	64	45	U	12, 34
48	1964 Mar 8	0135	44.30	167.87	12R	5	2	5.89A	5.80	C	S	F	262	83	147	12, 21
49	1965 Apr 11	0011	42.74	174.10	12R	16	13	6.09A	5.75	S ₆₀	N	M	110	55	297	12, 21
50	1965 Jun 15	0920	37.90	177.53	33R	50	47	5.62B	5.30	S ₆₀	N	M	37	U	U	12, 22

Table 1 continued:

No	Date	Time	°S	°E	h	h _c	h _t	M _w	M _s	T	FM	Rn	θ	β	γ	Refs
51	1965 Dec 8	1805	37.09	177.50	156	156	156	6.11A	-	S _D	U	D	40	U	U	12, 22
52	1966 Mar 4	2358	38.45	177.91	33R	24	21	5.64A	5.77	I	S	M	197	15	8	18, 22
53	1966 Apr 23	0649	41.63	174.40	22	19	16	5.75A	5.62	C	R	M	232	66	133	12, 21
54	1968 May 23	1724	41.76	171.96	15	10	0	7.23A	7.41	C	R	M	27	45	103	12, 21
55	1968 Sep 25	0702	46.49	166.68	12R	4	0	6.27A	6.15	C	S	M	60	85	182	18, 21
56	1968 Nov 1	0132	41.62	175.05	33R	25	23	5.35B	4.99	S ₆₀	N	M	37	U	U	12, 34
57	1971 Aug 13	1442	42.13	172.10	12R	9	6	5.70A	5.55	C	S	M	242	83	207	18, 21
58	1972 Jan 8	2133	37.57	175.69	12R	7	5	5.26B	4.94	C	N	N	335	45	U	12, 34
59	1973 Jan 5	1354	39.04	175.25	160	149	149	6.57A	6.27	S _D	R	D	40	U	U	12, 22
60	1974 Sep 20	1948	44.40	167.99	12R	25	23	5.55B	5.30	C	R	F	55	45	U	12, 34
61	1974 Nov 5	1038	39.65	173.63	12R	17	14	5.44A	5.45	C	S	M	195	41	213	18, 22
62	1975 Jan 4	2037	40.77	174.67	72	72	70	5.24B	4.56	S _D	U	D	40	U	U	12, 34
63	1975 Jun 10	1011	40.34	175.93	33R	38	36	5.62A	5.14	C	N	M	250	56	257	18, 22
64	1976 May 4	1356	44.67	167.45	12R	10	5	6.51A	6.38	C	R	F	48	47	130	12, 21
65	1976 Oct 27	2057	37.83	176.34	12R	5	3	5.0D	-	C	N	V	20	45	U	12, 34
66	1976 Dec 5	0457	38.17	175.51	1A	5	3	5.0D	-	C	N	N	335	45	U	12, 34
67	1977 Jan 18	0541	41.73	174.30	33R	34	30	6.02A	5.93	S ₆₀	N	M	213	68	251	12, 22
68	1977 May 31	1850	37.88	176.81	9	6	4	5.39B	5.15	C	N	V	30	45	U	12, 34
69	1979 Oct 12	1025	46.69	165.74	12R	12	5	7.23A	7.24	C	R	M	7	24	120	12, 21
70	1982 Sep 2	1558	39.74	176.93	46	31	29	5.38B	5.02	S ₆₀	N	M	35	U	U	18, 34
71	1983 Dec 14	2056	38.36	176.33	5R	3	1	5.06B	4.62	C	N	V	30	45	U	12, 34
72	1984 Mar 8	0041	38.31	177.29	75	80	76	5.91A	5.40	S _D	R	D	40	U	U	18, 22
73	1984 Jun 24	1329	43.60	170.56	5R	13	9	6.12A	6.07	C	S	M	62	83	206	18, 21
74	1985 Jul 19	1433	38.72	177.30	41	31	28	5.92A	5.90	S ₆₀	N	M	213	73	255	17, 22
75	1987 Mar 2	0142	37.88	176.84	10R	6	0	6.53A	6.62	C	N	V	49	45	257	12, 22
76	1988 Jun 3	2327	45.10	167.17	57	60	54	6.69A	6.50	S ₆₀	R	F	281	U	118	12, 21, 31
77	1989 May 31	0554	45.27	166.88	23R	24	20	6.33A	6.26	I	S	F	26	48	148	12, 21
78	1990 Feb 10	0327	42.25	172.65	13	8	2	5.93A	6.04	C	S	M	55	89	163	12, 21, 28
79	1990 Feb 19	0534	40.38	176.22	24	27	24	6.23A	6.45	S ₆₀	N	M	35	70	316	18, 22, 29
80	1990 May 13	0423	40.35	176.23	12	13	9	6.37A	6.40	C	R	M	40	40	111	17, 22, 30
81	1993 Apr 11	0659	39.74	176.52	35	24	22	5.63A	5.54	I	R	M	45	15	U	17, 23
82	1993 Aug 10	0051	45.21	166.71	5R	20	18	6.81A	6.98	I	R	F	32	15	U	18, 22
83	1993 Aug 10	0946	38.52	177.87	36	39	30	6.19A	6.17	S ₆₀	S	M	354	U	U	18, 22
84	1995 Mar 22	1943	41.05	174.18	90	90	86	5.83A	-	S _D	S	D	40	U	U	17, 25
85	1998 Jul 09	1445	30.75	180.86	U	146	146	6.91A	U	S _D	U	D	U	U	U	17, 25
86	Dummy A					7	0	7.0A		C	N	M				
87	Dummy B					7	0	7.0A		C	N	V				

Notes:

T	=	Tectonic Type (C, S ₆₀ , I)	θ	=	strike
FM	=	Focal Mechanism (N, R, S)	β	=	dip
Rn	=	Region (D, F, M, N, V)	γ	=	rake
U	=	Undefined or Unknown			

It is noted that while the actual isoseismals were of course not pure ellipses many of them were reasonably elliptical and nearly symmetrical about the fault strike (e.g. Figure 1).

The depth to the top of the fault rupture was estimated in a number of ways, depending on the information available, such

as surface rupturing (12 events), or from aftershock studies (Events 76, 78-83) [28-33]. For events where such information was not available, h_t was estimated from

$$h_t = h_c - 0.5 h_f \quad (2)$$

where $h_f = W \sin \beta$
 where W = fault width
 and β = fault dip.

For the purposes of the above calculations W was taken as 4, 7, 11, 14 km for M_w 5.5, 6.0, 6.5, 7.0 respectively, except that

for events of $h_c > 100$ km it was assumed that $h_i = h_c$. This model for W was considered to be reasonably representative of widths of different styles of faulting in interplate regions, but is not based on any specific model.

