VICTORIA UNIVERSITY

**A CONSISTENT
GEOLOGICAL-SEISMOLOGICAL
MODEL FOR
EARTHQUAKE OCCURRENCE IN
NEW ZEALAND**

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What we want is a story that starts with an earthquake and
works its way up to a climax.

Samuel Goldwyn

C O N T E N T S

Illustrations	xi
Tables	xiii
Acknowledgements	xv
Abstract	xvii
chapter 1	1
Introduction	
chapter 2	7
Seismological Occurrence Model	
1 The Historical Earthquake Catalogues	9
1.1 The Seismograph Network of New Zealand	9
1.2 Level of Completeness	11
1.3 Depth Distribution	14
1.4 Magnitude Consistency	17
1.5 Other Catalogue Related Problems	18
1.6 Australia	19
1.7 Data Preparation	20
2 Kernel Estimation	21
2.1 Basic Kernel Estimation	22
2.2 Adaptive Kernel Estimation	25
3 Temporal Sequence Filter	27
3.1 Coefficient of Variation	27
3.2 Filtering Sequences	28
4 Comparison of Different Kernel Models	29
4.1 Model Comparison	30
4.2 Scores	31

5 Results.....	33
5.1 Occurrence Model for Shallow Earthquakes.....	34
5.2 Occurrence Model for the Subduction Zones.....	37
5.3 Occurrence Model for Australia.....	39
5.4 Comparison of Integrated Errors.....	41
5.5 Comparison of Local Squared Errors.....	44
6 Summary	52
chapter 3	55
<hr/>	
Geological Occurrence Model	
1 Earthquake Source Parameters	56
1.1 Scaling Relationships	57
1.2 Frequencies.....	58
1.3 Summary	60
2 Parameters of Mapped Active Faults.....	62
2.1 Fault Parameter Sampling Biases.....	63
2.2 Characteristic versus Uncharacteristic Behaviour	67
2.3 Faults versus Earthquake Sources	68
3 Dynamic Stochastic Fault Model.....	77
3.1 Fault Linkage Based on Angle	78
3.2 Fault Linkage Based on Distance	79
3.3 Limits on Fault Linkage	80
3.4 Simulation of Fault Activity	82
4 Results.....	84
4.1 Magnitude Frequency Distributions.....	84
4.2 Deficiency of Large Strike-Slip Earthquakes.....	86
4.3 Magnitude Frequencies of Historical Earthquakes	88
4.4 Too many Large Dip-Slip Earthquakes	93
4.5 Spatial Representation of Fault Ruptures	94
4.6 Return Time Frequencies.....	102
5 Accumulated Deformation Rates	104
6 Summary	107

chapter 4	113
<hr/> Combined Occurrence Model	
1 Magnitude Frequency Distributions.....	113
2 Seismological and Geological Model Compared.....	116
2.1 Seismicity Features	117
2.2 Surface Faulting	129
3 Subduction Zones.....	137
4 b-values	140
5 Maximum Magnitude	144
6 Summary.....	151
chapter 5	155
<hr/> Conclusions and Future Work	
1 Conclusions	155
2 Future Work.....	158
2.1 Seismological Model.....	158
2.2 Geological Model.....	160
2.2 Other Studies.....	163
appendix 1	165
<hr/> Different Scaling for Earthquake Source Parameters	
appendix 2	181
<hr/> The Fault Compilation	
Works Cited	187

I L L U S T R A T I O N S

Number	Page
1.1: Turkey after the 1999 Izmit Earthquake	2
2.1: Upgrade of the seismograph network over time	10
2.2: Important seismometer types of the seismograph network	11
2.3: Magnitude frequencies during different time intervals.....	13
2.4: Depth distribution of recorded earthquakes	15
2.5: Depth distribution in the Eastern North Island.....	16
2.6: Magnitude comparison.....	17
2.7: Magnitude frequencies during different time intervals in Australia	20
2.8: Gaussian and inversbiquadratic kernel function for different bandwidths	23
2.9: Different kernel models for shallow earthquakes	35
2.10: Coefficient of variation for shallow earthquakes.....	36
2.11: Occurrence model for shallow earthquakes.....	37
2.12: Occurrence model for the subduction zones.....	38
2.13: Occurrence model for Australia	40
2.14: Integrated Errors	42
2.15: Gaussian kernel models	45
2.16: Inversbiquadratic kernel models	47
2.17: Comparison of kernel models	51
3.1: Source Parameter Frequency Distributions	59
3.2: Rupture Length to Displacement Ratio in Time	60
3.3: New Zealand Fault Parameters versus Source Parameters	71
3.4: Theoretical versus Measured Displacements for Strike-Slip Faults	73
3.5: Length Correction.....	74
3.6: Magnitude Frequencies for Different Fault Rupture Simulations	85
3.7: Magnitude Frequencies for Different Time Intervals	89
3.8: Magnitude Frequencies	90
3.9: Simulated Fault Activity for Different Parameters.....	96
3.10: Simulated Fault Activity for Different Magnitude Ranges.....	98
3.11: Fault Activity for Strike-slip Faults and Characteristic Faults	99
3.12: Return Time Frequencies	103
3.13: Simulated Slip Rates.....	106
4.1: Seismological Occurrence Model above 40 km depth for Different Time Intervals.....	119
4.2: Seismological Occurrence Model above 12 km depth.....	120
4.3: Seismological Occurrence Model since 1930	121
4.4: Large, Important, Historical Earthquakes in New Zealand.....	122
4.5: Earthquake Occurrence based on Simulated Fault Ruptures	131
4.6: Magnitude Frequency Distributions of the Subduction Zones	137
4.7: Earthquake Occurrence in the Subduction Zones	138

A1.1: Rupture Length Distribution.....	171
A1.2: Seismic Moment versus Rupture Length for Dip-slip Earthquakes.....	174
A1.3: Seismic Moment versus Rupture Length from Different Regions	174
A1.4: Seismic Moment versus Rupture Length for Strike-slip Earthquakes.....	175
A1.5: Rupture Length versus Rupture Width.....	176
A1.6: Distribution of Aspect Ratios for Different Slip-types	177
A2.1: Fault Map	181
A2.2: Geological Map.....	183

T A B L E S

Number	Page
2.1: Cut-off magnitudes for different time intervals.....	12
2.2: Cut-off magnitudes for Australia	20
2.3: PFC and POC scores for New Zealand	43
2.4: PFC and POC scores for Australia.....	43
 A1.1: Seismic Moment and Rupture Length Scaling	 173
A2.1: Fault Compilation	184

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A B S T R A C T

For the development of earthquake occurrence models, historical earthquake catalogues and compilations of mapped, active faults are often used. The goal of this study is to develop new methodologies for the generation of an earthquake occurrence model for New Zealand that is consistent with both data sets.

For the construction of a seismological earthquake occurrence model based on the historical earthquake record, ‘adaptive kernel estimation’ has been used in this study. Based on this method a technique has been introduced to filter temporal sequences (e.g. aftershocks). Finally, a test has been developed for comparing different earthquake occurrence models. It has been found that the adaptive kernel estimation with temporal sequence filtering gives the best joint fit between the earthquake catalogue and the earthquake occurrence model, and between two earthquake occurrence models obtained from data from two independent time intervals.

For the development of a geological earthquake occurrence model based on fault information, earthquake source relationships (i.e. rupture length versus rupture width scaling) have been revised. It has been found that large dip-slip and strike-slip earthquakes scale differently. Using these source relationships a dynamic stochastic fault model has been introduced. Whereas earthquake hazard studies often do not allow individual fault segments to produce compound ruptures, this model allows the linking of fault segments by chance. The moment release of simulated fault ruptures has been compared with the theoretical deformation along the plate boundary.

When comparing the seismological and the geological earthquake occurrence model, it has been found that a ‘good’ occurrence model for large dip-slip earthquakes is given by the seismological occurrence model using the Gutenberg-Richter magnitude frequency distribution. In contrast, regions dominated by long strike-slip faults produce large earthquakes but not many small earthquakes and the occurrence of earthquakes on such faults should be inferred from the dynamic fault model.

chapter 1

INTRODUCTION

I woke up at around 3 am in the morning with immense shaking. The bed I was on went up and down like a piece of paper for a couple of seconds. At first, I thought this was another dream, this time a bad one... I looked out of the window, electricity was cut and ambulance sirens got stronger and people started shouting “get out of your houses” I did not know what to do. First, I thought I will go under a table, then I felt maybe underneath the door was safer. Finally, I picked up a T-shirt and started running out of the apartment. I was safe outside but knew that I would not be able to say the same for many others who lived (used to live) in other parts of Turkey. I just thought, life is so simple and mankind is so hopeless and has no power against nature. You are all alive enjoying your cup of tea on a nice summer night, and the next few seconds you are buried under a huge rubble begging for a piece of air to survive. As we are unable to sign a contract with nature to predict when and where the earthquake will hit next, there is no solution other than just to surrender and to live with this devastating experience.

Nihat Ozen, 1999, Ankara

On the 19th August 1999, one minute after three o'clock in the morning, Turkey was struck by its most destructive earthquake disaster of the last century. It destroyed the homes of 600,000 people and more than 15,000 people were killed. With about 50,000 buildings heavily damaged or collapsed, the total estimated loss was US\$ 16 billion, which accounts for 7% of Turkey's GDB. Although the epicentre was located near the town Izmit, it affected an area hundreds of kilometres wide, including Istanbul and the capital Ankara. Apart from the damage inflicted from ground shaking, further damage was caused by tsunamis and seawater inundation.¹

¹ The information on this earthquake was found on an internet site maintained by the Bogazici University of Turkey (<http://www.koeri.boun.edu.tr/earthqk/earthquake.htm>).

Figure 1.1: Turkey after the 1999 Izmit Earthquake. *The fault ruptured through this set of apartment buildings.*

Apart from constructing earthquake safe buildings (or even ‘unsuccessfully trying to sign contracts with nature’), scientists are trying to predict the future location and time of destructive earthquakes. In principal, there are two ways to achieve this goal:

- The terminology ‘earthquake prediction’ is commonly used when earthquakes are predicted from some kind of information that directly precedes the earthquake (i.e. precursors). Some theories try to predict earthquakes deterministically from information about, for example, ground tilting, humidity changes, electrical currents, and magnetic field variations (e.g. Geller 1997, Kagan 1997a, Kirschvink 2000) or seismicity patterns (e.g. Feng *et al.* 1997, Eneva and Ben-Zion 1997, Li and Vere-Jones 1997). Other approaches use features of observed seismicity to make probabilistic predictions. For example, it is tried to estimate the probability of a large earthquake following precursors (e.g. Rhoades and Evison 1993) or seismic quiescence (e.g. Ogata 1988). To be useful, for any prediction method the rate of successful predictions has to be higher than

the rate of success based on random guesses. So far no prediction method has been proven to fulfil this criteria successfully (Geller 1997, Michael 1997, Kagan 1997a).

- The terminology ‘earthquake forecast’ is often used when the long-term behaviour of seismicity features is studied and this information is extrapolated into the future (e.g. Kagan 1997a, Kafka and Levin 2000). The assumption of this approach is that large earthquakes are related to the studied seismicity features; for example, large earthquakes often occur in regions where many small earthquakes occur (Kafka and Levin 2000). The goal of this approach is not to predict the time, size and location of a future earthquake, but rather estimate the probability of a large earthquake occurring in a given region. Studies that try to estimate such probabilities often present the results as an earthquake occurrence model, from which spatial and temporal probabilities for the occurrence of large earthquakes can be directly inferred.

The main objective of this study is to develop such an earthquake occurrence model for New Zealand. This model can then be used with attenuation models of New Zealand (e.g. Smith 1978a, Pancha 1997) to infer regional probabilities of expected ground shaking over a given time interval. From this the seismic hazard at a particular site can be estimated (e.g. Cornell 1968). Earthquake occurrence models are often developed with the help of historical earthquake occurrence, i.e. earthquakes that have been recorded historically. Simple models divide a region into zones of assumed homogeneous seismicity. The zones are

often adjusted to geologically different features. For each of those zones earthquake occurrence parameters can be calculated (e.g. Smith and Berryman 1986). Newer studies do not use seismic zonation but use more advanced methods like 'kernel estimation' (e.g. Vere-Jones 1992, Kagan and Jackson 1994, Frankel 1995, Cao *et al.* 1996, Woo 1996, and Jackson and Kagan 1999), which will be discussed in detail in chapter two. One disadvantage of the use of historical earthquakes is that the time interval of earthquake observation is in general very limited and often few large earthquakes have occurred during this time interval. Additionally, the behaviour of earthquake occurrence might change over time. Thus, it is possible that earthquake occurrence based on historical earthquakes is only valid for short time intervals.

Additional information about past earthquakes can be obtained from mapped , active faults. This information can also be used for the construction of earthquake occurrence models (e.g. Ogata 1999) and the resulting model can be expected to be valid for longer time intervals. However, the data of individual paleoearthquakes is obviously not as accurate as the data from instrumentally observed earthquakes and sampling biases have to be considered (e.g. the surface rupture length of an earthquake is often shorter than the subsurface rupture length, e.g. Wells and Coppersmith 1994).

Geodetic information about how the region of interest deforms can be used as well for the development of earthquake occurrence models (e.g. Ward 1994, Ward 1998). The deformation of a region can be measured with the help of geodesy or inferred from observed earthquakes (e.g. Haines and Holt 1993), but similarly to the historical earthquakes the observation time is rather limited. The main problem with the use of measured accumulated strain is that it is not

known how much is released in earthquakes and how much is released aseismically. Deformation rates are best used as a test, if an earthquake occurrence model for the region of interest matches the observed deformation or the theoretical deformation inferred from the plate movements.

A first attempt at developing an earthquake occurrence model for New Zealand was undertaken by Smith (1978b). The data used for this study were all recorded shallow earthquakes above magnitude four between 1940 and 1975. The model gave regional expected values of ground shaking (in MM Intensity) for different time intervals and concluded that the biggest hazard of ground shaking is in central New Zealand. This study was updated by Smith and Berryman (1986) incorporating paleoseismic data. For this study, New Zealand was divided into different zones and for each zone earthquake occurrence parameters were calculated. It also was the conclusion of this study that the biggest hazard of ground shaking is expected in central New Zealand.

In a recent study by Stirling *et al.* (1998) a new earthquake occurrence model for New Zealand was developed. This study also used historical earthquake and geological data. For historical earthquakes, kernel estimation rather than seismic zoning was used. With the use of attenuation models, several models of expected maximum ground shaking over certain time intervals were given. In contrast to the two earlier studies, the biggest expected ground shaking is not located in central New Zealand, but along the major faulting zones of New Zealand. All in all, this new model seems to have a better ‘resolution’ (i.e. more detail) than the two older models.

Apart from these three national hazard studies of New Zealand, there have been several studies of local seismic hazard (e.g. Berryman and Beanland 1989,

Dowrick 1992, Van Dissen *et al.* 1992, Berril *et al.* 1993, Van Dissen and Berryman 1996). An advantage of such local studies could be the use of more detailed historical earthquake catalogues (for example, the completeness magnitude of the Wellington earthquake catalogue since about 1976 is lower than the completeness magnitude for the whole of New Zealand) and the use of local fault studies. However, any local earthquake occurrence model should match the corresponding earthquake occurrence model for New Zealand in that particular region. Any mismatch demonstrates shortcomings of either the national or the local earthquake occurrence model.

Any earthquake occurrence model has to be consistent with historical earthquake occurrence, the geological record of active fault movements, and geodetic information of recent deformation rates (e.g. Field *et al.* 1999). Further, the model has to be in agreement with other earthquake behaviour such as the observed magnitude frequency distribution (e.g. Gutenberg-Richter distribution, e.g. Richter 1958, 359) or earthquake source parameter scaling (e.g. Kanamori and Anderson 1975). The main emphasis of this thesis lies on formulating a new earthquake occurrence model, i.e. developing new methodologies which can be used for the improvement of existing hazard models, that are consistent with all available information. The resulting model is only of secondary interest. It can be updated with the methodology developed in this study when new data become available.

SEISMOLOGICAL OCCURRENCE MODEL

Historical earthquakes are the most accurate observations (i.e. paleoearthquakes usually have high sampling errors) for developing earthquake occurrence models. The disadvantage of historical earthquake catalogues is the limited observation time (in general not more than one century) which means that earthquake catalogues mostly consist of small earthquakes and have few large earthquakes. Additionally, it is possible that the occurrence of earthquakes changes over time. One of the basic laws for earthquake occurrence is the Gutenberg-Richter law (e.g. Richter 1958, 359) according to which the distribution of the frequencies of all magnitudes is exponential (a detailed discussion of different magnitude frequency distributions follows in chapter four, section one). Therefore, the occurrence of large earthquakes can be predicted from small earthquakes. Historical earthquake catalogues are thus useful for the estimation of earthquake occurrence rates in the near future.

The earthquake catalogue of the Seismological Observatory of New Zealand, used for this work, is discussed in detail in the first section of this chapter. A short summary of the earthquake catalogue of the Australian Seismological Centre is given as well.

There are several different ways to develop earthquake occurrence models. The first possibility is the use of ‘parametric regression’ which implies that all the parameters needed to develop an earthquake occurrence model are known. As those parameters are currently not very well known, the use of ‘non-parametric estimation’ seems to be a better choice. The three most commonly used non-

parametric estimations are orthogonal series estimations, kernel estimations, and spline estimations (e.g. Härdle 1990). In orthogonal series estimation the model is estimated by estimating the coefficients of its Fourier expansion, and in spline estimation the local variation of the model is minimised by the introduction of a ‘roughness penalty’. Some work on spline estimation in connection with earthquake occurrence has been done by Ogata *et al.* (1991). There are other forms of non-parametric estimations that might be used for the estimation of occurrence models (e.g. Papazachos 1999). Every non-parametric estimation faces the problem of the choice of ‘estimation function’ (e.g. orthogonal series function, kernel function, spline function) and ‘bandwidth’ (i.e. the regional effect of the estimation function). Most earthquake occurrence models developed in the past suffer from ignoring the rather important effect of the bandwidth and the results reflect the particular choice of bandwidth rather than ‘true’ earthquake occurrence of the data used.

For this study, a form of ‘adaptive’ kernel estimation has been developed and is introduced in section two. Since single temporal sequences (such as aftershocks) are generally confined to small areas and time spans they might distort estimated earthquake occurrence models and should be taken out of the analysis. The filtering of temporal sequences is described in section three and in section four a comparison method for different kernel estimations is introduced. The results of adaptive kernel estimation are presented in section five.

Although this study is concerned with earthquake occurrence models for New Zealand the methodologies introduced here have also been tested on the Australian earthquake catalogue of the Australian Seismological Centre (e.g. McCue and Gregson 1994). This has been done to test the methods on an intraplate region with much lower activity level.

1 The Historical Earthquake Catalogues

A collection of data from historical earthquakes in New Zealand is available as a computer file (see Smith 1976 for details). Details related to the detection and determination of earthquakes in New Zealand can be found in the New Zealand seismological reports (e.g. Maunder 1999), Eiby (1970), Eiby (1975) and Adams (1979). The earthquake catalogue of the Australian Seismological Centre (e.g. McCue and Gregson 1994) is discussed in subsection 1.6.

1.1 The Seismograph Network of New Zealand

Since the number of operating seismographs, and hence the density of the seismograph network, determines the number and quality of earthquakes detected, a very short overview of upgrades of the New Zealand seismograph network during the last seventy years is given here. Further, changes in the network may produce artificial changes in the detected seismicity patterns (e.g. Habermann 1987). Details of the stations operating in the seismograph network can be found in the New Zealand seismological reports (e.g. Maunder 1999 and Smith 1981). In cases of doubt about the operating time of a particular station, information has been obtained (Diane Maunder, pers. comm.), who is currently in charge of the network.

The seismograph network in New Zealand consists of a main network together with additional stations for special research purposes (such as volcanological studies) and a set of smaller microseismicity networks, which have been installed regionally for different research purposes.

Figure 2.1: Upgrade of the seismograph network over time. *See discussion in text.*

Figure 2.1 shows the upgrade of the main network and the regional networks in time. Although instrumental detection began around 1900, the 1929 Buller and the 1931 Hawkes Bay earthquakes triggered the extension of the two then existing seismographs and thus the establishment of the national seismograph network. During the first three decades after this the number of operating stations did not change significantly, until the number of stations more than doubled during the sixties. The major rearrangement in 1984 is also evident, as well as the rearrangement of the network due to the change from analogue to digital recorders beginning 1987, which had its peak in the early nineties.

The first two regional networks (Wellington and Pukaki) were deployed in 1976. Towards the end of the eighties, two new networks, Hawkes Bay and Taupo, increased the total number of operating stations. In the nineties, several new networks (mainly for volcanological research purposes) were installed.

Figure 2.2: Important seismometer types of the seismograph network. *The percentage of operating seismometer types in different years is plotted.*

Figure 2.2 shows the different instruments used during the last seventy years. The most important feature is the move from Wood-Andersons to Willmore seismometers (Willmore and Mark Products seismometers have similar response features). Another important aspect is the closedown in the early nineties of the last Wood-Andersons, which were operated as a magnitude control over the seventies and eighties.

1.2 Level of Completeness

For the development of an earthquake occurrence model the lowest magnitude at which all earthquakes have been recorded (i.e. cut-off magnitude) is an important parameter. To find the cut-off magnitudes during different time spans, magnitude frequency plots (figure 2.3) have been produced for each year. The detection of earthquakes is concentrated in a region with defined limits around New Zealand (see New Zealand seismological reports for details, e.g.

Maunder 1999). Only earthquakes within this region are considered in this study. Further, the North and South Island have been analysed separately and different depth ranges have been studied as well. In some cases no clear cut-off magnitudes could be found, particularly in early years. Nevertheless, the values found here have been selected for their consistency over region, depth and time. Independently, similar values were found by Annemarie Christophersen (pers. comm.), who was using an automatic approach for magnitude cut-off detection. Any difference in our results could be explained by shortcomings of the automatic approach (i.e. for early years it is not possible to find stable results due to the small number of detected earthquakes).

A clear distinction in completeness above and below about 40 km depth has been found. The estimated cut-off magnitudes for the two different depth ranges are given in table 2.1. Magnitude frequency plots for these values are shown in figure 2.3. In some cases the chosen values may look a little conservative, but higher cut-off magnitudes were selected due to the slowly decreasing cut-off magnitude in time and depth (e.g. the cut-off magnitude in the early 80s is slightly lower than in the late 60s). For earthquakes before 1940 or deeper than 350 km no clear cut-off magnitude could be defined.

Year	Depth 0 - 40 km	Year	Depth 41 - 350 km
1991-1995	3.0	1988-1995	4.0
1987-1990	3.5	1963-1987	4.5
1962-1986	4.0	1948-1962	5.0
1943-1961	4.5		

Table 2.1: Cut-off magnitudes for different time intervals.

Figure 2.3: Magnitude frequencies during different time intervals. *The frequencies are given in events per year and are not cumulative. The grey line has a slope of minus one for reference (b -value = 1). The main reason for the high rate of crustal earthquakes in 1991-95 is a sequence of shallow aftershocks in offshore East Cape in 1995. Further, the data from the nineties is still not complete, as several aftershock studies are not finished yet. The exclusion of 1995 gives a much better fit to the grey line up to Mag 3.0. The b -value of the deep earthquakes is notably higher than 1.0.*

The year 1943 marks a change in earthquake determination in the computer catalogue (see Smith 1976). Thus, a change in the cut-off magnitude is to be expected. The change in 1962 and 1963 can be possibly explained by deployment of Willmore instruments, which was started in the late fifties (see figure 2.2). Since 1987 the data from the local networks have been used for the

main catalogue. Due to the upgrade starting in 1987 from analogue to digital recorders, which are capable of storing data for a week, seismographs in noisy places were dismantled and reinstalled in quieter places. In 1990 about half of the stations were equipped with digital recorders, which explains the lower cut-off magnitude in 1991. The abrupt changes are not only an effect of the upgrade of the seismograph network itself, but also reflect a policy change by the Seismological Observatory to increase the number of processed earthquakes.

1.3 Depth Distribution

Since the depth determination of earthquakes became more sophisticated in later years, the distribution of focal depths in different years has been analysed. The number of earthquakes per year above the cut-off magnitudes found earlier (table 2.1) are counted and rescaled to the highest cut-off magnitude (mag 4.5 for shallow and mag 5.0 for deep earthquakes) assuming a b-value of one for the crustal earthquakes and a b-value of 1.3 for the deep earthquakes, so that earthquake frequencies from different time intervals are comparable. In figure 2.4, the distribution for different depths over time is shown.

In the computation of earthquake depth, depths for crustal earthquakes have often been assigned restricted depths of 12 or 33 km (see New Zealand seismological reports for more detail, e.g. Maunder 1999) unless a station is near the earthquake. The proportion of restricted earthquake depths has decreased since the late seventies and, increasingly in the nineties. The annual number of lower crustal earthquakes (below 12 km) is higher in the nineties than in the preceding years. The annual number of upper crustal earthquakes is affected by aftershock clusters (e.g. Inangahua 1968), thus a possible decrease in the activity of the upper crustal earthquakes cannot be identified. It is possible that

earthquakes that formerly would have been assigned crustal depths of 12 km have been assigned greater depths in the nineties.

Figure 2.4: Depth distribution of recorded earthquakes. To normalize the activity, the numbers of earthquakes are given per km depth. For the deep earthquakes a higher magnitude cut-off (mag 5.5) has been plotted (b) as well as for mag 5.0 (a), to show that the small increase around 1988 is not related to the magnitude cut-off. The average behaviour of (a) and (b) is very similar.

A possible area for such a depth change might be the eastern North Island, where a lot of earthquakes have been located in the upper crust in earlier years, but much less activity in the upper crust was reported in the nineties (see figure 2.5). This change coincides with the installation of the Hawkes Bay network in 1987, which led to more precisely determined focal depths.

Figure 2.5: Depth distribution in the Eastern North Island. *The numbers of earthquakes are given per km depth. After the middle eighties the number of earthquakes in the upper crust decreased whereas the number of earthquakes in the lower crust increased.*

Whereas the annual number of earthquakes deeper than 250 km is reasonably constant through the last sixty years, the annual number of earthquakes between 40 and 250 km depth seemed to slightly increase around 1988. Since this increase coincided with the change of the cut-off level in 1988 for deep earthquakes (see table 2.1) and thus a possible change in processing, this increase of activity might be an artefact. It is possible that the determination of the magnitudes after 1987 is about 0.1 magnitudes higher than beforehand (Terry Webb, pers. comm.), which would correspond to an increase in annual number of events of about 0.015 and explain this discrepancy. Although the data are very scattered before 1963, the number of earthquakes seems to be higher during this time as well. It is possible that the time from the sixties to the eighties represent a period of (relative) seismic quietness, which would contrast with the uniform behaviour of the crustal and very deep earthquakes.

The presented results depend on the b-value that was chosen for the readjustment of the annual number of earthquakes during the different time

periods. If the b-values change with time that would weaken the conclusions drawn here.

1.4 Magnitude Consistency

Another important issue connected to earthquakes is the consistency of calculated magnitudes. Magnitudes, when they are simply measured from the maximum amplitude of a seismogram, are problematic, since they saturate for very big earthquakes. Kanamori (1977) and Hanks and Kanamori (1979) introduced the moment magnitude as a measure of how much moment is generated at the source of an earthquake. Relating maximum amplitude to the radiated energy is a difficult problem and has been extensively discussed in other studies (e.g. Brune 1968, Kanamori and Anderson 1975, Purcaru and Berckhemer 1982).

Figure 2.6: Magnitude comparison. *The local magnitude M_L compared to moment magnitude, the body and surface wave magnitude. The dotted lines show the standard deviation of the difference between the compared moments.*

To estimate the consistency of New Zealand's local magnitude of shallow earthquakes (above 40 km), it has been compared to the moment magnitude, reported by the Harvard group since 1977, and the body and surface wave magnitude, both reported by ISC and NEIC since 1964 (see figure 2.6). The local magnitude matches the moment magnitude reasonably well, except for the

largest earthquakes. The body wave magnitude is consistently lower and the surface wave magnitude is lower for magnitudes below 6.0 and similar to the seismic moment for the largest events (see also Harte and Vere-Jones 1999). The body and surface wave magnitude behave in a similar way, when compared to the seismic moment (Annemarie Christophersen, pers. comm.). There are not very many large events and the difference of the moment and surface wave magnitude from the local magnitude is still relatively insignificant, if one considers a standard deviation of the average difference of 0.5 (which can be inferred from the more numerous events in figure 2.6). Nevertheless, the local magnitude underestimates most bigger earthquakes compared to the moment and surface wave magnitude.

If there is an inconsistency in the local magnitude, it cannot be detected because of the big scatter in the data. There is the possibility of a small underestimation of the size of a few earthquakes above magnitude six, though. It is unlikely that this will have an impact on this work, since few earthquakes are affected.

1.5 Other Catalogue Related Problems

Other problems in earthquake determination that could possibly lead to artefacts include:

- The local magnitude defined by Richter (1935) was used in earlier years and is based on a Californian attenuation model. In 1977, the attenuation law was slightly modified by Haines (1981a). The expected effect on the magnitudes in the earthquake catalogues is explained by Haines (1981b). Earlier magnitudes have been recalculated from 1945. New attenuation laws that are currently being developed (e.g. Dowrick and Sritharan 1993, Pancha 1997) could be used to upgrade the magnitude calculation.

