Triggered Seismic Events along the Eastern Denali Fault in Northwest Canada Following the 2012 M_w 7.8 Haida Gwaii, 2013 M_w 7.5 Craig, and Two $M_w > 8.5$ Teleseismic Earthquakes

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Abstract We conduct a systematic search for remotely triggered seismic activity along the eastern Denali fault (EDF) in northwest Canada, an intraplate strike-slip region. We examine 19 distant earthquakes recorded by nine broadband stations in the Canadian National Seismograph Network and find that the 2012 M_w 7.8 Haida Gwaii and 2013 M_w 7.5 Craig, Alaska, earthquakes triggered long duration (> 10 s), emergent tremor-like signals near the southeastern portion of the EDF. In both cases, tremor coincides with the peak transverse velocities, consistent with Love-wave triggering on right-lateral strike-slip faults. The 2011 M_w 9.0 Tohoku-Oki and 2012 M_w 8.6 Indian Ocean earthquakes possibly triggered tremor signals, although we were unable to locate those sources. In addition, we also identify many short-duration (< 5 s) bursts that were repeatedly triggered by the Rayleigh waves of the 2012 M_w 7.8 Haida Gwaii earthquake. Although we were unable to precisely locate the short-duration (< 5 s) events, they appear to be radiating from the direction of the Klutlan Glacier and from a belt of shallow historical seismicity at the eastern flank of the Wrangell–St. Elias mountain range. The fact that these events were triggered solely by the Rayleigh waves suggests a different source mechanism as compared with triggered tremor observed along the EDF and other plate boundary regions.

Online Material: Description of Rayleigh-wave polarization detection; figures of spectrograms, S -wave amplitudes, polarization analysis, and static stress changes; tables of earthquake catalogs and velocity model; and animations of tremors and bursts.

Introduction

Dynamic stresses from large earthquakes are capable of triggering a wide range of seismic/aseismic responses at remote distances (Hill and Prejean, 2014). These include instantaneous triggering of microearthquakes (e.g., Velasco *et al.*, 2008; Jiang *et al.*, 2010), deep tectonic tremor (e.g., Peng and Gomberg, 2010), earthquake swarms (Gonzalez-Huizar *et al.*, 2012), slow-slip events (Hirose *et al.*, 2012; Zigone *et al.*, 2012), and near-surface icequakes (Peng *et al.*, 2014). Among these seismic/aseismic responses, tremor has been observed in a number of plate-bounding tectonic environments around the Pacific Rim (Peng and Gomberg, 2010; Beroza and Ide, 2011; and references therein). Compared with earthquakes, tremor is a nonimpulsive, emergent signal embedded with low-frequency earthquakes (LFEs) that are generated by shear slip, typically at depths greater than the seismogenic zone (e.g., Shelly *et al.*, 2007). Icequakes, on the other hand, occur as a result of many physical processes, such as calving, crevassing, and basal shear slip at the bed of the ice sheet (Nettles and Ekström, 2010). Systematic studies of triggered phenomena

such as these not only help us to understand how large earthquakes affect seismic/aseismic processes at remote distances but also improve our understanding of the necessary physical conditions responsible for the generation of the seismic activity.

In this study, we conduct a systematic search of remotely triggered seismic activity along the eastern Denali fault (EDF) in northwest Canada, where relative motion between the Pacific and North American plates changes from transform/strike slip in southeast Alaska to subduction in south-central Alaska (Fig. 1). The Denali fault is a major intraplate strike-slip fault that extends from central Alaska to western Yukon Territory, Canada. It is composed of the central and eastern sections that are segmented by the splay Totschunda fault. The 3 November 2002 M_w 7.9 Denali fault earthquake (Eberhart-Phillips *et al.*, 2003), which ruptured the Susitna Glacier, central Denali, and Totschunda faults, was the largest strike-slip earthquake to have occurred in the United States since the great 1906 San Francisco earthquake. Along

Figure 1. Map view of the study region. Black lines indicate fault traces, gray dots indicate background seismicity during 2010–2013 from the Advanced National Seismic System earthquake catalog, gray triangles indicate AK network stations, and black triangles indicate CN network stations. Focal mechanisms indicate moderate-size events ($M_w > 3.8$) occurring on or near the eastern Denali fault (EDF) since 1995 (see Table S1, available in the electronic supplement to this article). Focal mechanism of the M_w 7.9 Denali Fault earthquake is also marked. White arrow indicates relative Yakutat block–North American plate motion, numbered circles indicate average location of tremor triggered by the (1) Indian Ocean, (2) Haida Gwaii, and (3) Craig earthquakes. Location of study region (black box) with the black triangle marking HYT station. Numbered lines show the great circle path of the (1) Indian Ocean and (4) Tohoku-Oki earthquakes. Stars show locations of the (2) Haida Gwaii, (3) Craig, and (5) Nenana earthquakes. The color version of this figure is available only in the electronic edition.

the EDF, earthquake activity since 1995 has been of small-to-moderate size ($M_w < 6$) with both thrust and strike-slip focal mechanisms (see Table S1, available in the electronic supplement to this article). These focal mechanisms indicate the EDF is a transpressional environment.