Figure 2: Map of locations of earthquakes considered in this study (Except Event 85, which is at latitude 30.75°S). The regions considered for differences in attenuation, CVR, Fiordland and North-west are indicated, as are events of Crustal, Slab ($h_c \leq 60$ km), Deep Pacific Slab ($h_c \geq 70$ km) and Interface origins.

The values adopted for h_i and β are given in Table 1. The values for a , b , c and b/a are given in Table 2, weakly constrained values being omitted (except in a few cases where such data were found to be adequate for bolstering small subsets (CVR and Deep) or helping near source control of the models.)

It is noted that for calculating the source distance r_b for earthquakes classified as Deep, the approximation is made that the horizontal offset of the source from the Major Axis (Figure 5) is assumed to be zero.

The isoseismal map of the M_w 7.2 1979 Puysegur Bank earthquake (Figure 6) illustrates some issues relating to measurement of isoseismal dimensions a , b and c . In this case the epicentre is offshore and the strike angle $\theta = 7^\circ$ is

such that the whole of the strike vector lies well offshore. No values of a have been measured because to do so would have involved considerable extrapolation of the isoseismals. Thus the only dimensions deemed well enough constrained for use as data were b and c for MM5 and c for MM4. While the use of this map provides data from a valuable (large magnitude) event, the distance data is less well defined than for a comparable land-based event because b and c are measured from the epicentre (because it is offshore), rather than being half the isoseismal dimension. In general epicentres are not at the centre of their associated isoseismal patterns, although in many cases they are within small distances of it. In addition there is a locational uncertainty in the epicentre, which is likely to be 10 km or so. However, overall these sources of error amount to only 5-10% of the values of a , b and c .

Figure 3: Depth h_c , focal mechanism and M_w for the earthquakes considered in this study.

Table 2: Isoseismal distance and shape data used in modelling studies

[illegible]

No.	Date & Region	MM2				MM3				MM4				MM5				MM6				MM7				MM8				MM9				MM10				MM11		
		a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	b/a	c	a	b	c				
32	1951 Jun 28 M									241	135	0.6	181	131	71	0.5	96.2					#	#	#																
33	1952 Aug 28 M									112	112	1.0	112	62	67	1.1	65					#	#	#																
34V	1953 Jul 04 V					40	40	1.0	40											18				10																
34M	1953 Jul 04 M					158	158	1.0	158								32								#	#	#													
35	1956 Jan 30 V					157			162	120			129																#	#		#								
36	1956 Mar 02 V									27	27	1.0	27	12.5	12.5	1.0	12.5	7	7	1.0	7																			
37	1957 Feb 22 M									116	83	0.7	98	71	37	0.5	51																							
38	1957 Aug 11 M					290	92	0.3	163	#	#		#																											
39	1958 Jan 31 M													66	55	0.8	60	35	24	0.7	29																			
40	1959 May 22 M					181				125				75	87	1.2	81	#	#		#																			
41	1960 Feb 03 M									297	90	0.3	163	#	#		#																							
42	1960 May 24 F									288				125																										
43	1962 Jan 23 N									179	138	0.8	157	58	53	0.9	55					#	#		#															
44	1962 May 10 M												170				100				55																			
45	1962 Oct 15 M									143	127	0.9	134	51	65		58					#	#		#															
46	1963 Apr 12 M					208				131	125	1	128	67	54	0.8	60	35	29	0.8	32	#	#		#															
47	1963 Dec 22 N													33	25	0.8	29	19	14	0.7	16	12	6.9	0.6	9															
48	1964 Mar 08 F									225	233	1	229	166	141	0.9	153								#	#		#												
49	1965 Apr 11 M									176	134	0.8	153	91																										
50	1965 Jun 15 M					333				196				71	50	0.7	60																							
51	1965 Dec 08 M					778	192	0.25	386	579	100	0.2		#	#		#																							
52	1966 Mar 04 M									125	150	1.2	137	62	65	1.05	63	22	22	1.0	22																			
53	1966 Apr 23 M									200				113	95.0	0.84	104	50							#	#		#												
54	1968 May 23 M													341				162	112	0.7	135	112	58	0.5	81	56	34	0.6	44	26	21	0.8	23	12	8	0.65	9	3*	#	#
55	1968 Sep 25 M									300				142															#	#		#								
56	1968 Nov 01 M									187	204	1.1	195	60	87	0.7	73	29	29		29																			
57	1971 Aug 13 M									237				135	118	0.9	126								#	#		#												
58	1972 Jan 08 N									115	95	0.8	105	69	44	0.6	55	19	19	1.0	19	6.8	6.8	1.0	6.8	#	#		#											
59	1973 Jan 05 M									458	170	0.37	279	268	100	0.4	164	131	47	0.4	78	#			#															
60	1974 Sep 20 F					316				187	137	0.7	160									#	#		#															
61	1974 Nov 05 M									240	255	1.06	247	120	76	0.64	96																							
62	1975 Jan 04 M									204	137	0.7	167	125	65	0.5	90	#	#		#																			
63	1975 Jun 10 M													71	50	0.7	59	25	21		23																			

	Date & Region	MM2				MM3				MM4				MM5				MM6				MM7				MM8				MM9				MM10				MM11																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																					
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Note: * Two dummy events, see text

For earthquakes classified as Deep, the strike of the rupture is not relevant to the orientation of the isoseismal patterns. The need to model them separately from shallower events has previously been acknowledged by Smith [2, 3]. Earthquakes with sources in the subducting Pacific plate with $h_c \geq 70$ km have their isoseismal patterns controlled by the direction of the plate boundary. Figure 5 shows the isoseismal pattern of the largest of the Deep events in the data set, i.e. the M_w 7.3 Bay of Plenty earthquake of 1914. The source of this event was 300 km deep with its epicentre near Tauranga [11]. While the strike of the rupture is unknown, the major axis of the isoseismals is oriented at about $N40^\circ E$, as is the case for all of the other nine Deep events in the data set. For comparison, the rupture strikes of some of the Deep events are $\theta = 332^\circ$ (No. 51), 142° (No. 59) and 288° (No. 72). For the purposes of this study the strike of the Deep earthquakes has been taken to be parallel to the plate boundary, a value of

$\theta = 40^\circ$ being adopted throughout, and the dimensions a , b and c are related to this strike. As illustrated in Figure 7, a feature of these Deep earthquakes is that their effective surface strike (isoseismal major axis) is constrained by the highly attenuating mantle wedge above the dipping slab, to be in approximately the same location for all events, i.e. within about ± 30 km either side of a line passing through about Hicks Bay and Blenheim (line AA in Figures 2 and 7). It is noted that past earthquakes in the same depth range in the subducting Australian plate under Fiordland have not had narrow isoseismals, e.g. the 1965 May 20 (M_L 5.6, h 105 km) and the 1969 January 2 (M_L 5.8, h 135 km) earthquakes [12]. This has occurred because the Australian plate is dipping so steeply that there is little interference by the mantle to seismic waves propagating to the surface.

