Testing a Model of Earthquake Nucleation

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Abstract  Some laboratory models of slip find that a critical amount (or velocity) of slow slip is required over a nucleation patch before dynamic failure begins. Typically, such patch sizes, when extrapolated to earthquakes, have been thought to be very small and the precursory slip undetectable. Ohnaka (1992, 1993) has proposed a model in which foreshocks delineate a growing zone of quasi-static slip that nucleates the dynamic rupture and suggests that it could be large enough (~10 km across) to be detectable and thus useful for short-term earthquake prediction. The 1992 Landers earthquake (M 7.3) had a distinctive foreshock sequence and initiated only 70 km from the strain meters at the Piñon Flat Observatory (PFO). We use this earthquake to investigate the validity and usefulness of Ohnaka’s model. The accurate relocations of Dodge et al. (1995) show that the foreshock zone can be interpreted as expanding from an area of 800 m (along strike) by 900 m (in depth), to 2000 by 3200 m in the 6.5 hr before the mainshock. We have calculated the deformation signals expected both at PFO and 20 km from the foreshock zone, assuming either constant slip or constant stress drop on a circular patch expanding at 5 cm/sec over 6.5 hr. We find the slips or stress drops would have to have been implausibly high (meters or kilobars) to have been detectable on the strain meters at PFO. Slightly better limits are possible only 20 km from the source. Even though the distance from Landers to PFO is small compared with the average spacing of strain meters in California, we are unable to prove or disprove Ohnaka’s model of earthquake nucleation. This suggests that even if the model is valid, it will not be useful for short-term prediction.

Introduction

Laboratory studies of frictional sliding in rocks (Dieterich 1979, Ohnaka and Kuwahara 1990) show that subsonic slip is observed over an expanding area prior to dynamic failure. If the same type of slip occurs before earthquakes at detectable levels, it would be very useful for short-term earthquake prediction.

One motivation for observing crustal deformation is to search for such short-term precursors. Most observations of crustal strain have not detected any preseismic slip, however, and thus can only be used to limit the maximum preseismic moment release in the source region. The size of the limit depends on the time interval considered because the noise spectrum of these measurements increases with period (Agnew, 1992) and also on the distance from the source. Johnston et al. (1987) found the maximum preseismic moment over the final few minutes to be 2% or less of the seismic moment of the subsequent earthquake, for five earthquakes (M 3 to 5.5), with strain meters between 28 and 55 km away. Wyatt (1988) found preseismic moment to be limited to as little as 1% of the seismic moment for earthquakes in southern California. The tightest constraints come from observations of the 1989 Loma Prieta (M 7.0) and 1992 Landers (M 7.3) earthquakes, for which the limits over the final few minutes are less than 0.1% of the seismic moment for both events (Johnston et al., 1990, 1994; Wyatt et al., 1994).

These data suggest that preseismic slip is very small and not easily detected, perhaps not even with instruments directly over the source zone. Ohnaka (1992, 1993), however, has proposed a model for earthquake nucleation, implying that preseismic slip might occur over areas of order 10 km across and thus be readily detectable. This model interprets immediate foreshocks as occurring within the growing nucleation zone; since such foreshocks were observed for the 1992 Landers earthquake, we considered it worth testing using the available deformation data.

Ohnaka et al. (1987) and Ohnaka and Kuwahara (1990) developed a model to explain the slow but accelerating slip observed before dynamic rupture in laboratory experiments of frictional sliding. This model was extrapolated to the

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earth’s crust using both computer simulations (Yamashita and Ohnaka, 1992) and investigation of the physical characteristics of the crust (Ohnaka, 1992) to estimate possible values for slip and area of the nucleation zone. As noted above, this model allows the size of the foreshock zone to be used to infer the size of the nucleation zone; the foreshocks are due to small dynamic instabilities that do not propagate far.

Ohnaka (1992) suggests that earthquakes without foreshocks are nucleating in regions that are less heterogeneous so that no dynamic instabilities are triggered. For example, earthquakes nucleating at the base of the seismogenic zone may follow a nucleation process confined to the more ductile and more homogeneous lower crust.

Ohnaka (1993) considers the foreshock sequence of the Izu Oshima Island earthquake (7.2 M, 1978) and finds that the foreshock zone grows at a rate of 1 to 40 cm/sec, reaching a diameter of 10 km just before the mainshock occurred. He suggests that a nucleation zone size of 10 km may be typical for an M 7 earthquake. It is worth noting that the foreshocks of the Izu Oshima earthquake all occurred offshore; it is not clear how accurate the locations were. Since lower accuracies usually disperse the locations, the foreshock zone could have been less than 10 km across.

The Landers (1992) Earthquake Sequence

The 28 June 1992 Landers earthquake (Mw 7.3) was the largest event in California in 40 yr (Fig. 1) (Hauksson et al., 1992). It was preceded by a well-recorded foreshock sequence of 35 events (ML 1.5 to 3.6), 3 of which occurred 2 months before the mainshock, and 24 of which occurred within the last 7 hr. Dodge et al. (1995) relocated the foreshocks using a waveform correlation technique, to a relative accuracy of <100 m horizontally and <200 m in depth. This makes this sequence ideal for constraining Ohnaka’s model. As shown by Figure 1, the Landers earthquake is also only 70 km from the long-base strain meters at Pifion Flat Observatory, which, though running through the entire sequence, detected no preseismic strain (Wyatt et al., 1994).