We focus on the EDF for the following reasons. First, recent studies have shown that triggered tremor does occur in transpressive environments, such as major strike-slip faults near plate boundaries (e.g., Peng *et al.*, 2009; Shelly *et al.*, 2011; Aiken *et al.*, 2013; Peng *et al.*, 2013). Gombert *et al.* (2012) conducted an initial search of ambient tremor in the aftershock zone of the 2002 Denali fault earthquake without success. Hence, it was not clear whether triggered or ambient tremor could be observed at similar strike-slip faults in intraplate settings. Second, nine broadband seismic stations from the Canadian National Seismic Network (CNSN) were installed along the EDF and recorded continuously since August 2010 (Meighan *et al.*, 2013). Although the station density is sparse, they provide a useful dataset for us to conduct a systematic search for evidence of remotely triggered tremor in this region with the potential for locating possible triggered tremor sources. Third, we search for triggered tremor because (1) we know the potential for triggered tremor is greatest during large-amplitude surface waves of distant earth-

quakes and (2) triggered tremor generally has a higher signal-to-noise ratio than ambient tremor (e.g., Rubinstein *et al.*, 2007; Peng *et al.*, 2008).

Analysis Procedure

Our analysis procedure generally follows that of Aiken *et al.* (2013) and is briefly described here. First, we searched the Advanced National Seismic System (ANSS) catalog for distant earthquakes that occurred between August 2010 and May 2013 and had magnitude $M \geq 5.5$, depth ≤ 100 km, and an epicentral distance ≥ 100 km from station HYT. We select shallow, large-magnitude earthquakes, because these events generate large-amplitude surface waves capable of triggering seismic activity on critically stressed faults. We also choose earthquakes over 100 km from the study region so that the triggered seismic activity we are searching for is not obscured by body-wave coda. Using these parameters, we identified 1369 distant earthquakes. Next, we estimated the dynamic stress of each selected earthquake at station HYT using a surface-wave magnitude M_s relationship (van der Elst and Brodsky, 2010; Aiken *et al.*, 2013). The M_s relationship assumes that the magnitude listed in the ANSS catalog is equivalent to M_s . Because earthquakes that generate > 1 kPa of dynamic stress are known to have triggered tremor in other transform environments (Guilhem *et al.*, 2010), we used this threshold to reduce the selected earthquakes from 1369 to 19 events (see Table S2).

After identifying earthquakes of interest, we obtained 12 hrs of seismic data around the origin time of the distant earthquakes from the nine CNSN broadband stations. We removed the instrument response and rotated the horizontal components to the great circle path (GCP) (radial) and a 90° clockwise rotation direction (transverse). The rotation was performed using the *rotate to GCP* command in Seismic Analysis Code with a normal polarity, which is based on a left-handed coordinate system, with vertical up, radial away from the source, and the left face of the transverse component along the positive radial component. For seismic data without clipped or poorly recorded data, we applied two band-pass filters (1–10, 5–15 Hz) to events that occurred > 1000 km away to detect locally triggered activity and to avoid aftershock P -wave contamination. For events that occurred at distances < 1000 km, we applied a high-pass (> 20 Hz) filter (Guilhem *et al.*, 2010) to avoid contaminating P - and S -wave signals of the mainshock and its early aftershocks (e.g., Fig. 2). Finally, we visually inspected the filtered three-component data for seismic events triggered by surface waves of the distant earthquakes. We characterized long-duration (> 10 s) triggered tremor as emergent signals with no discernible P - and S -wave arrivals that are modulated by surface waves of distant earthquakes (e.g., see the Tremor Triggered by the 2012 Haida Gwaii and 2013 Craig Earthquakes section). In contrast, we characterized short-duration (< 5 s) triggered seismic events as bursts with sometimes sharp, distinguishable peaks that are modulated by surface waves of distant earthquakes (see the Short-Duration Triggered Events section). We also compared

Figure 2. Tremor triggered by the Haida Gwaii earthquake as recorded by station HYT: (a) Instrument-corrected broadband transverse (HHT) and vertical (HHZ) components, (b) high-pass filtered > 20 Hz transverse component, and (c) spectrogram. The dashed line in (c) indicates the 20 Hz corner frequency used in the high-pass filter to remove the *P*- and coda-waves from the distant mainshock. The narrow frequency band near 30 Hz recorded from roughly 350 to 375 s after the mainshock could possibly be either instrumental or local noise. This same frequency band is recorded periodically and has the same duration throughout the 12 hr waveform retrieved for analysis. The color version of this figure is available only in the electronic edition.

the occurrence times of short-duration events (SDEs) with the ANSS catalog and the Natural Resources Canada (NRCAN) earthquake catalog to check whether the SDEs were local earthquakes.