- In earlier years, the velocity model used for hypocentre determination was very simple and lead to systematic errors in location (Adams and Ware 1977). Revised velocity models have been in use since about 1987 and a difference between spatial distributions of older and newer hypocentres must be expected. As with attenuation models, velocity models are still the subject of ongoing research and inaccurate velocities and crustal depths in the current model have to be expected.
- In the early nineties, simulated Wood-Andersons seismograms were calculated and the last Wood-Anderson seismometers were closed down. Since the local magnitude is defined by Wood-Anderson seismometers, these seismometers were used as a magnitude control. If, as proposed by Uhrhammer and Collins (1990), the amplification of Wood-Anderson seismometers is different from that assumed, the reported local magnitude could have a systematic error of about 0.1 units.
- Although earthquakes are grouped into different time-intervals for the purpose of this study according to the detected cut-off magnitudes, the intervals do not have uniform accuracy of determination. For example hypocentres in the late sixties are less accurately determined than in the early eighties.

1.6 Australia

The Australian earthquake catalogue was available as a computer file for this study. Since Australia is a side study, only the completeness level of the recorded earthquakes has been studied here. Similar to the New Zealand catalogue, magnitude frequencies have been studied for different time intervals (see figure 2.7). The steps in the magnitude frequency distributions (figure 2.7) reveal that

the cut-off magnitude seems to vary regionally throughout Australia. Further, the activity of the nineties seems to be lower then the recorded pre-nineties activity.

The cut-off magnitudes are given in table 2.2.

Year	
1991-1996	3.0
1978-1990	4.0
1968-1977	5.0

Table 2.2: Cut-off magnitudes for Australia.

Figure 2.7: Magnitude frequencies during different time intervals in Australia. *The frequencies are given in number of events per year and are not cumulative.*

1.7 Data Preparation

For the construction of spatial earthquake occurrence models with the methods discussed below, it is of advantage if the earthquake catalogue is ‘binned’ into small cells. For this study, the earthquake locations have been binned into $0.1^\circ \times 0.1^\circ$ and $0.5^\circ \times 0.5^\circ$ cells for New Zealand and Australian shallow earthquakes respectively. For the two New Zealand subduction zones (e.g. Anderson and Webb 1994) a depth profile has been generated by projecting the hypocentres of the earthquakes onto a vertical plane along the main seismic activity trend running from Southwest to Northeast. The data have been binned into cells 12 km along strike and 5 km in depth. Within each cell, the number of earthquakes

above the cut-off magnitude has been summed and divided by the total number of years to calculate an annual rate of activity.

In this study, the probability densities of earthquake occurrence have been plotted as the probability of one or more earthquakes in one year, assuming a Poisson (time independent) process. This probability is given by:

$$P_{\geq 1}(\mathbf{x}) = 1 - \exp(-g_3(\mathbf{x})) \quad , \quad (2.1)$$

where $g_3(\mathbf{x})$ is the estimated average earthquake occurrence rate (see equation 2.8 below).

2 Kernel Estimation

Any seismicity catalogue for some region represents a sample in time drawn from an (unknown) parent distribution of seismicity. Thus, any appropriate model for the parent distribution (of earthquake occurrence) must pass the test that the actual earthquake catalogue has a reasonable probability of being drawn from the modelled parent distribution. My goal is to develop a model that is inferred from the data, rather than constructing a model with the help of external parameters, such as the use of seismic zoning, which dictates the boundaries of seismic activity in the model.

The methods in this study have been developed for the use in two dimensions to produce planar seismicity representations based on epicentres. In principle, they should also work for the production of three-dimensional seismicity representations using hypocentres, although the methods might have to be altered to achieve a similar performance.

2.1 Basic Kernel Estimation

Kernel estimation is commonly used for nonparametric density estimation, i.e. to estimate the parent distribution of a sample without the use of a parametric model. It redistributes the sample data using a kernel function, which controls the shape of the redistribution, and a bandwidth, which controls how much the sample data is redistributed over space - it controls the degree of ‘smoothness’ (i.e. amount of high frequency) of the resulting estimated continuous probability distribution. Kernel estimation has been suggested and used to transform discrete earthquake distributions into spatially continuous probability distributions (e.g. Vere-Jones 1992, Kagan and Jackson 1994, Frankel 1995, Cao *et al.* 1996, Woo 1996, and Jackson and Kagan 1999).

The two-dimensional kernel model (e.g. Silverman 1986, 76) is given by:

$$(2.2) \quad g(\mathbf{x}) = \frac{1}{Nc^2} \sum_{i=1}^N K\left(\frac{1}{c}(\mathbf{x} - \mathbf{x}_i)\right),$$

where K is the chosen kernel function, c is the bandwidth, N is the number of earthquakes, $\mathbf{x}_i$ is the location of each earthquake and $\mathbf{x}$ is the location in the spatial seismicity representation. The kernel model $g(\mathbf{x})$ gives the spatial probability density distribution of earthquake occurrence for the observation time and is equivalent to the (estimated) parent distribution of the spatial seismicity.

Different kernel functions have been suggested for the construction of representations of earthquake occurrences (e.g. Vere-Jones 1992, Kagan and Jackson 1994, Cao *et al.* 1996, and Woo 1996). The choice of kernel function should reflect the properties of spatial earthquake hypocentre distributions, but little research has been done in this field. For example, Woo (1996) suggests the

use of studies of the fractal distributions of hypocentres (e.g. Robertson *et al.* 1995, Otsuki 1998, Bour and Davy 1999, Clark *et al.* 1999). The detection of fractal distributions is, however, problematic (e.g. Marzocchi *et al.* 1997, Gonzato *et al.* 1998). Another problem can be the choice of a rotationally invariant kernel function, which assumes that the earthquake activity has no preferred direction. This is not the case for earthquakes located on well-defined long narrow fault systems, for example plate boundaries. Kernel functions with finite support (i.e. functions which have zero values outside a bounded region, e.g. Epanechnikov kernel) lead to locations of zero probability in regions of no earthquake activity. Since this is not realistic for hazard estimation, we suggest not using such kernels for the estimation of earthquake occurrence representations. Two commonly used kernels are the Gaussian kernel $g(\mathbf{x})$ and the inversebiquadratic (IBQ) kernel $q(\mathbf{x})$ (e.g. Vere-Jones 1992):

$$g(\mathbf{x}) = \sum_{i=1}^N A_i \exp\left(-\frac{(\mathbf{x} - \mathbf{x}_i)^2}{2c^2}\right) \quad (2.3)$$

$$q(\mathbf{x}) = \sum_{i=1}^N A_i \frac{1}{((\mathbf{x} - \mathbf{x}_i)^2 + c^2)} \quad (2.4)$$

where A_i is the normalisation factor. The main difference between these two kernel functions is the longer tail of the IBQ kernel (figure 2.8).

Figure 2.8: Gaussian and inversbiquadratic kernel function for different bandwidths.

Another commonly used kernel function reflects power law behaviour and is of the form (Woo 1996, Cao *et al.* 1996):

$$(2.5) \quad q(\mathbf{x}) = \sum_{i=1}^N A_i \left(\frac{c}{(\mathbf{x} - \mathbf{x}_i)} \right)^\alpha, \quad ,$$

where α is related to the fractal dimension (e.g. Robertson *et al.* 1995). This function has the disadvantage of diverging for $\alpha < 1$ (the integral of a kernel function over space has to be one, e.g. Silverman 1986) and is undefined for $\mathbf{x} - \mathbf{x}_i = 0$. Because of this, and its otherwise similar behaviour to the IBQ kernel, it is not further considered in this study.

Jackson and Kagan (1999) used a bimodal directional kernel that is based on the IBQ kernel. They were specifically interested in earthquake forecasting and the bimodal kernel was used to increase the probability of earthquakes in neighbouring regions of past events. For the estimation of the parent distribution of seismicity the bimodal kernel has the disadvantage of implying bimodal features in regions of little data, whereas in regions with more data the bimodality will, in general, vanish (Scott 1992, 138). Jackson and Kagan's choice of kernel direction is determined by the orientation of the P and T axes, and hence fault strike, of a previous event, but not by the local seismicity features itself (e.g. direction of plate boundaries). This can be a problem, for example, when the seismicity is only adjusted to the dominant fault strike direction of a fault zone, because regions with broad and narrow fault zones are incorrectly treated the same. If an oriented kernel were adjusted to the dominant slip direction of a fault zone, the seismicity of normal and thrust faults will be redistributed preferentially perpendicular to the strike of the faults whereas the seismicity of strike-slip faults will be redistributed preferentially along strike. Although the use of kernel functions that are not rotationally invariant can

provide an improvement in some circumstances, Jackson & Kagan's kernel is not further considered in this study, due to its complexity and the disadvantages noted above.

Different kernel functions produce qualitatively similar results, whereas the choice of an appropriate bandwidth is critical for good results (e.g. Silverman 1986). Bandwidths that are too high lead to very smooth representations, i.e. the sample data (earthquake catalogue) is redistributed strongly over space. This can lead to 'blurring' of local seismicity features (over-smoothing), i.e. main features that are present in the sample data cannot be identified in the estimated parent distribution. In representations generated by bandwidths that are too low, very small seismicity features are preserved, even if these features are generated by single earthquakes (under-smoothing), which would not be expected to be strongly represented in the parent distribution. In past studies (e.g. Vere-Jones 1992, Frankel 1995, Cao *et al.* 1996, Woo 1996, and Jackson and Kagan 1999) the bandwidth has been chosen to be 'global', i.e. the same bandwidth has been used for every location.

2.2 Adaptive Kernel Estimation

For spatial earthquake occurrence, it is to be expected that some regions are characterised by local clusters of high activity, whereas other regions consist of uniform activity. Such behaviour is better described by different degrees of 'local' (spatially varying) smoothness, which a global bandwidth does not allow for. In the spatial frequency domain, a local bandwidth allows for sharp borders between high and low activity levels in space, whereas sharp borders are filtered in regions where the data suggests a more uniform probability distribution. Since it is to be expected that the bandwidth is dependent on earthquake frequency and location, a local bandwidth may give a better representation of the

distribution of seismic activity in space. One form of kernel estimation that uses local bandwidths is the ‘adaptive’ kernel estimation (e.g. Silverman 1986).

In this study, the bandwidth has been chosen to be independent of time and magnitude, which has the advantage of permitting the identification of non-Poissonian time behaviour and deviations from the Gutenberg Richter magnitude-frequency law in the earthquake data set.

Adaptive kernel estimation is a three-step process (see Silverman 1986, 101):

$$\begin{aligned}
 \text{(I)} \quad g_1(\mathbf{x}) &= \frac{1}{2\pi N c_1^2} \sum_{i=1}^N \exp\left(-\frac{(\mathbf{x} - \mathbf{x}_i)^2}{2c_1^2}\right), \\
 \text{(II)} \quad c_2(\mathbf{x}) &= \sqrt{\frac{\mu}{g_1(\mathbf{x})}}, \\
 \text{(III)} \quad g_2(\mathbf{x}) &= \frac{1}{2\pi N c_1^2} \sum_{i=1}^N \frac{1}{c_2(\mathbf{x}_i)^2} \exp\left(-\frac{(\mathbf{x} - \mathbf{x}_i)^2}{2c_1^2 c_2(\mathbf{x}_i)^2}\right),
 \end{aligned}
 \tag{2.6}$$

where c_1 is a global bandwidth and μ is the global mean of earthquake activity per area (grid cell) during the observation period. In step one the ‘pilot estimate’ $g_1(\mathbf{x})$ is calculated, which is used to determine the local bandwidth $c_2(\mathbf{x})$ in step two. Relating $c_2(\mathbf{x})$ to the location of each earthquake $\mathbf{x}_i$, the probability distribution of earthquake occurrences $g_2(\mathbf{x})$ is estimated in step three. When the region of interest has no seismic activity on its borders, locations with no activity should be excluded when calculating μ , otherwise $c_2(\mathbf{x})$ is dependent on region size. Other ways of calculating $c_2(\mathbf{x})$ are possible and may improve the performance of this method.

3 Temporal Sequence Filter

A typical, unmodified historical earthquake catalogue includes short time activity features such as aftershocks and swarms, which are still present in the continuous representation $g_2(\mathbf{x})$. For development of an earthquake occurrence model, it is important to distinguish such sequences from regions of high activity that remain constant throughout the observed time period.

3.1 Coefficient of Variation

To detect temporal activity fluctuations in $g_2(\mathbf{x})$, the kernel estimates $g_t(\mathbf{x})$ of individual years can be compared to the kernel estimate $g_T(\mathbf{x})$ of the whole time period T , and their variance can be calculated. The coefficient of variation is defined by the variance over the mean:

$$V_T(\mathbf{x})^2 = \frac{\sum_{t=1}^T (g_T(\mathbf{x}) - g_t(\mathbf{x}))^2}{T \cdot g_T(\mathbf{x})} \quad (2.7)$$

To estimate each $g_t(\mathbf{x})$ the local parameter $c_2(\mathbf{x})$ from the kernel estimate of $g_T(\mathbf{x})$ should be used to avoid a time dependency of $c_2(\mathbf{x})$, i.e. $c_2(\mathbf{x})$ is the same for each individual year.

For a Poisson process, the expected value of $V_T(\mathbf{x})$ equals one (Johnson *et al.* 1992, 157). Thus, the measured coefficient of variation gives the deviation from a Poisson process, with a value above one indicating temporal clustering. Values below one are observed in regions where earthquakes decreased the probability of following earthquakes, which could be caused by stress relaxation. Under such circumstances, the seismicity pattern can be described as ‘repulsive’ in time.

Regions of low numbers of events usually have a high bandwidth $c_2(\mathbf{x})$ (see equation 2.6) and thus are highly influenced by neighbouring regions, which

leads to a similar level of activity for every year throughout the observation period. In this case kernel estimation usually leads to low temporal variances (i.e. small fluctuations of earthquake activity) and thus to a low coefficient of variation during the observation time. For longer observation times, more data become available, and it has to be expected that the temporal variance will increase. For these regions the small number of observed earthquakes is not sufficient to decide if the region behaves in a Poissonian or non-Poissonian way, due to the high uncertainty in the estimate of $V_T(\mathbf{x})$.

3.2 Filtering Sequences

The main interest of hazard studies is the probability of the occurrence of main shocks in the investigated region. By main shock occurrence I mean the representation of multiple sequences (i.e. aftershocks, foreshocks, swarms) by one main shock. Multiple sequences can distort this probability distribution, since they produce transient local activity features. Thus it is of advantage if such sequences are taken out of the analysis. After the probability distribution of main shocks is known, the hazard due to related earthquakes can be incorporated using an appropriate occurrence model for these sequences (e.g. Omori's law for aftershocks; Reasenber and Jones 1989).

Individual removal of earthquakes belonging to a sequence is often problematic (e.g. Davis and Frohlich 1991, Savage and dePolo 1993), because it is often not clear if an earthquake belongs to a sequence or not. An alternative way to estimate probability distributions of main shocks can be achieved by dividing the continuous representation $g_2(\mathbf{x})$ by a function of the coefficient of variation $V_T(\mathbf{x})$:

$$g_3(\mathbf{x}) = \frac{g_2(\mathbf{x})}{(V_T(\mathbf{x}) + 1)^2} . \quad (2.8)$$

The coefficient of variation has been amended to $(V_{\tau}(\mathbf{x})+1)^2$, otherwise the probability of earthquakes would rise in regions where the coefficient of variation is low. Since only temporal clusters should be filtered, and not temporal repulsion, a correction to using plain $V_{\tau}(\mathbf{x})$ is necessary.

The justification for $g_3(\mathbf{x})$ being a probability representation of main shock occurrence lies in the similar representations it produces for independent time intervals for New Zealand and Australia (see section five). Several approaches for the production of main shock representations have been tried and this form of filtering led to the ‘best results’, considering its proximity to the original earthquake catalogue and similarity between different time intervals. For coefficients of variation much larger than one the expression $(V_{\tau}(\mathbf{x})+1)^2$ behaves asymptotically as $V_{\tau}(\mathbf{x})^2$. For a coefficient of variation of one, $g_3(\mathbf{x})$ equals $1/4 g_2(\mathbf{x})$. As I stated beforehand, a coefficient of variation of one indicates Poissonian (cluster free) behaviour and thus should not be changed. As earthquakes produce aftershocks, though, it has to be expected that in reality locations with a coefficient of variation of one are not free of temporal clusters and the probability of earthquake occurrence has to be weighted down, as this approach correctly does. Nevertheless, a connection to physical earthquake behaviour would be helpful to justify $g_3(\mathbf{x})$ being an estimated probability distribution of main shock occurrence (e.g. Musmeci and Vere-Jones 1986, who try to relate spatial aftershocks features to the inverse-binomial distribution, or Chen *et al.* 1998, who find a fractal behaviour of temporal clustering).

4 Comparison of Different Kernel Models

Since the results of kernel estimation are dependent on the chosen kernel function and bandwidth it would be of advantage if different models could be

compared. Since an objective comparison does not exist for kernel models, a comparison has been developed that is suitable for comparison of different earthquake occurrence models that have been constructed with kernel estimation.

4.1 Model Comparison

To develop a useful earthquake occurrence model from a given earthquake catalogue, my goal is to find the parent distribution from which the catalogue has been drawn as a sample, and from which future seismicity samples arise. Earthquake activity rates in New Zealand and Australia were stable over the last forty and twenty years respectively (see section five). Thus I assume that, in general, the parent distribution changes little on a time scale of decades, and the activity of the near future (about a decade) can be forecast from the activity of the past.

To compare different kernel models, the investigated earthquake catalogues have been divided into two different time intervals. For each of the kernel estimation methods described above, the parent distributions of both the earlier and later time interval of the earthquake catalogue were estimated. Since I assume that the parent distribution changes little over the whole observation period, the earlier and later parent distribution estimates should look similar. The parent distribution representation from the earlier time interval was used as a forecast model for the parent distribution representation of the later time interval. These two representations were compared to estimate how well the earlier distribution forecasts the locations of the earthquakes in the later time interval. I call this kind of comparison ‘PFC’ (parent forecast comparison). Due to over-smoothing when high bandwidths are used, the difference between earlier and later

representations is expected to decrease with higher bandwidths, and the PFC fit will be consequently better.

Since an over-smoothed parent distribution representation is not able to describe all the detailed features inherent in the earthquake catalogue, I also compare the representation of the earlier time interval with the original earthquake catalogue for the same time interval. I call this kind of comparison ‘POC’ (parent original comparison). Because of their high degree of smoothing, higher bandwidths will produce a poorer POC fit. A low bandwidth will produce a better POC fit, but is likely to produce more differences between the earlier and later representation, due to its under-smoothing. Therefore, I define a ‘good’ model for the parent distribution to be one that optimises the fit for both the PFC and the POC together.

For ‘pure’ forecasting purposes, it would be better to directly compare the model derived from the earlier time interval to the distribution of earthquakes from the catalogue in the later time interval. Such a test can decide, for example, if the kernel of Kagan and Jackson (1999) performs better in terms of forecasting than a simpler kernel estimation. But here we are interested in the performance of the different kernel estimations in the generation of seismicity parent distributions rather than forecasting future seismicity distributions. The presented comparison method has the additional advantage of providing a test for the temporal sequence filter, since we choose not to remove individual aftershocks from the original earthquake catalogues.

4.2 Scores

Different ‘scores’ (performance criteria) to estimate the performance of different kernel estimations have been suggested (e.g. Scott 1992). The one most

commonly used is the ‘Mean Integrated Squared Error’ (MISE) (e.g. Silverman 1986; Scott 1992, 38; Wand and Jones 1995, 15). To calculate the MISE, the parent distribution is, in general, known and the mean difference from models, inferred from samples of the parent distribution, is estimated. Since the parent distribution of a given earthquake catalogue is in fact unknown, I prefer the ‘Integrated Squared Error’ (ISE) (e.g. Wand and Jones 1995, 15). The ISE is commonly used when the difference between two models is to be estimated, which is the case for the PFC and POC. It is given by:

$$(2.9) \quad ISE = \sum_{i=1}^N (g_A(\mathbf{x}_i) - g_B(\mathbf{x}_i))^2 ,$$

where $g_A(\mathbf{x}_i)$ and $g_B(\mathbf{x}_i)$ are the two models to be compared, N is the number of summed areas and $\mathbf{x}_i$ is the location of each area.

When used with a local bandwidth, the ISE has the disadvantage that it is dominated by the properties of the tails (Hall 1992). Hall (1992) suggests the use of the ‘Integrated Weighted Squared Error’ (IWSE):

$$IWSE = \sum_{i=1}^N w_i (g_A(\mathbf{x}_i) - g_B(\mathbf{x}_i))^2 . \quad (2.10)$$

The adaptive kernel estimation estimates the ‘localness’ of the features of the parent distribution, i.e. it controls the length of the tails. Thus I choose the weights to be $w_i = 1 / c_2(\mathbf{x}_i)$, where $c_2(\mathbf{x}_i)$ is the local bandwidth (see equations 2.6). The IWSE is the appropriate score to use for the PFC in conjunction with the adaptive kernel estimation. Since the original data distribution is discrete and thus does not have tails, the ISE was used for the POC. In the global kernel estimation each event is ‘weighted’ in the same way, and the ISE was used when comparing this type of kernel estimation. Since the IWSE and ISE are not inter-

comparable, the ISE was used for the comparison of the adaptive kernel estimation with the global kernel estimation.

Other scores like the integrated visual error (IVE) have been suggested (Marron and Tsybakov 1995). Whereas the ISE calculates distances in terms of classical mathematical norms, the IVE calculates a distance based on the ‘visual impression’. Although this score has certain advantages, Marron and Tsybakov (1995) state that the ISE should be preferred for forecasting purposes, thus the IVE is not used in this study. Another class of scores are based on information theory, like the Kullback-Leibler score (Scott 1992, 38). These scores are based on logarithms though, and cannot be used with the original data distribution, since areas with zero earthquakes are undefined in this case. Additionally, the Kullback-Leibler score is dominated by the properties of the tails (Hall 1987). There are other scores based on norms like the ‘Integrated Absolute Error’ (IAE) (e.g. Scott 1992, 38; Wand and Jones 1995, 16). I have experimented with the IAE, but found that the results are similar to the results using the ISE.

Any integrated score has the disadvantage of not showing the regional differences between the representations. Therefore, plots of the local squared error (SE) have been produced, which are helpful to show where different kernels produce similar features and where the features do not match well.

5 Results

I will first present results of the adaptive kernel estimation with the Gaussian kernel and a global bandwidth of $c_1 = 1.0$. This makes the reader familiar with the results that adaptive kernel estimation provides, and gives a first impression on earthquake occurrence models in New Zealand. During the work with different kernel estimations, the adaptive kernel estimation with a Gaussian

kernel and a global bandwidth of $c_1 = 1.0$ seemed to give the best performance when compared ‘by eye’ to other kernel estimations. Although the comparison of different models by eye is very common for general kernel estimation problems, I developed a comparison mechanism (see section four) to quantitatively compare earthquake occurrence models that are based on different kernel estimations. The results of this comparison confirm the assumption that adaptive kernel estimation with the Gaussian kernel and a bandwidth $c_1 = 1.0$ provides the best performance (see subsections 5.4 and 5.5).

5.1 Occurrence Model for Shallow Earthquakes

Figure 2.9 shows the different performance of kernel estimation with global bandwidth $g_1(\mathbf{x})$ compared to adaptive kernel estimation $g_2(\mathbf{x})$. Whereas a small global bandwidth (figure 2.9b) preserves high activity clusters, low activity offshore (e.g. around latitude -46° , longitude 171°), which in figure 2.9a appears to be regular in space, is clustered as well (‘under-smoothing’). A greater bandwidth (figure 2.9c) generates a higher degree of smoothness in low activity regions, but the high activity clusters are now distributed over a wider space, leading to a decrease of activity in their centres and an increase in activity in their immediate neighbourhood (‘over-smoothing’). The adaptive kernel estimation separates high activity clusters from low background seismicity and clusters of high activity can be readily identified. Another important improvement lies in the reduction of the uniform circularity of the activity features in the kernel estimation with global bandwidth, which is artificially imposed by the choice of a rotationally invariant kernel.

Figure 2.9: Different kernel models for shallow earthquakes. *a)* probability distribution inferred from the cumulative number of crustal earthquakes. *b)* and *c)* corresponding probability distribution generated by kernel estimation with global bandwidths 1.0 and 2.5, respectively. For comparison, the probability distribution generated by the adaptive kernel estimation is shown in *d)*. The colours give the probability per year of one or more earthquakes (see equation 2.1) above magnitude 4.0 and 40 km, from 1962 to 1997 in 0.1° squares. This simple form of projection has been chosen to avoid a ‘smoothing’ of the individual cells due to map projections.

Figure 2.10 shows the spatial coefficient of variation $V_T(\mathbf{x})^2$ between two different time intervals. Temporal activity clustering (figure 2.10a) can be mostly matched to spatial clusters of high activity (figure 2.9d). Although the exact location of most of the individual temporal sequences is different between the two different time intervals (compare figure 2.10a and 2.10b), most temporal sequences of the later time interval are located near the temporal sequences of the earlier time interval. This suggests that the area generating these temporal clusters is bigger than the spatial extent of any cluster. Some spatial clusters of medium activity in figure 2.9d (e.g. around latitude -40° , longitude 176°) indicate

little or no temporal clustering in figure 2.10a. They probably represent centres of persistent medium-rate activity.

Figure 2.10: Coefficient of variation for shallow earthquakes. *The coefficient of variation of crustal earthquakes in New Zealand from 1962 to 1990 above magnitude 4.0 (figure a) and from 1991 to 1999 above magnitude 3.0 (figure b). The colours give the amount of clustering (above one) and repulsion (inhibition; below one) above 40 km depth.*

Figure 2.11 shows the seismicity probability distributions $g_3(\mathbf{x})$ (equation 2.8) with temporal sequences filtered out, during two different time intervals. The two distributions reveal very similar spatial seismicity features (figure 2.11a and 2.11b), although they are independent, representing the activity of different time intervals and magnitudes. There are a few differences in detail, which could be explained by the shorter observation time for figure 2.11b compared to figure 2.11a. Whereas the uncertainty increases for shorter observation times, main shock activity is expected to undergo real fluctuations over a time interval of decades (e.g. latitude -40° , longitude 174°). Nevertheless, the similarity of figures 2.11a and 2.11b suggests that these temporal fluctuations are only of secondary order. Some differences may be due to the use of better instrumentation in the later years, which leads to more accurate hypocentres in figure 2.11b. The similarity of the two figures suggests that the main seismicity features in New Zealand have been stable over the last forty years.

Figure 2.11: Occurrence model for shallow earthquakes. The probability distribution of crustal earthquakes in New Zealand with sequences filtered out from 1962 to 1990 above magnitude 4.0 (figure a) and from 1991 to 1997 above magnitude 3.0 (figure b). The colours give the probability per year of one or more earthquakes (see equation 2.1) in 0.1° squares above 40 km depth.

5.2 Occurrence Model for the Subduction Zones

Figure 2.12 shows a depth profile of the seismicity probability distribution $g_2(\mathbf{x})$, the spatial coefficient of variation $V_T(\mathbf{x})$, and the seismicity probability distribution with temporal sequences filtered out $g_3(\mathbf{x})$. In contrast to the shallow activity, temporal clustering was nearly absent for the deep earthquakes in the subducted slabs during the observation period (compare figure 2.12a and 2.12b) and the coefficient of variation of the deep earthquakes looks very uniform.

Hence there is no need to filter temporal sequences out of the representation of earthquake occurrence probability (figure 2.12a and 2.12c look alike). One explanation for this result would be that deep earthquakes do not produce many temporally clustered sequences, such as aftershocks and swarms, as has indeed been observed for large deep earthquakes in the past (e.g. Robinson 1986). The low temporal variation further suggests that the spatial earthquake occurrence features have been very stable during the observation period.

Figure 2.12: Occurrence model for the subduction zones. *The cross section of New Zealand (in km) shows the probability distribution generated by adaptive kernel estimation (figure a), the index of dispersion (figure b) and the probability distribution with temporal sequences filtered out (figure c) for earthquakes down to 350 km. The strike of the cross section is shown in figure d).*

5.3 Occurrence Model for Australia

Figure 2.13 shows the seismicity probability distribution $g_3(\mathbf{x})$, with temporal sequences filtered out, during different time intervals for Australia. The two probability distributions (figure 2.13a and 2.13b) reveal similar seismicity features, although they do not match as well as the corresponding figures for New Zealand. This is not surprising, since the activity in Australia is much lower and a higher uncertainty in $g_3(\mathbf{x})$ is to be expected due to the lower sample size.

The region of highest activity occurred at the location of the Tennant Creek earthquake sequence (latitude -20° , longitude 134°) which commenced on 22 January 1988, which had one event above magnitude 6.5 and two other events above magnitude 6.0 (e.g. Gaull *et al.* 1990). Before these events, the area around this earthquake has been reasonably inactive (figure 2.13c). There is little doubt that this earthquake started a spatial cluster of high activity, which persists today. It is possible that other clusters of persistent high activity, which have produced large earthquakes in historical times, have been started by one or a few large events. On the other hand, no spatial clusters of persistent high activity in New Zealand and Australia decreased to background activity during the observation time.

The rate of activity in the nineties seems to be lower than the rate of activity before the nineties (compare figure 2.13a and 2.13b). This observation matches the magnitude frequency plots for those time intervals shown in figure 2.7.