To determine whether the Ohnaka model is consistent with the Landers foreshock sequence, we show in Figure 2 both the along strike and depth position of the foreshocks as a function of time. The three early foreshocks define a smaller zone than the immediate foreshocks, which is consistent with some growth of the zone. The size of the zone of immediate foreshocks, however, shows small sign of expansion either along strike or in depth (if the one outlier is removed). This conclusion is also reached by Dodge et al. (1995). It is possible, however, to interpret these last foreshocks as being part of an expanding zone, as shown in Figure 2. This interpretation is used here in order to investigate whether any preseismic slip over a nucleation patch delineated by the foreshocks (as suggested by Ohnaka) could have been large enough for detection. For this purpose, the foreshocks are considered here as defining a zone 0.8-km across by 0.6-km deep initially, increasing at about 0.5 mm/sec for 41 days, to a size of 0.9 by 0.8 km. In the last 6.5 hr, the zone would expand at about 5 to 10 cm/sec to about 2 by 3 km just before the onset of the mainshock.

If there were such an expanding zone, could subsonic slip on it be detected by the strain meters at Pinyon Flat? This depends on how much slip occurred in the zone, something Ohnaka’s model does not address. We consider two distinct models of slip. We assume in both cases that the radius over which some preseismic slip has occurred is \( r = vt \), where \( t \) is time and \( v \) is the velocity of growth, which is about 5 cm/sec for the last 6.5 hr. In one case, we assume that as the slip front passes, a slip of size \( s \) takes place, not
increasing thereafter so that the total moment depends only
on the total area $A$. The seismic moment ($M_0$) divided by the
shear modulus ($\mu$, taken to be $3 \times 10^{10}$ Nm) is the source
strength for creating crustal deformation, and this is then just

$$\frac{M_0}{\mu} = \sqrt{2} \sigma r^2. \quad (1)$$

Our second model is that of an expanding crack, in which
we assume the stress drop ($\Delta \sigma$) to be constant; we can then
use the result of Eshelby (1957) for a circular crack:

$$\frac{M_0}{\mu} = \frac{16 \Delta \sigma r^3}{7\mu}. \quad (2)$$

Figure 3 shows the moment increasing with time for these
two models for four possible values of stress drop and slip.

We used a standard dislocation program to compute the
deformation seen at Pinyon Flat from a N–S-striking slip
plane at the location of the foreshocks. Applying this to the
equations above, we calculate the expected NW–SE strain
at Pinyon Flat Observatory for $s = 1$ m and $\Delta \sigma = 100$ bars,
both values comparable to the seismic slip on this part of
the rupture (Wald and Heaton, 1994). The recorded strain
and GPS displacement data at Pinyon Flat Observatory are
also shown, and it is clear that even these large amounts of
preseismic slip would not have been detected. We cannot,
therefore, rule out Ohnaka’s model, but can provide no pos-
tive evidence in support of it.

Concluding Remarks

This investigation of Ohnaka’s model of earthquake nu-
cleation demonstrates that there is a lack of direct evidence
in support of it and that current data cannot disprove it. More
importantly, this study shows that even if Ohnaka’s model
of earthquake nucleation does apply to Californian earth-
quakes, the preseismic slip it predicts is not readily detect-
able at significant distance. At closer distances, matters are somewhat better. Figure 4 also shows the strain signals to be expected at a site 20 km from the foreshocks. In this case, strain data of the quality of those presented here would easily be able to detect large preseismic slip and perhaps place valuable bounds on what could have occurred. This is not, however, the case with all types of deformation data. We also show in Figure 3 the expected displacement 20 km from the foreshocks and show for comparison actual GPS data collected at Pinyon Flat. It is clear that even for such a favorable location, the sensitivity of the GPS system falls far short of what might be needed to detect plausible amounts of preseismic slip.

It is worth noting the growing number of observations of small seismic precursors to many large earthquakes, e.g., Landers (Abercrombie and Mori, 1994), Gulf of Corinth (Abercrombie et al., 1995), and earthquakes in Mexico (Anderson and Chen, 1995) and northern California (Ellsworth and Beroza, 1994). These precursors, or small onsets, to larger earthquakes may represent small events that trigger the mainshock or possibly slow but just seismic slip that nucleates the mainshock. In either case, as they are typically <1% of the mainshock moment ($M_w 4.4$, $\Delta \sigma \sim 10$ bars, for Landers, Abercrombie and Mori, 1994), their existence implies that any preceding aseismic slip would be even smaller (considerably smaller than that shown in Fig. 4), decreasing the chances of its being detectable, even at short distances, and so useful for short-term earthquake prediction.

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**Figure 3.** Time dependence of the preseismic moment predicted by the two slip models for different possible values of slip and stress drop.
Figure 4. The top panel shows the observed strain at Piñon Flat Observatory and that predicted by the slip front model (dashed, slip = 1 m) and crack model (solid, 100 bars). The second panel shows the strains for the same models predicted 20 km from the source. The third and fourth panels are the same as the top two but for displacement as measured by GPS. The GPS data are two minute samples of the JPLM-PINT baseline, processed with GIPSY/OASIS with daily multipath signals removed. In all four panels, the cumulative strain and displacement predicted by the models at the start of the mainshock are given on the right. The bottom panel shows the occurrence of the foreshocks in time.


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