Title
Empirical Estimation of Fault Directionality for Improved Non-parametric Estimation of Branching Models for Earthquake Occurrences

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Empirical Estimation of Fault Directionality for Improved Non-parametric Estimation of Branching Models for Earthquake Occurrences

A thesis submitted in partial satisfaction of the requirements for the degree Master of Science in Statistics

by

Joshua Seth Gordon

2013
ABSTRACT OF THE THESIS

Empirical Estimation of Fault Directionality for Improved Non-parametric Estimation of Branching Models for Earthquake Occurrences

by

Joshua Seth Gordon

Master of Science in Statistics

University of California, Los Angeles, 2013

Professor Frederic Paik Schoenberg, Chair

We implement a non-parametric estimation of an Epidemic-Type Aftershock Sequence triggering function with the directionality of the aftershock activity included in the estimation. To accomplish this we modify the triggering function \( v \) such that it is instead a function \( v(s, t, m, \theta) \), depending not only on the magnitude of previous events but also on the relative angular separation between events, relative to the primary fault plane associated with the prior event. These relative angles depend on estimates of the principal fault planes associated with each event. We estimate these fault planes using all seismic activity within 100km\(^2\) of a main shock event using \( M > 2.0 \) events in the entire earthquake catalog.
The thesis of Joshua Seth Gordon is approved.

Yingnian Wu

Nicolas Christou

Frederic Paik Schoenberg, Committee Chair

University of California, Los Angeles
2013
In memory of Ron Gordon . . .
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Besides my advisor, I would like to thank the rest of my thesis committee: Prof. Yingnian Wu, and Prof. Nicolas Christou.

My sincere thanks also goes to Eric Fox for his collaboration on this project.

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Last but not the least, I would like to thank my family Judith and Amanda Gordon for supporting me throughout my life, and Larry Bridgman for his inspiration and positive attitude.
0.1 Introduction

We implement a non-parametric estimation of an Epidemic Type Aftershock Sequence Model (ETAS) triggering function with the directionality of the aftershock activity included in the estimation. ETAS is a widely-used model in seismology which has been shown to fit rather well to various catalogs of earthquakes and their aftershocks including those in California (Ogata 1998), Japan, and worldwide (Marzocchi 2008).

In the standard model, conditional intensity $\lambda(s, t)$ is dependent on some background rate $\mu$, plus a triggering function $v(s - s_i, t - t_i, m_i)$ over all previous earthquakes at location $s_i$, time $t_i$, and magnitude $m_i$, such that $t_i > t_j$. We modify the triggering function $v$ such that it is instead a function $v(s, t, m, \theta)$, depending not only on the magnitude of previous events but also on the relative angular separation between events, $\theta_{ij}$, relative to the primary fault plane associated with the prior event, $\phi_j$. Fault planes are estimated from all seismic activity within 100km$^2$ of a main shock event using $M > 2.0$ events in the entire earthquake catalog.

Section 2 provides an overview of spatial-temporal point processes, Section 3 presents non parametric estimation techniques, Section 4 motivates the use of $\theta_{ij}$ in the triggering function, Section 5 contains a description of the data constraints, Section 6 outlines the methods applied to the data, Section 7 provides a discussion of further research areas.

0.2 Spatial Temporal Point Process

A spatial-temporal point process is a random collection of points. An example of this is origin times and locations of earthquakes. It is mathematically defined as a random measure $N$ on a space $S \subseteq \mathbb{R} \times \mathbb{R}^3$ of space-time, taking on non-negative
integers $\mathbb{Z}^+$. The measure of process $N(A)$ represents the count of points falling into the subset $A$ of $S$. We typically consider the case where $N$ contains a finite number of points on any bounded subset of $S$ (Daley and Vera-Jones 1998).

The conditional intensity $\lambda(t, x, y, z)$ can be thought of as the frequency with which events are expected to occur around a particular location $(t, x, y, z)$ conditional on what earthquakes have occurred previously $H_t$ of the point process up to time $t$. Formally it can be defined as the limiting expected rate of occurrence around a particular space time location, given the history $H_t$ (Daley and Vera-Jones 1988):

$$\lambda(t, x, y, z) = \frac{E[N\{(t, t + \Delta t) \times (x, x + \Delta x) \times (y, y + \Delta y) \times (y, y + \Delta y)\}|H_t]}{\Delta t \times \Delta x \times \Delta y \times \Delta z}$$

### 0.2.1 Conditional Intensity Models for Point Processes

The conditional intensity uniquely characterizes all finite-dimensional distributions of a point process (Daley and Vere-Jones 1988). A stationary Poisson process has constant conditional intensity and a non-stationary Poisson process has deterministic conditional intensity.