Figure 4: Geometry of the isoseismal distance parameters in relation to the earthquake source rupture.

Figure 5: *Isoseismal map of the 1914 Bay of Plenty earthquake ($M_W 7.3$, $h_C 300$ km), showing alignment of major axis with tectonic plate boundary, typical of all Deep ($h_C \geq 70$ km) events in the subducting Pacific plate.*

A further important issue relates to the CVR. Distances a , b and c for CVR events were only included if at least 90 percent of the source to isoseismal distance lay within the CVR (Figure 2). For events with sources located outside of the CVR, isoseismal distance data incorporating travel paths across the CVR were omitted, e.g., the MM4 b dimension in Figure 5 was not used.

To illustrate the range of the data in graphical form, the isoseismal source distance data as defined in Equation (1) are plotted against M_W in Figure 8, in terms of focal mechanism, intensity, and region. The total numbers of isoseismal source distance data are 300, 260, and 258, for r_a , r_b , r_c respectively. Of these, three isoseismals are defined by landslides only, i.e. MM10 for Events Nos 1 and 14 and MM9 for Event 12, adapted from maps by Hancox *et al.* [36]. In addition, in order to constrain the models at the centre of the isoseismal patterns, values of intensity at very small horizontal distances (a , b , $c = 0.1$ km) were assigned where such intensities were

likely to be higher than that of the innermost isoseismal of a given earthquake. This was only done where such intensities could be inferred as an upper bound to the epicentral intensity from the isoseismal map. For example MM11 was assigned at the centre of the 1931 Hawke's Bay earthquake in which MM10 but no MM11 intensities were observed, although MM11 could possibly have just been reached [14]. Also for the 1956 Jan 30 earthquake of M_W 6.3, MM9 was assigned at its centre which was well offshore in the Bay of Plenty. Such intensities were assigned for 55 of the earthquakes, as indicated by # symbols in Table 2. For 11 of these events the a dimension control adopted was larger than 0.1 km, e.g. for Event No. 1 at MM11 $a = 15$ was assigned as marked * in Table 2. The two dummy events (Nos. 86 and 87) of M_W 7.0 and Normal focal mechanisms were assigned MM10 at a , b , $c = 0.1$ km, on the basis that this data closely matched that modelled for strike-slip events which were well constrained at this magnitude.

Figure 6: Isoseismal map of the 1979 Puysegur Bank earthquake (M_W 7.2, h_c 12 km), showing isoseismal geometry in relation to an offshore epicentre.

Figure 7: Perspective view of the Australian plate and the subducting Pacific plate under the North Island, with the highly attenuating wedge of mantle protecting surface locations vertically above Deep earthquakes, and constraining the effective "strike" of the isoseismals.

4.0 DEVELOPMENT OF THE ATTENUATION MODELS

Attenuation models were developed for subsets of the data by applying multiple regression modelling to the MM intensity data with moment magnitude and functions of source distance being the primary explanatory variables, but with centroid depth, tectonic region, tectonic type, focal mechanism and rake also being considered for inclusion in the models. Terms were included or excluded based on the Akaike Information Criterion (*AIC*) statistic [37], given by

$$AIC = -2 \log L + 2k \quad (3)$$

where $\log L$ denotes the log likelihood of the fitted model and k the number of fitted parameters. In the *AIC* method, a term is

included in the model only if it lowers the value of *AIC*, i.e. if the improved fit produces an increase in the likelihood great enough to overcome the penalty of 2 for introducing an extra parameter. Typically a term needs to have an associated *t*-value (ratio of the parameter to its standard error) somewhat greater than one to be included using the *AIC* criterion.

Separate models were developed for the source distances r_a , r_b and r_c . For a given subset of the data, the same model was chosen for each of the distances wherever it was reasonable to do so. This meant an adjustment to the strict application of the *AIC* method, in which a term of marginal value was included in all three models if it lowered the value of *AIC* in two of them. In cases where a term clearly improved the fit for one distance, but clearly not for the other two, different models were allowed.

Figure 8: The data for source distance r_c plotted against M_w , showing (a) focal mechanisms, (b) intensities and (c) regions.

The regression model used was the random effects model [38] in which the error variance is decomposed into a within-earthquake component σ^2 and a between-earthquake component τ^2 . When considering adequately large subsets of the data, the enhancement of the random effects model to allow for individual earthquake magnitude uncertainties [39] was used. This model allows magnitude data of relatively low quality to be included in the analysis and given the weight that they are due. It extracts the effect of magnitude uncertainties from the between-earthquake component of variance. For class A, B, C and D magnitudes, standard errors of 0.1, 0.15, 0.3 and 0.3 respectively were adopted. The value of 0.1 for class A standard errors reflects the very small standard errors published for systematic estimates of M_w (e.g., Urhammer et al., [40]). The value of 0.15 for class B is based on an estimate of Dowrick and Rhoades [34] for this class. The value of 0.3 for class C is based on typical errors in Wells and Coppersmith [41] regressions of magnitude on fault rupture length and displacement. The value for class D is based on the authors' judgement.

A nonlinear function of source distance, denoted $\log D$, was applied in the regression modelling. It has the form

$$\log D = \log_{10}[(r^2 + d^2)^{1/2}] \quad (4)$$

where $r = r_a, r_b$ or r_c and d is a parameter to be estimated. The use of $\log D$ as a regression term rather than $\log r$ improved the fit to the near source attenuation intensity data, and a non-zero value of d proved to be useful for all subsets except the deep earthquakes considered by themselves.

Indicator variables were used to include the effects of factors, such as focal mechanism, tectonic type and tectonic region, in the regression models. For a given level L of a factor F , the indicator variable δ_L is defined by $\delta_L = 1$ if $F = L$ and 0 otherwise.

5.0 DISCUSSION OF THE ATTENUATION MODELS

5.1 Functional form

The random effects models fitted are of the form

$$I = \alpha + \beta M_w + \gamma \log D + \sum_{l=1}^m \theta_l x_l + \zeta + \varepsilon \quad (5)$$

where I is Modified Mercalli intensity, M_w is moment magnitude, $\log D$ is as defined above, x_l , $l=1 \dots m$ are additional regression terms, ζ is a normally distributed random variable which is determined by the earthquake and has variance $\beta^2 s^2 + \tau^2$ where s is the earthquake magnitude standard error, and ε is a normally distributed random variable with mean zero and variance σ^2 . The parameters to be estimated are d , α , β , γ , θ , $l=1 \dots m$, τ and σ .

Models were developed for three different subsets of the data, namely:

- (1) all data with known focal mechanisms, excluding only Deep events;
- (2) the Main Seismic Region, excluding Deep earthquakes;
- (3) Deep earthquakes.