Figure 2.13: Occurrence model for Australia. *The probability distribution of crustal earthquakes in Australia (above 50 km) with sequences filtered out from 1978 to 1996 above magnitude 4.0 (figure a), from 1991 and 1996 above magnitude 3.0 (figure b) and from 1978 to 1986 (figure c) above magnitude 4.0, in 0.5° squares. The colours give the probability per year of one or more earthquakes (see equation 2.1).*

5.4 Comparison of Integrated Errors

In this and the next subsection the performance of different kernel estimations is compared as discussed in section four. Figure 2.14 shows PFC (parent forecast comparison) versus POC (parent original comparison) for New Zealand and Australia for different kernel estimations. Values close the origin (i.e. in the lower left corner) show a better performance than values in the upper right corner. The ISE (integrated squared error) and the IWSE (integrated weighted squared error) scores are given in tables 2.3 and 2.4.

Figures 2.14a and 2.14c compare models generated by the adaptive Gaussian and adaptive IBQ kernel with different bandwidths c_1 . Although the model generated by the IBQ kernel forecast the future as well as the model generated by the Gaussian kernel (similar PFC scores), it did not stay as close to the original earthquake catalogue (bigger POC scores) and thus scores worse overall than the Gaussian model. In New Zealand and Australia, for the models generated by the Gaussian kernel the optimum bandwidth c_1 was about 1.0, whereas for the models generated by the IBQ kernel the optimum bandwidth was about 0.4.

For New Zealand, the models generated with the Gaussian kernel achieved a slightly lower PFC score with a bandwidth of 0.8 compared to 1.0, although (as discussed earlier) it is generally expected that the PFC score decreases with increasing bandwidth (figure 2.14a). It is likely that the temporal sequence filter derived from the model with a bandwidth of 0.8 filtered more aftershocks than the sequence filter derived from the model with a bandwidth of 1.0. This would lead to a lower PFC score.

Figure 2.14: Integrated Errors. *Parent Forecast Comparison* and *Parent Original Comparison* scores for New Zealand and Australia with sequence filtering. a) and c) compare the models generated by the adaptive Gaussian kernel and adaptive inversbiquadratic kernel. b) and d) compare the models generated by the adaptive Gaussian kernel and global Gaussian kernel. The numbers represent the bandwidths.

	PFC (IWSE)	PFC (ISE)	POC (ISE)
without sequence filtering			
Gaussian 0.8 adaptive	307.39		2.99
Gaussian 1.0 adaptive	243.90		4.78
Gaussian 1.2 adaptive	178.80		6.73
with sequence filtering			
Gaussian 0.5 global		1.95	52.03
Gaussian 1.0 global		1.81	52.13
Gaussian 2.0 global		1.55	52.76
Gaussian 0.6 adaptive	2.32	1.66	52.27
Gaussian 0.8 adaptive	1.08	1.61	52.18
Gaussian 1.0 adaptive	1.32	1.51	52.11
Gaussian 1.2 adaptive	1.04	1.53	53.59
IBQ 0.3 adaptive	2.45		52.85
IBQ 0.5 adaptive	1.12		53.32
IBQ 0.7 adaptive	0.69		53.69

Table 2.3: PFC and POC scores for New Zealand.

	PFC (IWSE)	PFC (ISE)	POC (ISE)
without sequence filtering			
Gaussian 0.8 adaptive	230.62		4.82
Gaussian 1.0 adaptive	188.26		8.47
Gaussian 1.2 adaptive	153.89		20.57
with sequence filtering			
Gaussian 0.5 global		1.30	165.43
Gaussian 1.0 global		1.17	167.39
Gaussian 2.0 global		0.80	170.73
Gaussian 0.8 adaptive	4.02	1.03	165.54
Gaussian 0.9 adaptive	3.10	1.06	165.57
Gaussian 1.0 adaptive	2.53	1.12	165.72
Gaussian 1.1 adaptive	2.17	1.18	166.02
Gaussian 1.2 adaptive	1.90	1.25	166.48
IBQ 0.2 adaptive	39.38	166.22	0.86
IBQ 0.3 adaptive	7.29	167.08	0.67
IBQ 0.4 adaptive	2.17	168.08	0.56
IBQ 0.5 adaptive	0.94	169.04	0.51
IBQ 0.6 adaptive	0.49	169.90	0.47

Table 2.4: PFC and POC scores for Australia.

Figures 2.14b and 2.14d compare models generated by the adaptive and global Gaussian kernel with different bandwidths. Whereas the adaptive approach succeeded in minimizing both PFC and POC, the global approach failed in doing so. For Australia, the PFC score increased with bandwidth for the adaptive case, contrary to expectation (figure 2.14d). This is due to the fact that the ISE is dominated by the properties of the tails of the kernel estimation. The IWSE for the adaptive case was decreasing with increasing bandwidth as expected (figure 2.14c).

Without sequence filtering the later seismicity model was not forecast very well by the earlier seismicity model and the PFC scores were large (see table 2.3 and table 2.4). On the other hand, the earlier seismicity model stayed close to the original earthquake catalogue and the POC score was very low. This reflects the fact that the temporal sequences were not removed from the original earthquake catalogue. However, they did not occur at the same locations in the later seismicity model as the large PFC score shows.

5.5 Comparison of Local Squared Errors

Figure 2.15 shows the seismicity representation for Australia modelled with the adaptive Gaussian kernel and two different bandwidths for different time intervals. As predicted, for the higher bandwidth there are less individual seismicity features in both models and the seismicity features of both models match better compared to the lower bandwidth, i.e. almost every feature in the earlier model has a corresponding feature in the later model. The difference between the earlier and later models for the lower bandwidth is reflected in the number of (mostly green) clusters in the PFC plots. There are less (but bigger)

Figure 2.15. Gaussian kernel models. The probability distribution of crustal earthquakes in Australia (above 50 km) generated by adaptive Gaussian kernel estimation from 1978 to 1990 (I) and from 1991 to 1996 (II) above magnitude 4.0 with bandwidths 0.8 (a) and 1.2 (b). The scale gives the probability of one or more earthquakes per year and area unit. (F) local Parent Forecast Comparison squared error scores. (O) local Parent Forecast Comparison squared error scores.

clusters for the higher bandwidth (as expected). Also, the biggest differences for the higher bandwidth are smaller than those for the lower bandwidth, i.e. they decrease from red to yellow and from yellow to green with increasing bandwidth. In the POC plot, there are more differences for the higher bandwidth compared to the lower bandwidth, which can be seen in the greater amount of green. The corresponding plots for New Zealand showed similar results.

Figure 2.16 shows the seismicity representation for Australia modelled with the adaptive IBQ kernel and two different bandwidths for different time intervals. The differences between the two models with different bandwidths are similar to the results with the Gaussian kernel. In contrast to the Gaussian kernel, the IBQ kernel smeared out the seismicity over space more, while maintaining very local features that are produced by single earthquakes (the POC plots give an impression of the corresponding original earthquake distributions if only red and yellow values are considered). Such behaviour is common for kernels with long tails (Vere-Jones 1992). For the lower bandwidth, many of the local clusters can be matched to individual earthquakes in figures 2.16 aI) and 2.16 aII). This leads to a high PFC score since the earlier model fails to forecast the locations of the earthquakes in the later model. In the POC plots, the regions of low background rate in the model show significant differences from the original earthquake catalogue, which can be seen by the high amount of green, especially for the higher bandwidth. Thus, the activity of low background rates is highly overestimated by the model generated by the IBQ kernel, which leads to a high POC score. The corresponding plots for New Zealand showed similar results.

Figure 2.16. Inversibiquadratic kernel models. The probability distribution of crustal earthquakes in Australia (above 50 km) generated by adaptive inversebiquadratic kernel estimation from 1978 to 1990 (I) and from 1991 to 1996 (II) above magnitude 4.0 with bandwidths 0.2 (a) and 0.6 (b). The scale gives the probability of one or more earthquakes per year and area unit. (F) local Parent Forecast Comparison squared error scores. (O) local Parent Original Comparison squared error scores.

Cao *et al.* (1996) stated that the power law kernel (equation 2.5), which is similar to the IBQ kernel, should be preferred over the Gaussian kernel for seismicity in California. Their reasoning is based on their choice of bandwidth for the Gaussian kernel, which produced an over-smoothed model from the earthquake catalogue. Whereas the power law kernel might produce better results than the Gaussian kernel for global kernel estimation, global kernel estimation does not forecast the future seismicity as well as adaptive kernel estimation. Further, global bandwidths have to be adjusted to different kernel functions, i.e. the same bandwidth value for different kernels gives, in general, different effective bandwidths (for further explanation see e.g. Wand and Jones 1995). This has not been considered by Cao *et al.* (1996).

Figure 2.17 shows the seismicity representation for New Zealand modelled with the adaptive and global Gaussian kernels and with and without temporal filtering. For the global Gaussian kernel, many local features in both models cannot be matched, whereas for the adaptive Gaussian kernel most local features can be matched for the two time periods, e.g. around latitude -46° , longitude 171° for the global kernel the five clusters in the earlier time period cannot be matched to the three clusters in the later time period. In the PFC plots, differences that appear in the global case (e.g. latitude -46° , longitude 171°) do not appear in the adaptive case. Thus the earlier seismicity representation modelled by adaptive kernel estimation better forecasts the corresponding later seismicity representation compared to the representations modelled by the global kernel estimation. The differences in the POC plot for the global model are very similar to the differences for the adaptive model.

Without temporal sequence filtering, the intensity of local seismicity features is often different in the earlier and later models. In the PFC plot, the differences

are mostly due to temporal sequences. For the filtered model, the differences are much lower and the earlier model can better forecast the later model. The differences in the PFC plot for the filtered representation reflect real fluctuations in the recorded seismicity (e.g. the region around Arthur's Pass, latitude -43° , longitude 172° , had higher seismic activity during the nineties). In the POC plot, the differences between the unfiltered model representation and the catalogue are small compared to the filtered representation and the catalogue, consistent with the non-removal of aftershocks from the original earthquake distribution.

All the models we used in this study are static in time and do not model any variations of either global or local seismicity changes in time. However, time-varying models can be tested with the presented methods as well. To test time-variance, two time periods of possibly different spatial activity have to be compared, e.g. it is likely that an earthquake occurrence model derived from the historical earthquake catalogue would be different from an earthquake occurrence model based on geological fault information (see chapter three and four).

Figure 2.17. Comparison of kernel models. *The probability distribution of crustal earthquakes in New Zealand (above 40 km) generated by adaptive Gaussian kernel estimation with temporal sequence filter (a), global Gaussian kernel estimation with temporal sequence filter (b), and adaptive Gaussian kernel estimation without temporal sequence filter (c) from 1962 to 1990 (I) and from 1991 to 1997 (II) above magnitude 4.0. (F) local Parent Forecast Comparison squared error scores. (O) local Parent Original Comparison squared error scores.*

6 Summary

The comparison of different kernel estimations provided the following results:

- Global kernel estimation performs worse than adaptive kernel estimation, because it cannot forecast the future seismicity as well as adaptive kernel estimation, while staying as close to the original earthquake distribution. Models produced with small global bandwidths fail to forecast the location of future earthquakes, especially in regions of low activity, whereas models produced with high global bandwidths do not stay close to the seismicity distribution of the original earthquake catalogue. It can be concluded that global kernel estimation is not useful for the construction of earthquake occurrence models.
- The Gaussian kernel seems to better forecast the location of future earthquake activity compared to the IBQ kernel, while staying closer to the original earthquake distribution. The long tails of the IBQ kernel dominate the calculation of $c_2(\mathbf{x}_i)$. As a result, local features which are produced by single earthquakes, and therefore not seen in other observation intervals, are still inherent in the seismicity representation even for high bandwidths, whereas the background seismicity of low activity regions is too high even for low bandwidths.
- The global bandwidth $c_1 = 1.0$ seems to provide the best results for the Gaussian kernel. The choice of bandwidth is connected to the choice of the size of the cells. The optimum value of one for the adaptive Gaussian kernel reflects the ‘sensible’ binning of the data.

The cell sizes for New Zealand and Australia have been chosen in a way that they are comparable to the sizes of the seismicity features.

- Models with temporal sequences filtered out forecast the locations of future earthquakes better than unfiltered models. Although the unfiltered models stay closer to the distribution of the original earthquake catalogue, temporal sequences should be filtered, because they lead to an over- or under-estimation of local activity rate. The closeness of the unfiltered model to the original distribution reflects the fact that the temporal sequences have not been removed from the original distribution. After a main shock occurrence model has been produced, temporal sequences may be incorporated using an appropriate occurrence model for these sequences (e.g. Omori's law for aftershocks).

Using adaptive kernel estimation with temporal sequence filtering on the historical earthquake catalogue of New Zealand provided the following results for earthquake occurrence:

- Whereas shallow earthquakes are mostly part of temporal earthquake sequences (such as main shocks with aftershocks), the subducting slabs in New Zealand has been nearly free from temporal activity variations for the last forty years, suggesting that deep earthquakes in New Zealand do not produce many multiple events, as observed in the past.
- The probability distribution of main events has been stable in New Zealand during the last forty years.

- No spatial high activity clusters have been observed to decrease to background activity during the observation time. I infer that there is a higher risk of the occurrence of medium to large earthquakes associated with these clusters during the short to medium term (a few years to decades).

These results will be discussed in further detail when the seismological occurrence model is compared to the geological occurrence model in chapter four.

GEOLOGICAL OCCURRENCE MODEL

Earthquake occurrence models inferred alone from historical earthquake catalogues can only be expected to be useful for short to medium time intervals (i.e. not longer than the historical observation time, which is typically about a hundred years). Since the return times of individual fault ruptures are generally longer than the time span during which earthquakes have been recorded (even very active faults have, in general, return times of a few hundred years), geological information has to be considered when an occurrence model for longer time periods is developed. Information about regional earthquake occurrences over the last few thousand years can be obtained from compilations of mapped, active faults. Such information only includes the largest mapped surface ruptures.

In fault compilations several measured parameters of individual faults, such as mapped fault trace and measured single event displacement are recorded (it is often assumed that faults have a similar displacement for each rupture, e.g. Schwartz and Coppersmith 1984). To test if the recorded fault parameters (that are mostly acquired by geological studies) match the observed source parameters of earthquakes from the last century, the source parameter scaling relationships of these earthquakes have been revised. The results of this study can be found in appendix one and are reviewed in section one of this chapter.

The fault compilation used for this study is discussed in section two. In this section the numerous problems of the parameters of mapped, active faults are

reviewed. Further, the source parameter scaling relationships discussed in section one are used to detect inconsistencies in the recorded parameters of the fault compilation. It is then explained how these inconsistencies can be corrected.

In section three a ‘dynamic stochastic fault model’ is introduced. This model has been developed in this study in an attempt to overcome the numerous problems connected with the geological fault parameters. These problems would bias any earthquake occurrence model if not taken into account. The results of using this model for the generation of earthquake occurrence models are discussed in section four.

Every earthquake occurrence model has to be consistent with the theoretical plate deformation rates. The moment release rates from the fault simulation have been converted into slip rates and compared to deformation rates predicted by plate tectonics in section five.

1 Earthquake Source Parameters

Since faults are the sources of earthquakes, the parameters of the fault compilation (i.e. fault length, single event displacement, etc.) have to match the earthquake source parameter relationships (i.e. seismic moment M_0 , rupture length L , rupture width W , and displacement d) that have been inferred from earthquakes from the last century. Since previous authors have come to different conclusions about how earthquake source parameters scale with each other (e.g. Scholz 1982, Romanowicz 1992, Wells and Coppersmith 1994), the scaling relationships have been re-examined in the framework of this project. For that purpose, the databases of several past studies have been combined and analysed with appropriate statistics. This new study has been published (Stock and Smith

2000) and can be found in appendix one. The results that are important for testing the recorded parameters of the fault compilation are summarized here. Some additional work on source parameter relationships is presented here as well.

1.1 Scaling Relationships

Kanamori and Anderson (1975) introduced the concept of self-similarity of earthquakes based on their observation that the mean aspect ratio L/W and mean strain drop d/L are constant (i.e. any deviation from the mean is due to scatter) for all earthquake sizes (see also Abercrombie 1995). Purcaru and Berckhemer (1982), Scholz (1982), and Scholz *et al.* (1986) found a breakdown of self-similarity (the aspect ratio grows with magnitude for large earthquakes) which can be explained by the finiteness of the width of the seismogenic layer, which limits rupture width (Pacheco *et al.* 1992). Whereas Scholz (1982) found that stress drop stays constant, Romanowicz (1992) inferred from her data set that rupture width limits displacement and thus stress drop is not constant for large earthquakes (see also Scholz 1994a, Romanowicz 1994, Scholz 1994b, and Pegler and Das 1996). The newer study of Wells and Coppersmith (1994), to the contrary, could not find any breakdown of self-similarity.

After re-examining the data, I found that large dip-slip and strike-slip earthquakes scale in a different way. Self-similarity breaks down for strike-slip earthquakes, but not for dip-slip earthquakes. Although this was observed earlier (e.g. Purcaru and Berckhemer 1982), it had been explained by a wider range of widths for dip-slip earthquakes (Scholz *et al.* 1986). I could not find any deviation from a mean constant strain drop for dip-slip earthquakes of all sizes and regions, and suggest that, for dip-slip earthquakes, the maximum possible

rupture width is limiting rupture length and displacement, and hence magnitude. In contrast, as shown in previous studies (e.g. Purcaru and Berckhemer 1982, Scholz 1982), the length of strike-slip earthquakes is not limited by width and thus they can grow long, which leads to increasing aspect ratios.

The resulting scaling relationships between seismic moment and rupture length are:

$$\begin{aligned} \log M_0 &\sim 3 \log L, \text{ for dip-slip and small strike-slip and} \\ (3.1) \\ \log M_0 &\sim 2 \log L, \text{ for large strike-slip.} \end{aligned}$$

Romanowicz's (1992) findings, that rupture width limits displacement, have been clearly rejected by my study. Nonetheless, there is the possibility that rupture area, rather than rupture length alone, limits displacement (e.g. Gross 1996). Another possibility would be a breakdown of strain drop constancy for very long faults (e.g. Scholz 1994b). Since the influence of rupture width on displacement is yet not completely resolved, I assume a constant strain drop for the rest of this study (following the principle of Ockham's razor by choosing the simplest model consistent with the data).

1.2 Frequencies

To be able to estimate earthquake magnitude from a given rupture length, the mean ratio $\log (M_0 / L^3)$ (for dip-slip and small strike-slip) and $\log (M_0 / L^2)$ (for large strike-slip) is used. For this purpose, I have studied the frequency distribution of these ratios (see figure 3.1a). All frequency distributions are approximately log-normal. The logarithmic mean ratio for dip-slip and small strike-slip earthquakes is 14.8 ± 0.8 , and for large strike-slip earthquakes $16.0 \pm$

0.4. For simplicity, I used 15.0 and 16.0 respectively to estimate earthquake magnitudes from rupture lengths in this study (see equation 3.2).

Figure 3.1: Source Parameter Frequency Distributions. *a)* Seismic moment [Nm] to rupture length ratio [km] distribution for different faulting mechanisms. n is 3 for dip-slip and small strike-slip earthquakes, and 2 for large strike-slip earthquakes. *b)* Rupture length [m] to displacement ratio [m] distribution for different faulting mechanisms. The black line is the log-normal distribution $N(4.6, 0.2)$, which is shown for reference. (The green and the red line show normal and reverse earthquakes respectively in both plots.)

Further, I have studied the rupture length to displacement ratio. The frequency distribution of this ratio is log-normal for all faulting mechanisms (see figure 3.1b). This strengthens the argument that displacement scales with rupture length. If displacement would scale with rupture width, there would be a tail in the distribution of the strike-slip earthquakes (similar to the aspect ratio, see figure A1.6). This tail would indicate a decreasing strain drop with rupture length. The mean of the logarithmic ratio $\log(L/d)$ is 4.6 ± 0.4 , which corresponds to a mean ratio L/d of about 4×10^4 .

The scatter of the L/d ratio is rather big (although reasonable for the logarithmic ratio). It is a mixture of the measurement uncertainty of each rupture length and displacement and true fluctuations in stress drop (e.g. regional effects, random behaviour). Under the expectation that the measurements of rupture length and displacement are more accurate in recent years and the measurement uncertainty decreases systematically with time, the decrease could be projected into the

future. When the predicted measurement uncertainty becomes very small at some point in the future, the true fluctuation of stress drop could be estimated. I have studied the logarithmic mean ratio and standard deviation for different time intervals (see figure 3.2). Whereas the mean ratio stays approximately constant, the standard deviation is slightly decreasing if very small events are omitted. The reasoning behind the exclusion of very small events is that the errors in the measurements of rupture length and displacement are proportionally bigger for smaller earthquakes. Since the proportion of small earthquakes is increasing in recent decades the inclusion of these events would lead to a bias in uncertainty. Since the scatter is very large for all time periods, the true fluctuations of strain drop could not be estimated, but it is not likely to be above ± 0.25 for the logarithmic ratio. It could still be below this value if it only reflects the scatter due to measurement uncertainties.

Figure 3.2: Rupture Length to Displacement Ratio in Time. *Logarithmic rupture length [m] to displacement [m] mean (a) and standard deviation (b) are plotted for different time intervals. The values indicated by the red squares include all events. For the green dots small earthquakes ($L < 10$ km, $d < 0.1$ m) have been excluded.*

1.3 Summary

A summary of the source parameter relationships I found in this study follows.

To estimate seismic moment from rupture length I used the following relationships:

$$\log M_0 [Nm] = 15 + 3 \log L [km] \quad , \text{ for dip-slip and small strike-slip and} \quad (3.2)$$

$$\log M_0 [Nm] = 16 + 2 \log L [km] \quad , \text{ for large strike-slip.}$$

The moment magnitude is defined by (Hanks and Kanamori 1979):

$$M_w = 2/3 (\log M_0 - 9.1) \quad . \quad (3.3)$$

To estimate displacement from rupture length I used the following relationship:

$$L [m] = 4 \times 10^4 d [m] \quad . \quad (3.4)$$

The rigidity of the elastic medium μ is defined by (e.g. Aki and Richards 1980, 53):

$$\mu = M_0 / L W d \quad , \quad (3.5)$$

From figure A1.6 a mean aspect ratio L/W of about 1.5 can be estimated. Together with the other scaling relationships this implies a mean rigidity of $\mu = 6 \times 10^{10}$ Pa. The rigidity of surface faulting is typically given by $\mu = 3 \times 10^{10}$ Pa (e.g. Bonilla *et al.* 1984), and rigidity increases with depth. Since the rigidity of $\mu = 6 \times 10^{10}$ Pa inferred from the estimated source relationships (equations 3.2 and 3.4) is reasonably close to the observed rigidities (considering that a small change in equations 3.2 or 3.4 would lead to a big change in μ), it can be concluded that the estimated source relationships fit the ‘true’ source relationships relatively well (i.e. within 0.3 in $\log \mu$). The source relationships of some studies (e.g. Wells and Coppersmith 1994) cannot be combined in this way. They rather give mean relationships between pairs of parameters and are not interested if the estimated source relationships fit together. The advantage of

my approach is that regional effects like higher stress drop or lower rigidities can be incorporated by correcting individual source relationships.

A very recent study of Mai and Berora (2000) revised the scaling relationships of Wells and Coppersmith (1994) and their source scaling relationships between magnitude, length, and width agree very well with my results within the error limits. As in this study, their rupture length to displacement relationship shows a very high uncertainty, but their findings also reject the scaling relationship found by Romanowicz (1992).

2 Parameters of Mapped Active Faults

To test the methodologies developed in this chapter a fault database of reasonable quality is needed. The perfect database would include independently observed fault parameters (such as mapped fault length and slip rates), a complete map of active faults that produce earthquakes above a certain magnitude (for example magnitude seven), and quantitative uncertainties for the observations. However, such a database does not exist for New Zealand. Given the complexity of the fault distribution in New Zealand, and the limitation that only surface ruptures are accessible, it has to be questioned if such a database can be produced within the near future.

However, to validate any hypothesis about earthquake occurrence, it has to be falsifiable and tested against a fault database. The recent national fault compilation of Stirling *et al.* (1998) has been used for this study (see appendix two). This fault compilation includes fault parameters for a majority of mapped, active faults in New Zealand. For each fault a simplified fault trace, magnitude, and mean return time (recurrence interval) are given. If known, slip rate and

displacement are also given. The data have been mainly compiled from other studies. If no observed values for magnitude or return time were known, those parameters were inferred from other known parameters. For a detailed description of each fault parameter see Stirling *et al.* (1998). For a brief discussion of the fault parameters see appendix two.

It should be noted that the database of Stirling *et al.* (1998) may have not recorded all fault traces of large earthquakes and also often inferred magnitude and return time from single event displacements and fault lengths (see Stirling *et al.*, 1998, and appendix two). The inference of magnitude and return time does not introduce any bias, if the regressions used to calculate these values are correct.

Each fault in this compilation is treated characteristically, i.e. it is expected that the recorded values of every individual fault are the same for each rupture (see subsection 2.2). I have removed the subduction interfaces from the compilation, since the main goal is the comparison of surface braking faults and recorded crustal earthquake activity. In the compilation, the Taupo Volcanic faulting zone has been simplified to two faults. This simplification will affect earthquake occurrences in this zone, and although the two faults are used in this study, the results should be treated with care (or ignored).

2.1 Fault Parameter Sampling Biases

When measuring fault parameters, there are a lot of difficulties involved that lead to biases in the recorded fault parameters. The major difficulties and sampling biases are discussed in the following:

- Wells and Coppersmith (1994) observed that for earthquakes in the last century the surface breaks are systematically smaller than the subsurface rupture length for all faulting mechanisms. This effect increases for smaller earthquakes, but is not linear since few small earthquakes produce surface breaks similar to the subsurface rupture length. If uncorrected surface lengths were used as an estimate for true rupture lengths, rupture lengths would be consistently underestimated.
- Since for smaller earthquakes the surface break decreases, the chance of not detecting individual faults increases with decreasing surface length / magnitude (even some earthquakes above magnitude seven show barely any surface breaks and can be missed; one good example is the Wairarapa, 1942, earthquake with a surface wave magnitude of 7.2 and no surface break, Downes *et al.* 1999). Additionally, old surface breaks could have vanished over time, for example by erosion. Although all faults that generate moderate earthquakes could possibly be detected (e.g. Lettis *et al.* 1997), resources for such studies are mostly lacking. Thus, it has to be determined above which magnitude (fault length) all faults in an area have been mapped (e.g. Cameron 1997).
- Not all faulting areas are equally mapped, with some areas being well mapped while others have been scarcely studied. This has an impact on the completeness of mapped faults. But also fault lengths can be in error, where some faults have been only mapped within a region

but not beyond it. There is always the possibility that such faults continue and are longer than mapped (e.g. Nicol *et al.* 1996).

- Single event displacements are often measured only on the surface. The displacement varies along the length of the fault with higher values to the middle and lower values towards the tips of the fault. Since the tips are usually buried under the surface, the geologically measured average displacement tends to be higher than the true average displacement of the whole fault trace. On the other hand, the maximum of the displacement lies under the surface, leading to a geological underestimation of average displacement. These sampling errors lead to a high uncertainty in the average displacement (Wells and Coppersmith 1994 studied the relationship between surface displacement versus subsurface displacement and found that the uncertainty of average displacement is indeed high). Similar sampling biases are attached to displacements that are measured in trenches dug at a fault.
- Another sampling bias occurs when one fault trace that produced one event in the past has been mapped as two separate faults (i.e. the middle of the fault has not been mapped). This will firstly have an effect on the estimated displacement, which will be too high for the two short faults. Further, the magnitude (inferred from a too short fault length) will be too low and it will be inferred that there will be more earthquakes produced in the same area and time interval. The linkage of two faults that produce separate earthquakes would have the opposite effect. The possibilities of individual mapped fault

segments joining or rupturing separately will be discussed in subsection 2.2.

- In the fault compilation used in this study earthquake magnitudes are mostly inferred from fault area (and thus fault lengths) and / or observed displacements. All sampling biases in the measured fault parameters will determine the (in)accuracy of the estimated magnitude. For example, since fault lengths are usually mapped too short, the estimated magnitude that the faults would produce will be too low, if inferred from the uncorrected fault length (e.g. Stirling *et al.* 2000).
- The distinction of whether a fault is active or inactive is usually made by reference to a fixed time interval (125,000 years for New Zealand, Mark Stirling, pers. comm.). As shown in subsection 4.6 the fault compilation does not include many faults with return times longer than 20,000. It is possible that many of the faults with return times longer than 20,000 have been classified as inactive (or are not mapped at all). For example, the White Creek fault, which has an estimated return time of 34,000 years), has not been recognised as an active fault until the Buller, 1929, earthquake, which ruptured this fault (Berryman 1980). If a lot of faults with long return times have been classified as inactive the fault compilation is under-sampled. This effect has the biggest impact in regions where there are numerous faults and the return times of these faults are long in general (e.g. North West Nelson). Such a region might have an

earthquake occurrence rate that lies well above the occurrence rate estimated from fault activity.

- Earthquake occurrence rates do not follow Poisson statistics, but are rather clustered in space and time (Kagan and Jackson 1991) and the variations in the individual return times are high as they are influenced by fault interactions (e.g. Becker and Schmeling 1998). This can have implications for the estimated return time for individual faults. If a fault has been more active than average over the time in which events have been dated, then the estimated return time will be too short. Similarly, too long return times will result for faults that have been less active than average over the time interval in which events have been dated.

