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Synthetic Seismicity of the Wellington Fault System

R Robinson, R Benites, IGNS

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Review of Project No 93/144
Synthetic Seismicity of the Wellington Fault System

Reviewer: Prof R I Walcott
Research School of Earth Sciences
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October 5th, 1995

Attainment of objectives:

The initial proposal listed 5 objectives. In my opinion they have all been met satisfactorily. A detailed assessment follows:

A) *Develop the theory and computer programs necessary to model the seismicity on interacting faults, of any orientation, in an elastic three-dimensional half space.*

Comment: The collection of computer programs that have been developed in the course of this project is the most general and appropriate available at the present time. Using the analysis developed by Okada (1992) and extending the simplified approach of Ben-Zion and Rice (1993), the project researchers have developed a three dimensional model appropriate to the problem set. The effect of approximations introduced into the model to attain a realistic computation time are uncertain but the researchers are aware of these and are proposing to remedy any shortcomings.

B) *Determine to what extent the interaction with other faults modifies the behaviour of a single fault under the same loading conditions.*

Comment: The results of this study seems to be quite unambiguous - interaction strongly modifies behaviour substantially increasing the variability in temporal distribution of major events within a region, and within the Wellington region in particular.

C) *Develop a fault model of the Wellington region and use the above results to simulate long-term seismicity under various initial conditions.*

Comment: Their second, as yet unsubmitted, paper addresses this objective. Any model for a region as complex as this must be a considerable approximation and it will take some time to determine which variants within the model are of first order and which of lesser importance and significance. The researchers have made a sensible approach to the problem using a few adjustable parameters that 'tune' the process to deliver (synthetic) earthquakes on the major faults. The various initial conditions examined differ only in the driving mechanism proposed for loading the model - the physical parameters of models are quite restricted and it is not clear from the text to what extent variation in these parameters will affect the calculated seismicity. There is little variation in output for different driving mechanisms.

D) *Investigate the implication of the final model for average and time variable hazard.*

Comment: A major conclusion of the study is that interaction in the Wellington fault system modifies the temporal distribution of large earthquakes on the major faults of the region.

E) *Prepare a paper(s) on all the results for publication.*

Comment: One paper has be accepted for publication in the Journal Geophysical Research - published by the American Geophysical Union. This is a refereed journal of excellent repute and a very appropriate journal in which to publish the results of the theoretical research. One paper is in preparation and attached as an appendix to the report. It is clearly nearly ready for submission and is a substantial piece of work. This last paper incorporates the body of study as applied to the Wellington region.

Does the report address the subject matter of the study:

The report, and the two papers appended to it, are directly relevant to the study. All the material is clearly written and well produced.

Appropriate use and good value for the grant:

This study is a fine example of basic research directed at a problem of some considerable relevance to earthquake hazard but nevertheless one that came about because technological developments and theoretical understanding of processes permits indicates that that particular research is likely to lead to some interesting conclusions. Because of recent technical developments in the calculation of stress/strain relations in three dimensions in fractured material of some considerable sophistication it became possible to undertake the construction of quite complicated models that could not be done before. The successful attainment of their objectives indicates several further lines of development of their model. It is now possible to think that a well constrained model could lead to a good understanding of the mechanical causes of earthquakes in a particular region and how the occurrence of an event in one part of the region may increase the risk within a short time on another part of the region.

Further research:

The researchers asked the question - just how important is interaction between faults, and in particular how important is it a region like Wellington. There study has show that faulting events on one fault do indeed affect the seismic behaviour of adjacent faults - and more importantly they show how to calculate what that affect may be. They have chosen to express the affects of the interaction by examining how seismic events are distributed in magnitude and time by computing a synthetic seismic sequence.

The consequences of this study are not null - it is a study with conclusions that indicate further lines of profitable investigation. It would now be well worthwhile to carry out strategic studies using the models developed to examine the effects on seismicity by varying specific aspects of the model. For example to try and determine what the likely consequences of a locked subduction zone will have on the seismic development of the region. What will happen when it is unlocked? More observation data is becoming available all the time - we now have a much better idea of the rate of strike-slip faulting in the Wellington region which can further constrain the model.

Additional recipients of the report:

I recommend that this report be sent to Departments of Geology at Otago and Canterbury and to Professor Smith at Victoria but not, at this stage, I should think much wider distribution. Most research people in this field here, and overseas, will see the papers when they appear. Insofar as the report presents the results of basic research it does not have a heavily applied component at this stage. It is my opinion that the results have a very high relevance and significance for the earthquake hazard but will need further development to have clear practical application of value to industry.

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Synthetic Seismicity of the Wellington Fault System

by

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Prepared for

Earthquake Commission

Final Report of

EQC Research Project 93/144

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SUMMARY

Our aim in this two year project has been to develop computer models of seismicity on multiple faults that interact elastically. The first year was devoted to development of the modeling techniques, and their application to the simple example case of two parallel strike-slip faults. The second year was spent in applying the model to the specific fault system of the Wellington region (about 100 km x 100 km). We give a brief summary here, but the bulk of this report is comprised of two research papers, to be published internationally, in which full details of our results are given and discussed:

(1) Robinson, R., and R. Benites, Synthetic Seismicity Models of Multiple Interacting Faults, Journal of Geophysical Research, in press (due out in the August issue);

(2) Robinson, R., and R. Benites, Models of Fault Interaction for the Wellington Region, New Zealand: Effects on the Temporal Clustering of Large Events, being submitted to Journal of Geophysical Research (currently under internal IGNS review).

Our model of the Wellington region includes the major strike-slip faults in the region (the Wairarapa, Wellington (treated as two separate segments), Ohariu-Shepherd's Gully (combined into one), and the Wairau faults; see Figure 1.). The parameters of the model were adjusted to reproduce the observed paleoseismic data on recurrence intervals, magnitudes, and slip during characteristic events. We also included the subduction interface that lies at about 20 km depth, because large events due to seismic slip on the interface are very common in other subduction environments worldwide. However, no such events have occurred in the Wellington region in historic times and there is little geological information with regard to their potential size and recurrence time. Thus a variety of scenarios were investigated that span what we consider as reasonable .

Our results indicate that, for the Wellington region, there is a strong enhancement in the risk of multiple large events within a short time due to fault interactions, as compared to the corresponding cases with no fault interaction. The magnitude of the increase is in proportion to the amount of seismic slip on the plate interface. For the case in which there is no seismic slip on the plate interface the probability of having two large events (magnitude 7.2 or more) within one year of each other (0.00027 per year) is 18 times that for the case of no interaction. For the case of 40% coupling, the increase in probability (to 0.00180) is by a factor of 93. To look at it a different way, for the case with least enhancement (model 1, no subduction thrust) there is a 7.6% chance that, given a characteristic event on one fault has occurred, the next characteristic event, on a different fault, will occur within 1 year. This compares to a 0.4% chance if interactions are suppressed. A magnitude 7.2 event on any of the faults we have included would produce a modified Mercalli intensity of 8 or more in central Wellington city (Dowrick, 1991).

The computer modelling technique we have developed has been successful and its application to the Wellington region has important implications for the Earthquake Commission, as well as other insurers. The model can be applied to other regions as well.

Reasons for the Project

The idea of investigating synthetic seismicity in this way was based on several observations:

1) statistical analyses of worldwide earthquake catalogues have indicated that some degree of clustering of large events in space and time is present (e.g. Kagan and Jackson, 1991);

2) when considering individual regions, which differ considerably in their geologic complexity, the available catalogues are too short and too inhomogeneous to provide any reliable information on clustering parameters, even when extended by paleoseismic observations;

3) simple models of the induced stresses due to large earthquakes can explain some examples of triggered events nearby (Hudnut et al, 1989; King et al, 1994);

4) the Wellington region is probably highly susceptible to such interactions with its complex system of sub-parallel strike-slip faults and the subduction interface directly below (VanDissen and Berryman, 1995); and

5) the possible clustering of large events has important implications for estimating time variable seismic hazard, and insurance strategies.

Thus our idea was that computer models of seismogenesis and faulting could be used to generate long, homogeneous, synthetic catalogues of earthquakes using parameters matched to the geology of a specific region, including many faults. The catalogues would be sufficiently long that statistical inferences about the degree of clustering of large events (or its lack) could be made. Such models have been developed in the past (e.g., Ward and Goes, 1993), but only for single, segmented faults.

The implementation of this idea was made possible by:

1) development of reasonably realistic models of fault failure and propagation of slip in a quasi-static manner, so avoiding having to solve the full elasto-dynamic equations (e.g. Ben-Zion and Rice, 1993);

2) the publication by Okada (1992) of a formulation for computing the induced stresses, anywhere in an elastic half-space, due to faulting of any type;

3) sufficiently (just) powerful computer resources; and

4) the growing amount of geologic information on the past behaviour of faults in the Wellington region (Van Dissen and Berryman, 1995).

LIMITATIONS IN RESULTS

We feel that several aspects of the model are still sufficiently uncertain that we are hesitant to apply any specific model's quantitative results to estimate time variable seismic hazard in Wellington. In particular, the critical role played by the amount of seismic slip on the interface between the subducting Pacific plate and the overlying Australian plate must be appreciated: we hope that research in progress (D. Darby, pers. comm.) will shed more light on this critical problem over the next few years. Also, we have included only the 6 major faults in the region in our model, and we recognise that events on smaller faults are important in estimating the total hazard. However, if it is accepted that the faults in the Wellington region rupture primarily during large characteristic events (as the geologic data strongly suggest) then we cannot envisage any future modifications to our models that would change our qualitative conclusion.

FURTHER WORK

We stress that our purpose has been to determine if fault interactions in the Wellington Region are likely to be important in determining time variable hazard: we think the answer is "yes". However, there are many possible modifications to our model that we would eventually like to explore, so that quantitative estimates of the risk of multiple large events, with well established error bounds, can be made with more confidence. In particular we would like to investigate the effect of more realistic driving mechanisms (how the faults are pushed toward failure) and a greater variety of mechanical properties of the faults. We don't think that any of these factors would increase the probability of multiple large events still more, but they conceivably could decrease the probability somewhat.

We would like to thank the Earthquake Commission for enabling us to undertake this interesting research.

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APPENDIX I

Synthetic seismicity models of multiple interacting faults

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Abstract.

A synthetic seismicity model for multiple, interacting faults, of any strike and dip in a three-dimensional elastic half-space has been developed. The ultimate purpose is to determine how hazard is modified by elastic interactions of events in complex fault systems and to examine the synthetic seismicity for unusual behavior before large events that might be observable in the real world. Each fault is subdivided into some number of equal-sized patches, and a coefficient of static friction, random within a specified range, is assigned to each. The patch static strength is then the product of friction and differential normal stress. Stress (both shear and normal) accumulates due to some predefined tectonic driving force until one or more patches fail and slip instantaneously. The slip can be in any direction within the fault plane and its magnitude is determined by a specified fractional stress drop. This initial slip induces shear stress and strength changes on all other patches and faults and perhaps results in other patches failing. Patches can slip more than once; once having failed, their strength is reduced, resulting in an overshoot that approximates the effect of dynamic stress enhancement. An event is over when all patches are stable. Although present computational constraints do not allow models with more than about 1500 patches, or a large number of faults, experiments with simple models of two parallel strike-slip faults (25 km length, 15 km depth, 694 m x 714 m patch size), driven at different rates and separated by 10 or 15 km, give some indicative results. A range of friction between 0.1 and 0.2, with a stress drop of 10% (about 3 MPa or 30 bars), produces a *b* value (500,000 events) close to 1. The models generate a distinct class of large "characteristic" events that rupture the full fault plane. The distribution of interevent times for the background seismicity is very similar to that for a Poisson process, while the characteristic events are quasi-periodic. Nearly half the larger events directly induce some activity on the other fault. The effect of a characteristic event on the occurrence of another characteristic event on the opposite fault is small and depends on the separation: at 10 km distance other events are retarded while at 15 km distance they are advanced. At 10-km separation the probability of a pair of characteristic events within 1 year is reduced to 0.3% from 1.7% for the case with no interaction. Consideration of interactions between slip on the faults and the driving mechanism, as well as higher stress drops and more overshoot, would increase the magnitude of the interaction effects.

Introduction

In most parts of the world the historical record of seismicity is too short to enable firm conclusions to be made about the repeat time of large earthquakes or possible clustering of large events in both space and time. Paleoseismic studies can usefully extend the record for specific sites, but the progress is slow and, of course, subject to large uncertainties. Likewise, it is difficult to make firm conclusions about possible precursory seismicity patterns from existing catalogs of instrumentally located earthquakes because of their short duration and

inhomogeneity. These facts justify the development of "synthetic seismicity" models, in which long catalogs of events are generated by computer models of seismogenesis. Such models can be "tuned" by requiring that they reproduce what is known of the statistics of past seismicity to a reasonable degree. The models can then be used to generate much longer, homogeneous, catalogs of events so that statistical inferences about the behavior of seismicity can be made.

Of course, the degree to which such synthetic catalogs can be taken as accurate descriptions of real-world seismicity depends on how physically realistic the underlying model is. Since many aspects of seismogenesis are still the subject of debate, some doubt must exist about how much weight to attach to any specific model's results. Also, computational limitations place severe restraints on how much detail can be included in a model. The expectation is that as both knowledge of seismogenesis and computational resources grow, the results of synthetic seismicity models will become more and more reliable sources of information.

In recent years, there has been a bewildering variety of synthetic seismicity models presented. Some general classes and representative examples are: (1) cellular automata models (usually nondeterministic) of two-dimensional faults which neglect the details of both elasticity and fault friction but which can reproduce seismic frequency-magnitude statistics as a manifestation of self-organized criticality [e.g., *Bak and Tang*, 1989; *Ito and Matsuzaki*, 1990; *Barriere and Turcotte*, 1994)]; (2) spring-block models of one dimensional or two dimensional faults with realistic frictional properties but which greatly simplify stress transfer to nearest neighbor interactions [e.g., *Burridge and Knopoff*, 1967; *Cao and Aki*, 1984; *Wang*, 1991; *Brown et al.*, 1991; *Carlson*, 1991a; *Carlson and Langer*, 1989; *Carlson et al..*, 1991]; (3) models of single two dimensional faults in which slip is discretized into patches which obey simplified frictional laws [e.g., *Mikumo and Miyatake*, 1979; *Rundle and Kanamori*, 1987; *Ward*, 1991, 1992; *Ward and Goes*, 1993; *Ben-Zion and Rice*, 1993]; (4) continuum models of single two dimensiaonal faults that utilize realistic constitutive friction laws and may consider details of slip nucleation and propagation, at least approximately [e.g., *Rice*, 1993; *Stuart*, 1986, 1988]; (5) actual physical, as opposed to computer, models of the spring-block type [e.g., *King*, 1975, 1991].

In addition, *Rundle* [1988a] developed perhaps the most general mathematical model for synthetic seismicity, based upon the hypothesis that earthquakes are fluctuations of plate motions. He applied this model to southern California [*Rundle*, 1988b]. *Harris and Day* [1993] developed a model of parallel, one dimensional, strike-slip faults in which details of slip initiation and propagation are considered via a finite-difference procedure and used it to study how slip can, or cannot, jump across fault jogs or offsets. *Dieterich* [1994] presented a model of seismicity based on a three dimensioanl distribution of point sources obeying realistic nucleation laws and subject to an imposed stress history but did not include elastic interactions.

Most of the studies referred to above were meant to model seismicity in a general sense, but some modeled specific faults. The model of *Ben-Zion and Rice* [1993] was developed to represent the San Andreas fault near Parkfield where a series of quasi-periodic magnitude 6 events have occurred. Their model showed that the temporal distribution of such events may be quite complex. In general, complex, perhaps chaotic, behavior is observed in many synthetic seismicity models [*Ito*, 1980; *Narkounskaia and Turcotte*, 1992; *Huang and Turcotte*, 1992], although periodic large events seem to occur in continuum models [*Rice*, 1993] (but see *Horowitz and Ruina* [1989]). *Ben-Zion and Rice* [1993] suggested that models with discrete areas of uniform slip are, perhaps, a better representation of the mechanical and geometrical complexity of real faults than continuum limit models.

Since our ultimate aim is to develop a realistic, three dimensional model to represent the system of faults near Wellington, New Zealand (subduction thrust plus overlying strike-slip faults in an oblique convergence environment), we have chosen to build upon the discretized, quasi-static class of models. Such models seem to offer the best combination of reasonably realistic dynamics and attainable computation time for the three dimensional case we wish to consider. Thus our initial model is similar in many ways to that of *Ben-Zion and Rice* [1993] but extended to a fully three dimensional case and including multiple faults of any orientation. However, we retain their quasi-static approach (the full elastodynamic equations, resulting in wave generation, are not used), simplified friction law (details of slip nucleation and propagation are neglected), and discretation into large patches instead of a continuum model with cells smaller in size than the critical slip distance [*Rice*, 1993]. Likewise, the effect of changes in pore pressure is not modeled as yet, nor are time-dependent strengths that would generate aftershocks [*Dieterich*, 1972]. Central to our approach are the results of *Okada* [1992], which allow the calculation of displacements, and their derivatives, anywhere in an elastic half-space due to slip of any orientation on a rectangular fault.