Magnitude uncertainties were included for subsets (1) and (2), but for subset (3) the number of data were insufficient to support the fitting of the uncertain magnitudes model. The parameter estimates, standard errors, residual standard deviations, and the number of data for each model are given in Tables 3-5.

5.2 Goodness of fit and robustness

The residual standard deviation of the models (calculated as $(\tau^2 + \sigma^2)^{1/2}$) is typically about 0.5 units of MM Intensity (listed in Tables 3-5), with τ being consistently less than σ . A small value of τ limits the extent to which the inclusion of further variables which are functions of the earthquake source, such as focal mechanism or tectonic type, can improve the fit, because such terms can serve only to reduce τ and not σ .

The standard errors of the parameters provide insight into how well the models are determined. These values are given in Tables 3-5. Overall, along with the residual standard deviation, they show that Models 1 and 2 are more robust than Model 3, and when considered as coefficients of variation (c.o.v. = Standard Error divided by Estimate) they indicate the robustness of the individual parameters. For example the least robust term in Model 1 is h_c (direction a) which has c.o.v. = 0.54, with the c.o.v values for the other terms ranging down to as little as 0.03 (for $\log D$).

One measure of the goodness of fit of the models can be gained by examining plots of the residuals against fitted variables, where the residuals are defined as the difference between the actual and fitted value of I , to test if the assumptions of linearity are approximately satisfied. This has been done exhaustively for all models and in no case is there strong evidence of a trend in the relation between residuals and fitted values of I , or between the residuals and regression variables. Examples of such residual plots are given in Figure 9. The four plots for Model 2 (Main Seismic Region) show the model fitting the data well, with the residuals spread reasonably symmetrically about zero over the full ranges of distance, magnitude and depth as well as fitted intensity. Figure 9 also shows residuals for fitted intensities in the direction of strike for the CVR, North-west region, Fiordland and Deep earthquakes. These plots show that these four subsets are reasonably well fitted by the respective models.

It is of much interest to examine the fit of the models to the data from individual earthquakes, especially in cases where more than one different model may be applied to a particular earthquake. Examples of such fits are presented in Figures 10 and 11. Figure 10 shows how Models 1 and 2 fit the data of four of the largest (M_w 7.1-7.8) shallow Main Seismic Region earthquakes. Three of these earthquakes are Crustal Reverse events, while the other occurred in the Slab ($h_c = 40$ km) with a Normal focal mechanism. All of these events have several data in both the a and b directions so that the fit with distance for each event can be seen individually. The fit in all cases is good, the residuals of most data points being less than the typical standard deviation of 0.5 of an intensity unit. Note that the MM10 data point for the 1929 earthquake is an estimate based on landslide data only. It is also of interest to observe that for these Main Seismic Region events the three models predict very similar intensities over the full distance range ($a = 1$ to 500 km).

Table 3: Model 1 "Focal mechanisms"

Parameter Estimates and standard errors in random effects regression models of MM Intensity (allowing for magnitude uncertainties) using only earthquakes with known mechanisms and excluding the Deep region. Model selection was based on the AIC statistic.

$$I = A_1 + (A_2 + A_{2R} \delta_R + A_{2V} \delta_V) M_W + (A_3 + A_{3S} \delta_S + A_{3V} \delta_V) \log_{10} \sqrt{(r^2 + d^2)} + A_4 h_C + A_5 \delta_I$$

where	δ_R	=	1 for Reverse focal mechanisms 0 for all other events
	δ_S	=	1 for Strike-slip focal mechanisms 0 for all other events
	δ_V	=	1 for CVR events 0 for all other events
	δ_I	=	1 for interface events 0 for all other events

Distance		a		b		c	
Term	Estimate	Std.Err.	Estimate	Std.Err.	Estimate	Std.Err.	
A_1	4.87	0.46	4.97	0.42	4.92	0.40	
A_2	1.25	0.07	1.03	0.07	1.12	0.07	
A_{2R}	0.048	0.022	0.058	0.020	0.061	0.019	
A_{2V}	0.21	0.06	0.18	0.06	0.20	0.05	
A_3	-3.77	0.09	-3.39	0.09	-3.54	0.09	
A_{3S}	0.16	0.06	0.23	0.06	0.25	0.06	
A_{3V}	-1.47	0.23	-1.35	0.23	-1.32	0.21	
A_4	0.0083	0.0041	0.015	0.004	0.013	0.004	
A_5	-0.68	0.23	-0.41	0.18	-0.63	0.19	
d	9.63	0.95	5.74	0.61	7.30	0.68	
τ	0.26	0.02	0.21	0.02	0.20	0.02	
σ	0.37	0.01	0.36	0.01	0.34	0.01	
Residual s.d.	0.45		0.42		0.39		
N**	262		231		230		
Neq*	73		74		72		

Notes: ** N = Number of source-to-isoseismal distance data
* Neq = Number of earthquakes

Figure 11 shows how various of the models fit the data of four small to medium-sized (M_W 5.3-6.6) earthquakes from four different regions. Our preferred models fit the data well in all four events. Event 58 (M_W 5.3, h_C 7, Normal) from the North-west Sub-region is seen to be fitted well by the Main Seismic Region Models 1 and 2 and not Model 1 volcanic, i.e. this event has the attenuation characteristics of the Main Seismic Region and not of the CVR. Event 80, (1990 Weber II) shows that the data of an M_W 6.4, h_C 13, Reverse Main Seismic Region earthquake is well fitted by Models 1-2. Event 59 (1973 Central Nth Island) M_W 6.6, h_C 149, is typical of Deep events (Figure 5), with data from both direction a and b being fitted reasonably

well by Model 3. The typical high attenuation of the CVR earthquake in Figure 11 (Event 68) is seen to be well modelled by Model 1 (Volcanic), but not by the model for Normal mechanism Main Seismic Region events.

Similar plots to those shown in Figures 10 and 11 have been made for all events in the database, an examination of which shows the fit to the preferred model to be satisfactory throughout. As is consistent with the functional form of the models, there is no sign either of magnitude saturation effects or that the attenuation rate varies with magnitude.

Table 4: Model 2 "Main Seismic region"

Parameter Estimates and standard errors in random effects regression models of MM Intensity (allowing for magnitude uncertainties) using data from the main seismic region (including the Northwest and Fiordland sub-regions) and excluding Deep earthquakes. Model selection was based on the AIC statistic.

$$I = A_1 + A_2 M_W + A_3 \log_{10} \sqrt{(r^2 + d^2)} + A_4 h_C + A_5 \delta_C$$

where $\delta_C = \begin{matrix} 1 & \text{for Crustal events} \\ 0 & \text{for all other events} \end{matrix}$

Distance		a		b		c	
Term	Estimate	Std.Err.	Estimate	Std.Err.	Estimate	Std.Err.	
A_1	4.25	0.46	4.20	0.43	4.12	0.42	
A_2	1.28	0.07	1.11	0.06	1.18	0.06	
A_3	-3.73	0.09	-3.27	0.09	-3.46	0.09	
A_4	0.017	0.005	0.021	0.005	0.021	0.005	
A_5	0.54	0.13	0.31	0.13	0.47	0.13	
d	10.05	1.10	5.65	0.70	7.31	0.75	
τ	0.24	0.02	0.20	0.03	0.21	0.02	
σ	0.39	0.02	0.40	0.02	0.36	0.02	
Residual s.d.	0.46		0.44		0.41		
N**	236		210		205		
Neq*	64		65		63		

Notes: ** and * see Table 3.