If the triggered tremor was visible on at least three stations, we determined the location of the tremor bursts using an envelope cross-correlation method (Chao *et al.*, 2013). For each station that recorded triggered tremor, we calculated the envelope of the filtered radial, transverse, and vertical components, and then stacked and divided by three to create an averaged envelope function. From the averaged envelopes, we manually identified peaks of each tremor burst and then cross correlated the envelopes to compute travel-time differences between events and station pairs based on these picks. Using a simple two-layer *S*-wave velocity model (see Table S3 for this region (Meighan *et al.*, 2013)), we computed the location of each tremor burst based on the minimum travel-time residual for all possible station pairs. Because depth is not well constrained in the envelope cross-correlation method, we set the depth to be 25 km, a depth similar to that used in previous studies of tremor on strike-slip faults (e.g.,

Shelly, 2009). With our envelope cross-correlation method, epicentral locations of the sources change only by ~5 km when focal depth is varied.

Tremor Triggered by the 2012 Haida Gwaii and 2013 Craig Earthquakes

For the 19 earthquakes examined in this study, we identified four (two clear and two possible) instances of tremor triggered by distant earthquakes (Table 1). Figure 2 is an example of tremor triggered by the 28 October 2012 M_w 7.8 Haida Gwaii earthquake. The mainshock occurred relatively close to the EDF region (~960 km), so the amplitudes and frequency contents of the *P* wave from the mainshock and its coda were quite high compared with other more distant events. Therefore, similar to Guilhem *et al.* (2010), we applied a high-pass (> 20 Hz) filter to reveal local tremor signals. We found more than 10 tremor bursts at station HYT that occurred during the large-amplitude, long-period surface waves. These bursts have relatively long duration (> 10 s) and show no clear *P*- and *S*- wave arrivals, similar to triggered tremor observed in other tectonic settings (Peng and Gomberg, 2010). The corresponding animation with sound is shown in Animation S1. Tremor occurred both during the Love and Rayleigh waves of the Haida Gwaii mainshock. For tremor triggered by the Love wave (~260–380 s after the mainshock), the bursts are clearly visible on stations HYT, YUK5, and YUK6 (Fig. 3a). The average location for this tremor source appears to be near station YUK7 (Fig. 1), but only short-duration (< 5 s) bursts (further examined in the Short-Duration Triggered Events section) were visible at that station (Fig. 3). No small-magnitude local earthquakes were listed in the ANSS or NRCAN catalogs during this time. We time shifted the waveforms to our tremor location using the velocity model of Meighan *et al.* (2013) and phase velocities of the Love and Rayleigh waves to visualize how triggering occurs at the tremor source. After time shifting the waveforms back to the tremor source, we find these tremor bursts are coincident with Love-wave velocity peaks, with the highest triggered tremor amplitude occurring around the onset of the large-amplitude (velocity) Rayleigh wave (Fig. 3b). Unfortunately, tremor triggered by the subsequent Rayleigh wave was only visible on the HYT station. Therefore, we were not able to locate those tremor sources.

Figure 4 shows tremor triggered by the 5 January 2013 M_w 7.5 Craig earthquake that occurred ~630 km away. Similar to the previous case (Fig. 2), we applied a 20 Hz high-pass filter to broadband waveforms to identify locally triggered activity (see Fig. S1). Tremor is observed at station HYT during the Love and Rayleigh waves of the Craig mainshock. These tremor signals were also observable on stations YUK5 and YUK6, but the tremor signals can be seen most clearly on station HYT (Fig. 4a). Unfortunately, there is no clear correlation of all tremor bursts among the three waveforms, and two earthquake-like bursts occurred immediately following the tremors (as seen on station YUK6). The first tremor burst, which occurred around 175 s, is located near the HYT station

Figure 3. Example of tremor and short-duration seismic events triggered by the 2012 Haida Gwaii earthquake. (a) 20 Hz high-pass filtered vertical (HHZ) waveforms at multiple stations. Circles indicate the tremor phase picks used for this location. The epicentral distance from the station to the tremor source is indicated. Gray boxes identify the segments used in the examples shown in Figure 7. (b) Transverse (HHT) and vertical (HHZ) component velocity (V) and displacement (D) waveforms recorded at the HYT station with 20 Hz high-pass-filtered waveform and envelope function. Waveforms are time shifted to the average tremor source location. The dashed lines indicate peaks of the tremor bursts (based on the envelope function) and show their correlation with the surface waves. The black vertical bar indicates the scale for broadband velocity waveforms, and the gray vertical bar indicates the scale for broadband displacement waveforms.

