Isolated cases of remote dynamic triggering in Canada detected using cataloged earthquakes combined with a matched-filter approach

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Abstract Here we search for dynamically triggered earthquakes in Canada following global main shocks between 2004 and 2014 with $M_S > 6$, depth < 100 km, and estimated peak ground velocity > 0.2 cm/s. We use the Natural Resources Canada (NRCan) earthquake catalog to calculate $\beta$ statistical values in $1° \times 1°$ bins in 10 day windows before and after the main shocks. The statistical analysis suggests that triggering may occur near Vancouver Island, along the border of the Yukon and Northwest Territories, in western Alberta, western Ontario, and the Charlevoix seismic zone. We also search for triggering in Alberta where denser seismic station coverage renders regional earthquake catalogs with lower completeness thresholds. We find remote triggering in Alberta associated with three main shocks using a matched-filter approach on continuous waveform data. The increased number of local earthquakes following the passage of main shock surface waves suggests local faults may be in a critically stressed state.

1. Introduction

Earthquake triggering by transient stresses from seismic waves of distant main shocks has been documented in many studies [Hill et al., 1993; Gomberg et al., 2001; Prejean et al., 2004; Brodsky and van der Elst, 2014]. Triggering propensity may be correlated with conditions such as volcanic or geothermal activity and extensional tectonics [e.g., Brodsky and Prejean, 2005, Moran et al., 2004; Harrington and Brodsky, 2006]. Most cases of observed remote dynamic triggering are where natural earthquakes tend to occur, but recent studies show triggering in regions with low levels of historical seismicity, such as near-active fluid injection sites. Locations where triggering occurs may therefore reveal regions of critical ambient stress conditions, independent of proximity to plate boundary faults [van der Elst et al., 2013]. In the following, the term induced seismicity refers to events that nucleate as the result of anthropogenic activity (e.g., fluid injection), whereas triggered seismicity refers to events triggered by natural stress perturbations (e.g., from passing seismic waves), regardless of how the seismogenic faults were brought to a critically stressed state.

Here we perform a broad search for triggering in Canada using earthquake catalog data, followed by a more detailed search in Alberta, where denser seismic station coverage monitors seismic activity associated with fluid injection related to oil and gas production [Wetmiller, 1986; Stern et al., 2013; Schultz et al., 2014, 2015b]. Our statistical tests of triggering using catalog data suggest that various locations throughout Canada have a propensity for triggering, particularly in western Canada (e.g., Cascadia Subduction Zone, along the border of the Northwest Territories and the Yukon), eastern British Columbia and western Alberta near regions of oil and gas production, southwest Ontario, and along the St. Lawrence rift system in eastern Canada (Figure 1).

Following the Canada-wide catalog study, we perform a more detailed catalog search combined with a matched-filter approach in Alberta. The reason for focusing on Alberta is that station coverage has become denser in the last 5 years, and the area is known to experience a significant amount of induced seismicity. The relative abundance of waveform data also means that the earthquake catalog is more complete. Waveform data analysis reveals evidence of both direct triggering in the surface wave train of one main shock in 2014, and evidence of delayed triggering within 12 h of the local arrival times of three main shocks. We first describe the catalog study for Canada, followed by the waveform study for Alberta and the results, followed by a discussion and conclusions section.
2. Catalog Study

We begin our search for triggering in Canada by identifying potential triggering main shocks between 2004 and 2014 with depths < 100 km and surface wave magnitude $M_S > 6$ in the International Seismological Center catalog (ISC). We use the $M_S$ scale as surface wave shaking is the dominant factor for dynamic triggering [Hill et al., 1993; Brodsky et al., 2000; Hill and Prejean, 2007]. We further restrict the list of main shocks to earthquakes that produced an estimated peak ground velocity (PGV) shaking $> 0.2$ cm/s in any part of Canada, where PGV values over much of the country could be well below that threshold. The 0.2 cm/s cutoff is based on documented cases of dynamic triggering in a variety of geologic and tectonic settings.
Using a lower threshold would likely not cause the rate of triggering to vary significantly given the sparse station coverage in Canada, and a higher triggering threshold may result in too few main shocks and thus too few events for a statistical analysis.

PGV is estimated using the following empirical ground motion regression [Lay and Wallace, 1995]:

$$\log A_{20} = M_c - 1.66 \log_10 \delta - 2$$ (1)

$$\text{PGV} = \frac{2 \pi A_{20}}{T}$$ (2)

where $\delta$ is epicenter-station distance in degrees, $T$ is surface wave period ($T = 20 s$), and $A_{20}$ is peak waveform amplitude filtered at 20 s. Table S2 in the supporting information lists the 19 main shocks satisfying the above criteria.