Table 5: Model 3 "Deep region"

Parameter Estimates and standard errors in random effects regression models of MM Intensity using Deep earthquakes under the main seismic region only. Model selection was based on the AIC statistic.

$$I = A_1 + A_2 M_W + A_3 \log_{10} r + A_4 h_C$$

Distance		a		b		c	
Term	Estimate	Std.Err.	Estimate	Std.Err.	Estimate	Std.Err.	
A_1	4.00	1.74	10.08	3.78	5.60	2.96	
A_2	1.63	0.35	1.77	0.58	1.83	0.51	
A_3	-4.03	0.34	-8.02	1.11	-5.56	0.62	
A_4	0.0044	0.0029	0.012	0.006	0.0058	0.0047	
τ	0.18	0.12	0.38	0.08	0.34	0.07	
σ	0.49	0.06	0.52	0.07	0.44	0.06	
Residual s.d.	0.52		0.64		0.56		
N**	32		25		24		
Neq*	11		10		10		

Notes: ** and * see Table 3.

Figure 9: Plots of residuals for the Main Seismic Region using Model 2 (considering fitted intensity, r_a , M_w and h_c), and residuals of fitted intensities for Models 1-3 (considering the CVR, Deep events the North-west region and Fiordland). The data for individual earthquakes are connected by lines and identified with the Event Number.

Figure 10: *Plots of Models 1 and 2 against a and b data for four large earthquakes, M_w 7-7.8, h_c 9-40 km. The header to each plot starts with event number and ends with region (M = Main Seismic Region) followed by focal mechanism (R or N). For Event 14 the MM10 distance data (circled) were estimated from landslides only.*

58 : 1972 Jan 8 Mw 5.26 hc 7 N N

58 : 1972 Jan 8 Mw 5.26 hc 7 N N

80 : 1990 May 13 Mw 6.37 hc 13 M R

80 : 1990 May 13 Mw 6.37 hc 13 M R

59 : 1973 Jan 5 Mw 6.57 hc 149 D R

59 : 1973 Jan 5 Mw 6.57 hc 149 D R

68 : 1977 May 31 Mw 5.39 hc 6 V N

68 : 1977 May 31 Mw 5.39 hc 6 V N

..... Model 1, - - - - - Model 2, - - - - - Model 3, - - - - - Model 1 Volcanic, - - - - - Model 1 Normal

Figure 11: Plots of Models 1-3, where appropriate, against a and b data for events of M_w 5.3-6.6 from four regions, Main Seismic Region (M), North-west (N), CVR (V) and Deep earthquakes (D).

A limitation of all of the models is that they only apply within the range of the data used in fitting them. Outside of that range, the errors are likely to be much increased and systematic. Except for the Main Seismic Region, the range of the existing data within any particular region is quite restricted. For example, the largest magnitudes and the number (N_{eq}) of events in some of the data sets are:

- MSR Crustal M_w 8.2, N_{eq} 51 (+ Dummy M_w 7.0 Normal)
- MSR Slab (excl. Deep) M_w 7.0, N_{eq} 19
- Deep M_w 7.3, N_{eq} 11
- CVR M_w 6.5, N_{eq} 10 (+ Dummy M_w 7.0)

Finally it is noted that the two dummy events, with intensity data at source only, were found to have the desired effect of controlling the near source modelling of large very shallow Normal events in the CVR and Main Seismic Region, without perturbing the fit of the models to the data.

5.3 Effect of depth

Centroid depth h_c was considered as a candidate explanatory variable and emerged as a significant term with a positive coefficient in all models. When the Deep earthquakes ($h_c \geq 70$ km) are considered on their own (Model 3), the coefficients for h_c for modelling a , b and c are all substantially smaller than those obtained for shallow earthquakes (for direction a , 59% and 28% of those for Models 1 and 2 respectively). In a recent study [35] of attenuation of peak ground accelerations in New Zealand, a term for h_c was also found to be significant.

The effect of depth on MM intensity attenuation is illustrated by Figure 12, in terms of Models 2 and 3, for Crust, Slab (S_{60}) and Deep earthquakes of M_w 6.5 and depths ranging from 10 to 150 km. The tendency for deeper events to have higher intensities for the same magnitude and source distance, as predicted by the h_c term, is clearly seen. For the same M_w and source distance, the models predict that the intensities in direction a are about 1.3 intensity units higher for events with h_c 150 km than with h_c 10 km, while the comparison between depths of 50 and 10 km gives an intensity increase of about 0.3 intensity units for events of the same focal mechanism.

5.4 Effect of tectonic type

Indicator variables for Crustal and Slab (S_{60}) tectonic types were included as candidate explanatory variables in Model 2 and Interface events in Model 1 (Crust, Slab and Interface are described in Figure 7). A positive crustal term appears in Model 2, the parameter being largest for the a direction. This helps the model reflect the small b/a ratios of Crustal earthquakes of large magnitude.

It seems likely that the positive Crustal term in Model 2 is accounted for by the fact that the Crustal events database contains proportionately more Reverse (22%) and Strike-slip (42%) events than the Slab (S_{60}) database (10% each), because of the higher intensities produced by events with those mechanisms as demonstrated by Model 1 (discussed later).

Regarding Interface earthquakes, a negative term appears in Model 1, predicting intensities about half an intensity unit

smaller than other events of the same mechanism, M_w , h_c and source distance. It is of interest that a study [35] of New Zealand peak ground accelerations found that the PGAs induced by the small subset of seven Interface events were weaker than those predicted for Crustal or Slab (S_{60}) events of the same M_w , h_c and distance, which is consistent with the findings of the present study. Using a large international database, Youngs *et al* [42] also found that PGAs from Interface earthquakes were weaker than those of Crustal and Slab events, for the same magnitude and distance.

5.5 Effect of focal mechanism

Indicator variables for Normal, Strike-slip and Reverse focal mechanisms were included as candidate explanatory variables for Model 1, and combination products with magnitude and distance variables (e.g. $\delta_R M_w$ and $\delta_R \log D$) were also included. Such terms allow for shape differences between the attenuation functions for different focal mechanisms to be captured.