2.2 Characteristic versus Uncharacteristic Behaviour

Schwartz and Coppersmith (1984) found that some faults rupture with similar earthquake source parameters every time they produce an earthquake, i.e. the same fault segment ruptures over and over again and produces similar displacements for each individual rupture. They called such behaviour ‘characteristic’ and proposed that many faults behave in this fashion. Such a model allows the construction of earthquake occurrence models directly by using the observed fault parameters without any modification (once sampling biases are corrected).

More recent studies question the characteristic earthquake model. Newer data shows that the faults studied by Schwartz and Coppersmith (1984) have variable rupture patterns and faults behave ‘uncharacteristically’ (e.g. Grant 1996). Schwartz (1999) has shown that large subduction zones behave

‘noncharacteristically’. For the rest of this study, I define the ‘uncharacteristic model’ to exclude any characteristic fault behaviour. In the uncharacteristic model, each segment changes its behaviour over time and sometimes joins other segments to produce bigger earthquakes with higher displacements and longer return times, rather than consistently producing similar source parameters. The ‘noncharacteristic model’ is defined here to allow a mixture of both characteristic and the uncharacteristic behaviour. The uncharacteristic and noncharacteristic models produce a different earthquake occurrence model from the characteristic model since they allow for a much greater variability of earthquake magnitudes and return times.

2.3 Faults versus Earthquake Sources

The fault parameters of the fault compilation used in this study have to be consistent with the earthquake source relationships discussed earlier (equations 3.2 and 3.4). Any deviations will indicate sampling biases in the fault compilation (or a consistent deviation of earthquakes in New Zealand from the world average). To develop an earthquake occurrence model based on fault information, magnitude, return time, and location (i.e. the fault trace) of a fault have to be known. The magnitude can be directly inferred from rupture length (equations 3.2 and 3.3). The mean return time T can be inferred from:

$$T = d / \sigma \quad , \quad (3.6)$$

where σ is the slip rate and the displacement d can be inferred from rupture length (equation 3.4). If the rupture length could be inferred from the mapped fault trace (i.e. the mapped length of the fault), the only parameters needed for the development of an occurrence model would be fault trace and slip rate, since the other parameters could be calculated with those two parameters. The slip

rate is a geological constraint that both the characteristic and the uncharacteristic model have to satisfy. Both models would treat mapped fault traces differently. The characteristic model would use the fault traces as they were mapped (apart from length corrections), whereas the uncharacteristic model would allow different fault traces to join or to break into separate segments.

It has to be noted that, by inferring return times and magnitudes from rupture lengths, it is assumed that all faults follow the mean behaviour of earthquake sources. This may not be appropriate for individual faults, since they might significantly deviate from the 'normal' behaviour. However, parameters for individual faults are mostly very uncertain (any deviation could reflect sampling biases), so it is not clear how much faults deviate from the mean fault behaviour. Random deviation from the mean source parameters could be modelled by including the uncertainties in the source parameter relationships. Since the main goal here is to establish a mean behaviour of fault rupture, the uncertainties have not been taken into account. Although regional differences in the source parameters relationships could not be found (see appendix one), there is still the possibility that faults in a particular region deviate systematically from the average behaviour. To take such a deviation into account, it firstly has to be clarified how big it is. There is not enough data for New Zealand to find a statistically significant difference from the average source parameter relationships.

The advantage of this approach is that rather than having three characteristic parameters (magnitude, return time, and fault trace), only one characteristic parameter is left (fault trace). It is fairly easy to convert characteristic fault lengths into uncharacteristic ones, by allowing different fault traces to join and to

break into separate segments. This approach has the additional advantage that only the sampling bias of fault length has to be corrected, which can be done with the help of the observed displacements. Both rupture length and displacement have to fit the scaling relationship found earlier (equation 3.4). Any deviation from this scaling relationship can be used to correct sampling biases of fault lengths.

Since the displacement is estimated from a corrected fault length it will also be ‘sampling bias free’ (meaning that it will fit the source scaling). Thus, the sampling biases of return times will only depend on biases in the slip rate and the omission of faults (due to ‘missed’ fault traces or faults of very long return times). Magnitudes will be free of sampling biases, because they are inferred from corrected fault lengths.

To study how well the compilation of New Zealand fault parameters fit the observed earthquake source relationships, I have inferred displacements from given mapped fault traces (equation 3.4) in the fault compilation and compared those values with the measured displacements. Figure 3.3a shows that the measured displacements are consistently higher than the displacements inferred from fault traces, independent of faulting mechanism. Further, when estimating return times from fault traces and slip rates, those estimated return times are consistently lower than the observed return times (figure 3.3b).

Figure 3.3: New Zealand Fault Parameters versus Source Parameters. *a) Theoretical displacements inferred from fault traces compared to measured displacements from the fault compilation (in meters). b) Theoretical return times inferred from fault traces and slip rates (theoretical displacements) compared to measured return times from the fault compilation (in years).*

This behaviour can be explained by two factors. Firstly, as noted beforehand, the mapped fault traces underestimate the true rupture lengths. This leads to estimated displacements that are too low if the length of the fault traces is not corrected. Secondly, for some short faults the observed displacement is very high (e.g. Alfredton, see appendix two). Such faults have probably ruptured in conjunction with other faults (e.g. the Alfredton fault might be connected to the Wairarapa fault). If such a segment ruptured alone it would most likely have had a significantly smaller displacement. Paleoseismologists might not be able to identify such small events. Thus it might not be clear how often such a fault joins another fault and how often it ruptures individually. Further, some small faults might have ruptured with very small displacements which could not be resolved as individual ruptures, and therefore several events might have incorrectly been interpreted as one single event. This would lead to measured displacements that are consistently too big. There is the possibility that strain drops are consistently too high for New Zealand, which would lead to a similar result. However, the strain drop would have to be about three times the average

strain drop worldwide. There is not enough data for New Zealand to decide if the strain drop is consistently higher than the world average.

To correct the underestimation of rupture length from mapped fault traces, I have firstly tried to match separate faults segments that are likely to connect to adjacent fault segments. Only faults with the same slip-type (or oblique faults) were allowed to join. Since the main interest in this study is to develop a fault model that can be used to generate an earthquake occurrence model, and the particular hazard model that results is only of secondary interest, I did this analysis judged on the spatial distribution of the faults alone, using common sense. For the generation of hazard models such an analysis should be undertaken by an experienced geologist. I was able to connect most of the strike-slip faults to at least one other close fault. After establishing all possible connections between strike-slip faults, I found that for most strike-slip faults the displacement inferred from the new fault traces to be relatively close to the observed displacement (see figure 3.4). Two faults with not so good fits are the Ohariu fault and the Pukerua-Shepherds fault. Both faults most likely continue offshore where they have not been mapped completely. The agreement between observed displacement versus displacement inferred from fault traces once faults that possibly produce ruptures together are connected suggests that mostly I have connected the faults properly. The remaining scatter could be due to variances in individual strain drop. This result suggests that the rupture lengths of strike-slip faults do not necessarily need to be corrected for New Zealand. It is still possible though that some strike-slip faults have been missed or are not fully mapped. Since the regions that have strike-slip faults have been mapped fairly well in New Zealand, the lengths of strike-slip faults are not corrected in this study. It is assumed that if individual faults are allowed to join, the resulting

earthquakes match the true occurrence rate reasonably well. Since the horizontal and vertical displacements are not recorded in the fault compilation, oblique faults have been treated similarly to strike-slip faults, because most of the oblique faults in the fault compilation have a significant strike-slip component. Oblique faults follow source parameter relationships that are in between the relationships of strike-slip and dip-slip faults (see figure A1.4).

Figure 3.4: Theoretical versus Measured Displacements for Strike-Slip Faults. *Theoretical displacements inferred from fault traces compared to measured displacements from the fault compilation of strike-slip faults after individual faults have been linked (in meters).*

Only a few normal and reverse faults could be connected with other fault segments. To correct rupture lengths for these faults, the observed lengths (fault traces) have been compared with the lengths inferred from observed return times and observed displacement (figure 3.5). With the help of equation A1.3 and A1.4 lines with the slopes 0.78 ± 0.47 for normal and 0.13 ± 0.71 for reverse faults could be fitted through the data points. For the normal faults a slope of $1/2$ has been chosen instead because most data point lie on this slope, which is within the error limit. The slope for the reverse faults is very steep and probably caused by the fact that even short faults have very large displacements and thus there are no data for small displacements. A slope of $1/3$ has been chosen instead, which seems to fit the data reasonably well (considering the quality of

the data, see page 74). For both slopes the uncertainty is very high. For these slopes corresponding intercepts of 0.91 ± 0.09 for normal and 1.59 ± 0.71 for reverse faults have been calculated.

Figure 3.5: Length Correction. *Measured length from the fault catalogue compared to length inferred from observed return time and displacement (in km). The dark red line corresponds to the correction relationships (equations 3.7).*

The following two linear relationships have been used to correct the fault lengths of normal and reverse faults:

$$\log \Lambda = 0.9 + 1/2 \log L \quad (\text{normal faults}) \quad (3.7)$$

$$\log \Lambda = 1.6 + 1/3 \log L \quad (\text{reverse faults}),$$

where Λ is the corrected length and L is the measured compilation length. These relationships are only valid for the correction of geologically measured fault traces in New Zealand. This result does not contradict the findings of Wells and Coppersmith (1994) that there are no linear relationships between observed surface and subsurface rupture lengths. The data from Wells and Coppersmith (1994) are collected from observed earthquakes and sometimes the surface rupture length from small earthquakes matches the inferred subsurface length very well. For the geological data of New Zealand small faults are represented by

a surface break that is systematically too low. Hence, the difference between this work and the study of Wells and Coppersmith (1994) is the difference in accuracy between observed rupture length and mapped fault length.

The need for such a correction is based on the assumption that the surface break is much smaller for small earthquakes than for larger earthquakes. An alternative interpretation of the data would be a high strain drop for New Zealand earthquakes. However, in figure 3.5 for the large earthquakes the observed surface break is relatively close to the estimated rupture length. It can be expected that for even larger earthquakes the match of the observed and inferred rupture length would be very close.

For the normal faults, apart from the five faults from the East North Island Fault Zone and the Edgumbe, 1987, earthquake, all faults follow the linear relationship very well (figure 3.5). The five faults from the east North Island fault zone probably continue offshore and thus are longer than mapped. The Edgumbe fault trace has been revised in a newer fault compilation and is mapped as longer now. It has to be noted that Edgumbe is the only ‘modern’ measurement. However, this shouldn’t be overemphasised given the scatter of modern observations (see figure A1.2). Additionally, the purpose of the correction is to match measured displacement and length of paleoseismic events. Since Edgumbe isn’t a paleoseismic event, it shouldn’t be included anyway.

The scatter is much bigger for the reverse faults and a linear trend is not as obvious as for the normal faults. The four shortest faults deviate the most. The Inangahua, 1968, earthquake has an observed average displacement of 40 cm. Since the earthquake had a surface wave magnitude of about 7.4 (Anderson *et al.* 1994) this displacement seems to be underestimated. The fault was probably

much longer and had a much higher subsurface displacement. For the Lyell fault, Blue Mountain fault, and Spy Law fault the fault trace has been used to calculate the return time in the fault compilation. Since the fault trace is likely to be too short, the return times are probably too short as well. Thus, the inferred fault lengths should be longer. When these four faults are taken out of the analysis the scatter is still large, but the general trend is that reverse faults are mapped much too short. For example, a rupture length of about 100 km has an average surface break of only 20 km (see figure 3.5). Thus, the chance of ‘missing’ reverse faults is very high, even for faults that produce big earthquakes.

Since the surface break of reverse earthquakes is so small compared to the subsurface rupture length it is possible that the measured displacement along the surface break is much bigger than the average displacement. This would mean that the lengths of reverse faults are overcorrected by my correction (equation 3.7). With the help of theoretical rupture models it could be possible to estimate a relationship between measured surface displacement and average subsurface displacement. This has not been done in this study since there seem to be further problems with this type of correction (see discussion in subsection 4.4).

The length corrections discussed above should only be used for New Zealand faults. In other regions of the world surface fault traces may be preserved better over time or they might have been mapped more extensively. It will probably be necessary to revise these relationships for an updated fault compilation for New Zealand. The length corrections may be related to sampling (e.g. area coverage, regional accessibility) rather than physical relationships (e.g. rate of erosion, stress drop).

In the fault compilation only magnitudes and return times are given for every individual fault. Since the return time changes when two different faults are combined to produce a larger event together, the slip rate is needed to estimate a corrected return time. The missing slip rates have been estimated from observed displacements or displacements inferred from rupture lengths and observed return times (equation 3.6, the new slip rates are shown in appendix two). This procedure is not as accurate as having an observed slip rate, since it has to be decided if the fault ruptures together with other faults to produce the observed return time.

3 Dynamic Stochastic Fault Model

One of the problems when developing areal earthquake occurrence models based on fault information is that earthquakes are represented by the projected faults onto the surface hence they are represented as line sources. Historical earthquake catalogues typically only record the hypocentres (point sources), and the projected epicentres are used for the development of occurrence models. To compare earthquake occurrence models based on fault information with occurrence models based on historical earthquake catalogues either the faults have to be converted to epicentres or the recorded epicentres have to be converted to faults. Similarly, for a three-dimensional study the fault planes have to be converted to hypocentres, or vice versa. For this study, I chose to convert fault lines to epicentres by simulating hypocentres occurring on the mapped, active faults over a fixed time interval (10,000 years).

It is general practice that the observed fault parameters are taken as the typical behaviour of a fault and left unchanged for every single rupture of the faults over the simulation period. Such a model can be called 'static' in time and

assumes that all faults behave characteristically. I use a ‘dynamical’ model where individual faults are allowed to produce ruptures together with other faults or where only part of a fault is allowed to rupture alone. This model is imposing noncharacteristic fault behaviour. Whether the dynamic fault model will produce characteristic or uncharacteristic fault ruptures depends on the properties of the considered faults and the imposed rules that allow or prohibit the linkage of faults.

3.1 Fault Linkage Based on Angle

To decide whether fault segments that are mapped as one fault in the fault compilation produce single or compound ruptures, the angle (orientation between both segments) and the gap (distance between segment tips) between pairs of segments is considered. There are several simulation studies that estimate the slip distribution for different angles between two fault segments that rupture together (e.g. Kase and Kuge 1998, Maerten *et al.* 1999). But there appears to have been no work investigating the probability for two segments rupturing together for a given angle between them. The lack of such studies can be explained by the complexity of fault tip zones. Fault tip zones include complications such as diffuse breakdown zones (multiple strands), significant fault bends (‘dog tails’ and ‘rainbows’), and stepovers (e.g. King 1986, Ward 1997b). Since there is no information about the details of the fault tips in the fault compilation, any complications have to be ignored. For this study, the orientation between two faults is assumed to be between 0° and 90° degrees. This choice imposes an equal chance of two faults joining whether the second fault continues to the ‘left’ or ‘right’ of the first fault with the same angle, i.e. the angle between two faults is treated symmetrically. This is not correct, because the dip direction and the vertical gap (stepover gap) relative to strike direction

between two segments would probably have an effect on the likelihood of linkage. Unfortunately, the influence of these asymmetries on the probability of fault linkage is not known.

To calculate the probability of two segments rupturing together depending on the angle between them, I use a one-sided standard normal distribution. Firstly, I draw a random number from a set of numbers between zero and infinity that are standard normally distributed. I multiply this number with a 'parameter angle' (e.g. 90°) and the product gives the 'cut-off angle'. If the angle between the two segments is less than or equal to the cut-off angle, the segments produce a linked rupture. The chance that two segments will produce a linked rupture when the angle between them is equal to the parameter angle (e.g. 90°) will be 68%. (By multiplying the drawn random number with the parameter angle I change the standard normal distribution to a normal distribution with sigma being the parameter angle). The probability of two segments producing a rupture together increases with the choice of a bigger parameter angle. In this study, I compared three different values of parameter angles: 22.5° , 45° , and 90° .

3.2 Fault Linkage Based on Distance

The maximum distance which allows two segments to produce joint ruptures has been well studied and is typically about 5 to 10 km (Harris *et al.* 1991, Harris and Day 1999, Magistrale and Day 1999) or 10% of the total length (An 1997). The probability of two faults rupturing together for a given distance between them has, to my knowledge, not been studied. Most likely, this probability depends not only on the distance (and angle) alone but also on other properties such as rigidity of the rock, amount of micro-cracks, etc. Since such properties are not known for faults in the fault compilation, the probability of two

segments producing joint ruptures has been calculated using a one-sided standard normal distribution. The calculation of the probability is equivalent to the calculation of the probability for a given angle. A drawn random number is multiplied with a 'parameter distance' (e.g. 10% of the total rupture length) to produce a 'cut-off distance'. The distance between the two segments has to be less than or equal to the cut-off distance or a 'maximum distance' (e.g. 20 km). A maximum distance has to be imposed to rule out the possibility that faults that are far apart could join (as they still would have a very small probability of joining). Parameter distances of 10% and 20% of total rupture length and maximum distances of 20 km and 40 km respectively have been used in this study. Those values may seem a little bit too generous (two faults with a gap of 10% of the total length in-between have a chance of 68% to produce a linked rupture although this gap length corresponds to the maximum gap length reported in the literature), but this choice should allow for the fact that the fault tips might be buried or that a subsurface connection exist between the two faults.

It should be noted at this point that the choice of parameter angles and distances may seem a bit generous. However, as discussed in section 4.1, the choice of the parameters had virtually no effect on the occurrence of earthquakes above magnitude seven.

3.3 Limits on Fault Linkage

To decide if two segments join, other criteria have been taken into consideration. Firstly, the fault produced by two segments has to continue into the same direction. Further, only faults of same slip types are allowed to join. Dip-slip faults are allowed to link with dip-slip faults that have a strike-slip component

(e.g. normal faults are allowed to join normal-oblique faults) and strike-slip faults are allowed to link with strike-slip faults that have a dip-slip component.

Faults with a logarithmic difference of 0.5 (base 10) in slip rate are allowed to join without any modifications (a logarithmic difference of 0.44 has been calculated as the scatter in the recorded slip rates in the fault compilation). Faults where the difference is bigger are allowed to join on a probabilistic basis, with the chance of linkage being one over the logarithmic difference. This compensates for the case where the slip rates of two linking faults are very different. If this would not be done one fault would produce too many earthquakes while the other one would produce too few earthquakes.

The limit of rupture length by maximum possible rupture width for dip-slip earthquakes has been considered (see appendix one). Normal faults are forbidden to have a rupture length over 70 km (which corresponds to a seismogenic width of 20 km and a dip-angle of about 25°; most normal faults have a dip-angle of about 50° and few normal faults have dip-angles of less than 20°, Jackson and White 1989). 70 km is the maximum recorded fault length for normal faults in the fault compilation (see figure 3.5). The maximum rupture length for reverse faults has been limited to 180 km (which corresponds to a seismogenic width of 20 km and a dip-angle of about 10°). Reverse faults generally have lower dip-angles than normal faults and thus bigger rupture widths (the global distribution of dip-angles of reverse earthquakes is bimodal with maxima at about 30° and 50°, Sibson and Xie 1998). The northernmost reverse fault in Hawkes Bay has a corrected rupture width of over 200km and would produce earthquakes bigger than magnitude 8.5. Since this fault lies offshore, it is plausible that it has been mapped in a very simplified way. I

divided this fault into two equal sized segments, to avoid the generation of overly large earthquakes from this fault. The limitation of length due to rupture width does not affect the connection of dip-slip faults very much, because for most dip-slip earthquakes that are mapped the strike directions are parallel, i.e. most dip-slip faults lie next to each other perpendicular to the strike direction rather than in strike direction. The only faults that are forbidden to join by the imposed length limit are the two simplified faults from the Taupo faulting zone and the offshore faults in Hawkes Bay. If a more detailed fault compilation were used, the imposed maximum lengths for dip-slip faults might have to be changed. It would be of advantage to use the dip-angle of the faults (which are not known very well) and the local seismogenic thickness to estimate maximum rupture length.

3.4 Simulation of Fault Activity

The simulation of a cumulated rupture distribution over 10,000 years is done by a computer program. For the simulation, a spatial distribution of faults is calculated for each year, i.e. the program decides which fault segments rupture together and which rupture alone for the particular year. To construct the annual distribution of fault ruptures, the program starts with the individual fault traces of the fault compilation. Since the fault traces in the compilation are simplified and individual segments have been joined, the program checks if the fault trace should be split into several segments depending on the angle between the segments of the fault. As explained above, the program draws a random number from a normal distribution and splits two segments if the angle between them is bigger than the cut-off angle. Once this has been done for all faults, the distance between all individual faults is checked. The decision, whether two segments join, includes several steps. Firstly, the total length of both segments is calculated

and a probability of both segments joining is calculated, as described above. If the segments would join, the two angles between each segment and the intermediate segment that would combine the original segments are calculated and tested to see if both segments join the intermediate segment. Segments that have been separated in the first angle test cannot re-join.

Once an annual rupture distribution has been established, the parameters for each fault are calculated. The slip rates of the joining segments are taken as the mean, and displacements, magnitudes, and return times are inferred from fault lengths (see earlier discussion). To separate small and large strike-slip faults (which has to be done to calculate an appropriate magnitude using equation 3.2), a fault length of 30 km has been used. The return time is then used to calculate the probability that the fault will rupture in this particular year. This probability corresponds to (approximate) Poisson behaviour over time. Faults rupture in clusters in time and space (Kagan and Jackson 1991), and thus the faults do not behave as Poisson. Since I am only interested in the mean behaviour of the faults over a long time (10,000 years), and not in fluctuations with time, Poisson behaviour can be safely assumed as an average behaviour.

If a fault ruptures, the event is equally distributed over the whole fault so that the integrated activity over the length of the fault equals one event. This enables the direct comparison of this model and the earthquake occurrence model based on the historical earthquake catalogue. Since all faults have different lengths and are mostly mapped too short, the results can look intriguing. Long faults appear to have a relatively lower activity over space than short faults. This is correct when it is assumed that earthquakes are generated by point sources. However, it appears as though long faults are less active than short faults which is not

correct. Further, due to the underestimation of rupture length, the activity of individual faults should have been distributed over a greater length. Since it is not known where the fault continues or where the unmapped parts are (otherwise they would be mapped!), the seismicity has been accumulated on the mapped part.

4 Results

For different simulations using different sets of simulation parameters, magnitude frequency distributions have been plotted. Those magnitude frequency distributions have been compared to the magnitude frequency distributions of the historical earthquake catalogue. To study the local earthquake occurrence distribution of different fault simulations, plots of the simulated fault distribution in New Zealand have been produced. Finally, the return time frequencies resulting from the simulations have been plotted.

4.1 Magnitude Frequency Distributions

Figure 3.6 shows the magnitude frequency distributions for several simulations with different simulation parameters (figure 3.6a) and the same simulation parameter (figure 3.6b). The magnitude frequency distribution of the simulated fault ruptures behaves approximately as predicted by the Gutenberg-Richter law. Angle parameters that are more likely to inhibit fault linkage produce more small earthquakes below magnitude seven than simulation parameters that encourage fault linkage. However, above magnitude seven the angle parameters seem to have little influence on the magnitude frequencies, and thus fault linkage. The two different gap parameters produce very similar results for all magnitudes. Figure 3.6b shows that the scatter of different simulations with the same simulation parameters is relatively small. Since the choice of simulation

parameters does not influence the results of earthquake occurrence above magnitude seven, the choice of simulation parameters seems to be of little importance for the construction of an earthquake occurrence model above magnitude seven based on mapped active faults.

Figure 3.6: Magnitude Frequencies for Different Fault Rupture Simulations. *a)* magnitude frequencies for different simulation parameters. Parameter angles of 22.5°, 45°, and 90° and parameter gaps of 10% and 20 km (0.1), and 20% and 40 km (0.2) have been studied. For comparison, the characteristic model is shown. The grey line shows a b-value of one. *b)* magnitude frequencies for a parameter angle of 45° and a fixed gap of 10% and 20 km for three simulations. The frequencies are given in events per year and are not cumulative.

For comparison, the magnitude frequency of the fault compilation if all faults are treated characteristically (and are not corrected for length) is shown. It can be seen that the characteristic model produces significantly fewer earthquakes than

the dynamic model. The overall higher activity of the dynamic model can be explained by the length correction. Since many faults have been corrected to be longer, they produce either bigger earthquakes, or more small earthquakes if fault segments rupture independently. The higher activity for large earthquakes could additionally be explained by the allowance that single faults join and produce bigger earthquakes. It has to be emphasised though, that the total higher activity can only be explained by the correction of sampling biases. It is not caused by the dynamic model alone.

Another important result is the observed maximum magnitude of the simulated fault ruptures. Over a time period of 10,000 years the dynamic model produces a maximum magnitude of 8.4, compared to 8.1 in the characteristic model. In total, the occurrence of magnitude eight and above earthquakes lies well above the occurrence of magnitude eight earthquakes in the characteristic model. Earthquakes of that size might be rare but are possible, if one considers a crustal thickness of 30 km (for example for offshore reverse faults in Hawke's Bay) and a possible dip angle of 15° , which would result in a magnitude 8.4 earthquake. Earthquakes with even lower dip-angles are observed in the Himalayas (Sibson and Xie 1998).

4.2 Deficiency of Large Strike-Slip Earthquakes

There seems to be a deficiency of strike-slip earthquakes between magnitude seven and a half and magnitude eight. Between magnitude six and a half and magnitude seven and a half the number of earthquakes seems to be too high (considering that some faults with magnitudes above seven have not been mapped). The gap between magnitude seven and a half and magnitude eight could possibly be filled in with earthquakes from regions where the fault model

does not show enough activity to match the activity inferred from the historical earthquakes (e.g. west Nelson as discussed in chapter four).

This mismatch could be caused by faults that have not been mapped or faults that have long return times and have been erroneously classified as inactive. Most likely, smaller faults would (at least) have been overlooked (and not included in the fault compilation) in the same proportion predicted by the Gutenberg-Richter distribution as large faults. Additionally, as small faults produce a smaller surface rupture than large faults, it is even more likely that they have been overlooked. Thus, the addition of earthquakes between magnitude seven and a half and eight to the magnitude frequencies would require the addition of earthquakes below magnitude seven and a half in the same proportion and raise the already high occurrence rate of earthquakes between magnitude six and a half and magnitude seven and a half.

It is more likely that some of the earthquakes between magnitude six and a half and magnitude seven and a half should be bigger, to fill in the gap between magnitude seven and a half and magnitude eight. This can be achieved if some faults in the dominant strike-slip regions (i.e. east North Island and Marlborough) have not been included in the fault compilation. Those faults would add to the probability of fault segments linking (as that would produce more possibilities for faults to connect with each other) and the number of single segment ruptures would decrease in favour of linked ruptures. For example, a previously unmapped fault could act as a bridge between two mapped faults that would individually produce magnitude six and a half earthquakes, and produce a compound fault rupture of magnitude seven and a half (see figure A1.4).

The increase in the probability of segment linkage does not necessarily require the addition of previously unmapped faults with short return times, but could also be explained by the inclusion of faults with long return times, that have been previously classified as inactive. If enough of such faults were added, more large earthquakes, at the cost of medium-sized earthquakes, will occur. This is an important result and it will be studied further in the rest of this chapter and the next chapter.

4.3 Magnitude Frequencies of Historical Earthquakes

Figure 3.7 shows the magnitude frequencies (non-cumulative) for different time periods of the historical earthquake catalogue compared to the magnitude frequencies of the dynamic fault model. The same time intervals as in chapter two section one have been used for the historical earthquakes (see table 2.1). For the earthquakes before 1942 magnitude frequency plots have been constructed for individual years equivalent to the methodology in chapter two, section one. A cut-off magnitude of 7.0 has been found from 1870 onwards. This time corresponds to the start of organised earthquake reporting in New Zealand. For the historical earthquake catalogue all magnitude frequencies have a b-value of about one. The activity level of all time intervals varies, with the nineties being very active and the earlier decades being less active. The activity of the simulated faults seems to approximately fit the magnitude frequency distributions of the historical earthquakes.

Figure 3.7: Magnitude Frequencies for Different Time Intervals. *The magnitude frequencies of different time intervals of the historical earthquake catalogue compared to the magnitude frequency of simulated faults. The simulation parameters are 45° for the angle and 10% and 20 km for the gap. The grey line shows a b -value of one. The frequencies are given in events per year.*

The historical earthquake catalogue contains aftershocks and earthquakes related to the shallow part of the Hikurangi subducted plate below the eastern North Island. The aftershocks have not been filtered in this analysis, because there is the possibility that large aftershocks have been recorded in the fault compilation (as fault offsets), leading to shorter return times for faults on which aftershocks have been identified as fault breaks. Further, the earthquakes below the eastern North Island have not been filtered either, because it is possible that they produce surface ruptures (i.e. the rupture might start at the base and progress upwards to the surface, e.g. Cole and Lewis 1981), and thus fault breaks would have been recorded in the fault compilation.

Figure 3.8: Magnitude Frequencies. *The magnitude frequencies (a) and cumulative magnitude frequencies (b) of the historical earthquake catalogue compared to the simulated faults and characteristic faults. The grey lines show a b -value of one. Strike-slip frequencies include oblique slip events as well. The frequencies are given in events per year.*

In figure 3.8 the magnitude frequency distributions of the historical earthquake catalogue have been joined together. The earthquakes for all years above the completeness magnitude have been added for that purpose (see table 2.1). The steps in the magnitude frequency indicate different activity levels for the different time periods (the b -value is still one as observed in figure 3.7!). The earthquake activity has increased over the decades since the forties and is highest in the nineties. The activity in and before the forties was high as well. It has to be noted that the activity of the historical earthquake catalogue above magnitude

seven might be slightly underestimated. Firstly, the local magnitude might be slightly underestimated for large earthquakes (see chapter two, subsection 1.4). Further, the only historically observed earthquake above magnitude eight (Wairarapa, 1855) has not been included, as only earthquakes after 1870 have been included.