We then count earthquakes in Canada using the NRCan catalog (completeness threshold of $\sim 3$) (Figure S1 in the supporting information) in 10 day windows before and after each main shock. Events are then grouped into $1^\circ \times 1^\circ$ spatial bins. The 10 day window is chosen to maximize the number of recorded earthquakes temporally associated with the main shock. Ideally, the size of the spatial bin should be smaller than the wavelength of the dominant triggering phase. However, we need to choose the $1^\circ \times 1^\circ$ bins to contain more than just a few earthquakes due to the low level of recorded seismicity [van der Elst and Brodsky, 2010].

The $\beta$ statistic is a widely accepted quantitative measure of the level of dynamic triggering, representing the standard deviation in the background seismicity rate following a remote dynamic stressing event [Matthews and Reasenberg, 1988; Reasenberg and Simpson, 1992; Hill and Prejean, 2007]. The $\beta$ statistic is given by the follow equation for each $1^\circ$ bin [Matthews and Reasenberg, 1988]:

$$\beta(N_1, N_2, t_1, t_2) = \frac{N_2 - E(N_2)}{\sqrt{\text{var}(N_2)}}$$ (3)

where $N_1$ and $N_2$ are numbers of earthquakes and $t_1$ and $t_2$ are time-window lengths before and after the main shock. For a Poisson process, $\text{var}(N_2) = N_2 \times t_2 / t_1$, where $t_1 = t_2$ is assumed here. $E(N_2) = N_1 \times t_2 / t_1$ represents the expected number of earthquakes after the main shock based on the premain shock seismicity rate. If no earthquakes occur in a given bin before the main shock, $N_1$ is set to 0.25 based on the equivalent range of the probability density function [Matthews and Reasenberg, 1988; Hough, 2005]. In general, a $\beta$ statistic > 2 indicates a significant increase in seismic activity at a 95% confidence level [Hill and Prejean, 2007].

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Figure 1 shows the $\beta$ statistic in Canada for all 19 main shocks with all earthquakes stacked in each bin, to compensate for the low seismicity levels. Figure S2 in the supporting information shows the $\beta$ statistic for each main shock without stacking. The apparent increase in triggering propensity after 2007 is likely an artifact of the decrease in catalog completeness magnitude from approximately $\sim 2.45$ for the period of September 2006 to December 2010 [Stern et al., 2013; Schultz et al., 2015a]. Four main shocks in Table S3 in the supporting information satisfy the criteria for triggering main shocks during the AGS catalog period. We also calculate the $\beta$ statistic following main shocks in 2014 using the Alberta Composite (AC) catalog running from January to November 2014 ($M_c \sim 2.8$, Figure S1 in the supporting information). Three potential triggering main shocks occurred during the time period of the AC catalog: the 25 July 2014 Ms 6.0 Southeastern Alaska earthquake, the 24 August 2014 Ms 6.1 South Napa earthquake, and the 14 October 2014 Ms 7.3 El Salvador earthquake. Note the three main shocks in 2014 generated the largest PGVs (0.07 mm/s, 0.10 mm/s, and 0.05 mm/s, respectively) in Alberta during the AC catalog period, but all fall below our initial threshold cutoff of 0.2 cm/s. Figures 1b and 1c show the stacked $\beta$ statistic using the AGS and AC catalogs, respectively ($\beta$ statistic for individual main shocks of the AC catalog are shown in Figure S3 in the supporting information). Shaded areas are indicative of triggering in parts of western Alberta, which we discuss in more detail in section 5.
3. Waveform Study

The TransAlta Dam Monitoring network (network code TD) became operational in 2014 in western Alberta. Given the high station density and the propensity for triggering near the TD network, we focus our waveform study around the TD array. Earthquake waveforms recorded at TD stations show that station TD012 has the highest signal-to-noise ratio (SNR) for most events; therefore, we use station TD012 in a matched-filter approach to detect uncataloged events.

3.1. Matched-Filter Approach

We apply a single-station matched-filter approach using continuous waveform data recorded at station TD012 to search for uncataloged earthquakes based on similarity to known events. Similar methods have been used to identify tectonic tremor in Japan and microearthquakes in the U.S. [Shelly et al., 2007; van der Elst et al., 2013]. The matched-filter approach searches for events by cross-correlating known events or event stacks, referred to as templates, with continuously recorded data. Detection is declared when the cross-correlation coefficient exceeds some predetermined threshold set by the user.