The differences indicated by Model 1 are illustrated in Figure 13. It is seen that for $h_c = 10$ km and the same magnitude and source distance, the model predicts generally that Reverse events give higher intensities than Strike-slip which in turn give higher intensities than Normal focal mechanism events, at most distances, and in both a and b directions, and always so near source. In addition the attenuation rate is higher for Reverse and Normal (i.e. dipping fault) events than for Strike-slip earthquakes.

Both of the above results are similar to those of the 1991 Dowrick intensity study [6, 7], except that Normal and Strike-slip events were combined in that study because of the small data set. In their study of New Zealand PGA's, Zhao *et al.* [35] also found that Crustal Reverse events gave stronger shaking than Crustal + Slab (S_{60}) Strike-slip + Normal events (combined). Other studies of strong ground motion, by Spudich *et al.* [43] and Abrahamson & Shedlock [44] suggest that crustal earthquakes with Normal focal mechanisms give smaller motions than Strike-slip. In addition Oglesby *et al.* [45] who carried out dynamic simulations of earthquakes on surface rupturing dipping faults, found that the ground motion from a Reverse fault is at least twice as strong as that from the equivalent Normal fault near the source. The difference between Reverse and Normal ground shaking predicted by our intensity model appears to be not quite so great. If the near source intensities from Model 1 for events of M_w 7.0 and $h_c = 10$ km are converted to PGAs, using a relation in which $\log_{10}$ PGA is proportional to $0.35I$, the ratio of PGA for Reverse to Normal events is approximately 1.5.

5.6 Isoseismal shape

Isoseismal shape varies with several factors, i.e. dip angle, focal mechanism, earthquake magnitude, depth, horizontal distance and MM intensity. Trends in isoseismal shape are best considered in terms of the ratio b/a derived directly from the isoseismal distance data, as given in Table 2. Box plots shown in Figure 14 show the relationship of the shape parameter b/a to other variables in the study. The white portion in the middle of each box plot indicates the median value, the solid black box shows the upper and lower quartiles while other horizontal lines and square bracket symbols indicate outliers. An outlier is any value more than 1.5 box-lengths beyond the end of the box.

Figure 12: Effect of depth and Tectonic location. Intensity plotted against source distance (a) in direction a, and (b) in direction b as predicted by Models 2 and 3 for Crust, shallow Slab (S_{60}) and Deep earthquakes of M_w 6.5 and depth $h_c = 10, 30, 60, 100$ and 150 km. Also note the high attenuation of Deep events in direction b.

Figure 13: *Effect of focal mechanism. Intensity plotted against source distance (a) along strike and (b) normal to strike, as predicted by Model 1 for Strike-slip, Reverse and Normal faulting earthquakes of M_w 5, 6, 7 and 8 and depth $h_c = 10$ km.*

Figure 14: Shape of isoseismals in relation to strike illustrated by box plots of b/a data from Table 2. Variation of b/a is depicted in terms of h_c , focal mechanism, intensity, region and dip angle, comparing "all data" cases with these restricted to the inner isoseismals of events of $M_w \geq 6$. The numbers of data in the subsets are listed above each plot. At the foot of Figures (a) and (b) the numbers indicate 10 km depth increments, e.g. 0 indicates $0 \leq h_c \leq 9$ km. The white portion in the middle of each box plot indicates the median (see text).

For earthquakes other than those classified as Deep, i.e. excluding events with $h_c > 70$ km in the subducting Pacific slab, a general observation regarding isoseismal shape is that they become more circular with increasing source distance, i.e. $b/a \rightarrow 1.0$. Also, the smaller the magnitude the more circular the isoseismals (Figure 14(h)), because small earthquakes approximate closely to being point sources. Thus, considering the innermost isoseismals of larger earthquakes ($M_w \geq 6$) in Figure 14(b) it is seen that the median values of b/a are 0.35 and 0.65 for $h_c < 10$ km and 10–19 km respectively, whereas for h_c in the 30–50 km range the median b/a is close to 1.0. Again in Figure 14(a), where all data are included, it is seen that the effect of the small earthquakes is to increase median b/a to 0.9 for $h_c < 10$ km.

In a semi-empirical modelling study of the spatial distribution of PGAs from fault ruptures of large earthquakes, Midorikawa [46] found similar b/a patterns, with $b/a \sim 0.35$ near the fault of an earthquake of magnitude 7.2.

Figures 14(c) and (d) show how Strike-slip events tend to have more elongated isoseismals than do Normal and Reverse events, particularly for the inner isoseismals. This tendency is an effect of fault dip, β , Strike-slip earthquakes commonly having β in the range 70–90°, while Normal and Reverse events have smaller β (see Table 1). This trend was confirmed in dip angle terms (plots not shown).

When considering b/a as a function of intensity, it is seen in Figures 14(e) and (f) that the median b/a reduces almost linearly from about 1.0 at MM3 to 0.3 at MM10. In contrast to the above cases, median b/a for the inner isoseismal and the "all data" cases is very similar for a given intensity.

Box plots of b/a are shown in Figure 14(g) for the four geographical regions of Figure 1 and the Deep earthquakes. The median b/a values for Fiordland, North-west sub-region and the rest of the Main Seismic Region are similar at about 0.8 to 0.9, and are not statistically significantly different. The CVR has a slightly higher median b/a of close to 1.0, which is artificially high because its b/a data are mostly for low intensities. Deep events have median $b/a \sim 0.35$ reflecting their consistently very elongated isoseismals, as discussed earlier.

The isoseismal shapes indicated by the b/a data, are of course reflected in the attenuation models. This is illustrated in Figure 15 where the isoseismal maps as predicted by Model 1 are shown for four earthquakes of different source characteristics. In general the outer isoseismals have large b/a , and the smaller and deeper events, Figures 15(c) and (d), also have more circular isoseismals. It is noted that the near source spatial distribution of ground motions for large shallow events which is demonstrated by this study, and that of Midorikawa [46], is very different from that implied by the more simplistic modelling currently adopted in standard strong motion attenuation expressions, such as Refs. [35], [42–44].

5.7 Regional effects

In the past, a number of attempts have been made by various researchers [2, 3, 6, 35] to quantify possible regional differences in attenuation. In the present study, comparisons were made of data relating to the four geographical regions and Deep

earthquakes shown in Figure 2. The events in these regions are identified in the Region columns of Table 1 as F (Fiordland), N (North-west), M (rest of the Main Seismic Region), V (CVR) and D (Deep). It was confirmed that there are strong regional attenuation effects differentiating the volcanic region, deep earthquakes, and earthquakes from the other regions. For that reason models were developed for the Central Volcanic Region and Deep earthquakes (Models 1 and 3). The bulk of the data are for what we call the Main Seismic Region and two models were developed for this region, Model 1 for the case where focal mechanisms are known, and Model 2 for the case where they are unknown or unspecified. No significant differences emerged between the Fiordland sub-region and the North-west sub-region and the rest of the Main Seismic Region. The above finding from our 10 Fiordland events is not consistent with that from a smaller data set (four events) for PGAs [35] which suggested that PGAs in Fiordland earthquakes are higher for a given set of parameter values than elsewhere in the country.