Consider a temporal point progress with events occurring between time 0 and time $T$. We can characterize $N$ as an ordered list of event times. A temporal point process $N$ can be viewed as a counting process $N(t)$. For any time $t$ between 0 and $T$, $N(t)$ is the number of points occurring at or before time $t$. $N(t)$ must be non-decreasing and right continuous and take only non-negative integer values.

A spatial-temporal point process can be viewed as a marked temporal point process. We have a random collection of points with an additional random variable associated with it called a mark. The mark in this case in the spatial information and any other information with which we want to use in estimation such as magnitude $M$. It represents activity of earthquakes of magnitude $M$ in a region during a specified period of time $[0, \infty)$, and location $x, y, z$. 

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0.2.2 ETAS

Epidemic Type Aftershock Sequence Model (ETAS) are multigenerational models which identifies aftershocks of earthquakes which trigger other aftershock sequences. ETAS models take on various parametric forms which attempt to model the dependency of time, space, and magnitude on seismicity above some background rate.

The Conditional Intensity of ETAS takes the form:

$$\lambda(t, x, y, z) = \mu(x, y, z) + \sum_{i: t_i < t} v(t - t_i, x - x_i, y - y_i, z - z_i; M)$$

With $t_i < t$ and we sum over all points $t_i, x_i, y_i, z_i$. Included is a background rate $\mu$ of activity which is a constant occurrence in time and $v$ which is the clustering density of aftershock events.

Ogata proposed a variety of triggering functions. One of which is the following isotropic cluster density (Ogata 1998):

$$v(t, x, y, m) = \frac{K_0 \exp[\alpha(m - m_0)]}{(t + c)^p(x^2 + y^2 + d)^q}$$

Where $t > t_i$, and $K_0, \alpha, c, p$ are all constants at each $i$ and represent some characteristics of seismicity in the region. $p$ is the decay rate of aftershocks, $\alpha$ is the efficiency of an earthquake at producing offspring. $M_i$ is the magnitude of earthquake $i$ and $M_0$ is the cut off magnitude of earthquakes included in the dataset.

0.3 Non-parametric estimation of Spatial Temporal Point Process

Various methods have been proposed for estimating the conditional intensity of a self-exciting point process. Such methods typically involve estimating a para-
metric model using maximum likelihood methods, or close variants. An alternative is non-parametric estimation, which can be desirable in order to avoid mis-specification.

Marsan and Langline 2008 proposed the following for non-parametric estimation of the background rate and the triggering function. Assuming the conditional intensity follows the form:

$$\lambda(|\Delta x|, \Delta t, m_i) = \mu + v(|\Delta x|, \Delta t, m_i)$$

where $|\Delta x|$ is the distance between earthquake $i$ and earthquake $j$. $\Delta t$ is the time between events. $\lambda_0$ is the intensity rate of background earthquakes (Marsan 2008).

Let $P$ be a $N \times N$ lower triangular probability matrix whose entries are given by $p_{ij}$, the probability event $i$ was caused (triggered) by event $j$ ($i > j$). $p_{ii}$ is the probability event $i$ is a background event and $p_{ij} = 0$ for $j > i$. $N$ is the number of events and the rows must sum to one:

$$\sum_{j=1}^{N} p_{ij} = 1 \quad \forall i$$

Initializing probability matrix:

$$p_{ij} = \begin{cases} 1/i, & i \geq j \\ 0, & \text{otherwise} \end{cases}$$

**Algorithm**

1. Initialize $P^{(0)}$

2. Estimate $v^{(k)}$ and $\mu^{(k)}$ at iteration $k$:

   $$v^{(k)}_n = \frac{\sum_{A_n} p^{(k-1)}_{ij}}{N \cdot \delta t \cdot S(|\Delta x|, \delta r)}$$