Western Alberta experiences a large number of what are likely hydraulic fracturing related events, such as the $M_s$ 4.4 earthquake near Fox Creek on 23 January 2015 [Schultz et al., 2015b]. A $M_s$ 3.8 earthquake southwest of Rocky Mountain House also occurred on 9 August 2014 near previously documented induced earthquakes [Wetmiller, 1986; Baranova et al., 1999; Stern et al., 2013], although the event was not spatially or temporally correlated with any reported hydraulic fracturing activity [Baranova et al., 1999]. We use the 9 August 2014 earthquake and seven other similar events recorded in the AC catalog to construct five templates for each channel of station TD012 (Figure S4 in the supporting information). (Note that some templates are stacks of events, while some are individual events). To create templates, we cut waveform data filtered between 5 and 12 Hz in 15 s time windows that include $P$ and $S$ arrivals of the known events. Templates consisting of stacked events are aligned on the $S$ wave arrivals. Figure S5 in the supporting information shows three-component waveforms of the five templates created for this study. We then cross-correlate all templates with three components of the continuous data at station TD012 to search for uncataloged events surrounding the three main shocks in 2014. We declare an event detection when the cross-correlation coefficient exceeds 5.5 standard deviations above the noise level. The 5.5 standard deviation cutoff value is determined by trial and error based on the tradeoff between detection sensitivity and the acceptable number of false detections. We then remove false detections by inspecting waveforms at all stations high-pass filtered at 1 Hz. Only events that can be identified at more than one station are regarded legitimate detections.

The delay between the origin time of a dynamically triggered earthquake and the imposed transient stress could range from seconds to days [Beresnev et al., 1995; Prejean et al., 2004; Hill and Prejean, 2007]. Therefore, we look at the changes in background seismicity in 12 h time windows on either side of the main shock for the waveform study. The primary reason for using the shorter time window compared with the catalog study is to avoid a tenuous causal link between dynamic stress and local seismicity rate with a longer time window [Hill and Prejean, 2007]. Second, most of the TD stations began operation on 23 July 2014, collecting just 2 days of waveform data before the 25 July 2014 Southeastern Alaska earthquake. The 12 h window guarantees we compare seismicity levels for equal time periods before and after the Alaska main shock.

Figure 2 shows the histograms of earthquakes detected with the matched-filter approach. After removing false detections, we observe 9 and 17 earthquakes in the 12 h time windows before and after the 25 July 2014 Southeastern Alaska earthquake ($\beta$ statistic of 2.67), 28 and 43 earthquakes before and after the 24 August 2014 South Napa earthquake ($\beta$ statistic of 2.83), and 8 versus 14 earthquakes before and after the 14 October 2014 El Salvador earthquake ($\beta$ statistic of 2.12). The $\beta$ statistic exceeding 2 suggest statistically significant triggering following all three main shocks. We also calculate the Poissonian $p$ values for each event, assuming the null hypothesis of a large $p$ value is consistent with no triggering. We calculate $p$ values of 0.063, 0.039, and 0.11, suggesting triggering at the 94%, 96%, and 89% significance level.

3.2. Dynamic Stresses Imposed Locally by Passing Surface Waves

In addition to the events detected with the matched-filter approach, the filtered main shock waveforms show two events that were directly triggered in the Rayleigh wave train of the 14 October 2014 El Salvador main shock (Events 1 and 2, Figure 3). We use the particle motion measured from the waveforms at the times of Events 1 and 2 to estimate the magnitude of triggering stresses [Lay and Wallace, 1995; Brodsky and Prejean, 2005; Peng et al.,...
et al., 2009, 2010; Hill, 2012). We remove the instrument response and rotate the low-pass filtered waveforms (corner at 10 s) to the great circle path. Using particle velocity values and assuming a crustal shear modulus rigidity ($\mu$) of 32 GPa and estimated Rayleigh wave velocity $V_R$ of 2.85 km/s for Alberta [Gu and Shen, 2015], we calculate the dynamic stress ($\tau$) and strain ($\varepsilon$) using the following equations [Love, 1927]:

$$\tau = \frac{PGV \times \mu}{V}$$

$$\varepsilon = \frac{PGV}{V_R}$$

The estimated peak transient stresses during Events 1 and 2 are 0.35 kPa and 0.16 kPa, respectively. Table S1 in the supporting information includes a complete list of peak strains and stresses due to all three 2014 main shocks. We discuss the particularly small values of peak stresses further in section 4.