One of the specific questions we hope to address is the effect of elastic interactions on triggering, or retarding, of large events by events on other faults [*Hudnut et al.*, 1989; *Harris and Simpson*, 1992; *Jaume and Sykes*, 1992; *Stein et al.*, 1992; *King et al.*, 1994; *Stein et al.*, 1994]. The observed clustering of earthquakes, beyond that caused by aftershocks [*Vere-Jones*, 1970; *Kagan and Jackson*, 1991; *Grant and Sieh*, 1994], as opposed to the quasi-periodic/characteristic event or random occurrence (Poisson) models, may have its roots in such interactions [*Ward and Goes*, 1993; *Du and Aydin*, 1993; *Robinson*, 1994; *Kagan*, 1994a]. *Reasenberg and Simpson* [1992] showed than induced stress changes (expressed as a change in the Coulomb failure function) as low as 0.01 MPa (0.1 bar), following the 1989 Loma Prieta earthquake, produced observable changes in the rate of small earthquake activity on nearby faults. *Cornell et al.* [1993] considered the approximate effect of interaction of fault patches on a stochastic model of event occurrence.

Another application of synthetic seismicity catalogs is to search for or test proposed precursory patterns in the low-level seismicity before larger events. *King* [1986] examined his laboratory model for such precursors, and *Carlson* [1991b], *Shaw et al.* [1992], *McCloskey and Bean* [1992], *Christensen and Olami* [1992], and *Pepke et al.* [1994] have used spring-block models in such a way, with mixed results. *Mikumo and Miyatake* [1983] examined the effect of various types of strength distributions on precursory seismicity patterns in their model of a two-dimensional fault. *Barriere and Turcotte* [1994] found only a weak precursory increase in rate of activity before large events in a cellular automaton model. *Gabrielov et al.* [1990] applied an "M8" type forecasting algorithm to a block model of seismicity with encouraging results.

Formulation of the Model in General Terms

We consider a homogeneous three dimensional elastic half-space in which there are a number of rectangular faults specified by their strike, dip, and position in geographical coordinates (east, north, up) and by their length L and width W. Each fault is subdivided into a large number of equal-sized rectangular patches. It is convenient to introduce two additional coordinate systems (Figure 1): (1) fault-referenced coordinates, x, y, z, as used by *Okada* [1992]; and (2) another set of fault-referenced coordinates useful for calculating the shear and normal stress in the fault plane, 1, along strike, 2, up dip, and 3, fault normal. Transformation of stress components from one frame of reference to another is done using the formula for second-order tensors.

Each fault patch is assigned a coefficient of friction, μ, and an initial normal stress, τ_{33}, based on lithostatic pressure. We then specify some external, constant, driving mechanism that loads the faults continously with time but is itself not affected by slip on the faults. This loading can be derived in several ways, as discussed below, and is expressed as changes of stress per unit time, $\delta\tau_{ij}$, on each patch. The loading pushes the fault patches toward failure due to the accumulation with time of the shear stresses τ_{31} and τ_{32} and possibly the normal stress τ_{33}. The sign of the accumulated stresses can be positive or negative.

The failure criterion for each patch is that the absolute value of the maximum in-plane shear stress (at the patch center) meet or exceed the static strength *S*, as given by the product of the coefficient of friction and normal stress . The maximum shear stress is easily calculated from the in-plane shear stresses.

Eventually a patch does fail and results in a stress drop that we take as a fixed fraction of the static strength. *Okada*'s [1992] computer routines are used to derive the slip (uniform over the patch) that would produce this stress drop at the patch's center. His formulation (which is too long to reproduce here) gives the displacements U_i and their spatial derivatives anywhere in an elastic half-space due to slip of any direction and magnitude on a rectangular fault (here a single patch). Strain, e_{ij}, is given by

$$e_{ij} = 0.5\left(\frac{\partial U_i}{\partial x_j} + \frac{\partial U_j}{\partial x_i}\right) \qquad (1)$$

and stress, τ_{ij} is given by the generalized Hooke's Law

$$\tau_{ij} = 2Ge_{ij} + \lambda\delta_{ij}\sum_{k=1}^{3}e_{kk} \qquad (2)$$

where λ and G are the media's Lamé constants (G is also called the rigidity) and δ_{ij} is the Kronecker delta function. We perform these calculations for a position at the patch center, assuming unit slip in the direction of maximum shear stress, to obtain shear stress change per unit slip and simply invert this for slip per unit shear stress change. The required patch slip is just this multiplied by the stress drop. This procedure amounts to a generalization of a patch's "self-stiffness" as defined by *Rice* [1993] for slip in the strike direction of a vertical fault. Note that we do not allow any slip in the fault normal direction.

The first patch failure marks the beginning of an "event," during which the loading mechanism and "time" are halted. The event may grow because slip on a patch induces stress changes on all other patches on all faults. These are again calculated using the formulation of *Okada* [1992] and Hooke's Law and transformed into fault-referenced shear and normal stresses. If these changes are sufficient to cause other patches to fail, they are allowed to do so as part of the same event or, if on another fault, as part of a separate event. If, indeed, more patches have failed, then their slips are calculated as above and the resulting induced stresses on all other patches determined. Patches may slip more than once, but eventually all patches are stable and the event, or events, are over. The loading mechanism is then allowed to resume, and the cycle repeats for as long as desired.

Because of the nonlinear rupture mechanism it is not possible to solve directly for the behavior of the system during one rupture episode, much less the entire future history. If the process were a linear one, we might be able to use use a boundary element formulation [e.g., *Gomberg and Ellis*, 1994]. As it is, we must do it iteratively.

It is important to note that an "event" is confined to a single fault and includes all patches on that fault that have failed. An event can be initiated by either the loading mechanism or induced stress from failures occuring on another fault. In the latter case the events occur at the same "time," but we refer to the second event as "induced" or "triggered" and to the first event as an "inducing" or "triggering" event.

In order to approximate the effects of slip dynamics on the induced stresses, the strength of a patch that has already slipped in an event can be held, during the remainder of the event, at some level below its static strength. This makes it easier for the patch to reslip, producing an "overshoot" effect as described by *Ben-Zion and Rice* [1993], and allows allows large "characteristic" events that rupture the entire fault to occur more easily. Once the event is over, all patch strengths are reset to their static values. There is no strightforward physical interpretation of the level of "dynamic strength": it is set by trial-and-error so that a realistic number of characteristic events are produced. It is also a weak function of patch size, becoming less necessary as size decreases.

Event moment, M_o (in Newton meters), is calculated as

$$M_o = G\sum_m Au_m \qquad (3)$$

where the sum is over all patches that failed during an event, A is the area of a patch (in square meters), and u_m is the absolute scalar slip on patch m (in meters). Moment magnitude, M_w [*Kanamori*, 1977], is given by

$$M_w = 0.67 \log M_o - 6.03 \qquad (4)$$

Our model differs from that of *Ben-Zion and Rice* [1993] in that fault heterogeneity is modeled by variations in the coefficient of friction, and hence strength, rather than by variations in stress drop. Kinematically, the two methods are much the same, and depending on the application, both types of variation can be included. Note that our model, like that of *Ben-Zion and Rice* [1993], does not generate aftershocks per se because of the lack of any time-dependent strength (except in so far as the normal stress changes) or slip nucleation details.

The specification of the loading mechanism can take several different forms. The most general scheme would be to take an unfaulted region of lithosphere and apply some boundary conditions (e.g., as dictated by plate tectonics) and then allow faults to form and evolve with time. However, this is an order of magnitude more difficult problem than that which we actually consider: how faults interact, given that they already exist and are observed to have slip in some direction. We do not try to treat the details of how they came to be. Thus one way to specify the loading is simply to take the stress increments to be constant along the fault and in the same direction as the desired slip (e.g., for a pure strike-slip fault, set $\delta\tau_{31}$ as constant for all fault patches and $\delta\tau_{32}$, $\delta\tau_{33}$ = 0). The magnitude of the increments can be adjusted so that the resulting slip rate matches the desired (observed) rate. However, due to interactions between patches this scheme results in total slip (over many events) that tapers from a maximum at the fault center to a minimum at the edges, plus some slip in other directions in the case of nonvertical faults. There may also be problems due to accumulation of fault-normal stresses so that some patches eventually become extremely weak or strong. It could be argued that in nature, such stresses would decay with time due to slip on nearby subsidiary faults that we do not model and incorporate such a decay into the model.

Another possibility (which we use in the example discussed below) is to take the loading mechanism as steady, aseismic slip along extensions of the fault planes both along strike and dip. The formulation of *Okada* [1992] can then be used to calculate the stress increments on each fault patch, which turn out to be larger near the edges than at the center of the fault. This results in a near-uniform slip distribution (over many events) and makes physical sense in the case of parallel faults but may be difficult to justify in other cases (e.g., faults at right angles).

A more general approach is to specify the ultimate desired slip distribution (e.g., uniform slip of x meters per unit time) on each fault and use *Okada*'s [1992] formulation to calculate the stress field that would result from that slip. That stress field is an approximation to the required stress increments that will eventually, after many events, result in the required slip. It is only an approximation because the slip accumulates over time through the complex, nonlinear, rupture process rather than in one go. Our experience is that it is an adequate approximation.

A Specific Example

The procedures outlined above can, in principle, be applied to any number of faults, subdivided into patches as small as desired. In practice, however, there are some severe constraints on our computational scheme. If the interaction terms between fault patches can be calculated ahead of time and stored in memory (random access memory (RAM), not virtual memory on disk!), then the computational time for each event is rather short. If the interaction terms are calculated anew as the event develops, the time required is extremely large. We choose the first scheme, for which the amount of memory required to store the interaction terms is large, since they must be specified for each of the six independent stress tensor components, for each component of displacement, and for each fault patch acting on all the others. For the moment, we have only considered cases where the number of patches in each fault is sufficiently small that the interaction terms can be stored in the memory of the most appropriate machine readily available to us (Digital Equipment Corporation AXP 3000 - 300, with about 150 MByte of available RAM).

As an example, we will describe the results for the case of two vertical, parallel faults, driven so that the slip is primarily strike slip. This case includes elements of both mutual slip enhancement and retardation, as will be seen. Each fault is 25 km long by 15 km deep (Figure 2), separated by either 10 km (model A) or 15 km (model B). Each fault is divided into approximately square patches, 694 m by 714 m (36 by 21 patches, 1512 in all). The driving mechanism is specified as steady slip, at 0.01 m/yr and 0.007 m/yr for faults 1 and 2, respectively, in a left- lateral sense, both below and along strike of each fault. Such a driving force associated with one fault does not produce any fault-normal stress changes on that same fault but does affect all the stresses on the other fault, both normal and shear. We have taken both the two Lamé constants of the half-space to be 3.18×10^4 MPa, and the density as $\rho = 2.65 \times 10^3$ kg/m^3; this is equivalent to a P wave velocity of 6.5 km/s, an S wave velocity of 3.46 km/s, and a Poisson's Ratio of 1/4. The initial normal stress on all fault patches has been taken as 195 MPa, equivalent to lithostatic pressure at 7.5 km depth. We have not included any depth dependence for two reasons: (1) below about 5 km depth, pore pressure is likely to increase at the same rate as the lithostatic pressure, so that the differential normal stress does not increase with depth [*Rice*, 1992, 1993]; (2) our model does not include any depth dependence of the elastic parameters. If a depth-dependent lithostatic stress were included, the model would produce very many small events at the shallowest depths. In reality, the crust at shallow depth does not fail seismically, in part because of a very low rigidity that we cannot model. Keeping the (initial) normal stress constant seems the best solution. However, changes in normal stress due to patch interactions are allowed to accumulate.

We have experimented with several different distributions of the coefficient of friction, as well as several different values of stress drop and dynamic strength/static strength ratio. The values chosen for this example are appropriate for a weak fault/low stress drop/slow rupture environment: the coefficient of friction ranging between 0.1 and 0.2, a stress drop of 10%, and minimal overshoot. This produces a reasonable frequency-magnitude distribution, consistent with observations. If no overshoot is included, then the largest events rupture only about 2/3 of the fault plane and their numbers drop off rapidly below that expected from an extrapolation of the magnitude-frequency distribution. We have included only just enough overshoot to generate a reasonable number of large "characteristic" events, rupturing the whole fault plane. This hardly affects the magnitude-frequency distribution in its linear range. The magnitude of the characteristic events is, of course, defined by the size of the faults. A stress drop of 10% is equivalent to about 3 MPa (30 bars), in accord with real world observations.

We will discuss the results for catalogs of 500,000 events (about 2 days CPU time; 28,900 years of simulated time). However, we will disregard the initial 25,000 events because of transient behavior as evidenced by very high *b* values and lack of characteristic events. As a check, the final average displacements across the two faults were compared with what would be expected from the driving mechanism acting over the duration of the simulation: the expected and actual were within 2% of each other, with only very minor vertical offsets.

For comparison with the above models, we have also generated catalogs with interfault elastic interactions suppressed (referred to as models A' and B'); all model parameters are otherwise the same as above.

Events produced using the parameters discussed above range in magnitude from 3.82 (for a single patch) to 6.16. The frequency distributions of numbers of events versus magnitude (Figure 3) are linear up to about magnitude 5.2, with a *b* value (negative of the slope) close to 1.0. A feature of the magnitude distribution is the smooth drop-off of the numbers of larger events below an extrapolation of the linear region, until the "characteristic" events appear. Such events appear naturally out of the mechanics assumed here and do not require any special distribution of strength, such as strong asperities. They tend to occur when the average shear stress on the fault is high, so that failure can more easily extend from patch to patch. The number of characteristic events is about what would be predicted by a *b* value of 1: if we had included more of an overshoot effect, then the number of such events would be higher. There are only small differences between the frequency-magnitude curves for models A, B, A', and B'. It should be noted that a *b* value of about 1 seems to be a nearly universal outcome of "self-organizing" seismicity models, and so cannot be used to claim that any specific model is more correct than another.

If events of all magnitudes are considered then the distributions of interevent times are very close to that for a Poisson process (Figure 4). However, the characteristic events (defined in this case as those with M_W 6.0 or more) on a given fault occur quasi-periodically (Figures 5, 6): the occurrence of such an event completely suppresses any reoccurrence, on the same fault, for some years. There is then a strong tendency for reoccurrence at about the "characteristic repeat time" (average characteristic event slip / loading rate). However, there are a significant number of much longer recurrence times. The results for models A and B are quite similar. If large events on both faults are considered together, the distribution of interevent times begins to converge to the

Poissonian form. We expect that if more faults with various characteristic repeat times were added, then the convergence would be more pronounced.

The effect of fault interaction on the characteristic events is easily observable in both models A and model B but of opposite sense. This can be seen by considering the occurrence of pairs of characteristic events within a given time interval as compared to the cases with interaction suppressed (Figure 7). For short time intervals the occurrence of pairs of events for model A (10 km separation) is reduced from the corresponding case with no interaction (i.e., a characteristic event on one fault tends to retard the occurrence of one on the other fault). The situation for model B (15 km separation) is the opposite (i.e., a characteristic event on one fault tends to advance the occurrence of one on the other fault). The reasons for this will be discussed below.

Another manifestation of interaction is that a significant fraction of the larger events (not just the characteristic events) on one fault immediately trigger events on the other fault (Figure 8a). The result is that 2.0% and 1.3% of the events in models A and B, respectively, are directly triggered (Figure 8b). While it is natural to expect, and it is indeed true, that larger events have a higher probability of inducing other events, it is also observed that larger magnitude events have a greater chance of being induced than smaller events.

Discussion

For the simple models considered above, the effect of a characteristic event on the occurrence of a similar event on the other fault depends on the distance between the faults. This can be understood by examining the slip distribution in a characteristic event and how its induced stresses vary with distance. The slip during a characteristic event induces changes in both the shear and normal stresses on the other fault. These changes can be conveniently combined as the change in a "Coulomb Failure Function" (CFF) [e.g., *Reasenberg and Simpson*, 1992] given by

$$\delta CFF = \delta\tau_{shear} + \mu(\delta\tau_{33}) \qquad (5)$$

where we have taken shear stress as positive if it favors slip with a horizontal component in the same sense as the driving mechanism (e.g., left-lateral in our models) and we use the sign convention that tensile normal stress is positive.

The slip during a typical characteristic event is maximum at about halfway along strike and 3/4 up dip, tapering off toward the edges and almost entirely along strike. The areal variation of the induced δCFF on a parallel fault for such a generalized slip distribution (Figure 9) shows that slip is almost entirely retarded (negative δCFF) for a separation of 10 km and at a depth of 7.5 km. At a separation of 15 km and at the same depth, the central part of the second fault undergoes a positive δCFF. In general, the distance out to which slip is entirely retarded, on a parallel fault, is roughly half the fault length. A fault with a more realistically longer length than we can now model would thus have a larger region in which slip is retarded.

A specific example of the induced stresses for model B (15 km separation, Figure 10) shows that the induced stress change is quite small (maximum of +0.03 MPa or 0.3 bars as compared to a fault strength of about 3 MPa). However, this is apparently enough to observably increase the chance of another characteristic event within a short time. It should also be noted that characteristic events can be directly induced by smaller (non characteristic) events on the opposite fault.