   $$\mu^{(k)} = \frac{\sum_{i=1}^{N} p^{(k-1)}_{ii}}{T \cdot S}$$

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3. Update probabilities $P^{(k)}$:

$$p^{(k)}_{ij} = \frac{v^{(v)}(x_i - x_j, t_i - t_j, m_i)}{\mu^{(k)} + \sum_{j=1}^{i-1} v^{(k)}(x_i - x_j, t_i - t_j, m_i)} \text{ for } i > j$$

$$p^{(k)}_{ii} = \frac{\mu^{(k)}}{\mu^{(k)} + \sum_{j=1}^{i-1} v^{(k)}(x_i - x_j, t_i - t_j, m_i)}$$

4. Repeat (2) and (3) until convergence

Where $T$ is the duration of the time series containing $N$ earthquakes and $S$ is the surface analyzed. $\delta r, \delta m$, and $\delta t$ as discretized parameters. $N_m$ is the number of earthquakes such that $m_i = m + \delta m$. $S(|\Delta(x)|, \delta r)$ is the surface covered by the disk with radii $|\Delta x| \pm \delta r$. $A$ is the set of pairs such that:

$$|x_i - x_j| = |\Delta(x) \pm \delta r|$$

$$m_i = m + \delta m$$

$$t_j - t_i = t + \delta t$$

0.4 Extending the triggering function with angular separation $\theta_{ij}$

Aftershock zones are thought to be elliptical, but estimation of the ellipses is fraught with difficulties (Ogata 1998). Instead of assuming the structure of aftershock zones, we can improve our estimate of the triggering function by adding information on the angular separation between an earthquake and its aftershocks. Previous analyses have used moment tensor estimated for earthquakes, however there is difficulty in using the information due to assumptions that must be made about the type of faults from which the moment tensor solutions are calculated (Wong 2009). Instead of relying on the moment tensors, we estimate the fault planes by looking at all the events of $M > 2.0$ that occurred in the vicinity of a main shock over a 30 year period.
Once we have an estimate of the angular separation between events, $\theta_{ij}$, there are a variety of ways we can use the information. One method of adding information on fault direction is to use a wrapped exponential description on a quarter circle which has the density:

$$f(\theta_{ij}) = \frac{\lambda e^{-\lambda \theta_{ij}}}{1 - e^{\lambda \pi/2}}$$

$\lambda$ is estimated locally by maximum likelihood successively in bins with $n$ pairs of points, sorted by the distance between a main shock and an aftershock (Wong 2009).

We can incorporate our estimate of $\theta_{ij}$ using a modification of the above algorithm:

Assume the conditional intensity follows the form $\lambda(\Delta t, \theta_{ij})$ with background intensity $\lambda_0$.

1. Initialize $P^{(0)}$

2. Estimate $\nu^{(k)}$ and $\mu^{(k)}$ at iteration $k$:

$$\nu^{(k)}_n = \frac{\sum_{A_n} P^{(k-1)}_{ij}}{N_{\theta_{ij}} \cdot \delta t}$$

$$\mu^{(k)} = \frac{\sum_{i=1}^{N} P^{(k-1)}_{ii}}{T}$$

3. Update probabilities $P^{(k)}$:

4. Repeat (2) and (3) until convergence

Where A is the set of pairs such that $t_j - t_i = t \pm \delta t$ and $N_{\theta_{ij}}$ is the number of earthquakes such that $\theta_{ij} = \theta_{ij} \pm \delta \theta_{ij}$. Further work is being done to develop a non parametric form for the triggering function of the conditional intensity $\lambda(|\Delta x|, \Delta t, m, \theta)$. 

6
0.5 Data

We took data from the ANSS Catalog using the RELM collection region which includes all of California and approximately 1.5° around it and is partitioned into 0.1° by 0.1° cells. The region included over 193,000 earthquakes with $M > 2.0$ between January 1, 1980 and January 1, 2013, with maximum depth 75km. In the estimation of $\phi$ we chose a minimum magnitude of 2. With a lower cutoff we might have some completeness issues though. Assessing the completeness of an earthquake catalogue is necessary to assess its accuracy. We can assess the completeness in terms of magnitude and a way seismologists typically look at this issue is by looking at the empirical survivor function, $1-F(m)$. Figure 1 shows that a sensible cutoff may be between $M=1$ and $M=2$ and that incompleteness is not a major problem above $M=2$ (Weimer 2000). Since our data set includes information from $M > 2.0$, we are fairly certain the catalog is complete.