3.3. Location of a Directly Triggered Event and Focal Mechanism Solution of a Nearby Earthquake

The directly triggered events following the 14 October 2014 El Salvador earthquake are too small to calculate focal mechanisms. Therefore, we calculate the focal mechanism for the cataloged 9 August 2014 $M_L$ 3.8 Rocky Mountain House earthquake and compare it to the ambient stress field (World Stress Map Project, GFZ, German Research Center for Geoscience) using the generalized Cut and Paste (gCAP) method, which is based on waveform inversion [Zhao and Helmberger, 1994; Zhu and Helmberger, 1996; Zhu and Ben-Zion, 2013]. The moment tensor

Figure 2. Histograms of earthquakes in Alberta detected near TD stations using a matched-filter approach 12 h before and after main shocks in Table S1 in the supporting information: (a) 25 July 2014 Southeastern Alaska earthquake, (b) 24 August 2014 South Napa earthquake, and (c) 14 October 2014 El Salvador earthquake. $N_1$ and $N_2$ represent number of earthquakes before and after main shocks, respectively. Red lines indicate local arrival time.
calculation models $P$ and/or $S$ waves in the frequency band of $0.01$–$0.5$ Hz, showing a variance reduction of $61.1$, with optimal source depth at $2.8$ km. (A variance reduction value above $50$ is considered a good fit.) As shown in Figure 4, the 9 August 2014 Rocky Mountain House earthquake occurred on a thrust fault consistent with optimal faulting orientation determined by the regional ambient stress field (Figure 4). While preliminary studies of the earthquake suggest no spatial or temporal link to local hydraulic fracturing activity, the shallow focal depth of $2.8$ km and history of induced earthquakes in close proximity makes it difficult to definitively rule out.

Of the two directly triggered events, only Event 2 was large enough to determine a location. We locate Event 2 using the open-source program Hypoinverse and manually picked $P$ wave arrivals at eight stations with the velocity model for Alberta [Klein, 2002] (Table S4 in the supporting information). Due to the low SNR at many

Figure 3. Two earthquakes directly triggered in the Rayleigh wave train of the 14 October 2014 El Salvador main shock. (a) Main shock phase arrivals indicated with red dashed lines on original radial, transverse, and vertical waveforms. Blue and green waveforms represent directly triggered events. (b) Directly triggered Event 1 (blue) on station TD012 and main shock surface wave particle velocities low-pass filtered at $10$ s (red). Peak transient dynamic stress is $0.29$ kPa. (c) Directly triggered Event 2 (green) on station TD012 and main shock surface wave particle velocities low-pass filtered at $10$ s (red). Peak transient dynamic stress is $0.13$ kPa. (d and e) Event 1 and Event 2 normalized waveforms, band-pass filtered between $0.9$ and $25$ Hz, on all stations with visible signal. Red lines in Figure 3e denote $P$ wave arrival time picks.
stations, we refine our $P$ phase picks using cross-correlation lag times relative to the arrival recorded at station TD012 (Figure S6 in the supporting information). The blue star in Figure 4 indicates the epicenter of Event 2. We calculate the magnitude of Event 2 using the equation below with a nearby cataloged $M_L$ 3.0 earthquake (epicenter shown by a purple star) as the reference event: 

$$M_1 - M_2 = \log_{10} A_1 - \log_{10} A_2.$$ 

where $A$ is the peak amplitude measured at the station in millimeters. Assuming a similar source-to-station distance between the two earthquakes and station TD012 gives an estimated magnitude $M_L$ of 1.0 (Figure 4).

### 4. Discussion

Figures 1b and 1c suggest that isolated locations are susceptible to triggering after the TD stations became operational. One reason triggering may appear isolated could be that many potentially triggered earthquakes fall below the completeness threshold. The observations of direct triggering exemplified in Figure 3 suggest that at least some triggered events have magnitudes of 1 or lower, while the estimated catalog completeness threshold in this region is currently at $M_L \sim 2.8$ (Figure S1 in the supporting information).
Western Alberta is not a region associated with elevated seismicity similar to British Columbia (a plate boundary) or the St. Lawrence seismic zone, yet it experiences comparable levels of triggering. The fact that dynamic stress perturbations of a fraction of 1 kPa (compared to values of 1–5 kPa in other studies) can trigger seismic events strongly suggests that local faults hosting triggered events are in a critically stressed state, which could be related to the presence of fluids [Brodsky and Prejean, 2005; Doan et al., 2006; Elkhoury et al., 2006]. If permeable structures allow fluid flow into preexisting fault zones, elevated pore pressure could effectively unclamp faults, resulting in seismic slip on those optimally oriented in the ambient stress field [Brodsky and Prejean, 2005; Hill, 2012]. A recent study of the Brazeau seismicity cluster near the Cordel Field, west central Alberta, found a strong temporal correlation of seismic events with monthly operations at a nearby disposal well [Schultz et al., 2014]. However, the detailed source characteristics (e.g., fracture energy and stress drop) of triggered and induced earthquakes are needed in order to quantitatively evaluate the fault stress state and further elucidate physical mechanisms. An interesting question would be to query how triggering propensity changes with time with respect to injection activity; however, given the recent changes installation in the TD network, it is difficult to quantify triggering rates prior to 2014 at the same level. Ideally, such a study would have to be conducted before and after oil and gas production commences.