While the effect of fault interaction on the characteristic events in our models is observable, it is not particularly large. For example, in model A the probability of getting a pair of characteristic events within 1 year is 0.3% with interactions versus 1.7% without interactions. This is in part due to the fact that the models represent an environment where induced stresses are minimal: the stress drops are low and dynamic effects are small. Moreover, the effect of interactions between the fault slip and the driving mechanism is neglected altogether. In the real world, interactions might be strong, serving to amplify the effect of direct elastic interactions over a period of time.

It is not possible to use specific models such as our example here to make conclusions about the magnitude or sign of interaction effects on the occurrence of large earthquakes in general. Each seismic region of the world has its own specific set of faults with widely varying geometries and so must be considered individually. Models such as ours that can be easily adapted to fit the circumstances will thus be important. The other option is to lump everything together and consider things in a statistical way [*Kagan*, 1994a]. Still, the models we discuss here do indicate that inhibition might be important in some cases. Of course, the same event that retards other events on one specific fault could advance events on faults of a different orientation. In the real Earth, there might often be such a favorably oriented fault.

Our model generates characteristic events without any unusually large regions of high strength, such as asperities. This is not to say that asperities do not exist in reality; in fact, we think they do and may have very important consequences with regard to the initialization of large events. Likewise, barriers would be important in models of faults extending over larger distances along strike then we can now consider. Future extensions of this work will hopefully consider these points.

One of the eventual purposes of this work is to examine the predictability of the large characteristic events, both through examination of the series of such events themselves using recently developed nonlinear forecasting techniques [e.g., *Weigend and Gershenfeld*, 1994] and through variations in the background seismicity. However, a full study is beyond the scope of this paper and will appear separately. Briefly, we find that simple seismicity statistics such as the *b* value and rate of activity are not of much value as precursors to characteristic events, nor is there much useful information in the time series of such events itself. If the time periods between events anywhere on a single fault are considerd as a time series, then our models behave chaotically in the sense that small changes in the initial condiations eventually produce large changes in the catalog. However, the overall statistics, like the *b* value, remain invariant.

Recent developments in the study of nonlinear dynamical systems have shown that chaotic systems with a low-dimensional attractor, embedded in a higher dimensional phase space, can mimic randomness such as we see. This has led to the hope that seismicity can be adequately explained by a similarly low dimensional attractor. However, attempts to determine an attractor dimension for real seismicity at Parkfield have had mixed results [*Horowitz*, 1989; *Beltrami and Mareschal*, 1993], and it seems that the seismicity is hardly distinguishable from randomness. Part of the problem may be that seismicity is in fact a spatio-temporal phenomenon, like turbulence, and has a very large phase-space dimension, of the order of the number of patches in our models. Techniques for studying such systems are only just being developed.

Our synthetic seismicity model has several obvious shortcomings, some of which we hope to remedy in future work. Among these are several that can be addressed by introducing time-variable strengths (lack of aftershocks, pore-pressure changes), whereas others are very difficult (vertical variation in elastic properties) or impossible in a quasi-static model (rate/state-dependent friction, more exact handling of dynamic effects). Intermediate in difficulty are two that we consider the most important for our purposes: (1) the lack of interaction between the driving mechanism and the fault slip; and (2) the size of the individual patches being too large. Patch size can, of course, be easily reduced but only at the expense of much larger computation time. For example, we did generate a catalog of 250,000 events for a model similar to those discussed above (but no overshoot) with a patch size of 500 m x 500 m. This took 35 days of processor time. The main effect was to decrease the magnitude of the smallest events generated, which would be useful if testing of precursory seismicity patterns was envisaged. Likewise, it would be possible, in principle, to introduce some interactions with the driving mechanism at the expense of computation time. For example, a large subduction thrust event is thought to induce further aseismic slip both updip and downdip, so leading to more rapid reloading.

Our model does not include any account of changes in fault geometry with time. However, by its nature the model can be used to investigate the effect of nonplanar fault surfaces. For example, each fault patch could be offset a small but random amount in the direction perpendicular to the fault, so simulating in some sense a "fractal fault surface" as envisaged by *Kagan* [1994b].

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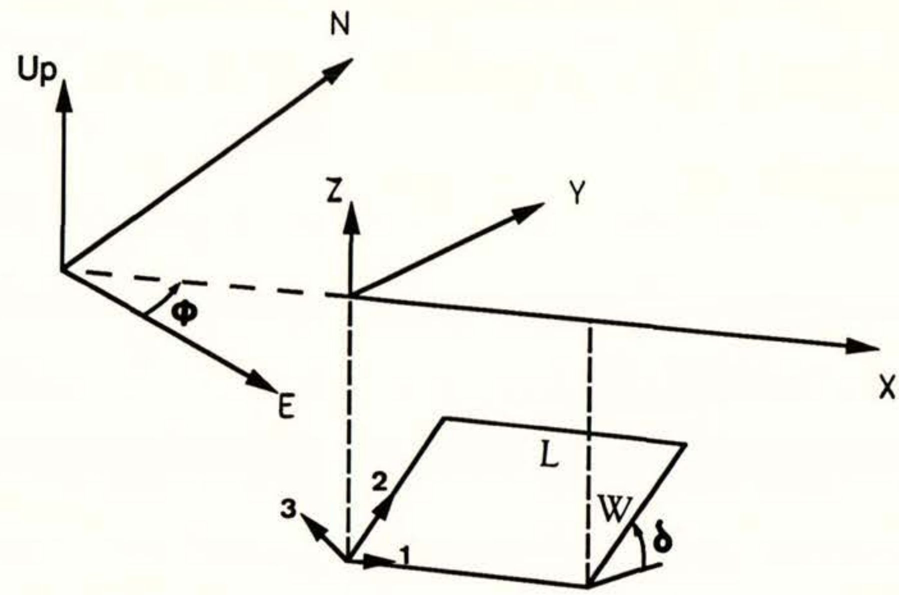

Figure 1. The three coordinate systems used in the synthetic seismicity model. E (east), N (north) and U (up) are the master coordinates in which each fault is specified; x, y, and z are the fault-referenced coordinates used by Okada [1992]; 1, 2 and 3 are our fault-referenced coordinates used for calculation of shear and normal stresses on the fault plane. Here δ is the fault dip, ϕ is the fault strike, and L and W are the fault length and width.

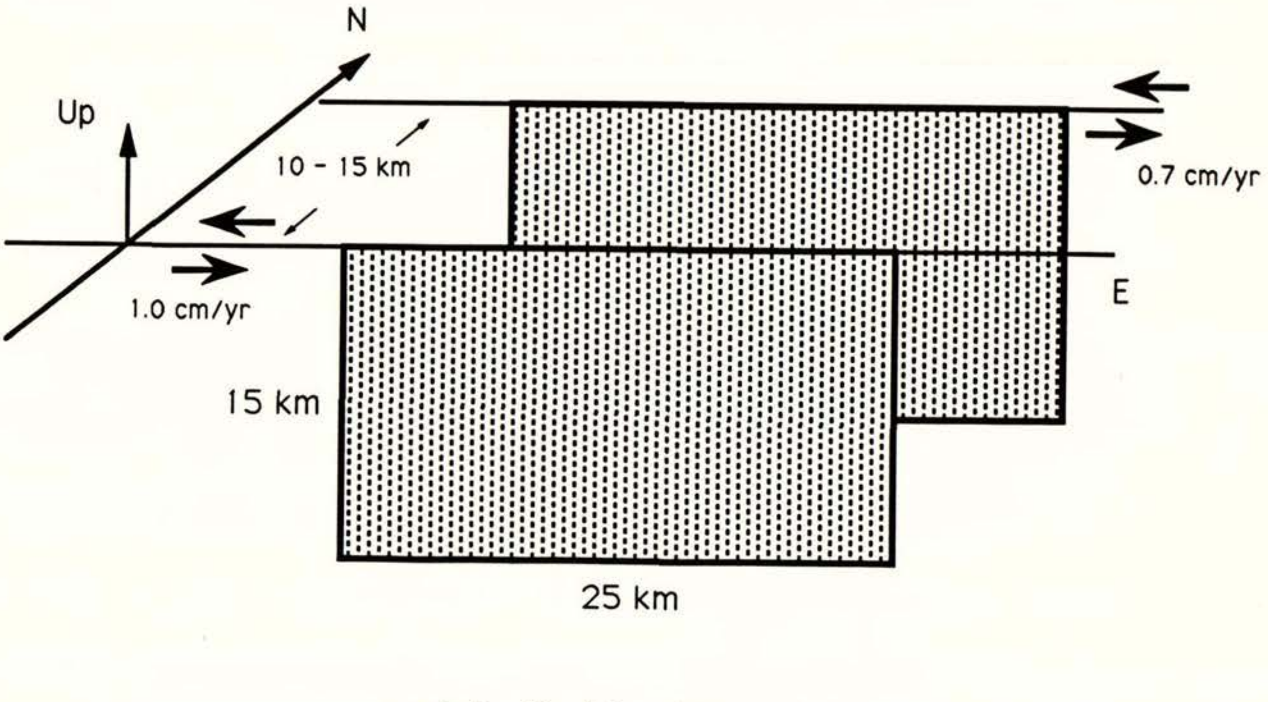

Figure 2. The two-fault model for which synthetic seismicity is calculated. The faults are vertical, 25 x 15 km, separated by 10 km (model A) or 15 km (model B), and driven by aseismic, left-lateral slip along an extension both along strike and downdip of 1 cm/year (fault 1) and 0.7 cm/year (fault 2). The two faults are subdivided into 756 square patches each, 694 m x 714 m in size.

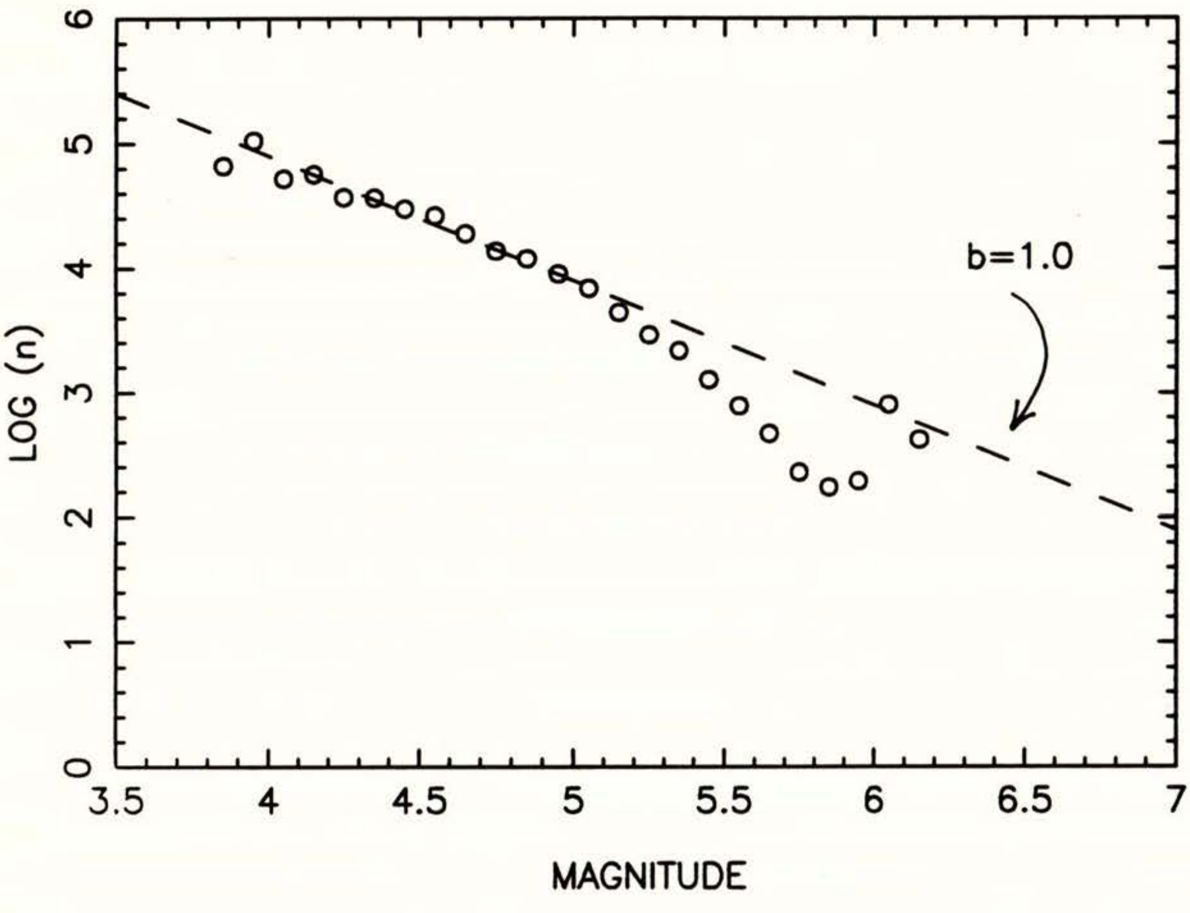

Figure 3. Frequency-magnitude distribution for model B. The total number of events is 475,000; *n* is the number per 0.1 magnitude interval. The dashed line indicates a b value of 1.0. Distributions for the other models are all very similar.

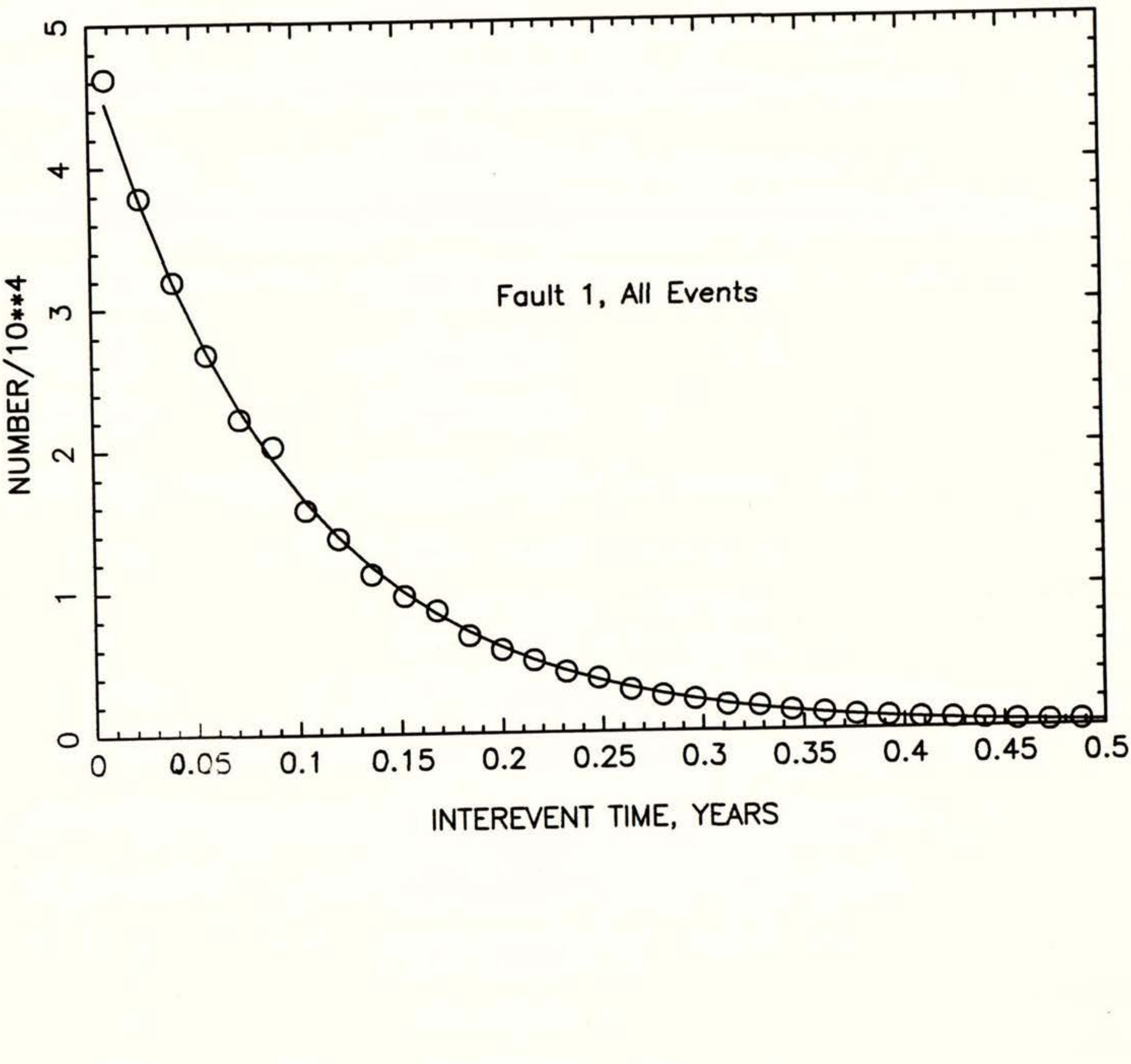

Figure 4. Distribution of interevent times for events on fault 1 of model B (open circles). The line is the distribution for a random (Poisson) process with the same mean and standard deviation. Distributions for other faults/models are all very similar.