The differences that exist between the Main Seismic Region, the Central Volcanic Region and Deep earthquakes are indicated using Models 1 and 3 in Figure 16, in which has been plotted intensity against source distances r_a and r_b for events of $M_w 6.5$. These plots show the much higher attenuation of the CVR compared to the Main Seismic Region and the major axis direction a of the Deep earthquakes. As CVR earthquakes generally have Normal focal mechanisms, a comparison is made using Model 1 of CVR events with Main Seismic Region Normal events in Figure 17. Two types of Normal mechanism event are modelled to have similar at-source intensities, and a large difference in attenuation rate.

The fit of the Deep earthquakes in direction a was not perturbed by the inclusion of the very distant 1998 earthquake near Raoul Island (Event 85), which is in itself well fitted by Model 3. This is important for modelling the influence on hazard in the East Cape, of the belt of Subduction zone earthquakes stretching NNE of East Cape. Also apparent (Figure 16 (b)) is the very high effective attenuation of the Deep earthquakes in the minor axis direction b (N50°W), which models the phenomenon shown in the isoseismal map presented in Figure 5. For Deep earthquakes located in the dipping Pacific slab where it has lost contact with the Australian plate, i.e. north-west of the line A-A on Figures 2 and 7, seismic waves travelling vertically or north-west have to traverse the highly attenuating mantle wedge to reach the ground surface. This effect occurs only on the north-west side of line A-A, attenuation on the south-east side apparently being normal, but insufficient distance data is available to quantify it because most of the area concerned is offshore.

Based on the difference found between attenuation inside and outside the CVR, our data being divided by the CVR boundaries (Figure 2), the geographical boundaries of the high attenuation volcanic region appear to lie close to those of the CVR, except that it is possible that the area north-west, west and south-west of Tauranga (i.e. NW of line B-B on Figure 2) may have normal (Main Seismic Region) attenuation characteristics. The latter proviso is made on the basis of the normal attenuation across that zone exhibited by the 1972 Te Aroha earthquake (Figure 11), and the high attenuation exhibited by the 1976 October 27 Te Puke earthquake in the zone just east of Tauranga.

Figure 15: Illustration of isoseismal dimensions and shapes predicted by attenuation Model 1, for earthquakes of various source parameters. The ratio b/a is seen to increase with (i) source distance, (ii) depth, and (iii) decreasing magnitude.

Figure 16: Effect of region. Intensity plotted against source distance (a) along strike and (b) normal to strike, as predicted by Models 2, 1 and 3 for earthquakes of $M_w 6.5$ in the Main Seismic Region ($h_c = 10$ km) CVR ($h_c = 10$ km) and Deep events ($h_c = 70$ km), respectively.

Figure 17: Comparison of Central Volcanic Region attenuation model with that for surface rupturing Main Seismic Region Normal faulting events.

6.0 COMPARISONS WITH OTHER NEW ZEALAND ATTENUATION MODELS

Because of their widespread use in recent years, it is also of interest to compare the models published in 1991 and 1992 by Dowrick [6, 7] with those of the present study. A direct comparison can be made in terms of focal mechanism, mean isoseismal radius c , and the centroid depth h_c (which equates to the effective depth h_e used earlier [6, 7]). Such a comparison is made in Figure 18 for earthquakes of M_w 5.0 to 8.0, all with $h_c = 10$ km. In comparing the models for Reverse focal mechanism earthquakes, it is seen that the 1991 and the present models are virtually identical near the source throughout the magnitude range, while at distances greater than about 5 km the 1991 model predicts greater intensities (by up to 0.5 intensity units) than the present model, for $M_w \geq 6$. It thus may be inferred from Figures 18(a) and 10, that the 1991 Dowrick model gives estimates of the size of the inner isoseismals of the two large shallow of 1929 and 1931 (Nos. 14 and 15) which are within the range of normal scatter of the better defined isoseismals now available. This is comforting in the light of some earlier concerns [3, 8].

Because the other focal mechanism model of Dowrick (1991) combined Normal and Strike-slip events, it is not fully comparable with the present Strike-slip model. The fact that the

two models are closest at high magnitudes (Figure 18 (b)) may be due to the fact that the weaker Normal events are represented in the 1991 dataset only in the smaller magnitudes i.e. M_s (max) ≤ 6.6 .

The attenuation models of Smith [3] were found to differ substantially from those of this study (by up to ± 1.0 intensity units), the differences apparently arising mainly from the different data used in the two studies, as noted in general terms in Section 1.0 above. Smith did not define depths explicitly, but for shallow earthquakes in the Main Seismic Region classified earthquakes as either (1) Upper Crust or (2) Lower Crust. Of 54 events in these two classes, the classifications for 21 earthquakes appear to be at variance with present information on centroid depths and aftershock patterns, e.g. the 1979 Oct. 12 event of h_c 12 km [21] and the 1990 May 13 event of h_c 13 km [22, 29] were both classed as Lower Crust. In addition the magnitudes used by Smith differed from those used in the present study, which are restricted to moment magnitudes. Of the 107 earthquakes in Smith's data set, the magnitudes of 68 were local magnitudes (M_L) which have been shown [34] to be poor indicators of moment magnitude. This is illustrated by the fact that the M_L values in Smith's data set differ from M_w on average by 0.3 and the differences range from +0.8 to -0.7.

Figure 18: Comparison of attenuation models of Dowrick (1991) with Model 1, for (a) Reverse faulting events, and (b) Strike-slip + Normal versus Strike-slip (Model 1), in terms of mean horizontal distance c , and $h_c = 10$ km.

7.0 ESTIMATING SOURCE DISTANCE

The regression analyses carried out in this study were designed for the estimation of the isoseismal intensity as a function of M_w , distance D , and depth h_c , as in Equation (5). Sometimes, however, it may be desired to estimate other parameters when I is given, e.g. the source distances may be required for drawing scenario maps. A convenient way of doing that is simply to invert Equation (5) to find distance, using the appropriate parameter values given in Tables 1-3.

Thus the horizontal distances a, b, c of Figure 3 may be estimated from an expression of the form

$$a = \left[10^{2(I - A_1 - A_2 M_w - A_4 h_c) / A_3 - d^2 - h_t^2} \right]^{1/2} \quad (6)$$

The above procedure involves an added approximation because strictly speaking the parameters A_1 - A_4 should be estimated from regressions of $\log D$ on I rather than I on $\log D$ as was done for obtaining the estimates in Tables 3-5.

However comparisons of the results obtained carrying out the two regressions, show that the estimates of I or distance are almost identical. This suggests that inversions of Equation (5) may be used to estimate any of the variables without significant extra error.