Figure 2 shows 658 main shock magnitudes $M > 4.5$ occurring since Jan 1, 1980. We can see that main shock events spread evenly spread through the time frame which relatively few events occurring above magnitude 6.5. In addition, Figure 6 shows the origin locations of main shock events occurring in California in the catalog. You can see clear clusters of events that occurred along fault planes.

Alternative datasets are possible such as taking only after aftershocks as in the STEP method. Other algorithms such as the magnitude based clustering could have been used (Ogata 1998). However, using both foreshocks and aftershocks seemed preferable because this would include additional information about the fault planes. For further information about alternative data sets see Table 1.

The ANSS dataset contained 658 earthquakes with $M > 4.5$. For each of these 658 large earthquakes, we use the entire catalog of earthquakes in the $M > 2$ range to estimate fault plane $\phi_j$. For each main shock, the dataset includes: origin time,
Figure 0.1: Survivor Function suggesting incompleteness is not a major problem above M=2
Figure 0.2: Main shock magnitudes

$t_i$, Magnitude $M_i$, longitude $x_i$, latitude $y_i$, depth $z_i$, and $\phi_j$ the estimated main direction of earthquake $j$’s fault.

0.6 Methods

0.6.1 Weighted Regression to estimate $\phi_j$

Weighted regression modifies the assumption that each data observation provides equally precise information about the variation in the model. We do not assume that the standard deviation of the error term is constant over all explanatory variables. This allows us to use additional information to influence the parameter estimates and improve our estimate.

With a reasonable guess of variance, you can re-weight the data. Consider the
standard model with unequal variance:

\[ Y_i = \beta_0 + \beta_1 X_i + \epsilon_i, \quad \epsilon_i \sim N(0, \sigma^2 X_i^2) \]

Set \( \tilde{Y}_i = Y_i / X_i \) and we have:

\[ \tilde{X}_i + \beta_1 + \epsilon^* \quad \epsilon^* \sim N(0, \sigma^2) \]

To fit the model we minimize \( \sum_{i=1}^{n} \frac{1}{X_i^2} (Y_i - \beta_0 - \beta_1 X_i)^2 \). Furthermore, weighted least squares takes the form:

\[ SSE(\beta, w) \sum_{i=1}^{n} w_i(Y_i - \beta_0 - \beta_1 X_i)^2 \]

Where \( Y_i \) is the observed value, \( w_i = \frac{1}{X_i^2} \) is the weighting factor for the \( i \)th observation (Taylor 2009).

Constraining OLS through the origin results in (Eisenhauer 2003):

\[ Y_i = \beta X_i + \epsilon_i \]

And the estimated slope of the regression line is:

\[ \hat{\beta} = \frac{\sum_{i=1}^{n} X_i Y_i}{\sum_{i=1}^{n} X_i^2} \]

Using \( M > 2.0 \) events occurring both before and after a main shock event in entire dataset we calculate \( \phi_j \), the estimated fault direction of each main shock. \( \phi_j \) is estimated within a spatial area of 100km\(^2\) surrounding each main shock using a weighted regression, constrained to pass through main shock location \( x_j, y_j \). That is, each earthquake \( k \) (\( M_k > 2.0 \)), within a square 10 × 10 km of a \( M_j > 4.5 \) event occurs at location \( x_k, y_k \). Each event is given a weight \( \frac{1}{d_{kj}} \), where \( d_{kj} \) is the Euclidian distance from (\( M > 2.0 \)) event \( k \) to (\( M > 4.5 \)) event \( j \).

Only included are aftershock events not occurring at \( x_j, y_j \). A small number of events were recorded at the same latitude and longitude of the main shock given
the prevision of the dataset. Since these events add nothing to the estimation of \( \phi_j \) they were removed from the weighted regression.