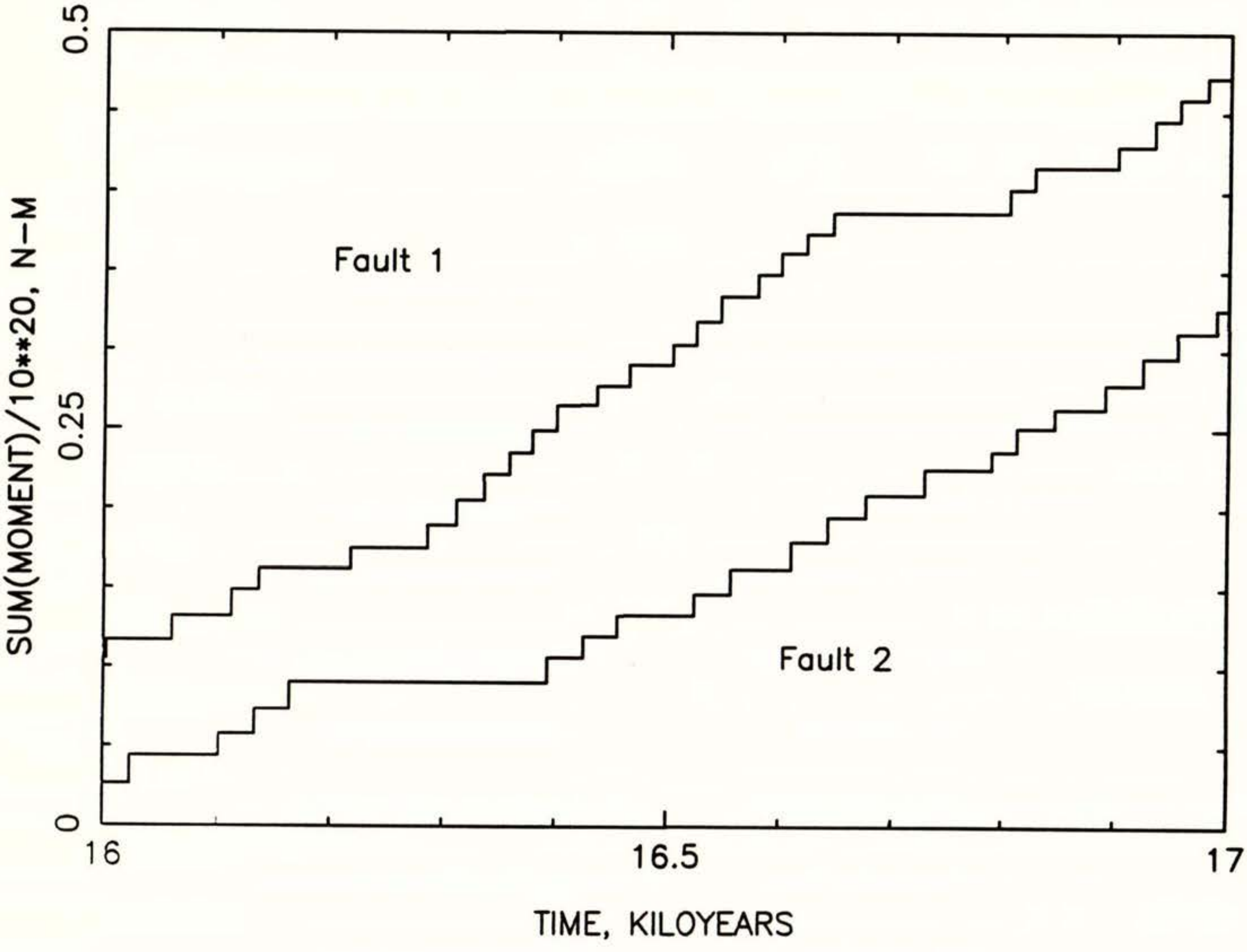

Figure 5. Cumulative moment-release curves for a portion of the synthetic catalog (model B), considering only "characteristic" events of magnitude 6.0 or more.

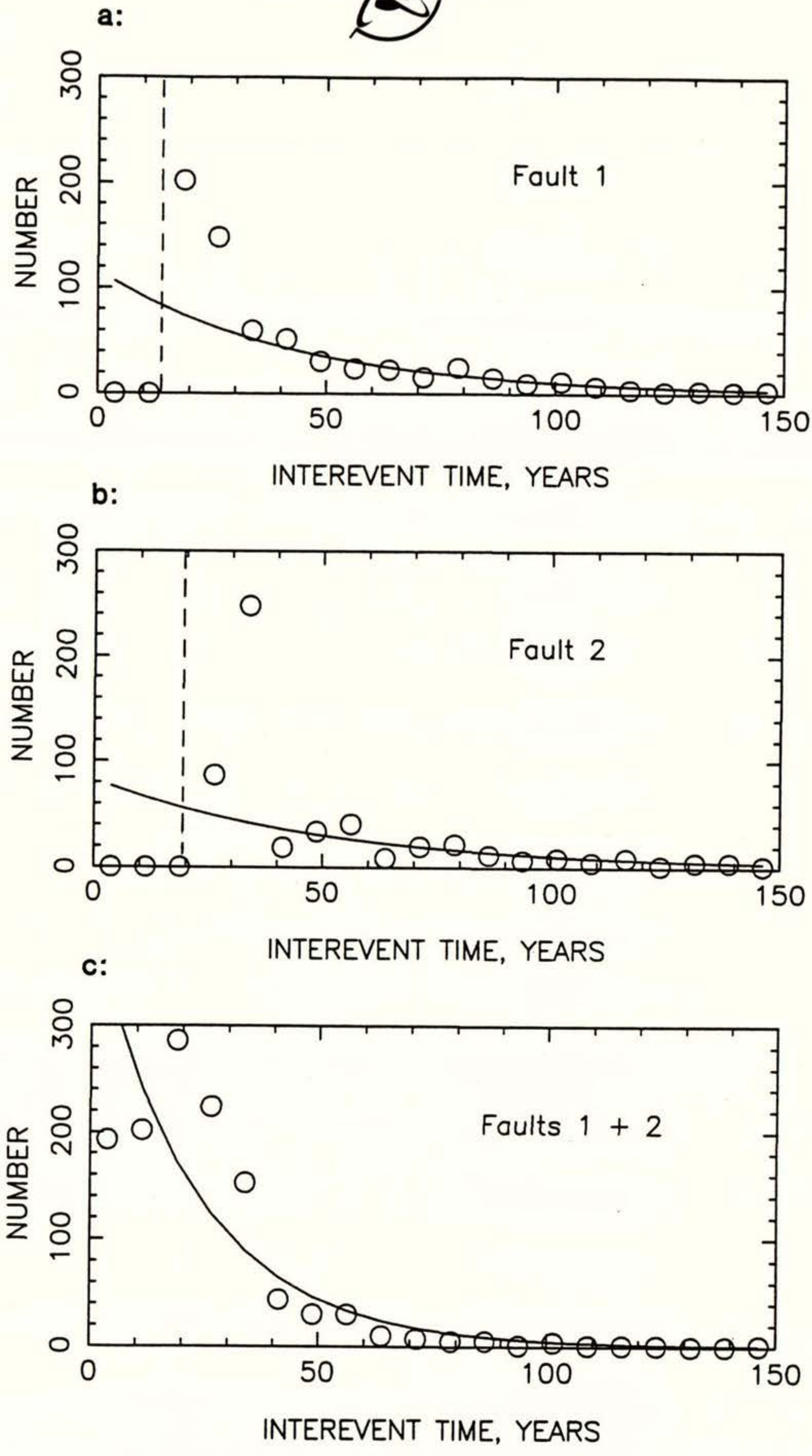

Figure 6. (a) Distribution of interevent times for characteristic events on fault 1, model B. The solid curve is the distribution for a Poisson process with the same mean and standard deviation. The dashed vertical line indicates the "characteristic repeat time" (see text). (b) As in Figure 6a but for fault 2. (c) As in Figure 6a but for faults 1 and 2 considered together.

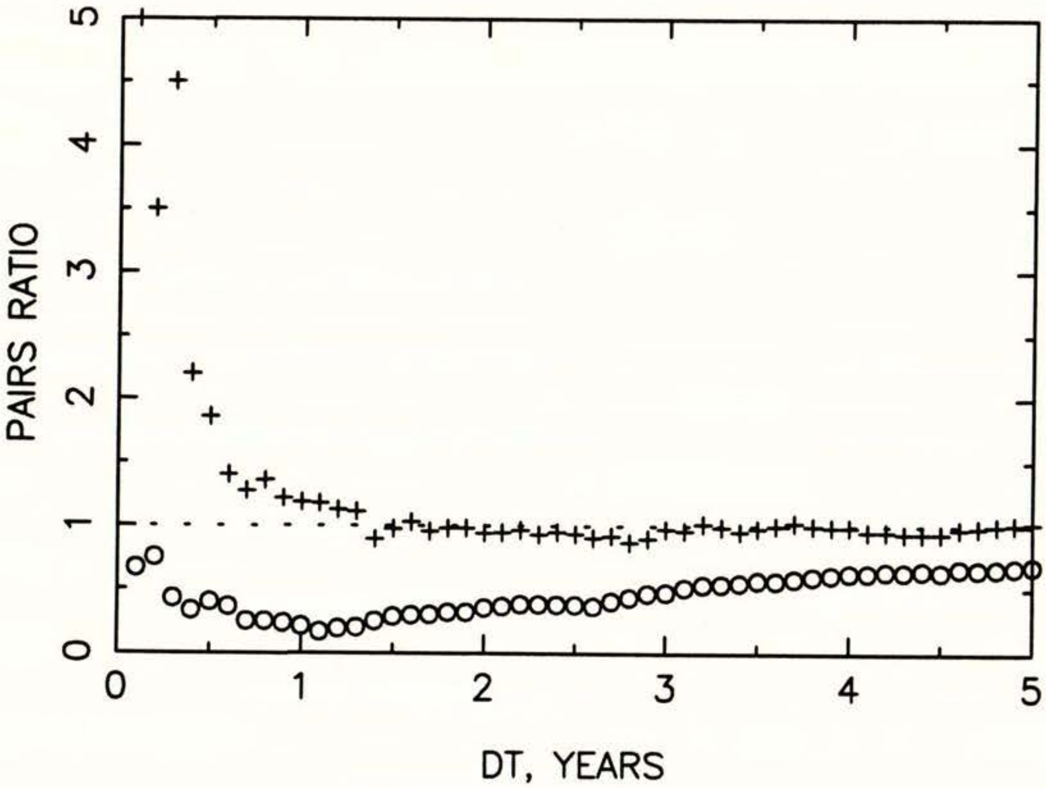

Figure 7. The number of times characteristic events occur within a given time interval, *DT*, for models with interaction, divided by the number for equivalent models with no interaction. Open circles are for model A (fault separation of 10 km); pluses are for model B (fault separation of 15 km). If there were no interaction effects, points would lie on the dotted line.

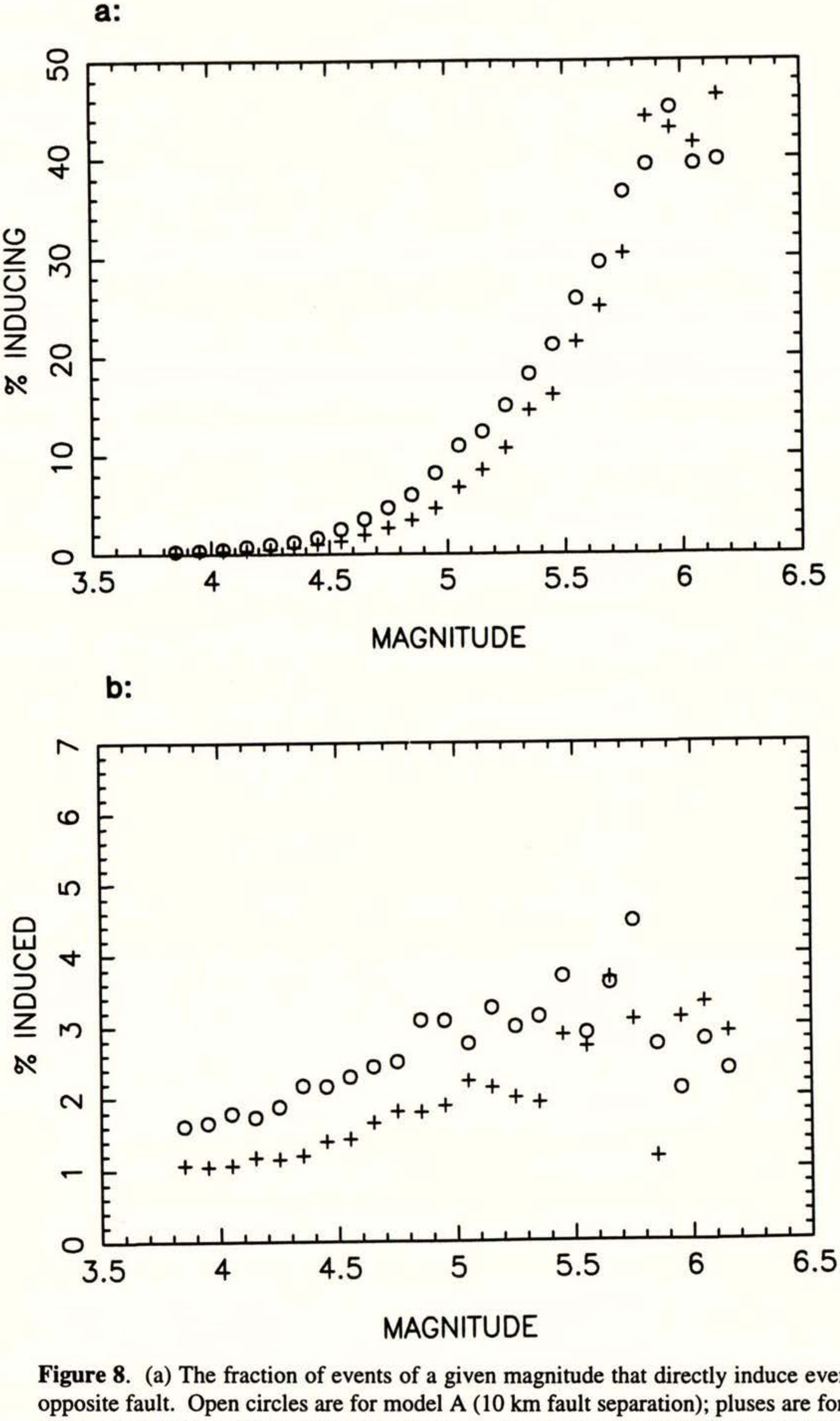

Figure 8. (a) The fraction of events of a given magnitude that directly induce events on the opposite fault. Open circles are for model A (10 km fault separation); pluses are for model B (15 km separation). (b) The fraction of events of a given magnitude that are induced events, again expressed as a percent. Open circles are for model A; pluses are for model B.

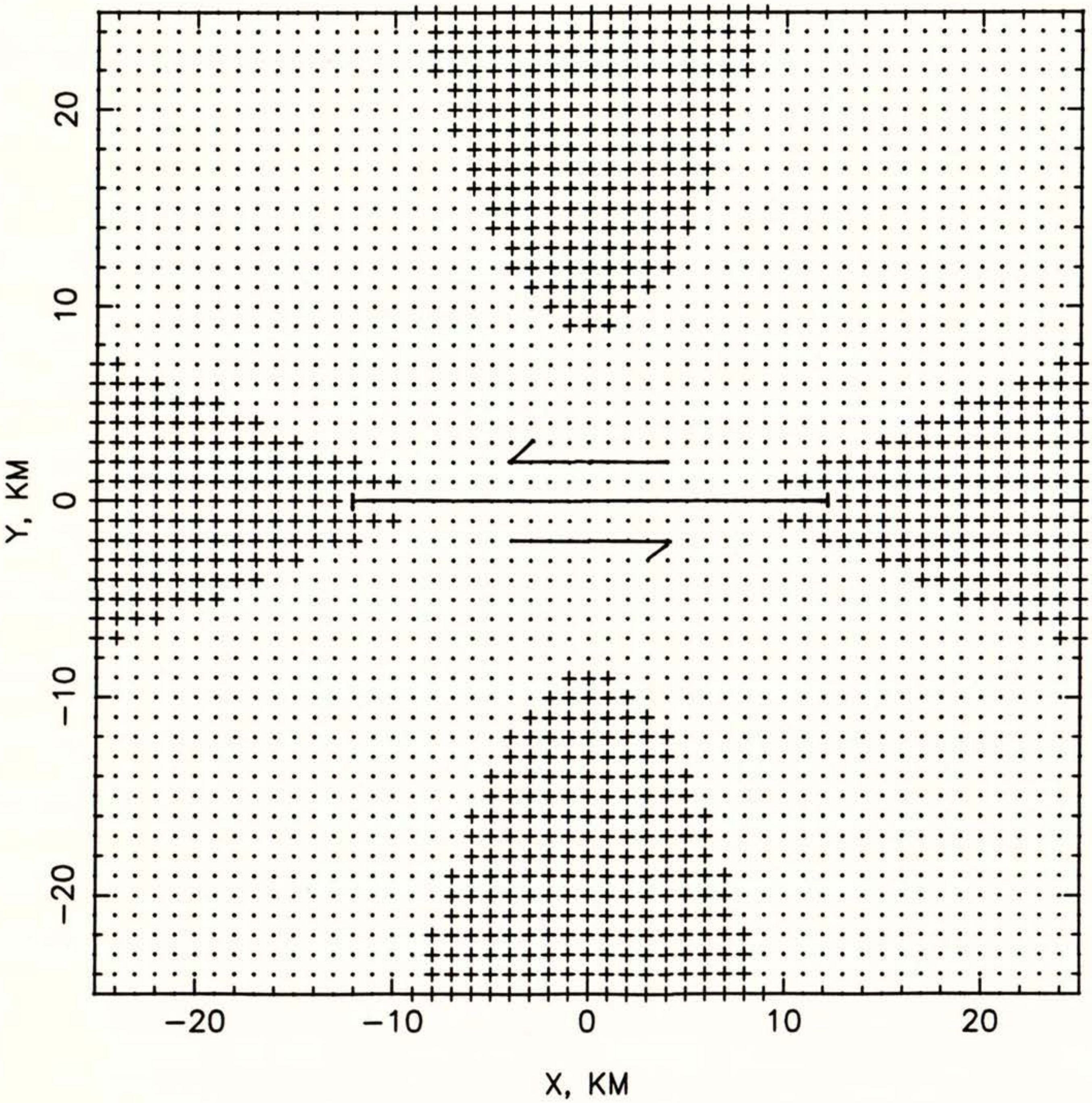

Figure 9. Map view of the induced changes in the Coulomb failure function, δCFF, due to a simplified "typical" characteristic event (as described in the text), for a parallel fault. The depth is 7.5 km. The pluses indicate regions of positive δCFF, while dots indicate regions of negative δCFF. Positive δCFF indicates that slip in the same horizontal sense as the loading mechanism is enhanced.