(Fig. 1). By time shifting the waveforms, the first two tremor bursts correlate well with Love-wave velocity peaks (transverse component) generated by the Craig mainshock (Fig. 4b).

Additional Observations of Triggered Tremor

We also observed tremor possibly triggered by two more remote earthquakes (Table 1). The first case is the 11 March 2011 M_w 9.0 Tohoku-Oki earthquake recorded by the YUK1

and YUK3 stations (Fig. 5). We filtered these recordings at the 5–15 Hz frequency range to avoid potential contaminations of the P waves of early aftershocks. Most of the stations appear to be noisy, but tremor triggered by the Rayleigh wave of the Tohoku-Oki mainshock can be seen on stations YUK1 and YUK3 near the United States–Canada border. Because these coherent tremor signals were observable on only two stations, we were not able to locate their sources. However, the tremor source must lie close to the YUK3 sta-

Figure 4. Example of tremor triggered by the Craig, Alaska, earthquake. (a) 20 Hz high-pass-filtered waveforms at multiple stations. (b) Transverse (HHT) and vertical (HHZ) component velocity (*V*) and displacement (*D*) waveforms recorded at the HYT station with 20 Hz high-pass-filtered waveform and envelope function. The symbols and notation are the same as in Figure 3b.

tion due to its earlier arrivals. The tremor is arguably near the fault trace, because a similar coherent signal was not observable on station YUK2, which is farther from the fault trace.

The second case is the 11 April 2012 M_w 8.6 Indian Ocean earthquake recorded by stations YUK5, YUK7, and HYT stations (Fig. 6). The Indian Ocean mainshock has

triggered many microearthquakes and tremor around the world (e.g., Wu *et al.*, 2012), as well as a transient global increase of $M_w \geq 5.5$ earthquakes (Pollitz *et al.*, 2012). Its Love wave also triggered an M_w 3.9 earthquake in central Alaska (Tape *et al.*, 2013), which was recorded by stations in our study region at ~ 2700 s after the Indian Ocean mainshock. We observed consecutive bursts of tectonic tremor

Table 1
Event Information for Triggering Earthquakes*

Date (yyyy/mm/dd)	Origin Time (hh:mm:ss)	Magnitude	Epicentral Distance (km)	Back Azimuth ($^{\circ}$)	Transverse PGV (cm/s)	Vertical PGV (cm/s)
2011/03/11	05:46:24	9.0	5876	283	0.213	0.249 [†]
2012/04/11	08:38:37	8.6	11790	307	0.195	0.079 [†]
2012/10/28	03:04:09	7.7	953	157	1.090	1.310 [‡]
2013/01/05	08:58:19	7.5	626	163	2.500	1.570 [‡]

*Distance, back azimuth, and peak ground velocity (PGV) measured at station HYT.

[†]Possible triggering.

[‡]Clear triggering.

Figure 5. Example of tremor triggered by the Tohoku-Oki earthquake. (a) Comparison between broadband and 5–15 Hz band-pass-filtered envelope. Vertical lines indicate the time window that is shown in more detail in (b)–(c). (b) 5–15 Hz band-pass-filtered north (HHN) component waveforms recorded at all stations, ordered from the most southern station YUK7 to the most northern station BVCY. Triggered tremor appears on stations YUK1, YUK2, and YUK3 during the Rayleigh waves. (c) Broadband transverse (HHT) and vertical (HHZ) components recorded at station HYT. Symbols and notation are the same as in Figure 3.

triggered by the long-period Rayleigh wave between 3500 and 4000 s after the Indian Ocean mainshock in the 1–10 Hz frequency band (Fig. 6). The seismic waves of the M_w 3.9 Nenana earthquake arrived in our study region around the arrival time of the Love wave of the Indian Ocean mainshock, and thus we were not able to detect tremor triggered during the Love wave strictly by visual inspection of the waveforms. We examined the spectrogram of the Indian Ocean mainshock but found no evidence of triggered tremor

signals during this time in a higher frequency band (Fig. S2). We also compared the predicted Nenana earthquake S -wave amplitude attenuation with measurements at seismic stations along the ray path but did not find any increase in the amplitude due to any additional local seismic source (Fig. S3).