The magnitude frequencies from the dynamic model drop below the activity of the historical earthquake catalogue between magnitude seven and a half and magnitude eight and rise to the activity level of the historical earthquake catalogue between magnitude six and a half and magnitude seven. This behaviour can be explained by missing strike-slip faults (or fault segments) as discussed in subsection 4.2. The magnitude frequencies of the simulated strike-slip fault ruptures show that the high activity level between magnitude six and a half and seven is indeed produced by the strike-slip faults. At about magnitude seven, the magnitude frequencies of the simulated strike-slip faults drop rapidly below the magnitude frequencies of the historical earthquake catalogue with a b -value higher than one. At magnitude eight the activity of the strike-slip faults is producing relatively more earthquakes than between magnitudes seven and eight, not following the trend of a b -value greater than one.

As the historical earthquake catalogue clearly follows a b -value of one in the magnitude range where the strike-slip earthquakes have a b -value of above one, it is likely that the high b -value is an artefact. If some faults that generated earthquakes between magnitude six and a half and seven in the simulation would rather produce earthquakes above magnitude seven and a half, the b -value could decrease to one, matching the b -value of the historical earthquake catalogue. It is even possible that some of those faults would sometimes connect in a way that

they produce earthquakes of magnitudes above eight. This would mean that the earthquakes simulated by fault ruptures are not complete above magnitude eight. Although the strike-slip faults that typically produce earthquakes above magnitude eight might be mapped completely, the incompleteness of faults that typically produce earthquakes below magnitude eight (but sometimes are still able to produce bigger earthquakes) could still cause incompleteness of simulated strike-slip earthquakes above magnitude eight. The magnitude frequencies of the simulated strike-slip fault ruptures support the result that there are strike-slip faults missing in the fault compilation (section 4.2).

The magnitude frequencies generated by the characteristic fault model do not match the magnitude frequencies of the historical earthquakes as it clearly generates too few earthquakes for all magnitudes. The magnitude frequencies from the characteristic model are very similar to the simulated magnitude frequencies from the strike-slip fault ruptures. This means that the characteristic earthquake model does not seem to be able to generate enough reverse earthquakes above magnitude seven and a half (i.e. the number of ruptures on mapped reverse faults does not match the historically observed activity of reverse earthquakes). The lack of large reverse earthquakes in the characteristic model will be studied further in the rest of this thesis.

The characteristic model seems to predict too few earthquakes above magnitude seven and a half. Although the characteristic model produces too few earthquakes between magnitudes seven and seven and a half it is possible that it produces too many strike-slip earthquakes if one adds dip-slip earthquakes at those magnitudes (as the characteristic model seems to not produce enough dip-slip earthquakes). This is a similar behaviour to the dynamic model which

predicts too many strike-slip earthquakes between magnitudes six and a half and seven (compared to the historical earthquakes which include dip-slip earthquakes) and too few earthquakes above magnitude seven and a half. This suggests that the problem of missing strike-slip faults exists also in the characteristic model.

An explanation for the cluster of too many strike-slip earthquakes being about half a magnitude lower in the dynamic model than in the characteristic model is the allowance in the dynamic model that strike-slip fault segments can rupture independently. This would lead to systematically lower magnitudes, if there are not enough possibilities for strike-slip faults to link and produce bigger magnitudes. Another reason for this difference is the different calculation of the magnitude for strike-slip faults. In the fault compilation the source scaling relationships of Wells and Coppersmith (1994) has been often used to calculate the characteristic magnitude. The source scaling relationships found in this study imply a lower magnitude for long strike-slip faults (see subsection 3.1). Thus, the magnitudes of long strike-slip faults are systematically too high in the characteristic model (for example, the rupture length of an earthquake of magnitude 7.5 after my source model would correspond to a magnitude of 7.6 in the Wells and Coppersmith (1994) model).

4.4 Too many Large Dip-Slip Earthquakes

The activity of the dynamic fault model over magnitude eight is higher than the activity of the historical earthquake catalogue (when extrapolated, figure 3.8). As the activity based on simulated strike-slip fault ruptures is much lower than the total simulated activity above magnitude eight and since the normal faults are not long enough to produce earthquakes above magnitude eight, the high activity

above magnitude eight is mainly generated by simulated reverse fault ruptures. The high activity level could possibly be explained by an over-correction of length for reverse faults, as discussed in subsection 2.3.

In chapter four it is suggested that dip-slip faults behave uncharacteristically rather than characteristically. Since the length correction will produce very long and straight reverse faults, the corrected faults will always produce characteristic ruptures in the simulation, i.e. the reverse faults will produce ruptures with a maximum magnitude (and displacement) every time they rupture. They should be allowed to produce smaller earthquakes as well as large ruptures, which would decrease the return times (and number) of above magnitude eight earthquakes and add earthquakes below magnitude eight to the magnitude frequencies. This behaviour of wrongly corrected reverse faults will be studied further in the rest of the thesis.

It can be concluded that the methodology introduced in this study generated too many reverse earthquakes above magnitude eight and too few reverse earthquakes below magnitude eight. There might be a similar effect for normal faults, but it will not be as pronounced as the length correction for normal faults is not as big as for reverse faults and normal faults are not very long (i.e. not longer than 70 km, see figure 3.5, which corresponds to a magnitude of about 7.0, see figure A1.2).

4.5 Spatial Representation of Fault Ruptures

Figure 3.9 shows the simulated fault activity for two different simulation parameters, for simulated ruptures over magnitude seven and a half. Both plots look very similar supporting the previous result that the simulation parameters do not have a big effect on simulated earthquake occurrence above magnitude

seven (compare to figure 3.6a). The only difference between both plots can be seen in the northeast of the North Island (onshore). There is less activity in this region when the chosen simulation parameters favour single fault segments to rupture individually (i.e. small parameter angle). Such parameters lead to fewer big and more small earthquakes, and therefore no earthquakes above magnitude seven and a half are observed in this region in figure 3.9a.

When comparing figures 3.9a and 3.9b one has to be aware that the dynamic fault model is a stochastic model and thus subject to random scatter. A good example is the different activity of the Alpine Fault in figures 3.9a and 3.9b. Since the Alpine Fault is very straight, the angle parameter has only a small influence on the rupture distribution of this fault. Therefore, the difference in activity of the Alpine Fault in both figures is due to random scatter. It is for example possible that in figure 3.9a the Alpine Fault has ruptured more often in individual segments. This would decrease the return time of individual ruptures and would increase the occurrence rate of ruptures on the fault. In contrast, in figure 3.9b the Alpine Fault produced less but bigger earthquakes over the same simulation period. All spatial representations of simulated fault ruptures presented in this study are subject to random fluctuations.

Figure 3.9: Simulated Fault Activity for Different Parameters. *a) fault activity above magnitude 7.5 with simulation parameters 22.5° for the angle and 10% and 20 km for the gap. b) fault activity above magnitude 7.5 with simulation parameters 90° for the angle and 10% and 20 km for the gap. The colour scale is in events per year.*

Figure 3.10 shows the regional simulated fault activity for different magnitude ranges. As predicted in subsection 4.2, the relative activity in the eastern North Island (onshore) is decreasing with increasing magnitude. Over a simulation period of 10,000 years no earthquakes above magnitude eight have been produced in this region. The decrease in the relative frequency of large earthquakes in that region could possibly be explained with the back-arc opening

of the Taupo Volcanic Zone (e.g. Smith *et al.* 1989). The strike-slip faults change their direction away from the strike of the plate boundary (e.g. Beanland and Haines 1998). The change in direction would increase the angles between single fault segments and lead to a lower number of segment linkages, which would decrease the number of large earthquakes. Additionally, since the faults are not orientated parallel to the strike of the plate boundary anymore, any ruptures on those faults would be less effective in decreasing the accumulated strain caused by the relative movement of the Australian and Pacific plate. This strain would rather have to be released by oblique faults to the east.

The activity of the dip-slip faults (e.g. Hawke's Bay, Otago) above magnitude eight (figure 3.10b) seems to be too high, as observed beforehand (compare to figure 3.8 and subsection 4.4). A likely simplification of the reverse faults in Hawke's Bay (they have been mapped very straight) and the correction of rupture length, which produces very long and straight reverse faults (see discussion in subsection 4.4), are responsible for this high activity. It is much more likely that a significant number of these earthquakes should have had a lower magnitude. The activity of the Taupo Fault Zone seems to be too high as well (figure 3.10a) which can be explained by the simplification of the whole Fault Zone to two faults in the fault compilation.

Figure 3.10: Simulated Fault Activity for Different Magnitude Ranges. *a) fault activity above magnitude 7.0 with simulation parameters 45° for the angle and 10% and 20 km for the gap. b) fault activity above magnitude 8.0 with simulation parameters 45° for the angle and 10% and 20 km for the gap. The colour scale is in events per year.*

Figure 3.11: Fault Activity for Strike-slip Faults and Characteristic Faults. *a) fault activity for strike-slip and oblique faults above magnitude 7.5 with simulation parameters 45° for the angle and 10% and 20 km for the gap. b) fault activity above magnitude 7.5 based on characteristic earthquakes. The colour scale is in events per year.*

Figure 3.11a shows the simulated fault activity for strike-slip and oblique faults. The activity of the simulated strike-slip fault ruptures seems to be fairly homogeneous in the South Island. It decreases progressively to the north. As the magnitude frequencies above magnitude seven and a half are too low for strike-slip fault ruptures (figure 3.8) this is probably the region where more fault

segments should produce compound ruptures (see subsection 4.2). It is possible that the increasing complexity due to the change in direction of the strike-slip faults on the North Island not only lowers the occurrence of large earthquakes, but increases the chance of missing active faults because of their shorter lengths. Another possibility is that the missing earthquakes are generated by faults with a longer return time since the stress is not mainly released by long mature faults that consistently produce large earthquakes. There could very well be a number of shorter faults with long return times that alternately produce earthquakes.

It could be that the number of large earthquakes expected in the eastern North Island is not as low as suggested by the simulated activity. However, the existence of unknown faults (or faults with very long return times) would increase the modelled earthquake occurrence in that region. In contrast, the strike-slip faults on the South Island are simpler and thus likely to be reasonably well mapped. The simulated activity of strike-slip fault ruptures on the South Island probably matches the 'real' occurrence rate fairly well.

Figure 3.11b shows the (unsimulated) activity above magnitude seven and a half if the faults were treated characteristically. The activity of the characteristic model matches the simulated activity of strike-slip ruptures (figure 3.11a) reasonably well. The activity of the characteristic model seems to be slightly higher than the activity of the simulated strike-slip fault ruptures, but this could be caused by the fact that the magnitudes in the characteristic model are often calculated with the relationships of Wells and Coppersmith (1994), which leads to too high magnitudes for strike-slip faults. Since the occurrence of characteristic strike-slip fault ruptures in the eastern North Island is similar to the activity of the simulated strike-slip fault ruptures, it is suggested that the

characteristic model cannot solve the problem of the low activity rate of large strike-slip earthquakes in that region.

Another important observation is that the dynamic fault model can produce strike-slip earthquakes above magnitude seven and a half where the characteristic model would never produce any. Apart from the south-eastern North Island, where both models produce large earthquakes, the dynamic fault model produces large earthquakes in the central-eastern North Island whereas the characteristic model produces no large earthquakes in this region. However, the characteristic model produces large earthquakes in the north-eastern North Island. As the strike-slip faults are likely to shorten progressively to the North (see section 4.2 and page 100), the dynamic model seems to produce a more realistic, continuously decreasing earthquake occurrence in the eastern North Island. The characteristic model produces a gap of activity in the central west of the North Island and large earthquakes to the north and south of this gap, although there is no apparent reason why this should be so in reality.

The characteristic model barely produces any dip-slip earthquakes above magnitude seven and a half. Further, the only earthquakes above magnitude eight (see appendix two) are produced on the Alpine fault (New Zealand's longest fault) and the Wairarapa fault (the only historically observed earthquake above magnitude eight). It has to be concluded that the characteristic earthquake model does not seem to be able to produce enough dip-slip earthquakes above magnitude seven and a half. Although the dynamic model produces too many dip-slip earthquakes above magnitude eight, it demonstrates that earthquakes above magnitude eight could possibly happen in any region dominated by reverse faults.

As the previous plots show, the dynamical model can produce earthquakes in places where the characteristic model is not able to produce any earthquakes. This makes the earthquake occurrence model based on fault information look more homogeneous, i.e. in areas where the characteristic model only produces magnitude seven earthquakes the dynamic model can produce bigger earthquakes. It has to be considered though that the dynamic model is still dependent on the fault compilation. One example is the eastern North Island where faults seem to be missing (see subsection 4.2). Another region where faults are likely to be missing is east of Cook Strait where faults from the South Island are not connected to the North Island. Although those faults are not connected directly (Carter *et al.* 1988) it is still possible that these faults rupture together with offshore faults and produce bigger earthquakes.

4.6 Return Time Frequencies

The frequencies of return times of the simulated fault ruptures above magnitude seven and eight are shown in figure 3.12. The two frequencies have been scaled to span the same range of frequencies. Both frequencies of return times increase until the return time is about 10,000 years and then they drop off rapidly. Although the frequencies are approximately normally distributed the drop in the frequencies of the high return times is too fast to be consistent with a normal distribution. The fault ruptures above magnitude eight have fewer short return times than the ones above magnitude seven, which reflects a higher minimum return time of magnitude eight earthquakes compared to magnitude seven.

For the fault compilation a fault is considered to be inactive if it has a longer return time than 125,000 years (Mark Stirling, pers. comm.). Since the number of faults with return times over about 20,000 years is very small it is possible that a

lot of faults with a return time above 20,000 years have been classified as inactive (e.g. when the return time is not known), have not been mapped, or are wholly buried. The sudden drop of return times over about 20,000 years could reflect this. It is questionable if this distribution should have such a fast decrease of long return times. A slower decrease similar to a normal distribution would be more reasonable.

Figure 3.12: Return Time Frequencies. *Because the two frequencies sets have been rescaled to span the same frequency interval the frequency values have been omitted. The return times are in years.*

As will be discussed in chapter four, the earthquake activity in west Nelson could be under-predicted by the fault model. The seismicity in this region is generated by faults with a long return time. This strengthens the argument that there are probably quite a few faults with return times longer than 20,000 that need to be included. If the number of these faults should increase in a similar way to the increase in the number of faults with shorter return times, there would be a lot more faults with long return times than faults with short return times (it should be noted, that at some stage the increase of numbers of faults with higher return times has to slow down). This would mean that these faults would contribute a significant part to the present earthquake occurrence and should not be ignored.

Such faults could be located in the eastern North Island where they would increase the occurrence of large strike-slip faults (see subsection 4.2).

5 Accumulated Deformation Rates

Every earthquake occurrence model has to be consistent with the theoretical deformation rates inferred from the plate movement. Deformation rates can be measured by observing the deformation over time, for example with the help of GPS (e.g. Bourne *et al.* 1998, Pearson 1998). This method has the disadvantage that the observation time has to be very long (at least several hundred years) to actually match the average long term rate. Otherwise the measured strain rate could be underestimated in a region where no big earthquakes have been observed, but they are expected to happen (e.g. Alpine Fault). For this study the estimated deformation rates from the rotation about the Euler pole for the Australian and the Pacific plates have been used (e.g. DeMets *et al.* 1990, DeMets *et al.* 1994).

The strain release rate from the simulated fault ruptures can be calculated from the measured moment release (Kostrov 1974):

$$(3.8) \quad \varepsilon = (\sum M_0 / T) / 2 \mu A W_s ,$$

where μ is the rigidity, A the surface area over which the strain rate ε is calculated, W_s the seismogenic width, and T the time over which the seismic moment M_0 is added. To calculate the strain rate over the whole of New Zealand all seismic moments produced by the dynamic fault model have to be added. For this study, it would be of advantage to compare the moment release of the dynamic fault model with the deformation along the plate boundary instead of the deformation distributed over the whole area of New Zealand. The

deformation along the plate boundary can be given as a slip rate. The sum of the slip rates (deformation release) of the modelled fault ruptures can be calculated with equation 3.5, when the seismic moment is substituted by moment release (i.e. the added moments divided by the simulation time) in this equation.

The plate motion vector can be split into two components. The plate boundary parallel motion is taken up by strike-slip faults and the plate boundary normal motion is taken up by dip-slip faults. Since the plate motion vector has a direction of approximately 45° to the plate boundary both deformation components (strike-slip and dip-slip) are approximately equal. It follows that if the slip rates of the modelled fault ruptures of both components are added the sum has to be divided by a square root of two to calculate the absolute slip rate.

Figure 3.13: Simulated Slip Rates. *Slip rates [mm] of the simulated faults, the simulated strike-slip and oblique faults, and the characteristic faults are compared to the theoretical slip rates inferred from the relative rotation of the Pacific and the Australian plates.*

Figure 3.13 compares the added slip rates of simulated fault ruptures, simulated strike-slip fault ruptures, and characteristic faults to the slip rates inferred from the relative rotation of the Pacific and the Australian plates per latitude. Since the dynamic model produces too many large dip-slip earthquakes, the added slip rates are sometimes higher than the theoretical slip-rates, especially in the

Hawke's Bay region (latitude $\sim 39^\circ$). Between latitude 41° and 44° the slip rates of the simulated faults match the theoretical slip rates very well. It has been suggested that the Marlborough faults take up 90-95% of the plate motion (Bourne *et al.* 1998). This area is reasonably free of reverse faults that produce too many large earthquakes.

The slip rates of the characteristic model are very similar to the slip rates of the simulated strike-slip fault ruptures. North of latitude 42° , the slip rates decrease progressively towards north, which corresponds to the region in which the number of large strike-slip earthquakes is decreasing. Although there are possibly more large earthquakes than observed in this region, which would increase the slip rate, it is likely that most of the accumulated strain is released by oblique faults to the east (e.g. Anderson and Webb 1998). It would also be possible that the strain is released mainly on the subduction interface. However, this doesn't match the typically low estimated slip rates of the subduction interface (e.g. Stirling *et al.* 1998).

Smith *et al.* (1989) suggested that the seismic quiescence is related to the back-arc processes in the west and that the seismic deformation takes part principally offshore. The slip rate decrease in the very south of New Zealand reflects the fact that no faults have been mapped in Fiordland. The slip rates of the Alpine fault (43° - 44° latitude) match the theoretical slip rates reasonably well, if one considers that the Alpine fault has a slip rate of 60-80% of the plate displacement rate (Sutherland 1999).

The mismatch of the slip rates of the characteristic model with the theoretical slip rates strengthens the argument that reverse earthquakes of magnitude eight and above can be produced anywhere along the plate boundary. The Alpine fault

and the Wairarapa fault alone cannot release all the accumulated strain along the plate boundary. Although a main part of the strain could be released by the subduction zones, the 1855 Wairarapa earthquake demonstrated that this strain can also be released on crustal faults, at least partly. Thus, it is plausible that the reverse faults in Hawke's Bay and Otago also produce earthquakes of about magnitude eight to release a significant part of the accumulated strain.

6 Summary

Here, the most important results of this chapter are briefly summarised:

- The recorded fault parameters do not match the historically observed earthquake source relationships. This mismatch can be explained by the systematic underestimation of fault length and the systematic overestimation of displacement.
- The dynamic fault model produces more large earthquakes than the characteristic model. Whereas the characteristic model does not produce enough large earthquakes to match the historically observed number of large earthquakes, the dynamical model roughly fits the historically observed magnitude frequency relationship.
- The maximum simulated earthquake magnitude over 10,000 years is 8.4. The characteristic model produces maximum earthquakes of 8.1. The only faults that produce earthquakes above magnitude eight are the Alpine fault and the Wairarapa fault for the characteristic model. The dynamic model allows earthquakes above magnitude eight anywhere along the plate boundary.

- Whereas the characteristic model produces too few large reverse earthquakes, the dynamic fault model produces too many. This mismatch of both models can be explained by the very short surface breaks of the reverse earthquakes. The characteristic model generates too small reverse earthquakes, which are inferred from the short surface breaks. In the dynamical model the rupture length has been corrected. The correction produces very long straight, and unrealistic fault traces for the reverse faults that create only characteristically large earthquakes above magnitude eight. It is more likely that these long faults produce only few large earthquakes and many more earthquakes below magnitude eight.
- Both the characteristic model and the dynamic model do not produce enough strike-slip earthquakes above magnitude seven and a half and too many strike-slip earthquakes below magnitude seven in the eastern North Island. It is likely that there are unmapped or wholly buried faults in this region that could connect single faults, which in the model produce magnitude six and a half earthquakes, to produce magnitude seven and a half earthquakes. Another possibility would be the existence of many faults with long return times in that region, which have been classified as inactive. These faults could similarly link smaller mapped faults.
- Overall, the dynamic model is an improvement over the characteristic model. It produces more homogeneous earthquake occurrence models and allows for a greater variation of possible fault ruptures. Nevertheless the dynamic fault model is still strongly dependent on the fault

compilation and cannot overcome inherent problems of the same. Areas where faults are missing from the compilation still pose a problem for the generation of earthquake occurrence models.

The above conclusions have been derived from various intermediate results and assumptions. The possible influence of these assumptions in regards to the conclusions are discussed in the following:

- It is assumed that the fault compilation used in this study is an accurate representation for the actual fault distribution in New Zealand. As discussed in section two, this is not the case. If another fault compilation was to produce similar results to the one used in this study, there would be a few questions that would have to be answered. For example, it would have to be explained why the magnitude frequencies of the historical earthquake catalogue (from 150 years) and the fault compilation do not match. Also, the deviation of the b-value to more than one above magnitude six and a half, and the high rate of fault ruptures between magnitude six and a half and seven would have to be explained (figure 3.8). Further, it would have to be explained why about 20% of all faults have return times between 10,000 and 20,000 years, but only very few faults have return times above 20,000 years, i.e. why is there a sudden saturation in return times (figure 3.12). Lastly, apart from the latitude band between 42° and 43° the fault compilation would not release enough accumulated strain (figure 3.13). While north of latitude 42° this strain could be released aseismically on the subduction interface, there is no subduction interface between latitude 43° and 44°.

- The scaling relationships between rupture sources found in this study are the ‘best’ relationships in a sense that they are based on a wide range of magnitudes and large number of observed earthquakes. However, it may be possible that the data has an unknown bias and hence that the scaling relationships appropriate for New Zealand differ from the ones used here. Further, it is possible that the average strain drop in New Zealand differs from the worldwide average strain drop. Also, the length correction for dip-slip faults used in this study could be in error. The possibility that the strain drop in New Zealand is higher than the worldwide average would not have much effect on the conclusions above. The length corrections were based on the observed single event displacements and would lead to systematically shorter faults if a higher strain drop would be used. To infer magnitude from fault length, the same higher stress drop were to be used, and in effect the short high strain drop faults would produce similar magnitudes as the longer average strain drop faults. Hence, the magnitude frequency relationships will not change significantly. As for different length corrections and rupture source relationships, the change in the magnitude frequency plots would be approximately a shift of intercept. For the strike-slip ruptures a shift to higher magnitudes would mean that many more earthquakes about magnitude seven would be produced than observed historically. A shift to lower magnitudes would mean that many more earthquakes above magnitude seven would have been observed compared to the numbers predicted by the fault compilation (figure 3.8). The other issues with the magnitude frequencies would remain.

- The model used in this study to determine how faults link is only theoretical. However, it has been demonstrated that the choice of linkage model seems to have virtually no effect above magnitude seven (figure 3.6). Also, the magnitude frequency distribution for strike-slip faults is practically the same for the characteristic and dynamic case (figure 3.8). However, as discussed, the location of fault traces has a big influence on fault linkage. There are several problems (e.g. the high number of earthquakes between magnitude six and a half and magnitude seven), which can be resolved if there are a number of active faults not included in the fault compilation.

COMBINED OCCURRENCE MODEL

An earthquake occurrence model based on historical earthquake catalogues has to be consistent with a corresponding earthquake occurrence model based on fault data. According to the Gutenberg-Richter magnitude frequency relationship (e.g. Richter 1958, 359) the distribution of the frequencies of all magnitudes (in a large enough area) is exponential. Thus, using the Gutenberg-Richter distribution, the numbers of large earthquakes generated by faults (in a large area) have to match the numbers of large earthquakes inferred from the smaller sized earthquakes from the historical earthquake catalogue. However, there are alternative models for magnitude frequency distributions. Commonly used models for magnitude frequency distributions are discussed in the first section.

The seismological and geological earthquake occurrence models are compared in the second section. This section also includes a short discussion of the seismicity features of New Zealand. In section three regional b-values are discussed and in section four maximum magnitudes are estimated for different regions.

The figure A2.2 in appendix two may be helpful to identify the local regions of New Zealand, that are discussed in this chapter.

1 Magnitude Frequency Distributions

Last century an exponential relationship between earthquake magnitude M_W and earthquake frequency N was found, which seems to be valid for a wide range of earthquakes (e.g. Richter 1958, 359):

$$N = a + b M_w .$$

'b-values' have been subject of several studies and are considered to be one of the important parameters in hazard modelling. More recent studies show that the b-value is most likely a universal constant of one for shallow earthquakes (e.g. Kagan 1997b, Godano and Pingue 2000) over a large range of magnitudes (e.g. Abercrombie and Brune 1994) and any deviation reflects the use of insufficient data (Godano and Pingue 2000). The only variation in b-value seems to be a systematic decrease with increasing depth (Mori and Abercrombie 1997). One explanation for the universal b-value of one is that crustal strains might organise along relatively discrete zones in way that regional fault slip rates and regional numbers of faults are connected to regional fault lengths (Wesnousky 1999).

The Gutenberg-Richter law cannot be valid for infinitely high magnitudes since the energy release would rise to infinity (Knoppoff and Kagan 1977). Kagan (1991) suggested a magnitude frequency behaviour that corresponds to a gamma distribution:

$$(4.1) \quad \Phi(M) = C^{-1} M_t^\beta M^{-1-\beta} \exp(-M/M_{xg}) ,$$

where C is a normalising coefficient, M_{xg} the maximum moment parameter (as a 'soft' limit), M_t the lower moment threshold, and b equals 1.5β . This magnitude frequency distribution is still exponential for smaller magnitudes until, at M_{xg} , the frequencies begin to decrease faster for higher magnitudes.

The existence of characteristic earthquakes (e.g. Schwartz and Coppersmith 1984) implies a characteristic magnitude frequency distribution (e.g. Wesnousky 1994, Stirling *et al.* 1996). For this type of distribution the smaller earthquakes still behave exponentially until they reach a characteristic magnitude at which the

frequencies cease to decrease or even increase for greater magnitudes. For magnitudes greater than the characteristic magnitude, the distribution falls off very fast.

Areas that typically show characteristic fault behaviour are regions of long, smooth strike-slip faults (e.g. Stirling *et al.* 1996). Over geological times, strike-slip faults often tend to get longer and less complex, i.e. the ratio of small to large earthquakes decreases with accumulated slip (e.g. Wesnousky 1988, Wesnousky 1990, Stirling *et al.* 1996). This matches the observation that the aspect ratio of these faults can get very big (see figure A1.6). It is possible that for these faults the accumulated strain is preferentially released in large earthquakes (as there are not many or only weak ‘barriers’ to stop a rupture), leading to a characteristic magnitude frequency distribution.

In contrast, dip-slip faults do not produce very large aspect ratios. This might be an indication that they do not grow in length nor smoothen in the same way as strike-slip faults do (see the discussion section of appendix one on possible differences between strike-slip and dip-slip earthquakes). Because of the greater complexity of dip-slip faults, it is possible that regions dominated by dip-slip faults are more likely to produce exponential or gamma distribution-like magnitude frequency distributions.

In some regional studies, the characteristic model has predicted too many earthquakes in comparison to the observed earthquake occurrence (e.g. Working Group on California Earthquake Probabilistics 1995, Field *et al.* 1999) and is consequently not undisputed (e.g. Kagan 1993, Kagan 1996, Wesnousky 1996). On the other hand, some theoretical earthquake models seem to indicate characteristic earthquake behaviour (e.g. Hamilton and Closkey 1997). It is

therefore important to further test the validity of the characteristic earthquake model.

The width limitation of earthquake ruptures due to the seismogenic thickness has an influence on the magnitude-frequency distribution (e.g. Pacheco *et al.* 1992, Romanowicz and Rundle 1993, Sornette and Sornette 1994, Romanowicz and Rundle 1994, Okal and Romanowicz 1994, Scholz 1997, Scholz 1998). The number of earthquakes changes above the magnitude for which the rupture widths begin to be limited and hence the magnitude frequency distribution should have a visible kink. However, as shown in this study, only strike-slip earthquakes (and possibly regions dominated by steep dip-slip faults) should be affected by this limitation (see appendix one). Thus, the size of the effect on the magnitude frequency would depend on the dominant faulting style in the region of interest. Further, observed breaks in magnitude frequency distributions are generally not statistically significant and it remains unclear if they are real (Main 2000). Breaks in the magnitude frequency distribution found in other studies (e.g. Triep and Sykes 1997, Molchan *et al.* 1997) may just reflect a roll-off at a maximum magnitude as described by the gamma distribution (equation 4.1).