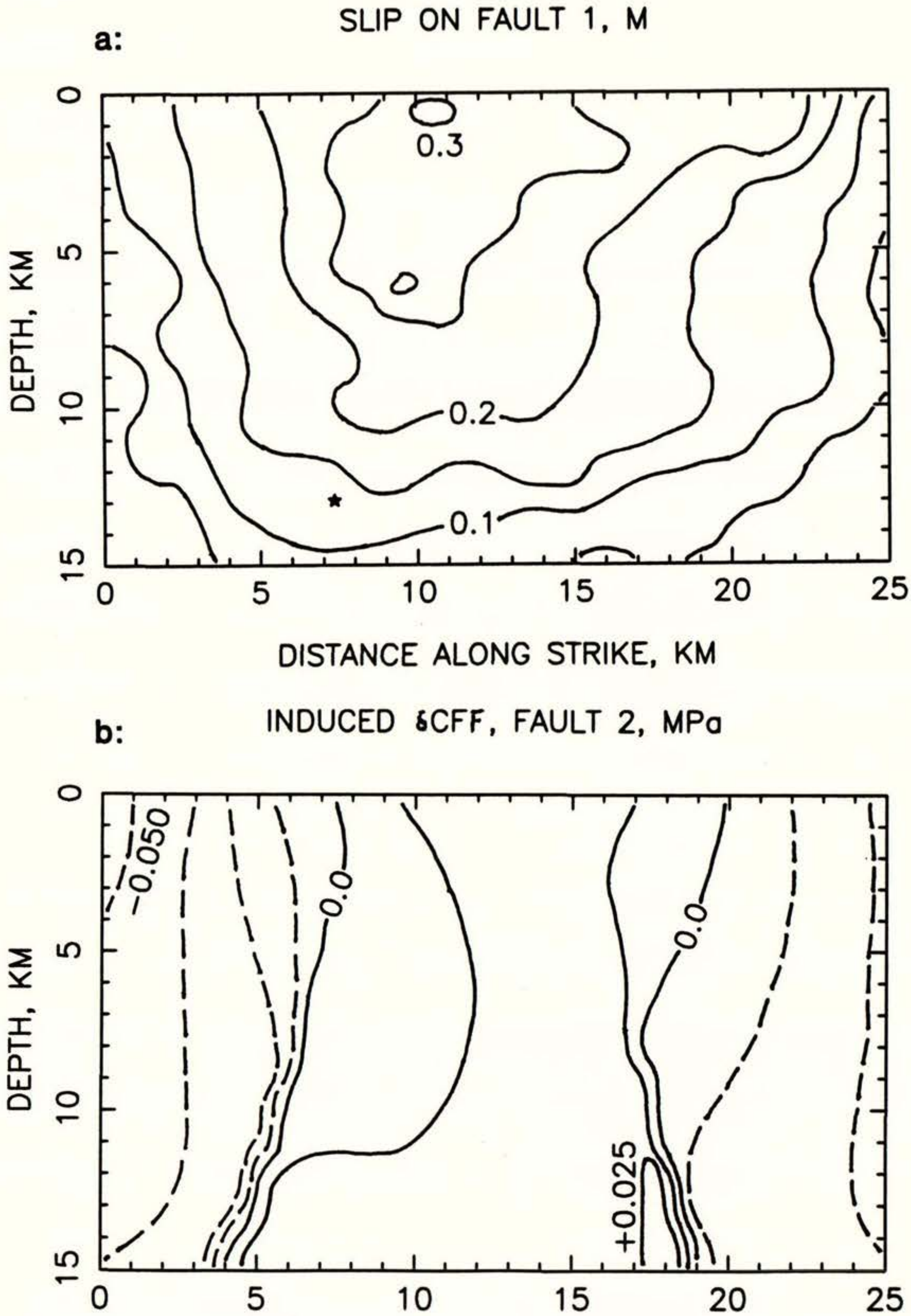

Figure 10. (a) Smoothed slip, in meters, for a particular instance of a characteristic event on fault 1, model B. The contour interval is 0.05 m. The hypocenter of the event (where failure initiated) is shown by the star. (b) The induced δCFF on fault 2 for a separation of 15 km. The contour interval is 0.0125 MPa (0.125 bars), dashed contours for negative values.

APPENDIX II

Models of Fault Interaction for the Wellington Region, New Zealand: Effects on the Temporal Clustering of Large Events

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Abstract

We have applied our previously developed synthetic seismicity model for multiple interacting faults in a 3-D half-space to the Wellington region, New Zealand. The model, which now includes the effect of induced pore pressure changes and static fatigue, generates long synthetic catalogues of earthquakes that can be used to make statistical estimates of the effects of interaction on the seismic hazard for various fault geometries, driving mechanisms and rates, and mechanical properties. The Wellington region is one of oblique plate convergence with the interface between the subducting Pacific plate and the overlying Australian plate at an average depth of about 20 km. Faults included in the model, besides the subduction thrust, are the four major arc-parallel, strike-slip faults overlying the plate interface. We take the driving mechanism to be steady increments in shear stress on each fault. The direction and magnitude of the increments, as well as mechanical properties of the faults, are adjusted to reproduce average recurrence intervals and offsets for large "characteristic" events derived from paleoseismic studies. However, there is no direct evidence for large events on the subduction thrust and a variety of scenarios have been investigated. In all models, the hazard of multiple large events (M 7.2 or more) within a short time period is increased compared to the corresponding cases with no fault interaction. The magnitude of the increase is in proportion to the amount of seismic slip on the plate interface. For the range of models we have examined (0 to 40% coupling) the probability of one or more events of magnitude 7.2 or more in a given year is between 0.00350 and 0.00500, and is independent of the interactions. The probability of 2 or more such events (with interactions) is between 0.00027 and 0.00180; the comparable figure for the case of no interactions is only about 0.00002. While several aspects of our model are uncertain or approximate, we feel that possible clustering of large events due to elastic interactions needs to be considered carefully when estimating seismic hazards.

Introduction

An important factor in seismic hazard analysis is the possible clustering of large events in space and time. Statistical studies of world-wide event catalogues have shown that some degree of clustering is common (e.g., *Kagan and Jackson,* [1991]), as opposed to purely random occurrence (the Poisson model). However, the historical record of earthquakes is too short and inhomogeneous to make firm estimates of how this clustering might affect the seismic hazard in a particular region. Paleoseismic studies can extend the record for some individual faults, but only with large uncertainties.

Recently developed synthetic seismicity models can generate long, homogeneous catalogues of events that can provide some insight into temporal variations in the recurrence time of large events. Such models for single, but segmented, faults have been used to study the Central American subduction zone [*Ward*, 1992a,b], the San Andreas fault in California [*Ward & Goes*, 1993] and the Parkfield region of the San Andreas fault in detail [*Ben-Zion & Rice*, 1993]. We have developed our synthetic seismicity model [*Robinson and Benites*, 1995] with the aim of extending such analyses to the case of multiple faults of any orientation, that interact elastically. Elastic interactions have been suggested as an important factor in the timing of large events in southern California (e.g., *King et al.*, 1994]). *Harris et al.* [1995] have shown that events of magnitude 5 or more, in southern California, tend to preferentially occur on faults favourably loaded by previous events on other faults.

In the present study we apply our synthetic seismicity model to the Wellington region, New Zealand, to determine if fault interactions are likely to be important in estimating earthquake hazards. Specifically, we ask "Do interactions significantly change the risk of multiple large events within a short time compared to what would be calculated assuming independent faults?". The Wellington region (Figures 1,2) is one of oblique plate convergence, with the interface between the subducting Pacific plate and the overlying Australian plate lying at an average depth of about 20 km. Above the interface there are multiple sub-parallel strike-slip faults, at about 5-25 km spacing, that are known to have ruptured in large events in the past [*Van Dissen and Berryman*, 1995], the most recent in 1855. Thus the region is one for which the seismic hazard is quite high, and for which interactions between faults might be important. We also note that much more of the North Island convergent margin is above sea level as compared to most other subduction zones, so that the section of the plate interface most likely to rupture in large events directly underlies population centres rather than lying offshore.

The Synthetic Seismicity Model

Our synthetic seismicity model is described in detail in *Robinson and Benites* [1995], where we also present a simple example of two parallel strike-slip faults. Here we give only a brief description, and indicate how the model has been extended to take account of pore pressure and to generate foreshocks, aftershocks, and swarms as a manifestation of static fatigue.

Each fault in our model is specified by its strike, dip, length, and width, and subdivided into some number (as large as computational constraints allow) of equal sized patches. As well as these geometric parameters each patch is assigned a coefficient of friction, random within a specified range, a percentage stress drop, and a dynamic strength/arrest strength ratio.

Each model must define how the faults are loaded towards failure. This can be done in several different ways, as described in *Robinson and Benites*, [1995]. The loading eventually causes the maximum shear stress (calculated from the two in-plane shear stresses) on some patch to reach the static strength level. Note that the normal stress, and hence strength, may be affected by the loading as well as the shear stress [*Sibson*, 1991]. This patch then fails and slips in the direction of maximum shear stress. The amount of slip is determined by the specified stress drop. This initial slip induces shear and normal stress changes on all other patches on all faults and perhaps results in other patches failing. These induced stress changes are calculated using the formulation of *Okada* [1992]. Patches can slip more than once; once having failed their strength is reduced to a lower dynamic level. A failure episode, which may involve events on more than one fault, is over when all patches are stable. Then all patch strengths are reset to their static level, and the loading mechanism is allowed to resume.

Event moments, and hence magnitudes, are easily calculated from the patch slips and the medium's rigidity. For cases with sufficiently small patch size our model generates a magnitude distribution similar to that for real seismicity (i.e. a "*b* value" close to 1), and also a distinct class of large "characteristic" events that rupture the whole fault and whose numbers are higher than expected from an extrapolation of the frequencies of smaller events. Such behaviour has been shown by *Wesnousky* [1994] to apply to most faults in southern California. Note that "characteristic" in this sense, does not necessarily imply periodic as well.

In our previous work we found that low coefficients of friction (about 0.2) and small stress drops (10%) produced synthetic events that matched observed earthquake rupture area-magnitude-slip statistics. However, laboratory studies of rock friction (e.g., *Byerlee*, [1978]), and some studies of induced earthquakes [*Raleigh et al.*, 1976], find coefficients of 0.6 -0.8. In our previous work the low coefficient of friction should probably be interpreted as reflecting the effect of high pore pressure. In the present model this is made more explicit by including pore pressure in the calculation of patch strength and setting the coefficient of friction close to laboratory values:

$$S_{static}(t) = -\mu \left(L + \delta\tau_{33}(t) + P_{base} + \delta P(t) \right) \quad (1)$$

where S_{static} is the static strength, μ is the coefficient of friction, L is the normal stress due to lithostatic pressure, $\delta\tau_{33}$ is the normal stress due to the loading mechanism and slip on other fault patches, P_{base} is some (unchanging) base pore pressure level (taken as a percentage of the lithostatic pressure), δP is the change in pore pressure due to slip on other patches, t is time, and we have suppressed patch indices. Our sign convention is that tensile stresses are positive, but we retain the common usuage of "pressure" (compression is positive). We assume that the medium is sufficiently permeable that the slow loading mechanism does not cause any changes in pore pressure. The change from our previous formulation

$$S_{static}(t) = -\mu \left(L + \delta\tau_{33}(t) \right) \quad (2)$$

is quite important because it means that for some given change in normal stress the effect on strength is now larger, by about a factor of 4. Thus our previous study probably under estimated the effect of interactions.

The base pore pressure level now becomes an additional variable, which we have taken as constant for all patches on a given fault but variable from fault to fault. As indicated above we now allow induced changes in pore pressure, although we can treat this only approximately. The induced changes are given by

$$\delta P = -\frac{\beta}{3}\delta\Delta \qquad (3)$$

where $\delta\Delta$ is the induced dilation ($\Sigma\tau_{ii}$) and β is Skempton's Coefficient which can range from 0 to 1 depending on pore geometry relative to the stress changes. We assume that all induced changes in pore pressure decay exponentially with a time constant τ_{pore}, which can be different for each fault, and do not solve the equations for flow explicitly.

Another change in our model from our previous work is the introduction of static fatigue. Previously, the induced stresses due to slip on one fault could immediately trigger another event on another fault, resulting in a complex stress interplay if both events continued rupturing. In actuality, as *Scholz* [1990] points out, there is often some delay between inducing and induced events due to static strength fatigue. We have now introduced such a phenomenon into our model by temporally increasing the static strength of a patch subject to an instantaneous increase in shear stress by a small amount. The strength decays exponentially to its normal level with a time constant $\tau_{fatigue}$. This also results in foreshocks, aftershocks, and swarms which were all absent before.

The specification of the driving mechanism and how it alters the stress on a fault can be done in several ways. However, a common problem is that the normal stress, due both to the driving mechanism and induced by slip elsewhere, sometimes accumulates (or decreases) after long periods, so making some patches extremely strong (or weak). Part of the problem is that we do not monitor the stress at points off, or along extensions of, the faults. In the real world such stresses would result in small earthquakes that serve, over time, to keep these stresses within reasonable bounds and, we suspect, keep the normal stresses on the major faults under control. To extend our model to take account of this would be computationally impossible. Thus we take the expedient step of allowing the normal stresses to decay back to the base levels (due to lithostatic and pore pressures) over a period long in comparison to the average time between events.

Application to the Wellington Region

In New Zealand (Figure 1) the Pacific and Australian plates are converging, resulting in the subduction of the former in the North Island and northernmost South Island, continent - continent type collision in the central South Island, and subduction of the Australian plate in the far south [*Walcott*, 1978a]. In the Wellington region (Figure 2), on the southern tip of the North Island, the convergence is at an azimuth of 85° (i.e., about east-west) and velocity of 0.04 m/yr (NUVEL-1A of *De Mets et al.*, [1994]). This convergence has resulted in a well defined Wadati-Benioff zone of earthquakes extending to a depth of about 250 km. The convergence direction is oblique to the strike of the subducted plate and to the strike of all major geologic structures in the region, which are about 40-45^{o} (i.e. northeast-southwest).

A number of sub-parallel, dextral, strike-slip faults in the region (Figure 2) are known to have ruptured in large earthquakes in the past [*Van Dissen and Berryman*, 1995]. The only such event in historical times was on the Wairarapa fault in 1855. Slip during that event extended from the south coast northeastward for about 100 km. The sense of motion was predominantly horizontal and up to 12 m [*Darby and Beanland*, 1992], although there are possible complications near the coast. Paleoseismic studies suggest that such events have occurred repeatedly in the past. In addition to the Wairarapa fault, the Wellington, Ohariu, and Shepard's Gully faults show clear evidence of repeated large events. It is generally assumed that the Wairau fault extends north from the South Island (where it shows clear evidence of large events) to lie 15 to 20 km off the northwest coast. About 60-90% of the total arc-parallel convergence (0.028 m/yr) derived from plate tectonics can be accounted for by seismic slip on these faults [*Van Dissen and Berryman*, 1995]: no evidence of fault creep has been observed on any of them.