We located the tremor sources triggered during the Rayleigh wave to be near the YUK6 station, for which no data were available. However, after time shifting to this location,

Figure 6. Example of tremor triggered by the Indian Ocean mainshock. (a) Comparison between broadband and 1–10 Hz band-pass-filtered envelope. Vertical lines indicate the time window that is shown in more detail in (b)–(c). (b) 1–10 Hz band-pass-filtered waveforms from multiple stations. (c) Transverse (HHT) and vertical (HHZ) component waveforms as recorded at station HYT with 20 Hz high-pass-filtered waveform and envelope. The filtered waveform is intentionally clipped so that weak tremor signals can be shown. (d) Zoom in to show the correlation between tremor bursts and surface-wave peaks. Symbols and notation are the same as in Figure 3.

these tremor bursts do not show consistent correlation during coda of the Love and Rayleigh waves on either the vertical or the transverse component of the broadband data. We did not observe tremor triggered by the M_w 8.2 aftershock, which occurred ~ 2 hrs after the M_w 8.6 mainshock, despite having a similar focal mechanism and occurring at similar epicentral distance (see [E](#) Fig. S4).

Short-Duration Triggered Events

In addition to triggered tremor signals, we identified three distinct episodes of short-duration seismic events triggered primarily by Rayleigh waves of the Haida Gwaii earthquake. In comparison with the long-duration triggered tremor, these brittle events produced short-duration (< 5 s) signals

Figure 7. Example of short-duration seismic events triggered by the Haida Gwaii earthquake. (a) Seismic activity recorded on station YUK7 on the southern end of the EDF. The top three traces show the 20 Hz high-pass-filtered velocity seismograms and an envelope function (log 10), and the bottom two traces show the broadband transverse (HHT) and vertical (HHZ) displacement waveforms. Dashed lines indicate event peaks (based on the envelope function). (b) Seismic activity recorded on station YUK4 near the middle of the EDF. Traces are plotted in the same way as in (a), except that the envelope is in linear scale. The top two traces show poorly recorded horizontal channels. (c) Seismic activity recorded on stations YUK2 and YUK3 on the northern end of the EDF. (d) Tremor triggered near the southern end of the EDF. (e)–(g) Short-duration events triggered near the southern, middle, and northern portions of the EDF. Time windows are marked by gray boxes in (a)–(c) of this figure, as well as in Figure 3.

and can be distinguished from tremor when converted to audio due to differences in their duration and power in higher frequencies (see Animations S1–S2). Peng *et al.* (2012) presented a similar analysis for distinguishing triggered events of the 2011 Tohoku-Oki mainshock by converting the seismic signals into audible sounds. One episode of short-duration triggered seismic events was recorded at station YUK7 (Figs. 3 and 7a). At this station, nine consecutive SDEs coincide with peaks of Rayleigh-wave displacement. A few occurred during the Love waves, but they did not show temporal correlations with the Love-wave velocity peaks. However, this could potentially be a result of propagation delays from the triggered source to the YUK7 station. We also note the two larger-amplitude SDEs occurred during the first and largest two cycles of the Rayleigh wave.

Figure 7b is the second example of short-duration seismic events triggered by the large-amplitude Rayleigh wave starting at ~375 s, as observed on the YUK4 station located near the middle of the EDF in the Yukon Territory. Five larger-amplitude SDEs occurred during the higher Rayleigh-wave amplitudes, and several smaller SDEs occurred just prior

to (5 events) and just after (12 events) the higher Rayleigh-wave amplitudes. Short-duration seismic events occurring after 375 s appear to be phase related to the vertical component of the Rayleigh-wave displacement, though we could only observe this correlation on that component due to significant instrumental noise on the horizontal components. Because this station is rather distant from the northern and southern cluster of stations (Fig. 1), it is clear that these short-duration seismic events represent a different source from those observed on the southern YUK7 station. Similarly, we were not able to locate the short-duration seismic sources, because these signals were only observable on the YUK4 station.

A final example of short-duration triggered events is shown in Figure 7c. Between 400 and 525 s, we identified short-duration signals visible on stations YUK2 and YUK3 located further north near the United States–Canada border. Specifically, we found nine consecutive SDEs (<5 s) occurring during the Rayleigh waves, which are phase correlated with the Rayleigh-wave displacement (see Animation S2). These short-duration brittle events were recorded at station