2 Seismological and Geological Model Compared

Before I compare the occurrence models based on the historical earthquake catalogue to occurrence models based on mapped, active faults, I will discuss the seismicity features revealed by the model based on the historical earthquake catalogue in more detail. Then I will compare these features with the seismicity features produced by the fault data.

2.1 Seismicity Features

Figure 4.1 shows earthquake occurrence models above 40 km depth inferred from the historical earthquake catalogue for different time intervals and completeness magnitudes. As discussed in chapter two, the main seismicity features seem to be stable over the whole observation time. The fluctuations in activity over time seem to be of only secondary order.

The seismicity features are progressively smeared out in earlier time intervals. This can be explained by the less accurate hypocentre determination in earlier years, as well as the fact that the number of earthquakes decreases exponentially for higher completeness magnitudes. For example, the number of earthquakes greater than magnitude 3.0 within a ten year time interval corresponds to the number of earthquakes bigger than magnitude 4.0 within a hundred year time interval (for a b-value of one). On the other hand, small fluctuations in the activity level may reflect expected spatial and temporal earthquake clustering which is only of secondary order. When comparing different decades in New Zealand, it can be seen that most seismicity features show more active and less active periods. No regions were observed that suddenly started to be active without having been active beforehand. Similarly, active regions may decrease in activity for a decade but will then increase in activity in the following decade.

Figure 4.1: Seismological Occurrence Model above 40 km depth for Different Time Intervals. *a) Seismological occurrence model above magnitude 4.5 between 1943 and 1997. b) Seismological occurrence model above magnitude 4.0 between 1962 and 1997. c) Seismological occurrence model above magnitude 3.0 between 1991 and 1997. The colour scale gives the probability of one or more events per year.*

Figure 4.2 shows the earthquake occurrence model inferred from the historical earthquake catalogue above 12 km depth. The only significant difference from the occurrence models spanning the whole crust (figure 4.1) is the absence of the seismicity band along the east coast of the North Island. As discussed in chapter two, section one, the opening of the Hawkes Bay network led to a systematically correction to a greater depth of the hypocentres in that region. Those earthquakes are located near the plate interface and are probably connected to processes in the subducted plate.

Figure 4.2: Seismological Occurrence Model above 12 km depth. *Seismological occurrence model above magnitude 3.5 between 1987 and 1997. The colour scale gives the probability of one or more events per year.*

Figure 4.3 shows the earthquake occurrence model above 40 km inferred from the historical earthquake catalogue from 1930 above magnitude 6.0. Earthquakes over magnitude 6.0 during this time interval have not been mapped completely, but should be reasonably complete in most regions, especially in central New Zealand. The seismicity features are still similar in later time intervals (figure 4.1) and it can be concluded that there is no evidence for any change in the main seismicity features since the systematic, instrumental observation of earthquakes started in the thirties (figure 4.3). Even the locations of the largest historical earthquakes in New Zealand (figure 4.4) mostly coincide with the areas that had above average activity in the nineties (figure 4.1c). Exceptions are the 1848 Marlborough ($M_L = 7.5$, Grapes *et al.* 1998) and the 1888 North Canterbury ($M_L = 7.3$, Cowan 1990) earthquakes which both occurred on long strike-slip faults. This seems to suggest that over the last 150 years the main seismicity features

were stable and that, apart from regions dominated by long strike-slip faults, the number of large earthquakes can probably be inferred from the observed number of smaller sized earthquakes. While it is not known how many small and medium sized earthquakes were located in Northwest Nelson prior to 1930. The area had a magnitude 7.5 earthquake in 1868, indicating that this area was active before 1929.

Figure 4.3: Seismological Occurrence Model since 1930. *Seismological occurrence model above magnitude 6.0 between 1930 and 1997 and above 40 km depth. The earthquakes for this time interval are not complete above magnitude 6.0. The colour scale gives the probability of one or more events per year.*

Although most of the main seismicity features observed after 1943 (figure 4.1a) can be seen in the occurrence model for magnitudes above 6.0 (figure 4.3), some of the features are missing (e.g. Taupo Volcanic Zone). This can be explained by the fact that the number of magnitude six and greater earthquakes after 1930 is so low that no earthquakes above this magnitude have been observed in less active regions (e.g. central Otago). For more active regions, like the Taupo

Volcanic Zone, this might be an indication that these regions do not produce earthquakes much larger than this, i.e. those regions might not produce many earthquakes above magnitude 7.0.

Figure 4.4: Large, Important, Historical Earthquakes in New Zealand. Taken from the GNS website: <http://www.gns.cri.nz/earthact/earthquakes/hist.html>.

Eastern North Island

One of the best expressed seismicity features in New Zealand is a long homogeneous band of seismic activity below the eastern North Island, which stops in the southwest under the Kaikoura Ranges (figure 4.1). This activity is mostly confined to a depth between 12 and 30 km (there is not much activity above 12 km, see figure 4.2; e.g. Reyners 1980) and it is related to the Hikurangi subduction zone. The sparser activity in the overlying Australian Plate is related

to the strike-slip faults in that region (e.g. Reyners 1980). Northeast, under the Raukumara Peninsula, the crust is thinning and the seismicity band shallows (see figure 4.1 and figure 4.2, Reyners and Mc Ginty 1999).

Above 12 km, the observed seismicity in this region consists of strike-slip earthquakes in the west and progressively changes its faulting style to reverse in the east (e.g. Reyners 1980). A band of strike-slip faults lies above this seismicity band at the southern end of the North Island and turns towards north, away from the seismicity band, in the northeast of the North Island. This might indicate that the strike-slip faults are not connected with the deeper seismicity band. Another indication for the separation of the strike-slip faults and the activity related to the subducted plate is the disruption of the strike-slip faults in Cook Strait (e.g. Robinson 1986, Carter *et al.* 1988, Eberhart-Phillips and Reyners 1997). Although the strike-slip faults from the southern North Island do not connect through the Cook Strait to the strike-slip faults in Marlborough, the deeper seismicity band is not disrupted (figure 4.1).

In the northeast of the North Island, the seismicity band should theoretically be dominated by normal faulting, which would be consistent with expected fore-arc spreading (Walcott 1987). To explain the dominant reverse faulting in this region, Walcott (1987) suggested ‘underplating’, i.e. the accumulation of sediments under the crust, which leads to the uplift of this area and thus reverse faulting.

Reyners (1989) suggested that part of this seismicity, at the location of the 1931 Hawkes Bay earthquake ($M_s = 7.8$, Hull 1990), is connected to regional relaxation after the stress release of this large earthquake. Figure 4.1 shows that the whole band consists of fairly homogeneous activity, which suggest that the whole

seismicity band has had constant activity along the whole plate interface during the whole observation time.

Central Volcanic Region

Further to the west another band of seismic activity is evident (figure 4.1) and can be connected to back arc spreading (e.g. Smith and Webb 1986, Anderson and Webb 1994). It coincides with the active volcanic front marked by Mt Ruapehu and White Island (Anderson and Webb 1994), with little activity to the east and west (Smith and Webb 1986). Most earthquakes are shallower than 10 km and this area is characterised by earthquake swarms (Smith and Webb 1986, Bryan *et al.* 1999). The region is dominated by normal faulting, but there might be strike-slip faults present with east-west orientation, that link the predominant normal faults (e.g. Smith and Webb 1986, Suggate 1986, Anderson and Webb 1994).

The band ends in the central North Island north of three seismicity clusters of which one lies under Lake Taupo and the other two to the west and east of Mt Ruapehu (figure 4.1). There has been no persistent activity observed under Mt Ruapehu itself (Olson 1985) and in the area of those clusters, there is little activity above 15 km (Reyners 1980). These three clusters may be connected to stress build-up related to the end of the backarc spreading. The western most of those clusters could also be stress induced by a sudden increase in crustal thickness towards south (Stern *et al.* 1987).

There is a weak seismicity band of medium activity northwest of the Taupo Volcanic Zone and to the southeast of Auckland. This activity is connected to the Hauraki Depression (Hochstein and Nixon 1979).

Cape Egmont Fault Zone

Whereas there is only little activity near Mt Taranaki (Cavill *et al.* 1997), to the southwest of the volcano there lies a short band of activity (figure 4.1, Anderson and Webb 1994). This band coincides with a seafloor scarp of steeply dipping normal and reverse faults (Nodder 1993). It matches the regional extension and it is possible that it is connected to the back-arc spreading in the Taupo Volcanic Zone, although the rate of extension is much lower (Nodder 1993, Anderson and Webb 1994). Alternatively, it could be part of the South Wanganui Basin, which is another back-arc basin related to the subduction process (Smith *et al.* 1989). Another explanation would be flexure of the crust caused by the sinking of the Wanganui Basin (Tim Stern, pers. comm.). It is unlikely that this region is connected to the reverse faults in North West Nelson (Webb and Anderson 1998), since the dominant faulting mechanisms are different. Nevertheless, earthquakes with a reverse component have been observed in the Cape Egmont Fault Zone (e.g. the 1974 Opunake earthquake, Robinson *et al.* 1976).

Wanganui

Another cluster of seismicity lies under the South Wanganui Basin (figure 4.1, Smith *et al.* 1989). The activity of this cluster seems to be tightly clustered in depth (Reyners 1980). This cluster consists of mostly small earthquakes and frequent, persistent swarms. The centre of the cluster is dominated by smaller, normal earthquakes, whereas the sides are dominated by larger, infrequent, reverse earthquakes (Garrick and Gibowicz 1983). The activity might be related to plutonic activity at depth or extensional faulting related to the subduction process (Garrick and Gibowicz 1983). An alternative explanation is the sinking of the Wanganui Basin, caused by loading of sediments (Tim Stern, pers. comm.)

Cook Strait

The Cook Strait seems to mark a disruptive line through the surface faults of Central New Zealand (e.g. Robinson 1986, Carter *et al.* 1988, Eberhart-Phillips and Reyners 1997). There is little activity above 12 km and the region west of the Marlborough Sounds has been completely aseismic over the last century (e.g. figure 4.1, Anderson and Webb 1994). It seems as if this spot might correspond to the aseismic corridor proposed by Hatherton (1970), but the aseismicity is limited to a smaller region. Additionally, this area seems to lack any signs of faulting (Les Singh, pers. comm.).

West Nelson

The activity band in west Nelson (figure 4.1) can be associated with reverse faulting (e.g. Suggate 1986, Anderson *et al.* 1993, Reyners *et al.* 1997). It has been interpreted as a cluster of aftershocks of the 1968 Inangahua earthquake (Anderson *et al.* 1994) and its high activity level has been considered to be an anomaly due to the relatively short observation period (Reyners 1989, Anderson *et al.* 1993). However, this seismicity band has been showing no decrease of activity during the whole observation time (figure 4.1 and figure 4.2). Reyners *et al.* (1997) conclude that the high activity level is consistent with the level of deformation and that this region is taking up a major part of the plate deformation. Thus, it is possible that the high activity level is not an anomaly due to the short observation time, but reflects activity which will persist for longer time intervals.

Marlborough

East of west Nelson earthquakes change their faulting style progressively to strike-slip (Anderson *et al.* 1993) and change back to reverse faulting towards the Kaikoura Ranges (Bibby 1981, Anderson *et al.* 1993). The central region of

Marlborough is dominated by larger, infrequent, strike-slip earthquakes (Anderson *et al.* 1993). The high uplift rate of the Kaikoura Ranges has been suggested to be connected to resistance in the subduction process (Smith *et al.* 1989).

Under Lake Tennyson there is a cluster of activity (figure 4.1), which extends to the southeast. Reyners *et al.* (1997) interpreted this cluster to be connected to the 1990 Lake Tennyson earthquake ($M_L = 5.8$, McGinty *et al.* 1997). However, this cluster has been persistent throughout the whole observation period (figure 4.1 and figure 4.2). It is located to the southeast of the connection of the Wairau fault and the Alpine fault. This connection shows a bend that is oriented differently from the connection of the three other long Marlborough strike-slip faults (Awatere fault, Clarence fault, and Hope fault) with the Alpine Fault in the south. I suggest that this cluster reflects local stresses related to the northernmost connection between the Marlborough Fault System and the Alpine fault.

This cluster extends towards the east and ends north of Christchurch (figure 4.1). The activity north of Christchurch is confined to a depth of greater than 15 km and might reflect a tear in the lower crust, where subduction ends (Reyners and Cowan 1993).

Central Canterbury

In the central South Island another short seismicity band is evident, which starts at about Arthur's pass and ends under Mt. Cook (figure 4.1). This region is dominated by young strike-slip faults that are probably associated with the nascent southern boundary of the Marlborough Fault System (Scholz *et al.* 1973, Anderson and Webb 1994) and probably reflect deformation stresses south of

the Hope fault due the rotation of the Marlborough strike-slip faults (Walcott 1998).

Alpine Fault

The Alpine Fault is the longest expressed surface fault in New Zealand (e.g. Wellman 1972, Wellman 1979). During the period of historical seismicity observation the central section of this fault was relatively aseismic (figure 4.1, e.g. Evison 1971, Anderson and Webb 1994). The Alpine Fault takes up most of the plate motion (e.g. Sutherland 1999), whereas the rest of the deformation is taken up by a broad zone to the east (Norris *et al.* 1990). It is not clear if the Alpine fault produces earthquakes corresponding to the whole length of the fault rupturing or if it rather breaks in small segments (e.g. Yetton 1998, Walcott 1998). A break along the total length of the fault would produce a very high aspect ratio, which has not been observed anywhere in the world (see figure A.6. Assuming a rupture length of about 300 km and a rupture width of about 8 km the aspect ratio would be about 40, 25% bigger than the largest observed aspect ratio). Since the Alpine fault is dominantly a strike-slip fault (e.g. Scholz *et al.* 1973) such rupture would not be limited significantly by rupture width and it cannot be excluded that such a rupture is possible.

Fiordland

Another activity band is located in Fiordland (figure 4.1) where it correlates to the Fiordland subduction zone (Reyners 1989, Anderson and Webb 1994). The faulting style in this region is dominated by reverse faulting and is oblique reverse further south and further inland (Scholz *et al.* 1973, Anderson *et al.* 1993). There is only a little activity to east in central Otago (Reyners 1989), which is characterised by reverse faulting (Suggate 1986).

2.2 Surface Faulting

Earthquakes observed during the last 150 years could generally not be connected to major Quaternary fault traces (e.g. Reyners 1989), apart from a few of the largest earthquakes which did produce surface faulting. Some of these large earthquakes occurred on faults that were not previously known. Nevertheless, it has been observed worldwide that large earthquakes occur in regions which have high numbers of small earthquakes (e.g. Kafka and Levin 2000). In this section the earthquake occurrence model based on the mapped, active faults is compared to the occurrence model based on the historical earthquake catalogue to see if the seismicity features discussed in the last section match the seismicity features of simulated ruptures produced by the mapped, active faults.

Figure 4.5 shows different earthquake occurrence models based on different fault models as discussed in chapter three. For comparison with the earthquake occurrence model based on the historical earthquake catalogue, earthquake occurrence models have been calculated based on the simulated fault activity and adaptive kernel estimation with a Gaussian kernel and a bandwidth of 1.0.

Figure 4.5a shows that the dynamic fault model produces too many earthquakes above magnitude seven and a half offshore from the eastern North Island and in central Otago when compared to the historical earthquakes (figure 4.1). As these regions are dominated by reverse faults this matches the observation, as discussed in chapter three, section 4.4, that the dynamic model produces too many large reverse earthquakes, which is due to the length correction. The simulated fault activity in the eastern North Island offshore parallels the deeper subduction plate activity slightly further west (figure 4.1), which could be an indication that those offshore reverse faults are connected to the subduction

zone interface, as suggested in other studies (e.g. Cole and Lewis 1981). The seismicity band of the simulated faults is interrupted at several places which reflects the fact that these faults are not mapped completely; a through going band similar to the observed deeper seismicity would be more realistic. The Taupo Volcanic Zone shows too much activity, which can be explained by the simplification of this faulting zone to two faults in the fault compilation.

Figure 4.5: Earthquake Occurrence based on Simulated Fault Ruptures. a) *Earthquake occurrence based on the dynamic fault model.* b) *Earthquake occurrence of strike-slip and oblique faults based on the dynamic fault model.* c) *Earthquake occurrence based on the characteristic fault model.* The colour scale gives the probability of one or more events per year above magnitude 7.5.

Figure 4.5b shows the earthquake occurrence model based on the dynamic model using strike-slip faults (including oblique faults). In the South Island, the

activity above seven and a half looks to be fairly homogeneous where there are long strike-slip faults. The simulated activity of the Alpine fault matches the historical activity to the north and south (figure 4.1) very well. Towards the north the activity is decreasing as observed in chapter three. The simulated activity in the southern North Island looks a little odd, but the same area looks more homogeneous in figure 4.5a. Since in this study the fault ruptures are modelled assuming Poisson behaviour, in figure 4.5b the model in this region probably reflects random fluctuations that can occur over a time interval of 10,000 years (see chapter three). The mean behaviour is probably smoother, close to how the same region looks like in figure 4.5a. The scatter caused by the randomness of the simulation could be reduced if fault ruptures would be simulated over 100,000 years rather than 10,000 years, but has not been done for this study as each simulation had a computing time of approximately 12 hours (for 10,000 years).

Figure 4.5c shows the earthquake occurrence model based on the characteristic fault model. In the South Island the activity is not as homogeneous as the dynamic model and the regional activity seems to be dominated by the individual faults. In the North Island the characteristic model seems to produce results that look more realistic compared to the dynamic model. One has to bear in mind though that the magnitudes of the strike-slip rupture events are systematically overestimated due to the use of the Wells and Coppersmith (1994) regression. If the southern North Island in figure 4.5c is compared to figure 4.5a rather than 4.5b, the differences between the two models in this region are not that big. On the other hand, the characteristic model does not seem to produce enough reverse earthquakes above magnitude 7.5 compared to the historical occurrence model (figure 4.1).

One region where both the dynamic and the characteristic models do not seem to produce enough earthquakes is west Nelson. Since the faults in this region have a very long return times it has been suggested that the activity observed during the last century is an anomaly (Reyners 1989, Anderson *et al.* 1993). However, since the activity of both small and large earthquakes is consistently high and since this activity is consistent with the level of deformation (Reyners *et al.* 1997), the historical activity could very well reflect the activity over a longer time span. The high level of activity could be produced by many faults with long return times that are currently classified as inactive or are wholly buried. This would explain why the simulated activity of the mapped, active faults is not producing enough earthquakes to match the historical record.

The shear belt of the eastern North Island does not have any apparent background activity, if one considers the seismicity band of the historical earthquakes to be connected to processes in the subducted plate. Apart from local seismicity clusters, which are presumed to be connected to the local stresses within the Marlborough fault system and the seismicity related to the uplift of the Kaikoura Ranges, the long strike-slip faults in the Marlborough region show no background activity and the Alpine fault does not either.

The lack of background seismicity in regions of long, smooth strike-slip faults is in agreement with the finding of Stirling *et al.* (1996) that such regions produce a characteristic magnitude frequency distribution. As suggested in section one these areas produce mainly large earthquakes, but very few small earthquakes. Characteristic fault behaviour has been suggested for these areas by Berryman and Beanland (1991). Berryman and Beanland (1991) show further that these regions do not behave characteristically in a strict way (i.e. only large earthquakes

occur in these regions), but that the observed fault displacements are more consistent with recurring constant large earthquakes and frequent moderate earthquakes. Such behaviour is modelled by the dynamic model, although both the dynamic and the characteristic fault model seem to produce too many moderate earthquakes at the cost of large earthquakes in the eastern North Island, as discussed in the chapter three, subsection 4.2.

The results of this study thus seems to indicate that the occurrence of large earthquakes in regions dominated by dip-slip or short strike-slip faults can be inferred from the Gutenberg-Richter law, up to a regional maximum magnitude. Such behaviour would produce magnitude frequency distributions similar to the gamma distribution (equation 4.1). The normal faults of the Taupo Volcanic Zone and the reverse faults offshore from the eastern North Island have been observed to follow such behaviour (Berryman and Beanland 1991). Cowen *et al.* (1996) have examined the Porters Pass Faulting Zone and have found Gutenberg-Richter behaviour for this region dominated by young, short strike-slip faults, when paleoseismic data is compared to historical earthquakes. Cowen *et al.* (1996) observed that the Gutenberg-Richter law predicts too many large earthquakes, which probably reflects a maximum magnitude between seven and seven and a half for this region and is in agreement with the gamma distribution (equation 4.1).

The only dip-slip regions where paleoseismic data suggest characteristic rather than Gutenberg-Richter behaviour are west Nelson and central Otago (Berryman and Beanland 1991). Berryman and Beanland (1991) suggested intermittently characteristic behaviour, i.e. faults rupture characteristically successively for certain time intervals, but there are periods of no activity in

between. The paleoseismic data in west Nelson is inferred from faults with very long return times. It is possible, that there are more faults with long return times that have not been taken into account by the study of Berryman and Beanland (1991). The addition of such faults possibly generates activity in times where the other faults are inactive.

As discussed earlier, it is possible that the historically observed activity in west Nelson is representative of longer time intervals. During the whole observation time the rate of small earthquakes matches the rate of large earthquakes and is consistent with the Gutenberg-Richter distribution. As there is not much fault data available for west Nelson, it is more likely that the historical earthquake data is representative of longer time periods, and inconsistencies with the fault data do not reflect an anomaly of the historical earthquake occurrence but rather the incompleteness of fault data. A similar argument may hold for central Otago.

Another indication for Gutenberg-Richter behaviour of regions dominated by dip-slip faults is the magnitude frequency distribution of simulated ruptures using the dynamic fault model (e.g. figure 3.8). The dynamic fault model produces too many reverse earthquakes around magnitude eight but produces not very many earthquakes below that magnitude, which corresponds to a characteristic magnitude frequency distribution. As discussed in chapter three, subsection 4.4, reverse faults are corrected in length to match the displacement data, which makes them too straight and too long, which results in characteristic earthquakes. If a smaller correction were used (for example to compensate for overcorrection of possibly too big displacements) characteristic reverse earthquakes of, for example, about seven and a half would be produced by the dynamic fault model. Reverse earthquakes of magnitude eight and above would

not occur anymore, although such large events are possible in New Zealand, given appropriate dip-angle and crustal thickness. Further, characteristic behaviour of reverse earthquakes at about magnitude seven and a half would raise the occurrence of all earthquakes of this size above the number observed over historical times. Thus, it is more likely that the number of reverse earthquakes greater than magnitude eight has to be scaled down in a Gutenberg-Richter-like fashion rather than just shifting all of these earthquakes to a lower magnitude.

Kagan and Jackson (1999) found that the recurrence times of doublets of large earthquakes (above magnitude 7.0) show power-law behaviour and conclude that large earthquakes cluster in time and space, similar to smaller sized earthquakes. Kafka and Levin (2000) show that areas of frequent smaller earthquakes do very consistently delineate areas for potential future large earthquakes. They conclude that misses are likely to be due to incompleteness and quirks in the earthquake catalogues. Combined with the results of this study, that no spatial clusters of persistent high activity in New Zealand decreased to background activity during the observation time, such clusters appear to be potential sources of future moderate to large earthquakes for at least a few decades, if not centuries.

In summary, there seems to be no observation that contradicts Gutenberg-Richter behaviour in regions of dip-slip faulting and in regions of short strike-slip faults which cannot be explained by possible shortcomings of the fault data. Regions dominated by long, smooth strike-slip faults show only little background activity and most likely produce preferentially large earthquakes. The rate of these large earthquakes cannot be inferred from the background activity. However, the dynamic model seems to generate a realistic model for the

occurrence of such earthquakes in the South Island. In the North Island however, the dynamic model seems to produce too many medium sized earthquakes at the cost of large earthquakes. Thus, the occurrence of large earthquakes has to be expected to be higher than predicted by this model.

3 Subduction Zones

In this section the two subduction zones of New Zealand are discussed in more detail. As it is unclear if the occurrence of large earthquakes can be predicted from the background activity, earlier studies are summarized. It is more likely that the observed background activity reflects stresses connected to processes in the subducted plates rather than reflecting the subduction processes itself. The observed activity followed a Gutenberg-Richter distribution for both subduction zones for the observation time (figure 4.6).

Figure 4.6: Magnitude Frequency Distributions of the Subduction Zones. *The numbers of earthquakes have been added for the individual completeness intervals between 1948 and 1995 and are given in events per year*

Figure 4.7 shows the earthquake occurrence model based on the historical earthquake catalogue as a depth profile of New Zealand (see also figure 2.12). For the construction of the model the adaptive kernel estimation with a Gaussian kernel and a bandwidth of 1.0 has been used. To the northeast the

Hikurangi subducted plate can be seen and to the southwest the Fiordland subducted plate can be seen.

Figure 4.7: Earthquake Occurrence in the Subduction Zones. *Earthquake occurrence model for earthquakes to a depth of 350 km above magnitude 3.5 after 1963. The distance units are in km and the colour scale gives the probability of one or more events per year.*

Hikurangi Subduction Zone

The earthquakes in this subduction zone occur in a dipping tabular volume (Hamilton and Gale 1968), which shape is consistent with that of a gently tapering cone (Ansell and Bannister 1996). The activity shallows towards southwest and ends under the northern South Island with a sharp termination of seismicity (figure 4.7, Reyners 1989, Anderson and Webb 1994). The earthquakes within this zone appear to be clustered in space (figure 4.7, Reyners 1989) of which some may correlate to volcanism (Anderson and Webb 1994). An interesting observation is that the clusters seem to orient in the same diagonal direction as the lower limit of seismicity (figure 4.7). There seem to be not many temporal sequences (e.g. aftershocks) deeper than 40 km (e.g.

Robinson 1986; chapter two). The subducted plate shows several interruptions and seems to be segmented (Reyners 1983). One of the disruptions lies under Wellington (Robinson 1986), although this disruption could also be interpreted as a sudden increase in thickness of the plate towards southwest (Eberhart-Phillips and Reyners 1997).

The faulting style is dominated by normal faulting with some earthquakes being oblique (e.g. Webb and Anderson 1998). The area is characterised by slip partitioning along the whole length of the margin (Webb and Anderson 1998). The plate interface has little transcurrent motion, which is mostly accommodated in the overlying Australian Plate (Webb and Anderson 1998). The downdip tension indicates that the plate sinks under its own weight (Harris 1982, Reyners *et al.* 1997).

It has to be expected that large earthquakes can occur along the whole Hikurangi Subduction Zone (Webb and Anderson 1998). The coupling of the plates seems to decrease towards the northeast (Evison and Weber 1986, Reyners 1998, Reyners and Mc Ginty 1999). It is possible that the plate interface is permanently locked under Marlborough. As the coupling decreases towards north, large earthquakes have to be expected under Wellington. The maximum magnitude is expected to decrease towards northeast (Reyners *et al.* 1997, Reyners 1998).

Fiordland Subduction Zone

The Fiordland subduction zone is defined by a thin, planar zone of earthquakes, that is nearly vertical due to the much thicker lithosphere of the Pacific Plate to the east. It is dominated by thrust faulting possibly with a small strike-slip component (Davey and Smith 1983, Smith and Davey 1984).

4 b-values

A common approach when developing earthquake occurrence models is the division of the region of interest into several zones that often are geologically different. For each zone the level of activity and a b-value are calculated. The advantage of this approach is that only two values for each region are needed to infer an occurrence rate for a given magnitude at a given location. As this study moves away from regional zonation, new ways of b-value calculation have to be found. There are a few studies that suggest methods for the calculation of regional b-values based on non-parametric regression (e.g. Wiemer and Wyss 1997, Papazachos 1999). Since the b-value is effectively the reciprocal of the mean magnitude (e.g. Aki 1965), kernel estimation models of the mean magnitude can be readily converted into models of the b-value. The mean magnitude can be calculated using kernel estimation as follows (Vere-Jones 1992):

$$(4.2) \quad \overline{M}(\mathbf{x}) = \frac{\sum_{i=1}^N M_i K\left(\frac{1}{c}(\mathbf{x} - \mathbf{x}_i)\right)}{\sum_{i=1}^N K\left(\frac{1}{c}(\mathbf{x} - \mathbf{x}_i)\right)} ,$$

where M is the magnitude, K is the chosen kernel function, c is the bandwidth, N is the number of earthquakes, $\mathbf{x}_i$ is the location of each earthquake and $\mathbf{x}$ is the location in the spatial seismicity representation (see chapter two). A regional map of b-values has been produced for this study, using adaptive kernel estimation and temporal sequence filtering, but these results have not been included, as there are several problems with the interpretation of a local b-value. The results showed that the b-value is less than one in west Nelson and central Canterbury and greater than one in the Taupo Volcanic Zone and southwest of the Taupo Volcanic Zone. As discussed in the following, it is not clear how these results have to be interpreted.

Firstly, it is not clear what local b -values represent, i.e. what happens if the area for which a b -value is calculated gets very small. There is no doubt that the Gutenberg-Richter relation is valid for bigger areas, but it is probably not valid, if the area gets so small that it is dominated by single faults. In this case the magnitude frequency distribution would be dominated by the rupture behaviour of the individual faults. For example, in a very small region which encloses a fault that typically produces a magnitude five earthquake, the magnitude frequency distribution of earthquakes smaller than magnitude two might still correspond to the Gutenberg-Richter distribution. For larger earthquakes, the magnitude frequency distribution would become characteristic. Although dip-slip faults are, in general, more complex than strike-slip faults and thus can produce earthquakes of different sizes, it is unlikely that a large dip-slip fault will produce very small earthquakes, as its complexity is due to the different fault segments that form the fault. Although self similarity of faults implies that small segments consist of microsegments, it is likely that at a certain size even dip slip fault segments show a degree of smoothing (and thus a decrease of complexity), similar to strike-slip faults, but on a smaller scale. This means that for very small regions, even dip-slip faults will probably show characteristic earthquake behaviour.