The arc-normal convergence that these strike-slip faults fail to account for (the Wairarapa fault takes up a small amount) must be taken up by slip on the subduction interface (seismic and/or aseismic) and/or on shallower, minor thrust faults to the east and offshore. While further north the presence of back arc extension in the Taupo Volcanic Zone [*Darby and Meertens*, 1995] may indicate that the seismic coupling on the shallow interface is low [*Scholz and Campos*, in press], there is no such reason to suspect that that is the case in the Wellington region. However, no large subduction thrust events have occurred in historical times and the paleoseismic record is unclear. There are uplifted coastal terraces along the southeast coast [*Berryman et al.*, 1989] but these are interpreted to be due to movements on imbricate thrust faults offshore and above the plate interface. There is evidence for a number of onshore thrust faults east of Lake Wairarapa [*Cape et al.*, 1990], but little is known of their recent history or length and we do not include any of them in our models. We note that the total theoretical uplift due to subduction thrust events is offset by the interseismic deformation so that there may be little evidence to be found in long-term vertical deformation data for their having occurred.

Instrumentally determined hypocentres in the Wellington region have been most accurately determined since the installation of the Wellington Seismograph Network beginning in 1976; prior to that the sparser National Network could not provide good depth control. *Robinson* [1986] used data from this network to study the seismicity, structure, and tectonics in some detail. Recent events, 1986-1994, when shown in cross-section (Figure 3), clearly show the presence of a northwest dipping band of seismicity, interpreted to lie within the subducted Pacific plate. The position of the plate interface is taken as lying along the top of this band, and can be defined on the basis of a change in focal mechanism when that is unclear. There have been few, if any, thrust events on the interface. The mechanism of events internal to the Pacific plate are predominantly dip-slip with a tension axis aligned down the slab dip, and so reflect "slab-pull". The epicentres of events above the plate interface (Figure 4) do not delineate the major faults, and their mechanisms are a mix of thrusting and strike-slip, reflecting east-west compression. Although this pattern of shallow epicentres may be transitory, the inference is that the major faults slip primarily during large characteristic events.

We take the area we wish to investigate as about 120 x 100 km in size, as shown in Figure 5. However, as described in our previous work, computational restraints place severe limits on the number of faults we can include in our model, and the number of patches into which each fault can be divided. For our present purpose, investigating the effect of elastic interactions on the temporal distribution of large events, we do not need to generate low magnitude events and so can take the individual patch size relatively large. However, we cannot take them too large (e.g., each fault composed of one patch) since we wish slip to be able to initiate at a large number of positions on each fault and also because details of the slip distribution become important for closely spaced faults. If we include only four major, shallow, strike-slip faults (with a 20 km depth extent and taking the close Ohariu and Shepard's Gully faults as one "Ohariu" fault), plus the subduction thrust, then a patch size of 5 x 5 km makes it possible to investigate a number of different variations in physical properties in a reasonable time. For example, a run that generates 10,000 events (magnitude 6.0 to 8.5) takes about 5 hours of computation time on the most appropriate machine available to us (Digital Equipment Corporation AXP 3000-300 with about 150 MBytes of available memory).

We approximate each fault with a single plane (Figure 5; Table 1). While the dip of the subduction thrust is well known from the seismicity, the dip of the overlying faults is uncertain because they are not defined by small events. From the appearance on the surface, they are usually taken as near vertical [*Van Dissen*, pers. comm.]. However, *Darby and Beanland* [1992] attempted to model the (limited) data on surface deformation due to the 1855 Wairarapa earthquake and suggested that a listric fault was possible, the dip becoming progressively more shallow at depth and joining the subduction interface. The models we have investigated all assume a vertical geometry for the two western most faults (Wairau and Ohariu), an 80° NW dip for the Wellington fault, and a 70° NW dip for the Wairarapa fault. In all cases the width has been taken as 20 km.

Because we cannot model the entire world, we have to impose a length limit on the faults, trying to set those limits to correspond to recognised fault segments. There is strong evidence that both the plate interface and overlying faults are disrupted in Cook Strait [*Robinson*, 1986; *Carter et al.*, 1988] so that we have imposed a southwest limit 15 to 20 km offshore (Figure 4), except for the Ohariu fault which terminates, as shown, onshore. Rupture on the Wairarapa fault in 1855 extended as far northeast as Mauriceville, giving in our model a length of 120 km. The Wellington fault has a 2 km right side-step near Kaitoke, at which the strike also changes significantly. Thus we have modelled the Wellington fault with two separate faults: the southern segment being 70 km long, and the northern being 52 km long so that its northeast extent matches that of the Wairarapa fault (there is no information available from geologic mapping re possible segment length for this northern part of the fault). The Ohariu fault can be traced as far northeast as Waikanae, giving a total length of 60 km in our model. Segmentation of the Wairau fault is unknown, and we have taken it simply as one single segment 100 km long.

In all models that we have considered the length of the subduction thrust has also been taken as 120 km (Figure 5), to match the length of the Wairarapa fault. *Tichelaar and Ruff* [1993] investigated the apparent maximum depth of seismic slip on subduction interfaces worldwide and found a typical maximum depth of 40 +/- 5 km. They also noted that the depth limit often matched the maximum depth of small events in the overlying plate. We have taken the maximum depth in our models as 35 km, which roughly corresponds to the limit of concentrated activity within the subducted plate and the maximum depth of overlying events. We have taken the upper limit as corresponding to the up-dip extent of that concentrated activity, i.e. about 18 km: this gives a total width of 75 km. We assume that slip above and below those limits is aseismic, in accord with the generic subduction zone model of *Byrne et al.* [1988]. Conceptually this makes sense, as the tensional mechanisms of the concentrated events in the subducted plate would then indicate that the slab pull force is being most strongly resisted along that region of the interface.

It would be appropriate to specify a driving mechanism based on our knowledge of plate tectonics. However, the stress field in a region of oblique convergence is difficult to envisage. Presumably the stress on the shallow strike-slip faults could be taken as due to uniaxial compression in the direction of convergence, but the situation for the plate interface is more complex. Immediately below the interface the focal mechanisms of small events indicate a tensional environment due to slab pull. Eventually we hope to use finite element modeling to help derive the driving mechanism, but for the moment we have simply specified it, for each fault separately, as due to steady increments in shear stress. For the shallow strike-slip faults the driving rates have been set so as to reproduce the long-term slip rate as derived from paleoseismic data and the direction specified as along strike. For the plate interface, we have investigated four models: (1) no seismic slip at all; (2) shear stress increments in the down dip (arc normal) direction; (3) shear stress increments in the direction of plate convergence; and (4) shear stress increments in a direction intermediate to the above two cases. For model 2 we have varied the rate of stress accumulation such that 10%, 20%, or 40% (models 2a,2b,2c) of the arc-normal convergence is taken up seismically. For model 3 we have set the rate such that 10% of the arc-normal convergence is taken up seismically. For model 4 we have set the rate such that 20% of the arc-normal slip is accommodated. The driving parameters of all the models considered are given in Table 2.

Mechanical fault properties (Table 3) have been adjusted, by trial-and-error, so that most of the slip occurs in large characteristic events. For faults considered in isolation this behaviour is easy to obtain. However, when there are interactions between faults it becomes more difficult to obtain this characteristic behaviour since interaction leads to heterogeneous stress fields which do not favour large events. The best results are obtained with a fairly uniform coefficient of friction and low dynamic strength. The large characteristic events that result usually have a slip maximum about half way along the length of the fault, tapering by about 1/2 at the ends. For the strike-slip faults the maximum is near the surface; for the subduction thrust the maximum is about 2/3 of the way updip. The characteristic event parameters, and equivalents from paleoseismic data, are shown in Table 4. We note that, because of interactions, the long-term slip of the Wairarapa fault includes a small (10% or less) thrust component even though the driving stress increments do not include any such component. There are smaller thrust components on the other strike-slip faults (northwest side up for those with vertical dips), except for the Wairau fault which has a small normal component (northwest side down).

We present our results, for each individual model and as averages of the 6 models, in terms of year histograms of interevent times for events of magnitude 7.2 or more for a period of 50,000 years (Figure 6, Table 5). We chose 7.2 because that is the magnitude of a characteristic event on the smallest fault in our models (Wellington North). The definition of the characteristic event magnitude is very clear for the subduction thrust, Wairarapa, and Wairau faults in that there is a large gap between the magnitude of the large events and the lesser events. For the other three faults the distinction is less clear, although they all exhibit a "low" in frequency-magnitude plots above which a larger number of characteristic events occur.

Characteristic events on any fault would produce a Modified Mercali Intensity of 8 to 10 at the centre of Wellington City (neglecting site effects) if we use the formula of *Dowrick* [1991] and take his "centroid distance" as the distance to the point of maximum slip. An intensity of 8 is about equivalent to a peak ground acceleration of 0.25 g [*McVerry et al.*, 1993] and implies significant structural damage, liquefaction at susceptible sites, and possible general panic [*Smith et al.*, 1992].

Discussion

Our results (Table 5 and Figure 6) indicate that, for the Wellington region, there is a strong enhancement in the risk of multiple large events within a short time due to fault interactions, as compared to the corresponding cases with no fault interaction. The magnitude of the increase is in proportion to the amount of seismic slip on the plate interface. For the case in which there is no seismic slip on the plate interface the probability of having two large events within one year of each other (0.00027 per year) is 18 times that for the case of no interaction. For the case of 40% coupling, the increase in probability (to 0.00180) is by a factor of 93. To look at it a different way, for the case with least enhancement (model 1, no subduction thrust) there is a 7.6% chance that, given a characteristic event on one fault has occurred, the next characteristic event, on a different fault, will occur within 1 year. This compares to a 0.4% chance if interactions are suppressed. However, about half (52%) of the cases of short interevent times in this model involve interaction between the Wellington North and Wellington South faults, and half (50%) of those occur instantaneously and would not normally be classed as separate events. The same is true for the other models, where it is also likely that synchronous events on the Wairarapa fault and subduction thrust would not be recognised as two events.

If our calculations of the risk of multiple large events is correct, then the subduction thrust plays a very important role and it becomes important to determine its mechanical properties. As mentioned above, there is no direct paleoseismic evidence for large subduction thrust events in the region, and none have occurred in historic times. *Walcott* [1978b] interpreted repeated triangulation surveys in the Wellington region as indicating a locked interface since about 1920 but unlocked before that (1855-1920); thus our models may need some temporal variations in properties added to them. *Darby and Beanland* [1992], in three study of surface deformation in the 1855 Wairarapa earthquake, suggested that there could have been slip on the subduction interface northwest of the Wairarapa fault, but not to the southeast. A shorter locked width for the subduction thrust in our models would decrease the interaction effects somewhat. We hope that recent geodetic measurements in the region using GPS technology (*D.Darby*, pers. comm.) will soon be able to shed more light on these important questions. In the mean time, we feel that model 4 is probably the most realistic because: (1) *Pacheco et al.* [1993] found the "coupling coefficient" for subduction zones world-wide to vary widely, with an average of about 25%; (2) the slip vectors of large thrust events in other oblique convergence subduction zones are usually rotated towards the arc-normal direction, away from the plate tectonic direction [*Fitch*, 1972; *McCaffrey*, 1992].

We can understand why large subduction thrust events in this model might enhance the chance of a large event on the overlying strike-slip faults by examining the induced stress in more detail. For example, we can examine the induced stresses on the Wairarapa fault due to a characteristic event on the subduction thrust, in terms of a change in the "Coulomb Failure Function" (δCFF) [*Reasenberg and Simpson*, 1992], defined in our models as

$$\delta CFF = \delta\tau_{shear} + \mu\,(\delta\tau_{33} + \delta P) \tag{4}$$

where we take τ_{shear} as positive if it favours slip in a dextral sense (as observed for large events on the Wairarapa fault)..For a typical subduction thrust event the maximum δCFF is +3.50 MPa (the average is 1.2 MPa), and is uniformly positive except for the extreme shallow/southwest corner of the fault. The maximum is about 8% of the average fault strength of 42.5 MPa. Since the 33% stress drop following a past large Wairarapa event would have left the shear stress still fairly high (30.0 MPa) the fault would be potentially "triggerable" by the induced stresses after about 1050 years (about 3/4 of the characteristic repeat time). The other strike-slip faults in our models have lower stress drops and so would be even more susceptible. Because we have tuned our fault parameters to produce characteristic events, the failure of any patch has a reasonable chance of cascading into full fault rupture.

In general, a large event on a vertical strike-slip fault will enhance the chances of another event along strike (as we observe for the Wellington North and Wellington South faults). They will inhibit similar events on nearby parallel faults (offset laterally) out to a distance of about half the rupture length; the effect is the opposite, but weaker, at greater distances. But the effect is sensitive to details of the slip distribution relative to the geometries of the other faults. All the interactions make it very difficult to envisage ahead of time what the end result in a situation with multiple faults may be. We also note that induced shear stresses need not be in the direction of final slip in order to trigger failure; a change in fault normal stress may do the trick.

Given the clustering of large events that our models imply, it is interesting to look at the record of historic seismicity in New Zealand for possible examples. The largest earthquake to occur along the North Island's convergent margin, except for the 1855 Wairarapa event, was the 1931 Hawke's Bay event, Ms 7.8. Surface deformation due to this event was interpreted by *Haines* [1988] to result from mixed strike-slip and thrusting on a northwest dipping fault above the plate interface, although the data do not tightly constrain the result and permit other interpretations [*Ishibashi*, 1987]. The next largest event was the 1934 Pahiatua event about 125 km to the southwest, Ms 7.6. The nature of faulting in that event is unknown: no surface rupture was identified so it is perhaps a candidate for a subduction thrust event. The distance between these events is based on the poorly determined hypocentre for the 1934 event, and given the likely length of its fault plane the effective distance might have been less. In any event we feel that the occurrence of these two large events so close in time and distance is probably more than coincidence. Interestingly, the largest historic event in the South Island, the 1929 Ms 7.8 Buller event, occurred 19 months before the Hawke's Bay earthquake and was itself preceded by 3 months with the Arthur's Pass event, Ms7.1, 120 km to the south. The period 1929-1934 was very busy indeed in New Zealand, and may indicate some mechanism at work in addition to elastic interactions. There are also examples of pairs of smaller events, the most relevant being two events in the Wairarapa region (east of the Wairarapa fault), Ms 7.2 and 7.0, with nearly identical epicentres but one above the plate interface and the other below, separated in time by 37 days (*Webb*, 1989; *Robinson* [1994] describes a similar pair of events further north, Mw 6.2 and 6.4, in 1990).

It can be seen from Table 5 that the number of large events, normalised to an equal time interval, is somewhat less for the models with interaction than for the case with no interaction. This is due to the fact that a larger proportion of the moment release occurs in non-characteristic events in the former case. In all models we observe that one effect of interactions is to make the stresses on all faults more heterogeneous in both magnitude and direction. This makes it much more difficult for slip to extend into large characteristic events. Indeed, we have had to adopt almost extreme mechanical properties (almost uniform friction, low dynamic strength, and highish stress drops) to obtain a characteristic event behaviour for all the faults and to reproduce the observed displacements. It may be that relaxation of these requirements would decrease the amount of clustering but only at the cost of disregarding the paleoseismic evidence. Lowering the stress drops, now about 9.0 to 14.8 MPa (90 to 148 bars), would decrease the induced stresses but also leave the faults closer to failure.

When elastic interactions in our models are suppressed the faults behave in the ideal characteristic way: large events of very similar slip at almost periodic intervals. With interactions, the characteristic events are still quite similar, but the time intervals between them become much more variable. For example, the return time of large events on the Wellington South fault in model 3 has a mean of 703 years with a standard deviation of 145 (404 minimum, 1350 maximum), compared to a mean of 797 years and standard deviation of 0.002 (i.e. nil) for the case with interactions turned off. (Figure 7). Even if our estimates of the increase in risk of short interevent times (considering all faults) turn out to be too large, we believe that this increased variability in repeat times for an individual fault has important implications for hazard estimation.

All our models have used the same static fatigue decay parameter, $\tau_{fatigue}$, of 10 days. We do not have a good feel for how realistic this is, although the observed duration of aftershock sequences could potentially provide some estimate. It could be that a much longer decay time is called for. As it is the value of 10 days does not really affect the histograms of interevent times very much. The amount of strength lost in our models is likewise quite small (1%): increasing this would decrease the number of instantaneous multiple events which now ranges from 30 to 80% of the 0-1 year interval bracket. However, it would also make it more difficult for single patch failures to cascade into characteristic events. Since we are trying to reproduce the characteristic behaviour inferred from paleoseismic data, we have chosen to keep it small. Likewise, the pore pressure decay time, τ_{pore} , of one year has been set in an *ad hoc* manner (and is the reason for our using one year as an indication of "short time"). Although induced changes in pore pressure do not have a large effect, changing the decay time would alter the details of the distribution of interevent times at the low end.

Concluding Remarks

Our results should not be used to estimate the seismic hazard at any particular location in the Wellington region because of the important contribution to that hazard from smaller, but close, events not included in our model. Also, it is quite likely that events of about magnitude 7.2 could be generated by faults we have not considered, in particular shallow thrust faults to the southeast and offshore, and normal faults in the interior of the subducted plate. We re-iterate that our purpose has been to determine if fault interactions in this region might be important in determining time-variable hazard: we think the answer is "YES". But work remains to be done in extending our work before we can provide tightly constrained quantitative results. In particular we eventually hope to: (1) include smaller faults; (2) make the driving mechanism more realistic (based on finite element modeling of oblique convergence and including interactions with large events); (3) make the patch size smaller so that possible precursory seismicity patterns can be investigated; (4) investigate the effect of quasi-periodic large events on extensions of the faults outside the Wellington region; (5) investigate more rigorously the effect of different material properties; and (6) convert the results into return times for both intensity and peak ground acceleration at specific sites so that estimates of dollar losses can be made. Given all that remains to be done our synthetic catalogues certainly cannot be used for specific predictions of where and when future events will occur.