YUK3 ~ 3 s earlier than at station YUK2, indicating their source is likely closer to station YUK3. The availability of only two stations limits our ability to ascertain an accurate location. However, the combination of Rayleigh-wave polarization analysis and simple geometry (see the [Rayleigh-Wave Detection Analysis](#) section in the electronic supplement) suggests the source may be south-southwest of station YUK3, near the Klutlan Glacier terminus (see Figs. S5–S6). These short-duration brittle events have similar characteristics to glacial tectonic sources (i.e., icequakes) detected in Antarctica that were also triggered primarily during Rayleigh waves of the 27 February 2010 M_w 8.8 Maule, Chile, earthquake (Peng *et al.*, 2014). Furthermore, recent evidence suggests that Alaska network seismic analysts routinely identify glacial events in this region, which are characterized by regionally observed signals deficient in high-frequency energy (> 3 – 5 Hz) (West, 2014). However, in this region, there also exists significant ambient tectonic seismicity at the Canadian eastern extension of the Totschunda–Duke River fault (Figs. 1, S6). We found no local earthquakes listed in the ANSS or NRCAN earthquake catalogs that coincided with the occurrence times of any of these triggered short-duration seismic event episodes, and the region was relatively quiescent prior to the arrival of the surface waves (see Fig. S7). Despite the limited quality of data and paucity of stations, our Rayleigh-wave polarization analysis and local seismic activity suggests the signals represent either small, shallow tectonic earthquakes or icequakes in the near surface.

Earthquake Triggering Potentials

Similar to previous studies (Aiken *et al.*, 2013), we analyzed the peak ground velocities (PGVs), seismic-wave incidence angles, and amplitude spectra of all earthquakes examined in this study to determine factors that promote triggering. When examining PGVs, we estimated the maximum dynamic stress each earthquake produced by

$$\sigma = \frac{(\mu)(\text{PGV})}{v_{\text{ph}}}, \quad (1)$$

in which σ is estimated dynamic stress, μ is an assumed crustal shear rigidity of 30 GPa, PGV is measured from each earthquake's waveform, and v_{ph} is an assumed Love-wave phase velocity of 4.1 km/s. Figure 8a shows dynamic stress σ computed from the measured PGVs of the instrument-corrected transverse component at station HYT as a function of back azimuth for each event (i.e., wave incidence angle on the EDF). To avoid contamination of the PGV measurement by local seismic activity, we applied a 1 s low-pass filter to the waveforms prior to measuring the PGV. We also examined the surface-wave amplitude spectra for each earthquake in our study (Fig. 8b). We cut the seismic data between the 5 and 2 km/s arrivals to capture the surface-wave energy and then computed the fast Fourier transform. Finally, these data were smoothed with a five-point sliding window (Peng *et al.*, 2009).

From our analysis of PGVs, wave incidence angle, and amplitude spectra, we find that large-amplitude (> 10 kPa), long-period (> 20 s) surface waves are responsible for triggering tremor and brittle seismic events along the EDF. The 2012 Haida Gwaii and 2013 Craig earthquakes triggered clear tremor with both the Love and Rayleigh waves, and we observed possible tremors that may have been triggered by the Rayleigh wave for the more distant 2011 Tohoku-Oki and 2012 Indian Ocean earthquakes. In addition, it is notable that 2012 Haida Gwaii and 2013 Craig earthquakes exhibit similar amplitude spectra and dynamic stress, but only the 2012 Haida Gwaii earthquake triggered short-duration seismic events.

Discussion and Conclusion

In this study, we provided clear evidence of remote triggering along the EDF. Tremor sources with durations (> 10 s) appear to be localized in the southern cluster of CNSN stations on either side of the fault near the epicenters of moderate-sized earthquakes (Fig. 1) and possibly near the northern cluster of stations (Fig. 5). In comparison, short-duration (< 5 s) seismic events—signals with sharp peaks—triggered only by the Haida Gwaii mainshock were recorded by several stations in the northern (stations YUK2 and YUK3), middle (station YUK4), and southern (station YUK7) portions of the EDF in the Yukon Territory (Figs. 1 and 7). Because fewer than three stations recorded these signals, we were unable to locate these sources. However, short-duration seismic events recorded by stations YUK2 and YUK3 appear to radiate from the vicinity of ongoing seismicity occurring along the eastern extension of the Totschunda fault, as well as near the Klutlan Glacier terminus (see Fig. S6).

Triggered tremor has been observed along several strike-slip faults in the Western Hemisphere (Fig. 9). It is evident that tremor triggered along the EDF has characteristics similar to that of tremor triggered in other regions (Guilhem *et al.*, 2010; Peng and Gombert, 2010); that is, they contain relatively long-duration (> 10 s) signals with no clear P - and S -wave arrivals and are coincident with Love- and Rayleigh-wave peaks. We observed tremor triggered primarily by the Love wave of the Haida Gwaii (Figs. 2 and 3) and Craig earthquakes (Fig. 4), with both events generating surface waves with near strike-parallel incidence. This is consistent with the Coulomb triggering model—maximum left-lateral shear stress induced by a Love wave with strike-parallel incidence on a vertical strike-slip fault (Hill, 2012). In contrast, it appears that tremor was triggered by only the Rayleigh wave for the Tohoku-Oki (Fig. 5) and Indian Ocean (Fig. 6) mainshocks, even though Love-wave triggering is also predicted by the Coulomb triggering model, because the waves from both earthquakes have near strike-parallel incidence on the EDF (Hill, 2012). Specifically for the Indian Ocean mainshock, we could not confirm Love-wave triggering visually due to overlapping seismic waves from the triggered M_w 3.9 Nenana earthquake in central Alaska (Tape *et al.*, 2013). Nor could we identify