Figure 3 shows an estimate of \( \phi_j \) for a Magnitude 5.8, Depth = 14.79km earthquake occurring at 1980/01/24, 19:00:08.58. This main shock used had 423 aftershocks (\( M > 2.0 \)). We can visually detect the fault plane by examining the aftershock activity centered around the origin. It is clear that fault plane \( \phi_1 \) provides a reasonable estimate of fault directionality. For additional estimates sees Figures 7 and 8. Figure 7 shows \( \phi_{47} \) based on fewer aftershocks and the fault direction is not as clear visually. In addition, Figure 8 shows \( \phi_{457} \) based on many aftershocks and the fault direction is also not as clear visually.

### 0.6.2 Determining angular separation \( \theta \)

We assume earthquake \( j \) comes first and therefore will be called the main shock and \( i \) the aftershock. We compute \( \theta_{ij} \), the angle made by the segment connecting point \((x_i, y_i)\) to point \((x_j, y_j)\), and the line through \((x_j, y_j)\) with direction \( \phi_j \) for each pair of earthquakes where \( t_i > t_j \).

\[
\theta_{ij} = \arctan \left( \frac{|m_1 - m_2|}{1 + m_1 \cdot m_2} \right), \quad \theta_{ij} \in [0, \pi/2]
\]

Where \( m_1 \) is the slope of the line passing through \((x_j, y_j)\) and \((x_i, y_i)\). \( m_2 \) is estimated he fault direction \( \phi_j \). Figures 3 and 4 show the relationship between various \( \phi_j \) and \( \theta_{ij} \). In Figure 3, we can the same fault plan as shown in Figure 2 in addition to the slope of the line between mainshock event 1 and 3. The angle between the two slopes is 47.00106°. Figure 4 shows a much smaller angular separation between events which implies they likely occurred on the same fault.
Figure 0.3: Fault plane $\phi_1$ for Mainshock 1: magnitude 5.80 and occurred on 1980/01/24 19:00:08.58 at location (37.8400, -121.7678) and Depth 14.79km based on 423 $M > 2.0$ aftershocks.
Figure 0.4: $\theta_{31} = 47.00106^\circ$

Mainshock 3: magnitude 4.80 and occurred on 1980/01/24 19:03:18.44 at location (37.8342, -121.7810) and Depth 3.92km.

Mainshock 1: magnitude 5.80 and occurred on 1980/01/24 19:00:08.58 at location (37.8400, -121.7678) and Depth 14.79km.
Figure 0.5: $\theta_{52} = 13.84518^\circ$

Mainshock 5: magnitude 5.1 and occurred on 1980/01/27 02:33:35.34 at location (37.7490, -121.7063 14.43) and Depth 14.43km.

Mainshock 2: magnitude 5.40 and occurred on 1980/01/24 19:01:01.54 at location (37.8110, -121.7750) and Depth 6.9km.
0.7 Discussion

New approaches for estimating $\phi$ and $\theta$ can be made. Each estimate of $\phi$ is given equal weight in the estimation of $\theta$ even though their associated mean squared errors are significantly different. Not requiring $\phi$ to pass through a main shock epicenter is another option. Combining information from seismic moment tensors into the estimation of $\phi$ and $\theta$ might prove useful.

0.8 Conclusion

We overcome the limitation that in ETAS, $\lambda(s, t)$ cannot depend on what happens after time $t$. To accomplish this we modify the triggering function $v$ such that it is instead a function $v(s, t - t_j, m_j, \theta_j)$, depending not only on the magnitude of previous events but also on the angular separation between events, $\theta_j$. $\theta_j$ are based on estimates of fault planes, $\phi_j$ which are estimated from all seismic activity within a mainshock region.
Tables and Figures

Figure 0.6: Origin Locations of 658 ($M > 4.5$) events within California RELM Collection Region from January 1, 1980, to January 1, 2013. The Collection region includes all of California and approximately 1.5° around it and is partitioned into 0.1° by 0.1° cells.
Figure 0.7: Fault Plane Estimates for Magnitude 7.2, Depth 14.64km event occurring at (41.0842, -124.6157) on 11/08/1980 at 10:27:33.20 based on 38 $M > 2.0$ aftershocks within $100km^2$. 
Figure 0.8: Fault Plane Estimates for Magnitude 7.1, Depth .02km event occurring at (34.5940, -116.2710) on 10/16/1999 at 09:46:44.13 based on 2575 $M > 2.0$ aftershocks within 100km$^2$
Table 0.1: Summary of Data Constraints

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<th>days since main shock</th>
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References