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Table 1: Geometry of the Synthetic Faults

Fault	Strike, Deg.	Dip, Deg.	Length, km	Width, km	Num. Patches
1: Subduction Thrust	45.0	13.5 NW	120	75	24 x 15
2: Wairarapa	45.0	70.0 NW	120	20	24 x 4
3: Wellington South	56.5	80.0 NW	70	20	14 x 4
4: Wellington North	42.0	80.0 NW	52	20	10 x 4
5: Ohariu	36.5	90.0	60	20	12 x 4
6: Wairau	45.0	90.0	100	20	20 x 4

Table 2: Driving Mechanism Models

All Models:

	Along Strike Stress Increment (Pa/Sec)	Along Dip Stress Increment (Pa/Sec)
Fault 2	-2.7E-04	0.0
3	-3.0E-04	0.0
4	-3.0E-04	0.0
5	-1.7E-04	0.0
6	-1.5E-04	0.0

....

Model 1: No Subduction Thrust: Results in about 65% of the arc-parallel plate tectonic rate; 0% of the arc-normal rate.

....

Model 2:

A: Results in about 65% of the arc-parallel plate tectonic rate; 10% of the arc-normal rate.

Fault 1	0.0	0.5E-04

B: Results in about 65% of the arc-parallel plate tectonic rate; 20% of the arc-normal rate.

Fault 1	0.0	1.0E-04

C: Results in about 65% of the arc-parallel plate tectonic rate; 40% of the arc-normal rate.

Fault 1	0.0	2.0E-04

....

Model 3: Stress Increments on the Subduction Thrust in plate tectonic direction, resulting in about 78% of the arc-parallel rate and about 10% of the arc-normal rate.

Fault 1	-0.5E-04	0.5E-04

....

Model 4: Stress Increments on the Subduction Thrust in an intermediate direction, resulting in about 78% of the arc-parallel rate and about 20% of the arc-normal rate.

Fault 1	-0.5E-04	1.0E-04

Table 3: Mechanical Properties

Fault	Coeff. of Friction	Base Pore P./ Lithostatic P.	Static Stress Drop	Dynamic Strength/ Arrest Strength
1: Subduction Thrust	0.70/0.72	0.90	20%	1.04
2: Wairarapa	0.80/0.81	0.80	33%	1.03
3: Wellington South	0.80/0.81	0.80	20%	1.05
4: Wellington North	0.80/0.81	0.80	20%	1.05
5: Ohariu	0.80/0.81	0.80	25%	1.04
6: Wairau	0.80/0.81	0.80	25%	1.04

All: Tpore = 1 year; Skempton's Coefficient = 0.5; Tfatigue = 10 days;

Medium: The two Lame constants = 3.18 x 10**4 MPa
Density = 2.65 x 10**3 kg/m**3
(the above are equivalent to Vp = 6.5 km/s, Vs = 3.46 km/s, and a Poisson't ration of 0.25).

Table 4: Average Synthetic Results and Comparison to Paleoseismic Data

Fault		Average Characteristic Magnitude	Average Mean dT years	Average Min. dT years	Average Max. dT years	Average Max. Slip, m H	V
1: Subduction Thrust	(Synth.)	8.05	1381	1009	1760	0/5	4/11
2: Wairarapa	(Synth.)	7.90	1316	934	1573	10.3	0.9
	(Paleos.)	8.15	1160-1880			12.1	
3: Wellington South	(Synth.)	7.35	735	366	1661	4.4	0.2
	(Paleos.)	7.60	500-770			4.2	
4: Wellington North	(Synth.)	7.30	1134	395	4371	4.0	0.4
5: Ohariu	(Synth.)	7.30	4602	2327	8195	4.2	0.3
	(Paleos.)	7.60	1500-5000			4	
6: Wairau	(Synth.)	7.65	2336	2102	2485	7.6	-0.1
	(Paleos., South Is.)	7.60	1000-2300			6	

Slip for the Subduction Thrust depends on the model driving rate.
Paleoseismic data from VanDissen and Berryman, 1995

Table 5: Distribution of Interevent Times (dT) for Events of Magnitude 7.2 or more, normalized to 50,000 years.

Model 1;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	13.47	0.76
1 - 3	(1.04)	1.27
3 - 10	(2.07)	4.32
10 - 30	6.74	12.24
30 - 100	18.13	38.72
100 - 300	60.10	82.69
300 - 1000	75.65	67.61
>1000	0.00	0.00

Model 2a;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	49.25	0.71
1 - 3	(0.95)	1.51
3 - 10	(0.95)	5.17
10 - 30	(2.84)	15.74
30 - 100	13.26	42.81
100 - 300	49.25	86.33
300 - 1000	75.77	66.38
>1000	(2.84)	0.00

Model 2b;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	45.31	0.75
1 - 3	(0.65)	1.82
3 - 10	(1.94)	5.93
10 - 30	7.77	16.63
30 - 100	18.12	48.69
100 - 300	64.72	89.68
300 - 1000	72.49	64.88
>1000	0.00	0.00

Model 2c;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	89.92	0.97
1 - 3	(1.75)	1.78
3 - 10	(0.58)	7.59
10 - 30	(1.75)	19.67
30 - 100	12.85	57.13
100 - 300	67.74	100.80
300 - 1000	77.08	60.65
>1000	0.00	0.00

Model 3;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	49.46	0.62
1 - 3	(0.44)	1.23
3 - 10	(1.32)	4.47
10 - 30	5.30	12.52
30 - 100	26.50	41.90
100 - 300	69.78	81.26
300 - 1000	71.54	57.99
>1000	0.00	0.00

Model 4;

dT, years	Number (with interaction)	Number (without interaction)
0 - 1	60.75	0.81
1 - 3	(0.47)	1.42
3 - 10	(1.40)	5.88
10 - 30	(4.67)	14.94
30 - 100	14.02	48.58
100 - 300	67.29	95.83
300 - 1000	76.64	64.14
>1000	0.00	0.00

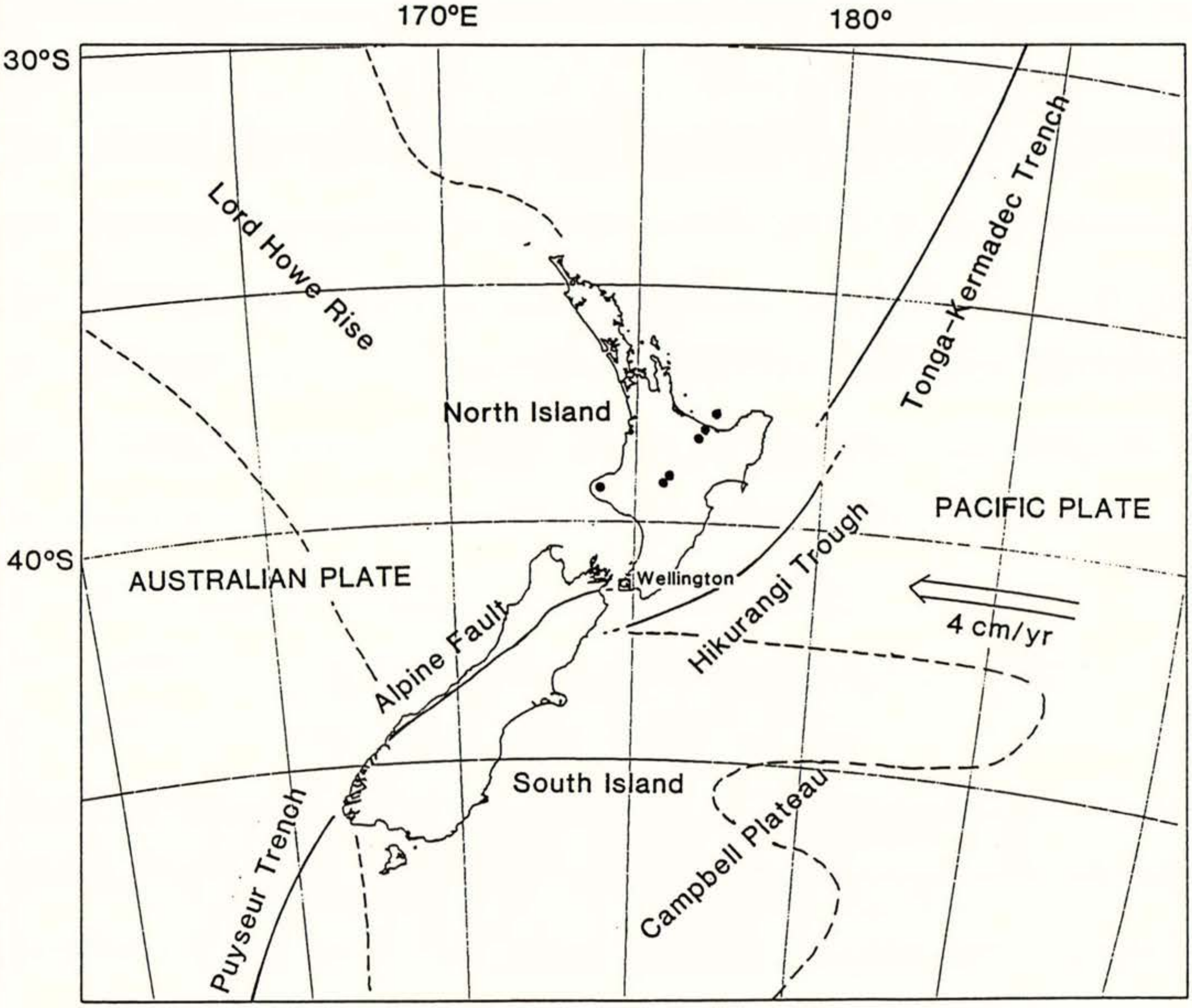


Figure 1: The New Zealand region and the major tectonic features. Dots are andesite volcanoes. The Wellington region lies at the Southern tip of the North Island

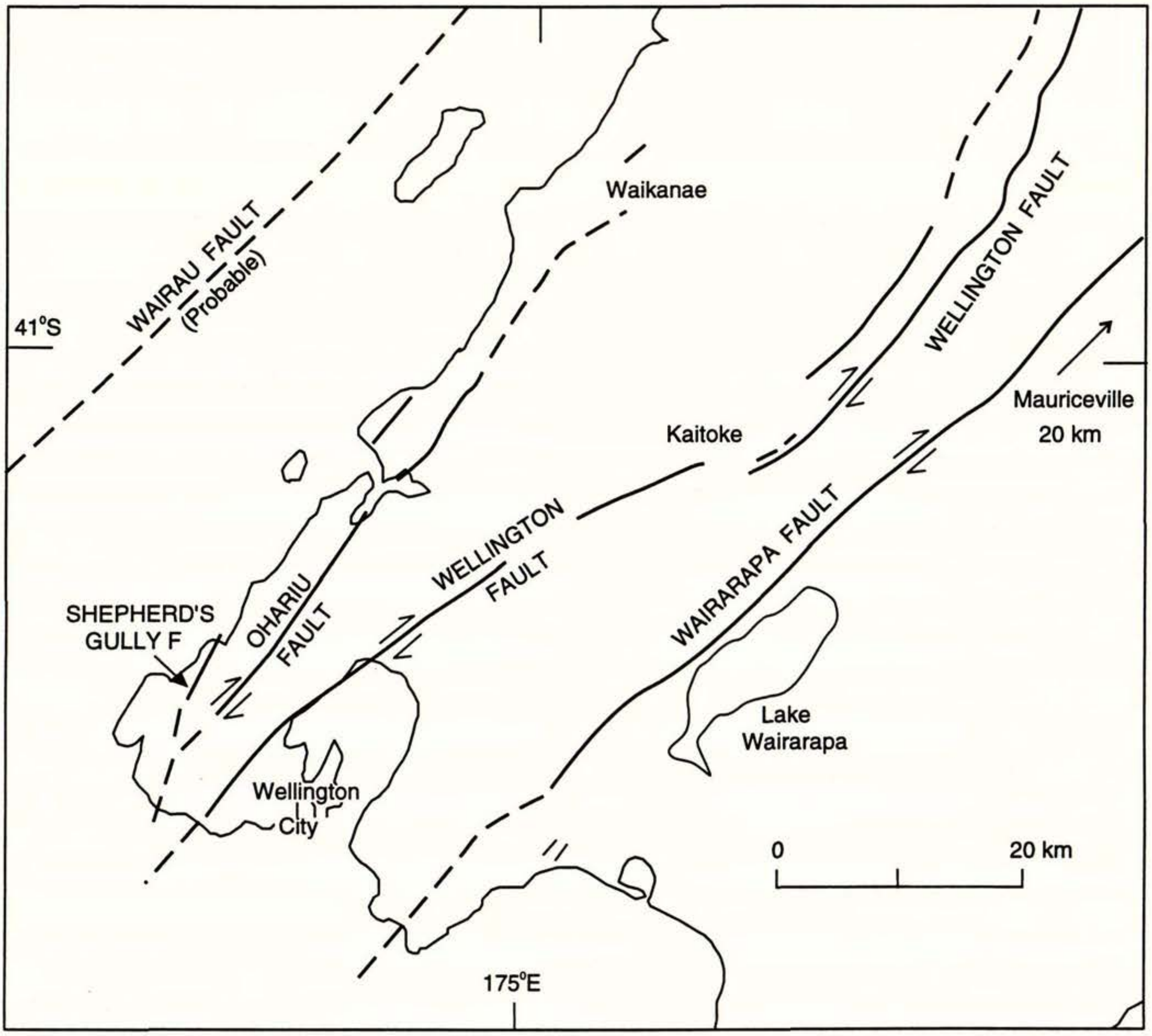


Figure 2: Major active faults of the Wellington region, from Van Dissen and Berryman (1995).

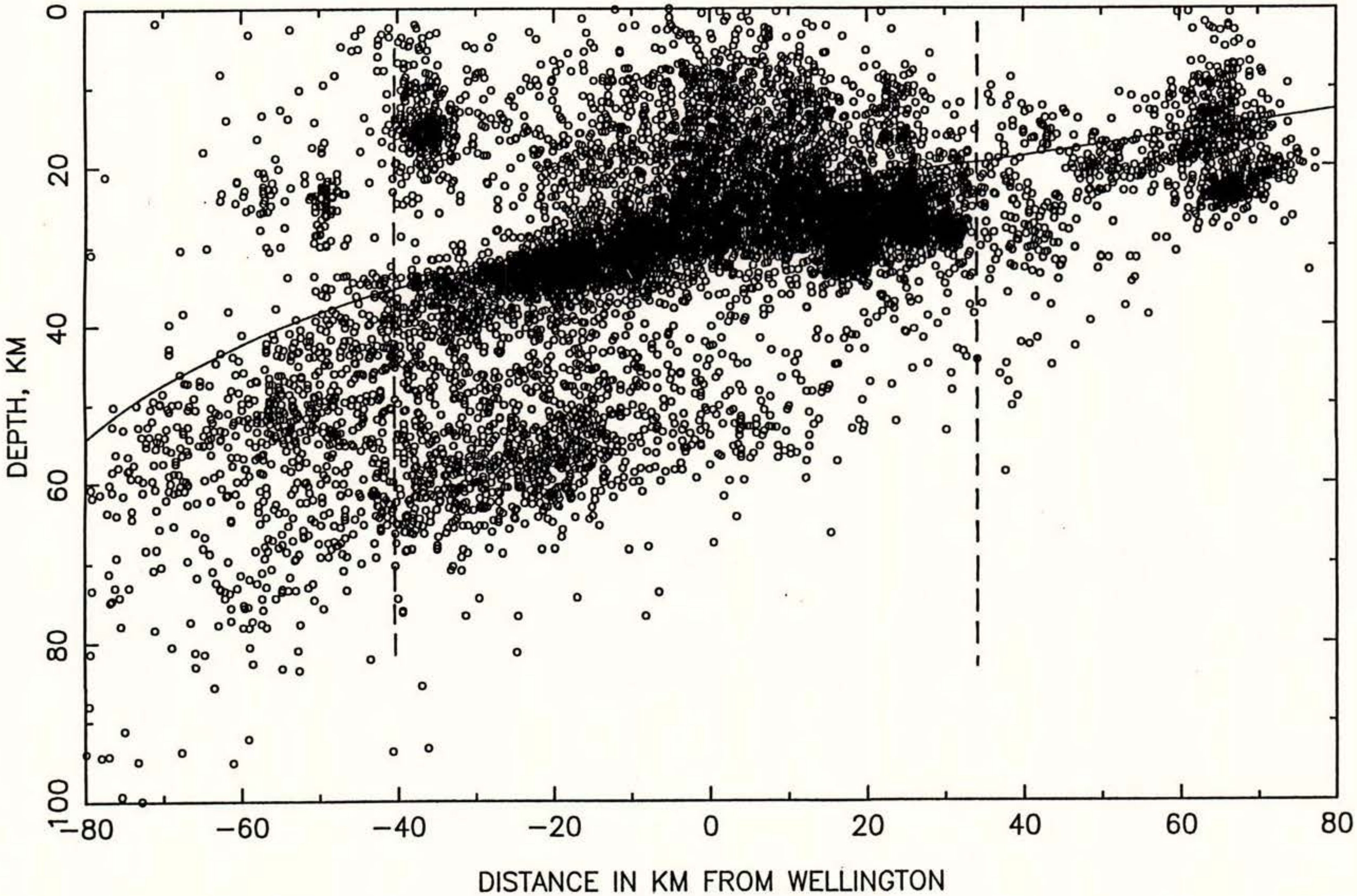


Figure 3: Hypocentres of events, located by the Wellington Seismograph Network, 1986 - 1994, projected onto a plane striking northwest - southeast. The solid line represents the top of the subducting Pacific plate, and the vertical dashed lines indicate the limits of seismic slip in our models.