Figure 8. Triggering potential (dynamic stress, wave incidence angle, and frequency dependence) of distant earthquakes. (a) Transverse dynamic stress versus back azimuth for all earthquakes examined in this study. Squares indicate triggering and possibly triggering earthquakes. Circles indicate nontriggering earthquakes. Arrows indicate normal and parallel seismic wave incidence on the fault surface. (b) Transverse velocity amplitude spectra for all earthquakes examined. Thin lines indicate nontriggering earthquakes, and thick numbered lines indicate triggering earthquakes. Number indicates the event name and corresponds to the tremor locations in Figure 1 and squares in part (a) of this figure. The color version of this figure is available only in the electronic edition.

Figure 9. Comparison of triggered tremor observations along four strike-slip faults. Each small panel shows the surface waves of an earthquake (broadband) and of triggered tremor (filtered) with the station and component marked. Triangles indicate station where earthquake and tremor was recorded. The gray lines indicate plate boundaries. (a) 2012 Haida Gwaii, Canada. (b) 2011 Tohoku-Oki, Japan. (c) 2012 Haida Gwaii, Canada. (d) 2010 Maule, Chile.

evidence for triggered tremor in a higher frequency band (see ⑤ Fig. S2) or from comparisons of the predicted Nenana earthquake *S*-wave amplitude attenuation with measurements at seismic stations along the ray path (⑤ see Fig. S3).

One may argue that the plate structure in this region is complex and that the triggered tremor could be radiating from the subduction interface rather than the Denali fault

(Fig. 1). Recent studies have detected ambient and triggered tremor in south-central mainland Alaska (Peterson and Christenson, 2009; Gomberg and Prejean, 2013). The tremor discovered in those studies lie where the Yakutat terrane is transitioning from flat-slab subduction in the west to a zone of collision in the east, resulting in a fold-and-thrust belt (Worthington *et al.*, 2012). This region is > 300 km from

where we observed triggered tremor in the southern portion of the EDF in the Yukon Territory, and thus the low-amplitude tremor signals we observed cannot be radiating from this part of the complex subduction zone. This does not rule out the possibility that the observed tremor signals may be nucleating from the fold-and-thrust belt responsible for the St. Elias orogeny. In this region, the strike of the thrust faults mostly trend east–west, as the strike-slip Fairweather fault system transitions to subduction to the west. In this case, triggering waves would have a near strike-normal incidence on the thrust faults. Hill (2012) demonstrated that tremor triggered on low-angle thrust faults best correlates with strike-normal incident Rayleigh-wave displacements. However, we found the triggered tremor best correlates with the Love-wave particle velocity, rather than Rayleigh-wave displacements from the Haida Gwaii and Craig earthquakes (e.g., Figs. 3 and 4), which previous studies also observed on a vertical strike-slip fault (Peng *et al.*, 2008; Hill, 2012). Moreover, the distance to the fold-and-thrust belt is still >100 km, which is likely too far to generate the low-amplitude tremor observed in this study. Given these observations, we surmise that Love-wave displacement gradients with strike-parallel incidence induce shear-stress changes on the EDF, resulting in the small, shear-failure events that we recognize here as tremor.

The four triggering earthquakes (two clear and two possible) produced the greatest dynamic stresses at longer periods (>20 s) than did nontriggering events (Fig. 8b). Although the time period of our study is limited to a few years, it appears that ~ 10 kPa of dynamic stress is needed to trigger tremor in this region. This triggering threshold is greater than that observed in Parkfield, California (e.g., 1–3 kPa) (Peng *et al.*, 2009; Guilhem *et al.*, 2010), but similar to that observed along the Queen Charlotte margin near Haida Gwaii (Aiken *et al.*, 2013). Because these regions are tectonically similar, the triggering threshold may reflect the current strength of a fault, which likely varies over time. However, because station density is sparse in Haida Gwaii and surrounding the EDF and because the data at EDF are only available since 2010, it is possible that such a disparity in observed triggering thresholds between transform and strike-slip regions simply reflects differences in data quality and/or data availability.