For small time intervals the magnitude frequency relation might reflect local stress build up on the faults, or stresses related to a rupture that happened not long ago. For example, the b -value changes for short time intervals after large earthquakes (e.g. Gibowicz 1973a) and during earthquake swarms (Gibowicz 1973b). Earthquakes are clustered in time and space (e.g. Kagan and Jackson 1991), and a cluster of large earthquakes would change the observed b -value to be less than one. In the periods between the clustered occurrences of large

earthquakes the b-value would be above one (as observed in New Zealand after the 1968 Inangahua earthquake, Robinson 1979, Imoto 1987).

Godano and Pingue (2000) have shown that the crustal b-value has a universal value of one and any deviation reflects the fact that too few earthquakes have been used for the calculation of the b-value (even for mid-oceanridges). This would mean the deviations from the b-value do not, in general, reflect physical earthquake processes (apart from short time variations and limitations in area size, as discussed). For example, if the estimated maximum magnitude for a region is too high, using a simple Gutenberg-Richter distribution for the calculation of the b-value would result in a b-value that is too high. If a gamma distribution (equation 4.1) with a correct maximum magnitude were used the b-value will be lower than the b-value inferred from the Gutenberg-Richter distribution. This could be the case for the Taupo Volcanic Zone, which has an estimated high b-value (e.g. Smith and Berryman 1986) and a low maximum magnitude.

A further weakness of an occurrence model that just uses total activity and b-value to infer occurrences of large earthquakes is the fact that the local magnitude frequency distribution has to be Gutenberg-Richter like to be a valid model. However, the regions dominated by long strike-slip faults behave characteristically and the use of background seismicity to infer the occurrence of large earthquakes does not work for these regions.

In conclusion, the use of a b-value of one is probably the best choice as any deviation from one probably indicates that the observation period has not been long enough. Additionally, a gamma distribution with an appropriate maximum magnitude should be used. Although it is possible that local individual b-values

might deviate from one over long time intervals, it still has to be proven that there are significant deviations. However, the use of a b-value of one to infer the occurrence of large earthquakes only works in regions of dip-slip faulting and regions dominated by short strike-slip faults. For those regions it is recommended to use the historical earthquake catalogue and adaptive kernel estimation with temporal sequence filtering (e.g. figure 4.1) to infer the occurrence of earthquakes. The use of a long time period for the construction of an occurrence model (e.g. figure 4.1a) has the advantage of being most stable in time, because variations that reflect fluctuations in seismicity within individual decades are averaged. For later times periods (e.g. figure 4.1c) the hypocentre determination will be best and features of the earthquake occurrence model will be more accurate.

For the occurrence of large strike-slip earthquakes in the South Island, the dynamic fault model should be used (e.g. figure 4.4). For the North Island, the dynamic model does not produce enough large strike-slip earthquakes and too many medium sized earthquakes, as discussed in chapter three. More studies have to be done on how to best compensate for the missing strike-slip faults.

5 Maximum Magnitude

Another important parameter for earthquake occurrence models is the regional maximum magnitude. One way to estimate a maximum magnitude is the use of the past distribution of magnitudes (e.g. Pisarenko *et al.* 1996, Kijko and Graham 1998). However, the maximum magnitude for a certain region could perhaps not always be inferred from the past, as multiple faults, which usually produce a certain maximum magnitude individually, might produce a compound rupture that is much larger (Jackson 1996). However, it is not undisputed that such large

rupture can be produced on faults which are expected to produce smaller sized earthquakes (Ward 1997a, Stirling and Wesnousky 1997).

Similar to the definition of regions for which seismic activity parameters are determined, representative regions for the maximum magnitude have to be selected. This could be done with kernel estimation, but there are a few technical problems to be solved first. Firstly, when using kernel estimation to calculate a regional activity rate, the activity rate is smeared out over space. This lowers the activity in cells of observed peak activity and raises the activity in cells of lower activity. If this is done correctly (as in the case of adaptive kernel estimation) the regional activity over several cells does not change as the integrated activity stays the same for this region. However, the maximum magnitude for single cells should not be lowered and redistributed over neighbouring cells. In other words, the integrated maximum magnitude over several cells is expected to increase, since the historical time interval is not long enough for the occurrence of an earthquake of maximum magnitude in each individual cell. Thus, conventional kernel estimation cannot be used to produce regional maximum magnitudes.

Another problem is the determination of the area over which a single observed maximum magnitude should be representative. The observed activity level could be used to solve this problem. For example, if the activity level of a region is very high and limited in space, then a low observed maximum magnitude will indicate that very large earthquakes are not very likely to occur in such region. Similarly, for a region of very low activity the uncertainty of the maximum magnitude will be high and the chance of the occurrence of large earthquakes cannot be excluded. However, it is obvious that the observed maximum magnitude itself cannot be used to infer the extent of the region which has the same maximum

magnitude, i.e. a region of earthquakes with a maximum magnitude of seven can be bigger than a region of earthquakes with a maximum magnitude of eight.

Despite the technical problem on how to develop a model for the regional occurrence of maximum magnitudes based on non-parametric regression, the interpretation of a local maximum magnitude, similar to the interpretation of a regional b-value, has to be clarified. For a local hazard study, it makes sense to study the local faults to get an estimate of the possible maximum earthquake size. However, there still would be the possibility of unmapped, larger faults in the same region or the chance that some faults might link to produce a larger compound earthquake. To estimate regional maximum earthquakes, I would recommend producing a kernel estimation model using the historical earthquake catalogue (figure 4.1). This model should be compared to the earthquake occurrence model based on mapped, active faults (figure 4.5). In the regions dominated by characteristic faults the maximum magnitude can be estimated from the model based on mapped, active faults. With the use of characteristic fault lengths and source scaling relationships (section 3.1) maximum magnitudes can be estimated. With further assumptions on dip angle and crustal thickness, the maximum magnitudes in uncharacteristic regions can be estimated.

In the Taupo Volcanic Zone there has been no earthquake observed historically above magnitude six in the southern part and magnitude seven in the northern coastal and offshore part. The occurrence of earthquakes bigger than magnitude 7.5 in the dynamic fault model (figure 4.5) is based on simplified faults and is not realistic. Since the crustal thickness limits normal faulting (appendix one), this region would produce maximum magnitudes of 7.2, assuming a dip angle of 30° and a crustal thickness of 15 km. In the Cape Egmont Fault Zone the fault

model produces earthquakes bigger than magnitude 7.5, but this could be due to the Cape Egmont fault which is mapped very straight. The true maximum magnitude for this faulting zone could be lower, but the large 1868 Cape Farewell earthquake has been located near this zone (see figure 4.4). It is possible that this earthquake belongs to the faulting zone in North West Nelson, which might not be connected to the Cape Egmont Faulting Zone (Webb and Anderson 1998). The Wanganui seismicity cluster probably behaves similarly to the Taupo Volcanic Zone and the Cape Egmont Faulting Zone, but two above magnitude seven earthquakes (Gaye Downes, pers. comm.) have been reported in this region in the last two centuries. Anyhow, one of these earthquakes was probably located deeper than previously assumed and was connected to the subduction zone and there is evidence that the other one could have occurred in Hawke's Bay (Gaye Downes, pers. comm.). In conclusion, there is no hard evidence that these three dip-slip zones produce earthquakes much bigger than about magnitude 7.2.

In the eastern North Island the maximum magnitude connected to the strike-slip faults probably decreases towards north as the complexity of the faults is rising, but is probably still about magnitude 7.5 (a little bit less in the north and a little bit more in the south, see figure 4.4). The reverse faults probably produce earthquakes bigger than magnitude 8.0, depending on the dip angle, which is decreasing towards east (see discussion in chapter three). However, in between the pure strike-slip faults towards the west and the pure reverse faults offshore to the east, earthquakes are produced on oblique faults. These faults possibly produce the largest earthquakes in this region as they are not firmly limited by the seismogenic thickness, i.e. they can increase in length and produce greater aspect ratios compared to pure dip-slip earthquakes. Additionally, they will

produce bigger earthquakes than pure strike-slip faults since they can have bigger rupture widths than strike-slip faults (see figure A1.4). These faults could possibly produce earthquakes as big as magnitude 8.5.

Such events would not be very frequent, but are consistent with the historical data, as the three largest earthquakes in New Zealand occurred on faults with a reverse component (1855 Wairarapa, magnitude 8.1, Darby and Beanland 1992; 1931 Hawke's Bay, magnitude 7.8, Hull 1990; and 1929 Buller, magnitude 7.8, Dowrick and Smith 1990). Of these three events, the only one with nearly pure reverse faulting was the 1929 Buller earthquake (Anderson *et al.* 1993, Doser *et al.* 1999). These three events, together with the 1934 Pahiatua, magnitude 7.6, event (Dowrick and Smith 1990), were the only earthquakes above magnitude 7.5 in the last 150 years. The 1855 Wairarapa, the 1931 Hawke's Bay, and the 1934 Pahiatua earthquakes are all classified as being dominantly strike-slip, with only a small (1855 Wairarapa and 1931 Hawkes Bay) or virtually no (1934 Pahiatua) reverse component (Terry Webb, pers. comm.). However, the surface break for all three events was too short to be consistent with the source scaling relationships found in this study (chapter three, section one).

The source model for the 1855 Wairarapa earthquake is a steeply dipping fault on the surface, but less steeply to the Northwest towards the subduction interface (e.g. Darby and Beanland 1992). This increases the rupture width of the fault significantly, in a way that the observed magnitude is compatible with the rupture area. However, faults with dominant strike-slip faulting and rupture widths above forty kilometres have not been observed worldwide (see figure A1.4). Faults that have a large strike-slip component commonly occur on nearly vertical faults and thus have only limited rupture widths. It is possible that the

1855 Wairarapa event had a significant reverse component at greater depths. However, the source models of the 1931 Hawke's Bay and 1934 Pahiatua earthquakes suggest steeply dipping faults and only a small reverse component for the 1931 Hawke's Bay earthquake (Terry Webb, pers. comm.). If those earthquakes are consistent with the average source scaling relationships they must have had a rupture length much longer than observed on the surface. However, the rupture length of the 1934 Pahiatua event has been estimated to be about 65 km, based on observed aftershocks (e.g. Downes *et al.* 1999). The scaling laws found in this study would predict a rupture length of about 170 km. This means that the stress drop of this earthquake has to be about four times as much as the average stress drop for large strike-slip earthquakes. It is possible that the stress drops of the 1855 Wairarapa and the 1931 Hawke's Bay events were also very high, which would explain the small dip-slip components and the short surface breaks. The relatively high return times (about 2000 years) of the faults on which the 1855 Wairarapa and the 1931 Hawke's Bay occurred are consistent with above average stress drop (Kanamori and Allen 1986).

The dynamic fault model produces earthquakes above magnitude 7.5 on the strike-slip faults in the Marlborough region. A possible 200 km fault rupture corresponds to a magnitude 7.7 earthquake. For comparison, the biggest observed earthquake in this region was the 1948 Marlborough, magnitude 7.5, event (Grapes *et al.* 1998). A rupture of 400 km on the Alpine fault would produce an earthquake of magnitude 8.1. However, such a rupture would be exceptional worldwide as the rupture width would be very small (Walcott 1998), i.e. the aspect ratio would be about 40% bigger than the observed maximum world wide (see figure A.4). If the lengths of the seismicity features from historical earthquakes (figure 4.1) are representative of a maximum rupture on

the strike-slip fault regions near Lake Tennyson and in Canterbury, those regions would produce an estimated maximum rupture between 100 and 150 km which corresponds to a maximum magnitude of 7.3 to 7.5. For comparison, the 1929 Arthur's Pass earthquake had an estimated surface wave magnitude of 7.1 (Dowrick and Smith 1990).

In west Nelson and central Otago, the maximum magnitude for a reverse earthquake would be 8.1 given a 15° dipping fault and 20 km seismogenic thickness. The whole length of the seismic activity band in west Nelson is about 200 km (figure 4.1) which would even be consistent with a 8.5 magnitude earthquake. The dip-angle for such an event would have to be very shallow though, which might not be realistic for this region. This region has produced three large earthquakes above magnitude seven in the last 150 years of which the 1929 Buller earthquake had an estimated surface wave magnitude of 7.8 (Dowrick and Smith 1990). Reverse earthquakes above magnitude eight have been observed in the Himalayas. The Assam, 1897, earthquake may have even reached a magnitude as big as 8.7 (Sukhija *et al.*, 1999). Also, there is the possibility of two parallel faults rupturing at the same time, which could easily lead to earthquakes above magnitude eight.

For hazard studies, Fiordland is not of very much interest, and this region has the sparsest fault data in New Zealand. The combination strike-slip and reverse faulting in this region could produce earthquakes above magnitude eight, similarly to the eastern North Island. However, since the subduction zone is nearly vertical (Davey and Smith 1983, Smith and Davey 1984) the surface faults might be fairly steep as well. In that case, the rupture width would be limited and

a smaller maximum magnitude would be likely compared to the eastern North Island.

In conclusion, given the right dip-angles, the regions of reverse faulting can produce earthquakes above magnitude 8.0. Oblique faults are able to produce bigger magnitudes than pure strike-slip and dip-slip faults, because they are only partially limited in length and width. The only strike-slip fault that might be capable of producing earthquakes above magnitude eight is the Alpine fault. There still would be the possibility that the other strike-slip faults would occasionally produce one long through going rupture, as Jackson (1996) suggested, but such an event would not be very likely if possible at all. The strike-slip faults are more likely to produce maximum magnitudes of 7.7 in Marlborough decreasing to the north to maybe 7.5 at the northeast of the North Island. This seems to be consistent with the observed earthquakes in historical times. The only three pure strike-slip earthquakes above magnitude seven over the last 150 years are the 1948 Marlborough, magnitude 7.5, event (Grapes *et al.* 1998), the 1888 Canterbury, magnitude 7.3, event (Cowan 1990), and the 1934 Pahiatua, magnitude 7.6 event (Downes *et al.* 1999).

There is no evidence that the normal faults in New Zealand will produce magnitudes much bigger than 7.2. For these faults, the dip-angle and seismogenic thickness limits the maximum possible magnitude.

It has to be emphasised that these maximum magnitudes are estimated from average earthquake source scaling laws. If the stress drop on many faults in New Zealand (presumably faults with high return times) is significantly higher than the average global stress drop, as observed for the 1934 Pahiatua event, then the maximum magnitude has to be increased.

6 Summary

Here follows a short summary of the comparison of the earthquake occurrence model based on the historical earthquake catalogue and the earthquake occurrence model based on the dynamic fault model:

- Areas dominated by dip-slip faulting and short strike-slip faults are consistent with Gutenberg-Richter behaviour. Earthquake occurrence models based on the historical earthquake catalogue for different time intervals can be used to infer earthquake occurrence for all magnitudes up to a maximum magnitude. The use of longer time intervals has the advantage of producing occurrence models that are stable over time, whereas the use of more recent time intervals results in more accuracy in the detailed locations of individual features of earthquake occurrence.
- Areas dominated by long strike-slip faults behave characteristically. They show little background seismicity, but are episodically able to produce large earthquakes. The accumulated strain is preferentially released in large earthquakes as those faults have grown long and smooth over geological times.
- Regions of oblique reverse faulting possibly produce the largest earthquakes over longer time periods, which may reach magnitudes of up to 8.5. Long strike-slip faults do not produce earthquakes as large, as they are limited to near vertical faults and rupture widths of a maximum of 40 km. Those faults probably do not produce earthquakes above magnitude eight, with the possible exception of the Alpine fault. Pure reverse faults can produce earthquakes that reach magnitudes above eight, but are limited in length (and thus size) by the seismogenic

thickness of the crust. Normal faults produce the lowest maximum magnitude as they are limited in dip-angle and seismogenic thickness.

They are not likely to produce earthquakes bigger than magnitude 7.2.

The conclusion that strike-slip faults and dip-slip faults follow different magnitude frequency relationships was inferred from the following results:

- The inferred magnitudes from reverse fault ruptures are quite close together. This reflects the similar mapped lengths of all reverse faults. If rupture length is corrected for observed single event displacements, the dip-slip faults will generate more magnitude eight earthquakes than observed historically. Additionally, these earthquakes would release too much strain when compared to the rotation rate of the plates. If the single event displacements were lowered to match the mapped fault lengths, the reverse faults would barely produce any earthquakes above magnitude seven and a half. However, larger earthquakes are needed to explain the strain release in the northern and central South Island. A Gutenberg-Richter like magnitude frequency would produce regular medium sized earthquakes and few large earthquakes. The more complicated fault distribution, for example in central Otago, would agree with a model that individual fault segments would rupture more often alone, and occasionally link with other fault segments to produce larger earthquakes.
- The low background seismicity in regions dominated by long strike-slip faults suggest characteristic behaviour for these regions, otherwise not enough earthquakes would be produced to release the accumulated strain. A higher rate of large earthquakes is also suggested by the return

times of long fault ruptures estimated from the fault compilation. It may be reasonable to argue that earthquake rates may change over centuries. However, it has to be noted, that in all dip-slip and strike-slip regions activity rates have not changed over at last seventy years (if not one hundred fifty years). If one was to divide New Zealand into about eight geological sub-regions, the probability would be high that at least one of these regions would have showed a change during the observation time, if indeed the activity did change over the timescale of centuries.

CONCLUSIONS AND FUTURE WORK

We were just having tea, for we do not have supper here, when a tremulous motion was felt which quickly increased to a most terrific shaking. The piano came into the middle of the room and the crashing of earthenware and glass, the creaking and rocking of the house caused a most fearful and terrifying noise – this continued nearly two minutes.

George Richardson, 1855, Wellington

The only earthquake above magnitude eight in New Zealand that has been recorded historically, happened on the 21st January 1855, sixteen minutes after nine o'clock in the evening. It was felt over nearly the whole of New Zealand and substantially damaged many buildings in central New Zealand. Luckily, there were only a few human casualties, despite the size of this earthquake. The question of where in New Zealand such events can be expected and how often they occur is of high social interest.

In this study, methods for the construction of earthquake occurrence models have been developed, which can give an answer to this question. In this chapter the main results have been summarized and an outlook for possible future studies is given.

1 Conclusions

The results of this study can be summarised as follows:

- There is no evidence for any changes in the observed seismicity patterns over the last 150 years. For example, the locations of earthquakes of

magnitude seven and above over the last 150 years match the locations of earthquakes above magnitude three for the last ten years.

- Seismicity features inferred from simulated fault ruptures based on geological fault information do not match the occurrence of historical earthquakes in regions that are dominated by long strike-slip faults. Although these regions have been relatively seismically quiet in the last two centuries, the fault information indicates that frequent large earthquakes can happen in these regions. Additionally, two large earthquakes have been observed in such a region (Marlborough) in the nineteenth century. It is concluded that these regions behave characteristically, i.e. they produce large earthquakes, but not many medium sized earthquakes.
- Any other mismatch between the earthquake occurrence model based on historical earthquakes and the model based on geological data can be explained by shortcomings of the fault data. For example, the global average earthquake source scaling relationships suggest that the mapped fault traces of dip-slip faults do not seem to be representative of the rupture lengths these faults typically produce.
- Earthquake occurrence in regions dominated by dip-slip faulting and short strike-slip faults is best represented by adaptive kernel estimation models based on historical earthquake catalogues with temporal sequences filtered out. The occurrence of larger earthquakes can be calculated using a Gutenberg-Richter distribution (or gamma distribution) with a b-value of one.

- Earthquake occurrence rates in regions dominated by long strike-slip faults cannot be estimated from the historical earthquake record. These faults behave characteristically and ruptures of these faults are best modelled with the help of a dynamic fault model, i.e. individual fault segments are allowed to rupture individually or link with other fault segments to produce a compound rupture. The dynamic fault model seems to produce realistic earthquake occurrence models for the South Island. In the North Island the dynamic fault model seems to produce too few large earthquakes and too many medium sized earthquakes.

The geological modelling process was tested with a particular fault compilation. Future studies may show that the fault compilation used in this study is indeed a good representation of New Zealand's fault distribution, which would have an impact on the conclusions drawn here. However, if it is true that dip-slip faults behave in a Gutenberg-Richter like way, it may be impossible to produce a fault representation of sufficient quality for dip-slip faults based on surface features in the near future, because this fault compilation would have to have complete coverage of at least all fault traces that typically would produce six and a half magnitude earthquakes. Also, several inconsistencies with other observations, which were discussed in detail in this thesis will have to be explained (e.g. the inconsistent magnitude frequency distribution inferred from fault ruptures).

2 Future Work

Although the methods developed in this study are an improvement over existing methods, there are several issues that should be addressed in future studies.

2.1 Seismological Model

A few shortcomings of the seismological model, which have to be addressed in future studies, and some suggestions to further improve the methodology introduced in this study are listed in the following:

- A well-known class of nonparametric regression methods are splines. It would be useful to develop earthquake occurrence models based on splines and to compare the performance of this approach to the performance of adaptive kernel estimation with the comparison methods introduced in this study.
- The suggested filtering of temporal clusters is not connected to any physical processes. It would be of advantage if the filtering could be connected to physical processes. This may result in the introduction of local filtering, where different regions get filtered by a different amount (matching the underlying physical processes in that region).
- A further improvement of the introduced kernel estimation would be the use of non-radial kernel functions. Non-radial kernel functions have the advantage of not imposing radial symmetrical features on the data. An implementation of non-radial kernel functions during this study proved not to be successful. The main reason for the failure of this non-radial approach was that the local asymmetry of each seismicity feature was not weighted by the number of earthquakes within this feature. A

method has to be developed that can relate the amount of asymmetry in a seismicity feature to the number of earthquakes, i.e. if only few earthquakes form a fairly linear feature it should be down-weighted, as it could just be a random feature.

- Although this study has demonstrated that earthquake occurrence in New Zealand has been stable over the last 150 years, the largest earthquakes in New Zealand have been observed to occur in temporal clusters (e.g. Smith 1994). Short-term temporal clusters have been removed for the construction of earthquake occurrence models in this study. Longer term clustering of earthquakes (e.g. Kagan and Jackson 1991, Pérez and Scholz 1997) can be seen if earthquake occurrence models of different decades are compared to each other, i.e. individual seismicity features had different activity levels in different decades for New Zealand. Unfortunately, the time interval for which sufficient historical earthquakes have been observed for the development of an earthquake occurrence model that varies in time is very limited (i.e. about fifty years for New Zealand). Smith and Smith (1995) and Robinson and Benites (1996) compared different temporal clustering models to simulate earthquake occurrence over a longer time interval. Several non-Poissonian occurrence models have been suggested (e.g. Kimberly and Kiremidjian 1995, Wu *et al.* 1995, Papazachos and Papadimitriou 1997), but it is not known which model truly describes temporal clustering of earthquakes in New Zealand and there is probably not enough data for New Zealand to incorporate reliable temporal clustering into earthquake occurrence models. A good starting point for the incorporation of temporal clustering into earthquake occurrence models might be the

finding of Kagan and Jackson (1999) that temporal earthquake clustering can be described by a power law distribution. It has to be considered that local temporal clustering might deviate from a global average behaviour. Nevertheless, the average seismicity over individual decades still has to match the total average.

2.2 Geological Model

The geological earthquake occurrence model developed in this study is not satisfactory. The shortcomings of this model are not due to the proposed dynamic fault simulation but due to sampling biases in the fault compilation and the way they were corrected. For future studies the following issues should be considered:

- Displacement has been assumed to directly scale with rupture length in this study. Although the data do not contradict this scaling behaviour, it is still possible that displacement scales with rupture area rather than rupture length alone. Such a scaling relationship would change the correction of rupture lengths in this study and thus the results of the developed earthquake occurrence model. There are not enough data to decide which is the correct scaling relationship, but once more data are available, the scaling relationship should be revised.
- The strain drop in this study has been assumed to be the same in New Zealand as the rest of the world. Although there is no indication that earthquake source relationships differ in different regions of the world, it cannot be excluded that the average strain drop in New Zealand is higher than in the rest of the world (see chapter three and four). Once there are more data for New Zealand earthquakes, it should be tested

whether the average strain drop is significantly different, i.e. whether any difference cannot be explained by sampling biases of the data. It is possible that especially faults with long return times have higher strain drops, thus it would be of advantage to develop earthquake source relationships dependent on return time.

- For the dynamic fault model developed in this study, very basic assumptions have been made about when fault segments rupture individually and when they link with other segments to produce compound ruptures. It would be advantageous to study the probability of fault linkage dependent of the angle and gap between two segments. The results of such a study could then be used for a revised dynamic fault model.
- The way lengths of dip-slip faults were corrected for the geological earthquake occurrence model lead to too many large reverse earthquakes. The relatively short mapped fault traces of the dip-slip faults have been lengthened to match the relatively high single event displacements. This correction resulted in very long straight dip-slip faults that produce too frequent, large earthquakes. However, it is more likely that dip-slip faults often rupture in single segments and less often the whole fault ruptures. To correct the mapped length of dip-slip faults a method has to be developed that divides the corrected fault trace into several segments which are allowed to rupture individually. Since the amount of segmentation is not known, the segmentation has to be based on assumptions. One assumption would be that the resulting magnitude frequencies of a region dominated by dip-slip faults would match a

Gutenberg-Richter distribution. Another possible assumption would be that the accumulated slip rate matches the deformation rate inferred from the plate movements.

- It should be tested whether there should be a revision of the classification of faults as ‘inactive’ based on a recurrence time longer than 20,000 years. The incorporation of faults with longer return times might lead to an earthquake occurrence model that better matches the historical record of earthquakes (e.g. in the eastern North Island and in west Nelson). If there still is a mismatch after the addition of such faults, it is likely that there are still faults missing, and a model has to be developed to add extra faults. Such an addition of extra faults would not be an easy task and would have to be based on several assumptions, e.g. that the accumulated slip rates of the faults match the deformation rate inferred from the plate movements (or a fixed part of the deformation rate, if aseismic strain release is to be allowed for).
- Another possible future fault model would allow for assigning probabilities of fault linkage to individual faults in an interactive program. While the development of such a program would be fairly easy, there are not enough data to decide how big the probability is for individual faults segments to produce compound ruptures with other segments. However, it is possible that the individual segments cannot be treated with a general model that considers all faults to be equal.

2.3 Other Studies

It would be an advantage to produce representations of local b-values and maximum magnitudes similar to the representations of earthquake occurrence rates produced in this study. The problems connected with the development of such representations have been discussed in chapter four, section four and five.

It would be useful to calculate uncertainties of the local modelled earthquake occurrences. This can be done for adaptive kernel estimation, for example, with the bootstrap method (e.g. Wand and Jones 1995). However, the uncertainty estimation has to also incorporate an error for the temporal sequence filtering. The difficulty for such an uncertainty estimation is that it is not known what part of the temporal variation is due to earthquake processes (e.g. aftershocks) and what part is due to random scatter. The estimation of uncertainty of the fault model has other complications. Firstly, the uncertainties due to sampling biases have to be taken into account. Further, it would have to be estimated how many faults have not been mapped, i.e. how high the likelihood is of large earthquakes occurring where no faults have been mapped. This uncertainty can only be estimated with assumptions, for example that the earthquake occurrence rate based on the fault model has to match the one based on historical earthquakes.

Overall the methods for the generation of earthquake occurrence models developed in this study have produced an improvement over existing methods. However, there is still a lot of room for improvement. The main difficulty in developing earthquake occurrence models is that the physical processes that generate earthquakes are not sufficiently understood. The better the understanding of the physical processes is, the better the resulting earthquake occurrence model will be.

EVIDENCE FOR DIFFERENT SCALING OF EARTHQUAKE SOURCE PARAMETERS FOR LARGE EARTHQUAKES DEPENDING ON FAULTING MECHANISM

1 ABSTRACT

Scaling relationships between seismic moment, rupture length, and rupture width have been examined. For this purpose, the data from several previous studies have been merged into a database containing more than 550 events. For large earthquakes, a dependence of scaling on faulting mechanism has been found. Whereas small and large dip-slip earthquakes scale in the same way, the self-similarity of earthquakes breaks down for large strike-slip events. Furthermore, no significant differences in scaling could be found between normal and reverse earthquakes and between earthquakes from different regions. Since the thickness of the seismogenic layer limits fault widths, most strike-slip earthquakes are limited to rupture widths between 15 to 30 km while the rupture length is not limited. The aspect ratio of dip-slip earthquakes is similar for all earthquake sizes. Hence, the limitation in rupture width seems to control the maximum possible rupture length for these events. The different behaviour of strike-slip and dip-slip earthquakes can be explained by rupture dynamics and geological fault growth. If faults are segmented, with the thickness of the seismogenic layer controlling the length of each segment, strike-slip earthquakes might rupture connected segments more easily than dip-slip events, and thus can produce longer ruptures than dip-slip events of the same width.