While two triggered earthquakes occurred during the wave train of the 14 October 2014 El Salvador earthquake, the other two main shocks in 2014 failed to trigger events during surface wave shaking, despite higher peak dynamic stress amplitudes (Table S1 in the supporting information). It is possible that the three main shocks generated a variation in Coulomb stress changes on the recipient faults, depending on orientation. However, given the difficulties in knowing triggering fault orientation, we look to other factors that may be related to triggering stress magnitudes, such as low-frequency shaking [Brodsky and Prejean, 2005; Peng et al., 2010; Hill, 2012]. Figure S7 in the supporting information shows the 14 October 2014 El Salvador earthquake has larger amplitude shaking at periods below 20 s, indicating that the long-period waves could be more important for triggering. Several studies also indicate that dilatational strain generated from passing seismic waves can alter pore pressure, flow velocities, and permeability at a given frequency [Brodsky et al., 2003; Elkhoury et al., 2011; Candela et al., 2014, 2015]. Enhanced permeability could drive flow flushing temporary blockages from fractures, thereby redistributing the pore pressure and reducing the effective normal stress, bringing faults to failure.

We also examine the duration and cumulative energy of the main shocks to see if they provide plausible explanations for triggering consistent with the observations [Hill et al., 1993; Sturtevant et al., 1996; Brodsky et al., 2000]. Figure S8a in the supporting information shows that the two directly triggered earthquakes occur nearly 1800 s and 2000 s after the onset of shaking from the El Salvador earthquake, while shaking from the other two main shocks lasts less than 1500 s. Thus, we cannot exclude duration as a criterion for direct triggering. Figure S8b in the supporting information indicates the 24 August 2014 South Napa main shock has the highest cumulative energy and an absence of directly triggered events, suggesting cumulative energy may not be an important criterion for triggering.

The observed stress perturbations associated with direct triggering are lower than typical ranges of 1–5 kPa observed in other studies [Brodsky and Prejean, 2005; Aiken and Peng, 2014]. One reason could be that other studies focus on geothermal and volcanic areas often in extensional or transtensional tectonic regions while our study focuses on fluid injection (Figures 1b and 1c) and gas extraction sites in a compressional tectonic regime [Wetmiller, 1986]. This study provides only limited observations of triggering, but a correlation between triggering capability and tectonic regime has been suggested in previous work [e.g., Harrington and Brodsky, 2006]. Alternatively, the matched-filter approach could be more efficient at finding small, triggered earthquakes, if earthquake magnitude is somehow correlated with triggering stresses as suggested by Hill et al. [1993]. Although the triggering stresses/strains observed here are small, they are consistent with observed cases of triggering with strain as low as 3 × 10⁻⁹ (equivalent to 0.1 kPa triggering stress) [van der Elst and Brodsky, 2010].

5. Conclusions

Here we provide evidence of remote dynamic triggering in isolated areas of Canada using catalog data and in Alberta using both catalog and waveform data. Our catalog study suggests that much of the observed
propensity for triggering occurs near active faults, but that in some cases, triggering occurs outside of active seismic zones. The propensity for triggering may be correlated with the propensity to induce earthquakes. A more localized waveform study of triggering near the TransAlta network in Alberta area shows statistically significant increases in seismicity in 2014, after station coverage increased significantly. The waveform study also provides two clear examples of directly triggered events within the surface wave train of the 14 October 2014 El Salvador earthquake. Transient stresses measured at the time of triggering were 0.35 kPa and 0.16 kPa, and transient dynamic strain amplitudes were $3.9 \times 10^{-9}$ and $4.0 \times 10^{-9}$, respectively. Although the observed dynamic stress and strain amplitudes are lower than those in many previous studies, the dynamic strain is 1 to 3 times of $3 \times 10^{-9}$, which is interpreted to be an observational limit of strain perturbation for earthquake triggering in California [van der Elst and Brodsky, 2010]. The triggering sensitivity to such low strain and stress perturbations suggests preexisting faults in the study area may be critically stressed. Finally, the waveform study of triggered events suggests that long-period shaking below 20 s may be more important for triggering than cumulative shaking or cumulative energy.

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