8.0 VARIABILITY OF LOCAL OBSERVATIONS OF INTENSITY

In New Zealand, standard seismological practice is to draw isoseismals as lines enclosing most of the local observations of intensity of the isoseismal concerned. Thus considering the isoseismal map of the 1979 Puysegur Bank earthquake (Figure 6) it is seen that the MM5 isoseismal encloses all but one MM5 observation, and the MM4 isoseismal contains all of the MM4 observations. As the variability of the local observations is sometimes of interest it was decided to examine the statistical distributions of local observations of intensity from a sample of the events considered in this study. The 17 events chosen were rich in data, having a total of 1130 intensity observations, and ranged in magnitude from 5.3 to 7.8. The number of observations of each level of intensity within each intensity zone were counted. Intensity zones are defined as the areas between adjacent isoseismals, e.g. the MM4 zone lies between the MM4 and MM5 isoseismals (and hence contains mostly MM4 local observations as seen typically in Figure 6).

For each intensity zone estimates were made of

- the mean deviation from the nominated intensity for the zone.
- the standard deviation of the observations from the mean and RMS deviation from the nominated intensity for the zone.

The results of the above calculations are given in Table 6 and an example plot of the deviations from the mean are given for MM8 in Figure 19. First it is seen that the mean of the local intensities is generally slightly lower than the zone intensity (about 0.2 units lower on average). This is a natural

consequence of the way the isoseismals are located in New Zealand practice. In Table 6 it is also seen that the standard deviation of the local observations is 0.64 in relation to the full data set, with no significant trend with intensity. This additional local variability needs to be taken into account if the attenuation relations presented here are used to estimate intensities at particular locations, and not just mean isoseismal intensities, in scenario events.

The statistics in Table 6 incorporate the effects of all ground conditions from the reporting areas, which are alluvial much more frequently than rock in New Zealand. While the data does not allow us to differentiate between different ground classes, it is of interest to note a finding from a study [47] of seismic hazard of the city of Christchurch which has large areas of alluvium of varying degrees of responsiveness. In the 22 intensity maps having isoseismals of MM4 or stronger affecting Christchurch, the local intensities reported from Christchurch alluvial areas (which are mostly firm-stiff) were on average (only) 0.1 of an intensity unit higher than the isoseismal value of the corresponding intensity zone. This supports the contention that New Zealand intensity attenuation models relate to firm to stiff soil sites.

9.0 CONCLUSIONS

As a result of this study, the following conclusions were drawn:

1. Attenuation expressions have been derived for Modified Mercalli intensity isoseismals in New Zealand earthquakes in terms of magnitude M_w and source distance. The expressions allow for the influence of depth, focal mechanism, tectonic type and region, and attenuation along and normal to the strike of the source rupture (a new procedure).
2. All of the models were found to be reasonably robust in statistical terms, and plots of models against the data for each event showed the fit to be good in most cases. The shape of the attenuation curves over long distances is well supported by those events having three or more data. However the models are not recommended for application outside of the range of the data.
3. Distance is accounted for as the logarithm of a function of the distance from the top of the fault rupture to the isoseismal. Three distance measures were modelled, i.e. radius along strike, radius normal to strike and the mean radius.
4. There is no sign either of magnitude saturation effects or that the attenuation rate varies with magnitude despite having data with a large magnitude range, $4.9 \leq M_w \leq 8.2$.
5. Depth is accounted for explicitly as a function of centroid depth. For events with depths down to about 60 km the depth effect is moderate, eg events with depths of 60 km cause predicted mean intensities approximately 0.5 units higher than those of events with depths of 10 km, when magnitude, source distance and focal mechanism are held equal.

Table 6: Deviation of local observations of intensity from the "nominated" intensity of the zone

MMI Zone	Mean dev.	Std dev.	RMS dev.	n
4	-0.07	0.55	0.56	295
5	-0.18	0.66	0.68	361
6	-0.19	0.67	0.70	174
7	-0.24	0.83	0.86	164
8	-0.16	0.51	0.54	111
9	-0.24	0.52	0.57	25
All	-0.18	0.64	0.66	1130

Figure 19: Probability of deviations of local intensity from the nominated intensity of their zone.

6. Shallow earthquakes ($h_c \leq 60$ km and excluding CVR and Interface events) with different focal mechanisms were found to have different attenuation characteristics. Reverse events were found to cause predicted mean intensities that were mostly higher than those of Strike-slip events, which in turn were higher than those of Normal events, for the same magnitude, depth and distance. Reverse and Normal events were found to have higher attenuation rates than Strike-slip events.
7. The intensities of the four Interface earthquakes in the data set are lower on average by about 0.5 intensity units than those predicted by Model 1 for events of the same magnitude, depth and focal mechanism. This is consistent with the predictions for PGAs from a models based on New Zealand data [35] and on world-wide data [43].
8. The rate of attenuation of intensity was found to be substantially higher within the Central Volcanic Region than in other parts of the country.
9. The geographical boundaries of the high attenuation volcanic region appear to lie close to those of the CVR, except that (on the evidence of one earthquake) the area north-west, west and south-west of Tauranga may have normal attenuation characteristics.
10. Attenuation related to both the Fiordland and the North-west sub-regions was found to be consistent with that of the rest of the Main Seismic Region.
11. Deep earthquakes ($h_c \geq 70$ km) in the subducting Pacific slab have a very high attenuation rate in the N50°W direction, while in the major axis direction N40°E the attenuation rate is about the same as for shallow Main Seismic Region events (Model 1). Their attenuation rate in the south-east direction is apparently not highly attenuated as it is in the opposite direction, but has not been modelled due to lack of data.
12. Iseismal shape, measured by the ratio b/a (radius normal to strike over radius along strike) has strong dependencies on magnitude, source distance, intensity and fault dip. The ratio b/a is about 0.4 near the source of large magnitude shallow strike-slip earthquakes with near vertical fault rupture; b/a tends to 1.0 with decreasing magnitude and increasing source distance, except for Deep events in the subducting Pacific plate.
13. The near source spatial distribution of ground motions for large shallow earthquakes differs substantially from that predicted by standard strong motion attenuation models.
14. The mean value of local observations of intensity between adjacent isoseismals (i.e. within an intensity zone) was found to be about 0.2 intensity units lower than the nominated intensity for the intensity zone. The standard deviation of the local observations was found to be 0.64 across the full range of intensity zones considered (MM4 to MM9).
15. The expressions presented here for estimating intensity may be inverted for estimating distances to isoseismals of given intensities, without significant added error arising from using MMI as the dependent variable in the regression.
16. The 1991 attenuation model of Dowrick [6, 7] is quite similar to that of this study for shallow ($h_c = 10$ km) Reverse faulting earthquakes. The 1991 lumping together of Strike-slip and Normal earthquakes, resulted in under- and over-estimations of the intensities of these two mechanisms respectively when compared to the present model for shallow Strike-slip ($h_c = 10$ km) earthquakes.

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