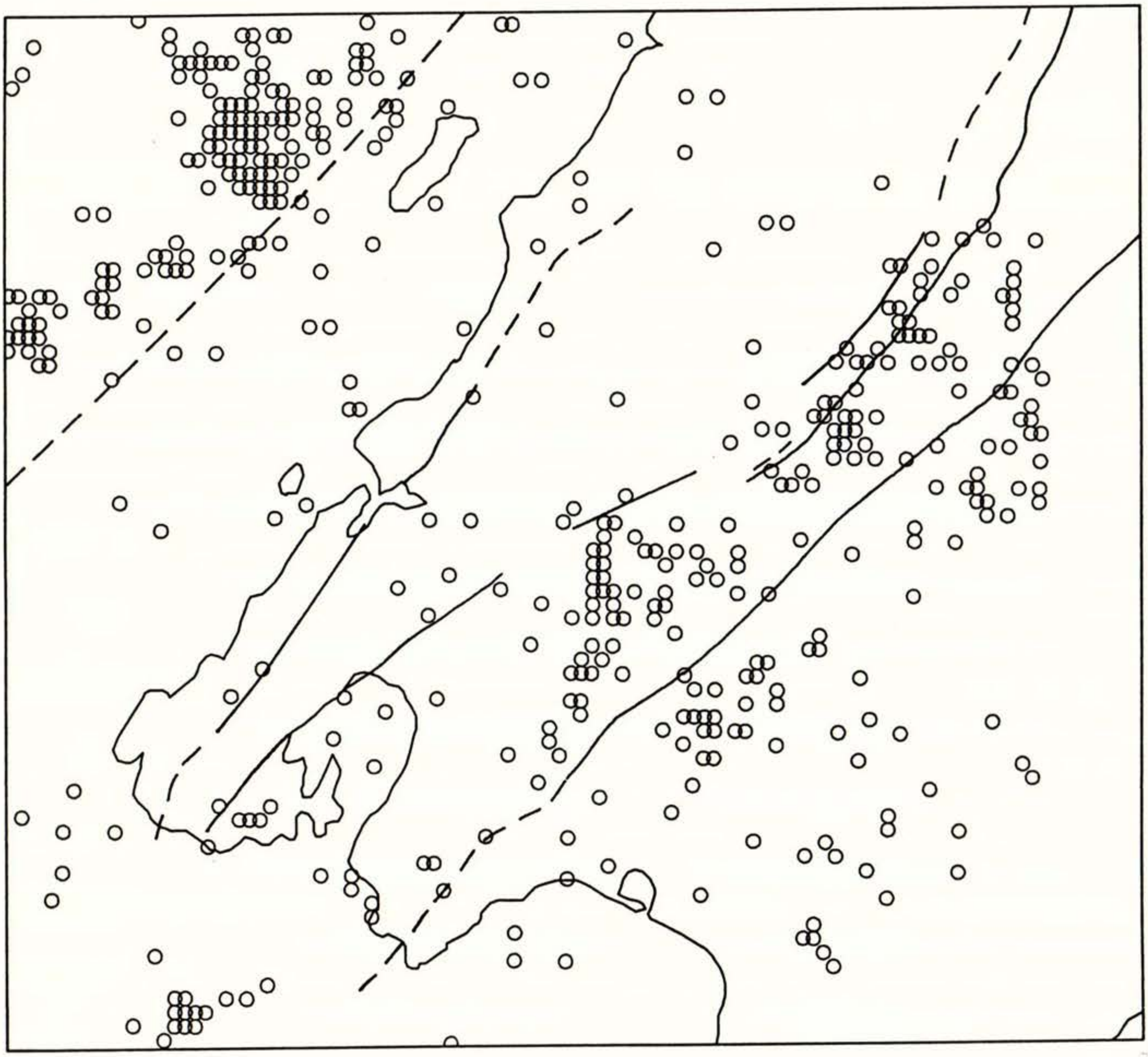


Figure 4: Epicentres of events, 1986 - 1994, with depths of 15 km or less.

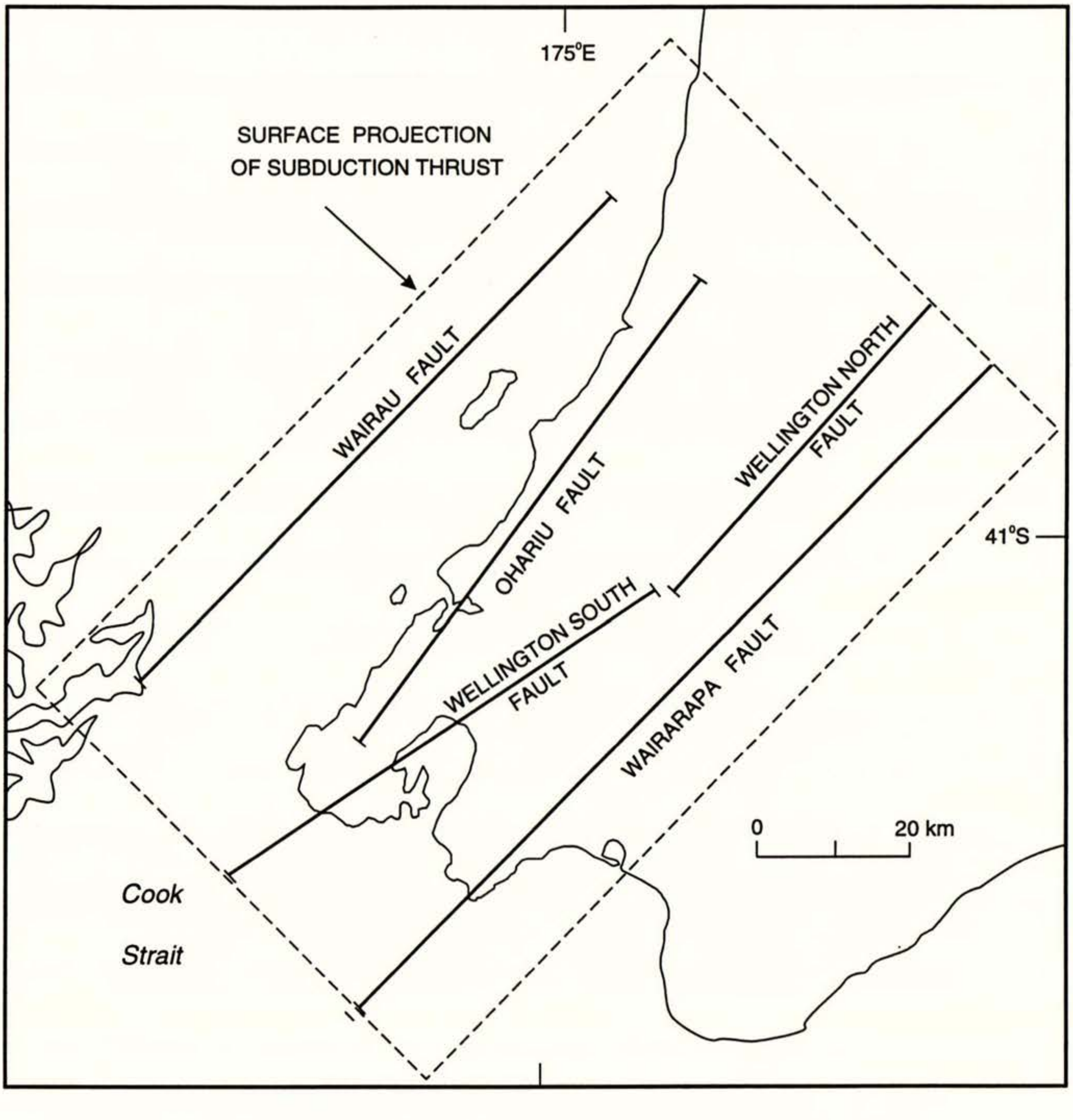
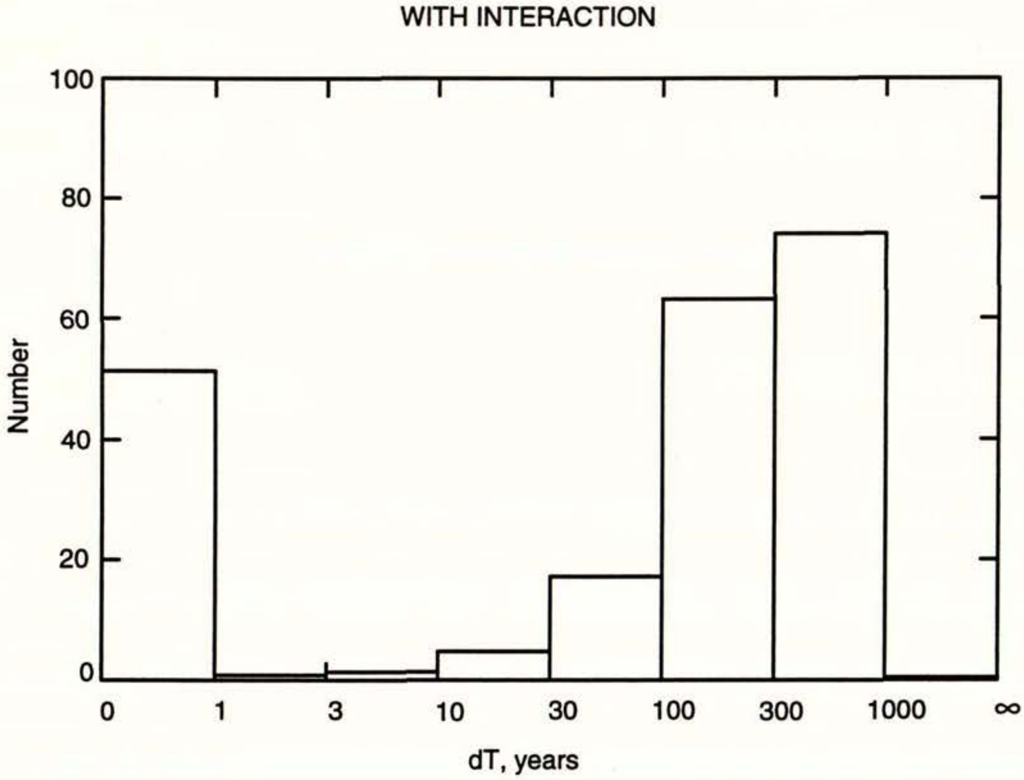


Figure 5: Representation of the faults in our synthetic seismicity model. The dashed box is the projection onto the surface of the subduction thrust.



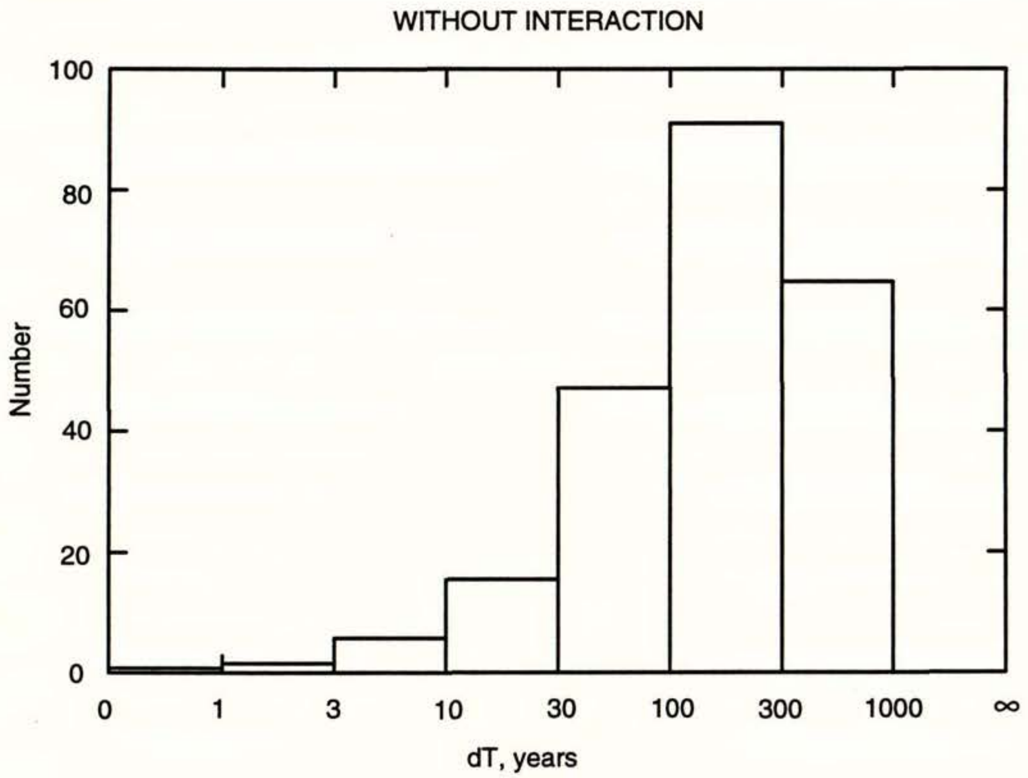


Figure 6: Histograms of times between "characteristic" events, averaged over our 6 models. The total number of events correspond to a period of 50.000 years

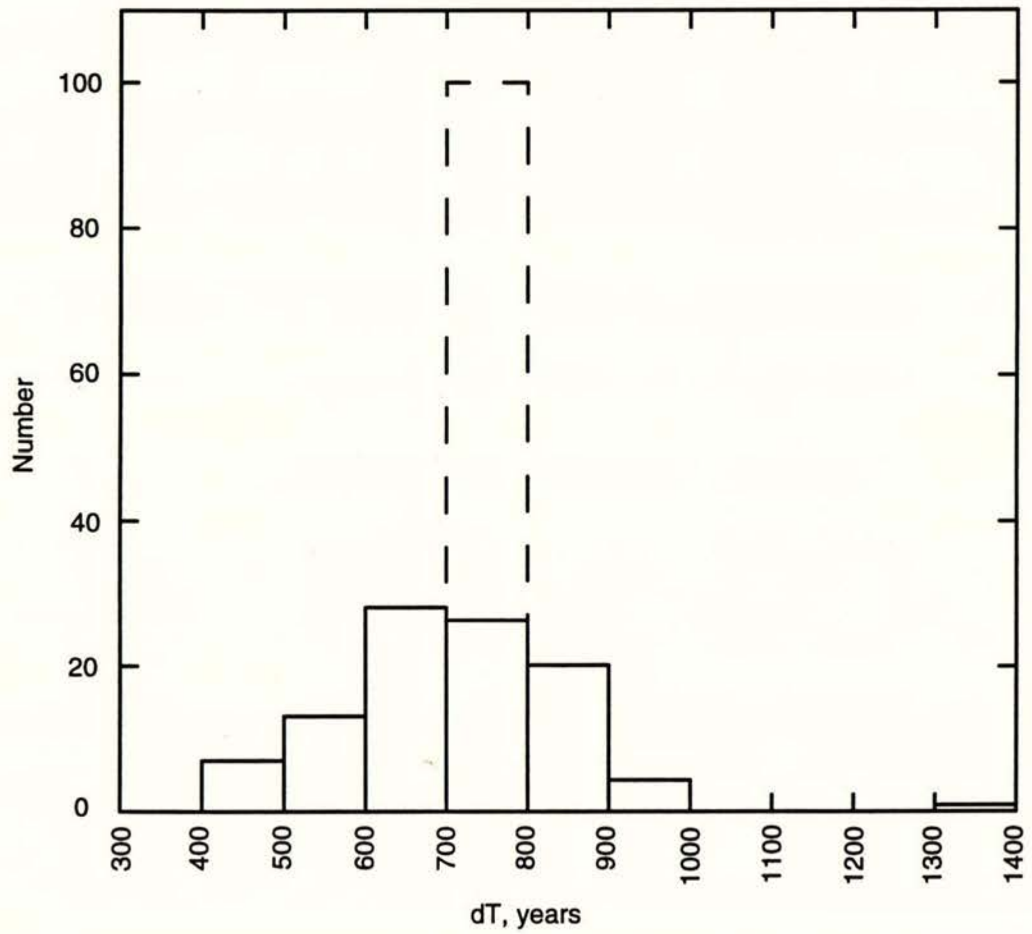


Figure 7: Histograms of interevent times for "characteristic" events on the Wellington South fault, model 3. The dashed lines indicate the corresponding distribution for the case of no elastic interactions.