Finally, we computed the static stress changes on the EDF based on the finite-fault source slip distribution from Lay *et al.* (2013) for the 2012 Haida Gwaii and 2013 Craig earthquakes, considering their proximity to the region. The resulting static stress changes are less than 0.03 kPa (see Figs. S8–S9), two orders less than the corresponding dynamic stress changes (10–100 kPa). This indicates that the dynamic stresses from the surface waves of the Haida Gwaii and Craig earthquakes are more likely to trigger the tremor and short-duration seismic events, rather than the much smaller static stress changes.

In comparison with the Love- and Rayleigh-wave triggering of tremor, the short-duration seismic events were primarily triggered by the Rayleigh wave of the Haida Gwaii

earthquake (Fig. 7). These short-duration seismic events occurred during the upward motion of the Rayleigh-wave displacement when the dilatation is positive (Miyazawa and Brodsky, 2008; Rubinstein *et al.*, 2009). One could argue that the triggered, short-duration signals reflect single, LFE events known to partially comprise tremor (e.g., Shelly *et al.*, 2007). Although both the triggered tremor and short-duration signals are brittle, shear-failure events, their particle motions differ. For example, the deep tremor signal is dominated by the *S* waves of many LFE events. However, the short-duration signals we observed here behave more like Rayleigh waves (see Fig. S5). Also, although the tremor bursts and short-duration signals have similar frequency contents (up to 30–40 Hz), triggered tremor has weaker power in the higher frequencies over a longer duration (Animations S1–S2). Although we were unable to accurately locate these short-duration sources, their waveform characteristics and similarity to Rayleigh waves allow us to identify the source as south-southwest of station YUK3 (see Fig. S6c), near the Klutlan Glacier terminus, where there is also a historical record of shallow (<10 km depth) seismicity and regional events flagged as glacial events by Alaska Earthquake Center analysts (West, 2014). Because of the poor data quality, it is difficult to distinguish whether the sources are icequakes similar to those observed in Antarctica triggered by the M_w 8.8 Chile earthquake (Peng *et al.*, 2014) or shallow earthquakes like those triggered near active volcanoes or geothermal regions (Prejean *et al.*, 2004; West *et al.*, 2005). Deducing which type of events actually occurred—icequakes or shallow earthquakes—would require more stations in the region.

Previous observations of triggered and ambient tremor in this region were primarily associated with the Alaska–Aleutian subduction zone (Peterson and Christensen, 2009; Brown *et al.*, 2013; Chao *et al.*, 2013). Gomberg *et al.* (2012) also conducted a systematic search but did not find any precursory tremor signals before the 3 November 2002 M_w 7.9 Denali fault earthquake or in several time intervals after the mainshock. Permanent stations near the central Denali fault are predominately located near the initial rupture point of the Denali fault earthquake (e.g., fig. 1 of Gomberg *et al.*, 2012). We searched for tremor on stations surrounding the central Denali fault, and the triggered tremor generally arrives at station KLU prior to PAX (e.g., Fig. S10), suggesting that they mostly occur further south in the south-central Alaska sweet spot (Chao *et al.*, 2013; Gomberg and Prejean, 2013). No tremor signal was observed on station DDM (seismic data archived for only 1 yr, from 2009–2010), located north of the central Denali fault. Although we did not find any tremor signals radiating from the central Denali fault in this short-time frame, tremor may still occur there. However, it is also possible that the tremor-generating conditions are different in our study region surrounding the EDF than the central Denali fault where the M_w 7.9 earthquake occurred. Nevertheless, our observation of triggered tremor along the EDF provides additional evidence of tremor occurring in strike-slip environments. It is important to continue investigating tremor

activity along this intraplate fault in the hope that we can better understand the physical mechanisms responsible for not only tremor generation, but also its role in the seismic cycle of large earthquakes along the EDF.

Data and Resources

The Advanced National Seismic System earthquake catalog is made available by the Northern California Earthquake Data Center via www.ncedc.org (last accessed June 2013). The Natural Resources of Canada earthquake catalog is available at <http://www.earthquakescanada.nrcan.gc.ca/stndon/NEDB-BNDS/bull-eng.php> (last accessed July 2013). Seismic waveforms and response files recorded by the Canadian National Seismic Network are accessible by the Incorporated Research Institutions for Seismology Data Management Center (www.iris.edu; last accessed June 2013) and by the Natural Resources of Canada AutoDRM (http://www.earthquakescanada.nrcan.gc.ca/stndon/AutoDRM/autodrm_req-eng.php; last accessed January 2014). Focal mechanism data in Figure 1 and Table S1 were provided by Honn Kao at Natural Resources Canada. Figures were made using Generic Mapping Tools version 4.5.1 (www.soest.hawaii.edu/gmt, last accessed February 2015; Wessel and Smith, 1998) and proprietary MATLAB (<http://www.mathworks.com/products/matlab/>; last accessed February 2015) software version 7.11.0.584 (R2010b).

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