2 INTRODUCTION

Over the last three decades, scaling relationships between different earthquake source parameters (e.g. rupture length, width, and displacement) and the seismic moment or magnitude have been the subject of many studies. Early studies found similar scaling relationships for earthquake source parameters throughout the whole range of earthquakes, and the concept of self-similar earthquakes (e.g. Kanamori and Anderson 1975) was proposed. Later studies found a breakdown of self-similarity for larger strike-slip earthquakes (e.g. Purcaru and Berckhemer 1982, Scholz 1982, Romanowicz 1992). This can be explained by the finite thickness of the seismogenic layer and the near verticality of strike-slip faults, which limits their rupture width (e.g. Shimazaki 1986, Pacheco *et al.* 1992). No breakdown of self-similarity has been found for dip-slip earthquakes (except for the largest subduction zone events, e.g. Purcaru and Berckhemer 1982), which has been explained by the much greater range of possible rupture widths in geological dip-slip environments (e.g. Scholz 1982, Romanowicz 1992). Scholz *et al.* (1986) excluded subduction zone events from their data set and found that large dip-slip earthquakes scale in the same way as large strike-slip earthquakes, which implies a corresponding breakdown of self-similarity for large dip-slip events. In another study, Romanowicz (1992) suggested that the seismogenic depth limits rupture slip as well and inferred scaling relationships for large earthquakes that were different from those of Scholz (1982), Scholz *et al.* (1986) and Scholz (1994a), whereas Gross (1996) suggested that the rupture area limits rupture slip. The recent studies from Wells and Coppersmith (1994) and Wang and Ou (1998), in contrast, could not find any difference between the scaling relationships of source parameters for earthquakes of different sizes and faulting mechanisms.

The two main reasons for the disagreements lie in two limitations of the data used in past studies. First, most studies only included a small number of earthquakes (20-50 events). These numbers are not sufficient for statistical studies if the data are further separated and examined for dependence on earthquake size, faulting mechanism and region. The second limitation lies in the range of source parameter values. For example, Wells and Coppersmith (1994) analysed a data set of nearly 250 events, large enough for a good statistical analysis. However, they included only those rupture lengths less than 500 km and those rupture widths less than 35 km, although rupture lengths and rupture widths up to 1000 km and 300 km are known for the largest earthquakes. This limitation resulted from their exclusion of subduction zone events, which led to the omission of nearly all large dip-slip earthquakes.

In our study, a large data set has been constructed and analysed with appropriate statistical methodology to clarify whether there is a difference in scaling behaviour between large dip-slip and strike-slip earthquakes, which cannot be simply explained by the wider range of possible rupture widths of dip-slip earthquakes. Further, the scaling relationships between seismic moment and rupture length have been calculated.

3 DATA AND METHODS

The data from selected past studies have been merged together into one database containing more than 550 events (Kanamori and Anderson 1975, Geller 1976, Purcaru and Berckhemer 1982, Scholz 1982, Bonilla *et al.* 1984, Kanamori and Allen 1986, Scholz *et al.* 1986, Shimazaki 1986, Romanowicz 1992, Wells and Coppersmith 1994, Anderson *et al.* 1996, Yeats *et al.* 1997, Margaris and Boore 1998). Obvious typographical errors have been corrected. If

different values for a source parameter of an earthquake are given in different studies, the lowest and highest values and the mean have been included. For the calculation of scaling relationships between earthquake source parameters, only the mean value has been used. The individual faulting mechanisms (slip-types) have been adopted from the different studies, with Wells and Coppersmith (1994) as first choice. The final database contains seismic moments between 10^{15} and 10^{24} Nm (M_W : 4.0 - 9.6), rupture lengths between 0.1 and 1000 km, and rupture widths between 1 and 300 km. Our database can be downloaded from <http://www.geo.vuw.ac.nz/~stock/scaling.txt>.

Since we have simply adopted all the source parameters (including faulting mechanism), we depend on the accuracy of the data from the past studies we used. A good and full description of acquisition and possible errors and biases of source parameters can be found in Wells and Coppersmith (1994). Few of the studies used discuss the possible error range of their data, and most do not explain how the faulting mechanism was defined (and how well it is defined). Figures A1.2 to A1.5 show that the range of the individual source parameters is rather large.

An earthquake fault is mathematically described by a displacement discontinuity (dislocation) across a surface in an elastic medium. This displacement is equivalent to the distribution of double couples on the same surface (e.g. Burridge and Knopoff 1964) and is represented by a tensor – the seismic moment tensor (e.g. Aki and Richards 1980, 53). The total scalar moment of the double couples is given by:

$$(A1.1) \quad M_0 = \mu A d ,$$

where μ is the rigidity of the elastic medium, A is the area of the surface and d is the average dislocation (displacement). The surface area A is often simplified as the product of the average dislocation length L along the strike of the surface and the average dislocation width W down the dip of the surface. If the aspect ratio L/W and the strain drop d/L are constant, earthquakes are self-similar (Kanamori and Anderson 1975).

For scaling relationships between seismic moment and rupture length of the form

$$\log M_0 \sim n \log L \quad (\text{A1.2})$$

a slope of $n=3$ indicates self-similarity. The limitation of the rupture width of large earthquakes by the finite thickness of the seismogenic layer (e.g. Shimazaki 1986, Pacheco *et al.* 1992) leads to a breakdown of self-similarity and different values for the slope n . If the strain drop d/L stays constant for large earthquakes, the slope will change to $n=2$ (e.g. Scholz 1982). In contrast, if the displacement scales with width, the slope will change to $n=1$ (Romanowicz 1992).

The slope n of the ‘best’ fitting line through two observables, when both have a statistical uncertainty, can be estimated as given in Casella and Berger (1990, 589):

$$n = \frac{-(S_{xx} - \lambda S_{yy}) + \sqrt{(S_{xx} - \lambda S_{yy})^2 + 4\lambda S_{xy}^2}}{2\lambda S_{xy}}, \text{ where} \quad (\text{A1.3})$$

$$S_{xx} = \sum_{i=1}^N (x_i - \bar{x})^2, \quad S_{yy} = \sum_{i=1}^N (y_i - \bar{y})^2, \quad S_{xy} = \sum_{i=1}^N (x_i - \bar{x})(y_i - \bar{y}), \text{ and } \lambda = \sigma_y^2 / \sigma_x^2.$$

The variances σ_x^2 and σ_y^2 give the deviation of the data points from the best fitting line in x and y directions. They are a combination of the individual measurement errors of the two earthquake source parameters and the scatter from an ideal linear behaviour due to different stress drops, aspect ratios, etc. for each earthquake. From estimating the scatter in figures A1.2 to A1.4, the approximate relation $\sigma_x^2 = 2 \sigma_y^2$ was inferred for this study, where x and y are given by the rupture length and the seismic moment, respectively. Other values of λ have been tried, without significant differences in the results.

For a useful interpretation of the goodness of fit, the data should be (log)normally distributed. Since the earthquake source parameters in this data set have been measured using different methods (surface length, aftershocks, etc.) this assumption may not hold true. To demonstrate that the data are approximately log-normally distributed, the frequencies of all differences between predicted rupture lengths, inferred from the estimated scaling relationship, and measured rupture lengths have been calculated (see figure A1.1). The frequency distribution passes the Kolmogorov-Smirnov test for the normal distribution $N(0,0.15)$ at the 95% level. Further, scaling relationships have been calculated for the minimum and maximum values of the source parameters, as well as for the mean values. The relationships obtained lead to the same conclusions about earthquake source parameter scaling. The variance of the differences between predicted rupture lengths and measured rupture lengths has been found to be the same for all magnitudes.

Figure A1.1: Rupture Length Distribution. *The frequencies N of the difference between measured rupture lengths L and rupture lengths Δ modelled by the orthogonal least-squares regression. For comparison, the normal distribution $\mathcal{N}(0,0.15)$ is shown.*

The estimator of the variance of the slope n is given by (Casella and Berger 1990, 592):

$$\sigma_n^2 = \frac{(1 + \lambda n^2)^2 (S_{xx} S_{yy} - S_{xy}^2)}{(S_{xx} - \lambda S_{yy})^2 + 4\lambda S_{xy}^2} \quad (\text{A1.4})$$

The goodness of fit for the scaling relationships has been tested using the confidence interval with a confidence level of 95%, which is given by (Casella and Berger 1990, 593):

$$n - \frac{t(N-2, 0.025) \sigma_n}{\sqrt{N-2}} \leq n' \leq n + \frac{t(N-2, 0.025) \sigma_n}{\sqrt{N-2}}, \text{ where} \quad (\text{A1.5})$$

n' is the slope to be tested, N the number of data points, and $t(N-2, \alpha/2)$ is the Student-t distribution.

4 Large and Small Earthquakes

Earthquakes rupturing the whole seismogenic layer are classified as ‘large’ and are limited in rupture width by the thickness of the seismogenic layer. ‘Small’

earthquakes, in contrast, do not rupture the whole seismogenic layer. Since the thickness of the seismogenic layer differs regionally (e.g. 15 km in California, and 25 km in East Africa; Jackson and White 1989) separation of the data using a fixed rupture width does not lead to good results. Instead, large and small earthquakes are here separated by an aspect ratio (rupture length to width) of three, although for small earthquakes an aspect ratio of two has been given in past literature (Geller 1976, Purcaru and Berckhemer 1982). After trying a range of values it was concluded that a higher ratio leads to better separation because of the scatter in the measured values of rupture length and rupture width. With this type of separation, large and small earthquakes are still not perfectly separated, but the method nevertheless leads to good results.

As shown later in the discussion, the frequencies of the aspect ratios of the dip-slip and small strike-slip earthquakes follow a log-normal distribution with a mean of about two (figure A1.6). Very few dip-slip, and large strike-slip, earthquakes have an aspect ratio of more than three. This is the basis for our decision to separate small and large earthquakes by an aspect ratio of three.

5 Results

Scaling relationships between seismic moment and rupture length have been calculated for various earthquake sizes and faulting mechanisms. Dip-slip events from Japan and eastern Russia, and strike-slip events from California, have each been compared to corresponding events from the rest of the world (see figure A1.3 and A1.4). The results are given in table A1.1 and figures A1.2 to A1.5. In table A1.1, N is the number of events used for the calculation, n and σ_n are given in equations A1.3 and A1.4. In figures A1.2 to A1.5, the mean values of the

source parameters are plotted as well as an error bar, which is the range between smallest and greatest measured values.

Slip-Type	Size	N	n	σ_n
normal	small	32	3.1	0.8
	large	6	4.1	2.1
reverse	small	77	2.9	0.9
	large	9	2.9	0.9
dip-slip (Japan)		21	3.2	1.1
dip-slip (Eastern Russia)		16	2.9	1.3
strike-slip (California)	small	27	2.8	1.4
	large	9	2.1	0.8
strike-slip (other Regions)	small	33	2.9	1.1
	large	25	2.3	1.4

Table A1.1: Seismic Moment and Rupture Length Scaling.

No difference in the scaling behaviour, including mean aspect ratio, between normal and reverse earthquakes could be found. Furthermore, there is no statistical evidence for a different scaling behaviour between small and large dip-slip earthquakes. All dip-slip earthquakes lie approximately on a line with the slope of one third (see figure A1.2). The tested slope of $n = 3$ lies within the 95% confidence limits for all dip-slip data sets. The tested slope of $n = 2$ lies outside the 95% confidence limits except for the large normal earthquakes. For the few large normal events, and the largest dip-slip events (see figure A1.2), a breakdown in $\log M_0 \sim 3 \log L$ scaling might be possible. No statistical difference between the scaling behaviour of dip-slip earthquakes in different regions could be found (figure A1.3).

Figure A1.2: Seismic Moment versus Rupture Length for Dip-slip Earthquakes. *The solid lines have a slope of 1/3 and 1/2. Both sets of events lie closer to the line with the slope of 1/3 than to the one with a slope of 1/2.*

Figure A1.3: Seismic Moment versus Rupture Length for Dip-slip Earthquakes from Different Regions. *The solid lines have a slope of 1/3 and 1/2. The three sets of events lie closer to the line with the slope of 1/3.*

In contrast to dip-slip events, small and large strike-slip earthquakes scale differently (see figure A1.4). Whereas the small earthquakes scale with $n = 3$, the large earthquakes scale with $n = 2$. There is no statistical evidence for a different scaling between Californian strike-slip earthquakes compared to earthquakes throughout the rest of the world. The tested slope of $n = 3$ lies within the 95% confidence limits for both data sets of small strike-slip events, whereas the tested slope of $n = 2$ lies outside the 95% confidence limits. The tested slope of $n = 2$

lies within the 95% confidence limits for both data sets of large strike-slip events, whereas the tested slopes of $n = 3$ and $n = 1$ lie outside the 95% confidence limits. The only exception are large strike-slip events in California, where the tested slope of $n = 3$ lies just inside the 95% confidence limits. This can be explained by imperfect separation of small and large earthquakes.

Figure A1.4: Seismic Moment versus Rupture Length for Strike-slip Earthquakes. *The solid lines have a slope of 1/3 and 1/2. Between the seismic moments of 10^{19} and 10^{20} Nm the strike-slip events change slope from 1/3 to 1/2.*

Figure A1.5 shows the different rupture length to width scaling behaviour for large dip-slip and large strike-slip earthquakes. For strike-slip earthquakes, the rupture width does not increase for rupture lengths bigger than about 75 km. However, there is no statistical difference in rupture length and rupture width scaling for large and small dip-slip events. Only the five dip-slip events with the longest rupture lengths possibly show a breakdown of self-similarity. Most of these events occurred in the Aleutian Arc before 1965 and may be poorly estimated (e.g. Sykes 1971, Johnson and Satake 1997). Thus, a breakdown in $n = 3$ scaling even for very large earthquakes is not evident statistically.

Figure A1.5: Rupture Length versus Rupture Width. *The solid lines have a slope of 1. The three lines give a mean aspect ratio of 1, 2 and 3.*

Figure A1.5 shows, in addition, that earthquakes with oblique mechanisms lie between the pure slip mechanisms. At the low end, some earthquakes seem to deviate from the linear rupture length and rupture width scaling, but these earthquakes have rupture lengths and rupture widths of only a few kilometres. Their values may not be very well determined (source parameters for such earthquakes are often given as a radius).

6 Discussion

The frequencies of the aspect ratios for dip-slip earthquakes seem to follow a log-normal distribution (see figure A1.6), which explains why not many dip-slip events are classified as large (see table A1.1). Only one normal event (31/6/1970, Columbia, M_W 7.7) and four reverse events (09/03/1957, Aleutians, M_W 8.25; 22/05/1960, Chile, M_W 8.5; 04/02/1965, Rat Island, M_W 8.25; 29/09/1973, Vladivostok, M_W 7.75) have a rupture length to width ratio larger than four (see figure A1.6). It is possible that they are not very well determined: not only could the source parameters be in error, but also each of these earthquakes could have had an oblique, rather than a pure dip-slip, mechanism.

This behaviour combined with the $n = 3$ scaling for all, except perhaps the largest dip-slip events, suggests that there is no breakdown of self-similarity for dip-slip events. The aspect ratios of strike-slip earthquakes, in comparison, are not log-normally distributed, with the biggest events reaching aspect ratios up to thirty. For large strike-slip earthquakes, $n = 3$ scaling breaks down for strike-slip events for widths between 15 to 30 km, as observed in previously studies (e.g. Purcaru and Berckhemer 1982, Scholz 1982, Romanowicz 1992). The result of Romanowicz (1992), that large strike-slip earthquakes scale with $n = 1$ and, thus, that the rupture width controls slip, could not be confirmed. However, an intermediate value of n between 1 and 2, as suggested by Gross (1996), is possible.

Figure A1.6: Distribution of Aspect Ratios for Different Slip-types. *The aspect ratios of the dip-slip events are approximately log-normally distributed. The strike-slip events show a tail to higher aspect ratios. The oblique events lie in between the dip-slip and strike-slip events.*

The reason for this breakdown of the self-similarity of strike-slip earthquakes is the limitation of rupture width of near vertical strike-slip faults imposed by the thickness of the seismogenic layer (e.g. Pacheco *et al.* 1992). Large pure strike-slip events mostly occur in regions where the maximum thickness of the seismogenic layer is 10 to 30 km. In contrast, the largest dip-slip events occur in subduction zones where the shallow dips of the fault planes allow rupture widths up to 300

km. Accordingly, for these events no breakdown of earthquake source parameter scaling is apparent. Compared to large strike-slip earthquakes, even the largest subduction zone events do not reach as high aspect ratios. Their large moments may cause adjacent fault segments to rupture, and thus may produce bigger aspect ratios than for smaller dip-slip events, but there is no evidence that they can produce aspect ratios as high as those produced by strike-slip earthquakes.

The set of dip-slip events includes intraplate and other non-subduction zone events, and the question arises why no dip-slip events with small rupture widths and long rupture lengths are observed - no dip-slip events appear among the width-limited strike-slip events in figure A1.5. Jackson and White (1989) studied normal fault traces from different parts of the world and found that they consist of small segments with a maximum length of about 25 km. They concluded that the thickness of the seismogenic layer controls the length of these segments. Hayward and Ebinger (1996) also related the length of normal faults in the East African Rift system to the width of the seismogenic layer. Scholz and Contreras (1998) give an explanation for this behaviour for faults in continental rifts. The absence of high aspect ratios in the set of all dip-slip earthquakes suggests that dip-slip earthquakes in other tectonic regimes (e.g. backarc spreading, continental uplift) are also limited in length by the width of the seismogenic layer.

In contrast to dip-slip faults, strike-slip faults lengthen over geological time (e.g. Wesnousky 1988). The rupture direction could provide an explanation for the difference in dip-slip and strike-slip faults, and thus the difference in large earthquake scaling. Dip-slip events may rupture dominantly along dip and strike-slip events along strike (see for example the Landers, California, earthquake of 1992, e.g. Kanamori *et al.* 1992; and the Northridge, California, earthquake of

1994, e.g. Wald *et al.* 1996). If so, it is dynamically easier for strike-slip events to rupture successive segments, compared to dip-slip events (for an example of dynamic rupture modelling see Madariaga *et al.* 1998). Rupture in the direction of the strike may cause adjacent segments to break, and consequently the rupture length may continue to grow until very long rupture lengths are produced. A rupture direction perpendicular to the strike of the fault is less likely to cause adjacent segments to rupture and thus the rupture will not propagate very far along strike. Hence, strike-slip faults may grow in length during single and successive ruptures, while dip-slip faults are less likely to. As a consequence, the width of the seismogenic layer controls the maximum size of dip-slip events, as it effectively limits both rupture width and rupture length (and therefore slip).

7 Conclusion

This study has shown that rupture length and rupture width scale in the same manner for small earthquakes, independent of faulting mechanism. For large earthquakes with strike-slip mechanisms, the self-similarity of earthquakes breaks down due to the limitation of rupture width caused by the thickness of the seismogenic layer. However, the rupture length is not limited for these events. In contrast, $n = 3$ scaling does not break down for earthquakes with dip-slip mechanisms. Only for the very largest dip-slip events might a breakdown of self-similarity occur, but the data are too few to make this statistically significant.

Two other important results are the absence of any difference in scaling behaviour between normal and reverse events, or even any evident difference between the aspect ratios of the two groups, and the absence of any observable difference of scaling behaviour between different regions of the world.

The different behaviour of large dip-slip and strike-slip earthquakes cannot be explained by the wide range of possible rupture widths for dip-slip earthquakes. For dip-slip earthquakes the rupture length is controlled by the width of the seismogenic layer, whereas strike-slip earthquakes can produce longer ruptures. This behaviour could be explained by the rupture direction relative to the strike of the fault.

a p p e n d i x 2

THE FAULT COMPILATION

The fault compilation from Stirling *et al.* (1998) is listed here for reference (see table A2.1 and figure A2.1). A detailed discussion on how individual values were obtained and further references can be found in the study of Stirling *et al.* (1998).

Figure A2.1: Fault Map. Taken from Stirling *et al.* (1998) and slightly modified.

The following is taken from Stirling *et al.* (1998):

“The fault data are obtained largely from published and unpublished sources. The fault traces have been simplified to reduce computation time. Magnitude and rupture length are preferentially taken from observed ruptures. If no rupture has been observed on a fault, magnitude is inferred from estimates of single event displacements and fault area. If single event displacements are not available, the magnitude is inferred from fault length using Wells and Coppersmith (1994). Return time is preferentially taken from geological estimates. If such estimates are not available, return time has been inferred from slip rate and single event displacement. If single event displacements are not known, the return time is inferred from magnitude and moment release rate.

Despite the large magnitudes of some observed earthquakes, only few of them have produced clear evidence of surface rupture. The combination of sparse historical records and rapid modification of surface rupture in the humid New Zealand climate may have prevented the identification of some surface ruptures.”

The values reported here are mean values. The slip rates listed in bold are not part of the original database and have been estimated from return times and displacements, or from fault lengths using equation 3.4 where displacements are nor available.

Figure A2.2: Geological Map. Taken from Stirling et al. (1998).

Fault Name	Faulting Mechanism	Displacement	M _{max}	Slip Rate	Return Time
Ohariu	strike-slip	5.0	7.5	2.7	3250
Pohangina	reverse	2.5	6.8	0.30	6500
Levin	reverse	2.5	6.4	0.30	6500
Marton	reverse	2.5	6.6	0.75	6500
Dry River	reverse		6.9	1.18	3000
Wellington SW	strike-slip	4.2	7.4	7.10	600
Wellington NE	strike-slip	4.2	7.6	7.10	800
Wellington W	strike-slip		6.9	7.10	1000
Feilding	reverse	2.5	6.9	0.30	6500
Awatere	strike-slip	6.0	7.7	8.00	2930
Wairua	reverse-oblique		7.0	1.25	1745
Wairau	strike-slip	6.0	7.9	4.90	1650
White Creek	reverse	6.0	7.6	0.20	34000
Lyell	reverse		6.7	0.20	4661
Brunner	reverse		6.9	0.38	15000
Inangahua	reverse	0.4	7.4	0.10	4400
Clarence	strike-slip	3.6	7.5	4.00	900
Elliot	strike-slip		6.9	1.20	1064
Kekerengu	strike-slip	7.0	7.2	9.00	778
Jordan	reverse	4.7	7.2	2.00	2583
Fowlers	strike-slip		6.6	2.50	338
Hope SW	strike-slip	3.0	7.2	17.50	200
Hope 1888	strike-slip	2.0	7.2	17.50	120
Hope NE	strike-slip	3.0	7.2	17.50	200
Kakapo	strike-slip	3.0	7.3	7.00	500
Kelly	strike-slip		6.5	0.78	650
Torlesse	strike-slip		6.6	0.17	3500
Rolleston	strike-slip		6.5	0.13	3500
Harper	reverse		6.9	1.07	3500
Porters Pass	strike-slip	3.5	7.4	5.50	700
Lake Heron	reverse	4.0	7.1	0.10	5000
Foxs Peak	reverse	4.0	7.2	1.00	5000
Alpine SW	reverse-oblique	10.0	8.1	25.00	400
Alpine NE	reverse-oblique	10.0	7.7	10.00	800
Alpine Bends	reverse-oblique		7.2	2.80	580
Pisa	reverse	3.0	7.1	0.37	30000
Nevis	reverse		6.8	0.30	3677
Spylaw	reverse		6.3	0.50	1300
Blue Mountain	reverse		6.4	1.06	800
Akatore	reverse	1.5	6.8	1.88	900
Esk	strike-slip		7.0	0.42	3500
Alfredton	strike-slip	8.0	7.3	2.00	2000
Wairarapa 1855	reverse-oblique	12.1	8.1	9.45	2000
Mohaka S	strike-slip	2.0	7.1	4.65	1000
Mohaka N	strike-slip	2.0	7.1	4.65	1000
Ruahine N	reverse-oblique	3.5	6.9	1.25	3500
Ruahine C	reverse-oblique	3.5	7.3	1.25	3500
Ruahine S	reverse-oblique	3.5	7.2	1.25	3500
Napier 1931	reverse-oblique	2.5	7.8	1.06	2500
Waimana	strike-slip	3.5	7.3	0.55	3500
Whakatane	normal-oblique	3.5	7.4	3.20	3500
Waiohau	normal-oblique	4.5	7.5	0.80	3500
Edgecumbe	normal	1.9	6.6	1.00	1190
Waikaremoana	strike-slip	3.5	7.0	0.88	3500
Awaiti	normal		5.9	1.00	387
Kerepehi S	normal	1.8	6.7	0.50	3600
Kerepehi N	normal	1.8	7.0	0.50	3600
Inglewood	normal-oblique		6.5	0.50	1552

Ararata	normal		6.2	0.02	28210
Waverley	normal		6.4	0.03	24046
Nukumarū	normal		6.2	0.07	12500
Mt. Stewart	reverse	2.5	6.8	0.30	6500
Pukerua-Shepherds	strike-slip	3.8	7.2	1.90	3750
Galpin	normal		6.3	0.04	15828
Leedtown	normal		6.3	0.70	916
Upokongaro	normal		6.6	0.01	82793
Moumahaki	normal		6.2	0.19	2900
Ridge Road	normal		6.6	0.59	650
Waitotara	normal		6.1	0.71	694
Himatangi	reverse	2.5	6.9	0.30	6500
Rurima	normal		6.8	0.60	1970
Pukehoko	normal		6.1	0.50	1173
Nukuhou	normal		6.2	2.40	251
Rangitaiki	normal		6.4	2.30	306
Ohiwa	normal		6.5	0.70	1433
White Island	normal		7.0	0.70	2482
Aorangi	reverse		7.0	4.00	335
Pa Valley - Makuri	strike-slip	6.0	7.4	2.25	2500
N Weber	strike-slip	3.0	6.7	0.75	2000
Saunders-Weber	strike-slip	3.0	7.1	0.75	2000
Ruataniwha	reverse-oblique	3.0	6.7	0.32	4000
Oruawhoro	strike-slip	3.0	6.9	0.54	4000
Waipukurau-Poukawa	reverse-oblique	3.0	7.1	0.21	5000
Kaweka	strike-slip	3.5	7.1	0.50	3500
Patoka	strike-slip	4.0	7.0	3.46	1000
Rangiora	strike-slip		6.4	5.2	899
Cape Kidnappers W	normal	2.8	6.8	0.33	4000
Cape Kidnappers E	normal	2.8	6.7	0.27	4000
Clear Hills	strike-slip		6.7	0.76	3500
Silver Range W	normal	2.8	6.7	0.27	3500
Silver Range C	normal	2.8	6.7	0.33	3500
Silver Range E	normal	2.8	6.8	0.33	3500
Mangaoranga	normal-oblique		6.1	0.05	5000
Waitawhiti	normal-oblique		6.4	0.09	4000
Mauga	normal-oblique		6.8	0.19	5000
Poroutawhao	reverse	2.5	6.8	0.17	20000
Ruahine Range	reverse		6.5	0.14	20000
Tararua Range	strike-slip		6.1	0.01	15000
Mataroa	normal		6.2	0.06	10000
Hihitahi - Waiouru	normal		6.3	0.58	1250
Kariori	normal		6.3	0.49	1500
Ohakune	normal		6.5	2.48	370
Raurimu	normal		6.6	1.54	650
Taupo Fault Belt S	normal	1.5	6.8	7.99	150
Taupo Fault Belt N	normal	1.5	7.0	10.47	150
Wanganui Offshore	normal		6.8	0.24	5000
Coastal Zone	reverse		7.0	1.99	2000
Hawkes Bay 1	reverse	5.0	7.3	2.80	1250
Hawkes Bay 2	reverse	5.0	7.2	2.74	1250
Hawkes Bay 3	reverse	5.0	7.4	4.39	1250
Hawkes Bay 4	reverse	5.0	7.4	4.39	1250
Masterton	normal-oblique		6.3	0.55	1231
Carterton	normal-oblique		6.7	1.00	1012
Ostler N	reverse	3.0	7.1	1.00	3000
Ostler C	reverse	3.0	6.9	1.00	3000
Ostler S	reverse	3.0	6.9	1.00	3000
Ahuriri River	reverse	2.5	6.7	0.25	10000
Irishman Creek	reverse	4.0	7.0	0.26	15000
Quartz Creek	reverse	2.5	6.7	0.26	5000
Rangitata Range	reverse	3.0	7.0	0.31	10000
Dryburgh	reverse	2.5	7.0	0.17	22000

Waitangi	reverse	2.5	6.7	0.10	35000
Wharekuri	reverse	4.0	7.1	0.24	20000
Lindis Pass	reverse-oblique	3.0	6.9	1.01	3000
Otematata	reverse	2.5	6.7	0.24	20000
Grandview	reverse	3.0	6.9	0.17	30000
Cardrona S	reverse	2.0	7.0	0.25	7500
Cardrona N	reverse	2.0	7.1	0.25	7500
Blue Lake	reverse	3.0	6.9	0.75	5000
Dunstan N	reverse	4.0	7.2	1.00	8000
Dunstan S	reverse	4.0	7.0	1.00	8000
Raggedy Range	reverse	3.0	6.9	0.36	8000
N Rough Ridge	reverse	3.0	6.9	0.36	8000
Rough Ridge	reverse	3.0	6.9	0.47	8000
Ranfurly S	reverse	3.0	6.9	0.47	8000
Ranfurly N	reverse	3.0	6.9	0.47	8000
Hyde	reverse	3.0	7.0	0.20	15000
Kirkliston	reverse	3.0	6.9	0.20	15000
Tukituki	reverse	5.0	6.7	0.55	5000
Matata	normal		5.8	0.80	466
Raetihi	normal		6.8	0.49	1500
Shannon	reverse	2.5	6.5	0.13	20000
Avoca	strike-slip		6.7	0.13	3500
Moonlight	reverse-oblique		7.0	0.17	10000
Cape Egmont	normal		7.1	0.50	2915

Table A2.1: Fault Compilation.

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