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UNEXPECTED RESPONSE OF MT. WRANGELL VOLCANO, ALASKA, TO THE SHAKING FROM A LARGE REGIONAL EARTHQUAKE: A PUZZLE FOR INTERMEDIATE-TERM EARTHQUAKE-VOLCANOES INTERACTIONS.

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ABSTRACT

On November 3, 2002 three segments of the Denali fault in interior Alaska ruptured during a Mw 7.9 earthquake, offering a unique opportunity to study earthquake-volcano interactions. Out of the 24 volcanoes that are seismically monitored by the Alaska Volcano Observatory (AVO) only Mt. Wrangell, the closest volcano to the epicenter (247 km), showed a clear response to the shaking in the intermediate-term (weeks to months) time scale. The response was unexpected because it consisted of a decline by at least 50% in the volcano's seismicity rate (mostly low-frequency events) that lasted for five months. Because most well documented previous instances of short-term (minutes to days) responses of volcanic centers to the passing waves of distant earthquakes, have all been seismicity increases, the decline in seismicity at Mt. Wrangell poses a controversial puzzle. By using several independent methods to measure the seismicity rate at the volcano from before to after the main shock, and applying rigorous statistical testing, we conclude that the change in seismicity at the volcano was a real effect of the Denali earthquake. We suggest that a depressurization of the volcanic plumbing system took place either as a result of sudden decompression (static stress changes) or because of creation of new pathways resulting from the strong shaking (dynamic stresses). At present we cannot distinguish between these two possibilities.

Key Words: Alaskan volcanoes, Denali fault earthquake, Mt. Wrangell, seismicity decreases

RESUMEN

En Noviembre 3 de 2002 durante el terremoto de 7.9 Mw, se quebraron tres segmentos de la falla Denali en Alaska, ofreciendo una única oportunidad de estudiar los terremotos por interacción volcánica. Aparte de los 24 volcanes monitoreados sismicamente por el Observatorio Volcanológico de Alaska (AVO), sólo el Mt. Wrangell, el volcán más cercano al epicentro (247 km), mostró una respuesta clara ante el movimiento en un término de escala de tiempo intermedio de semanas a meses. La respuesta fue inesperada porque consiste en un declive de por lo menos 50% en la velocidad sismológica del volcán (frecuencia de eventos cada vez más lenta) que duró hasta cinco meses. Puesto que muchos casos de corto término (minutos a días), muy bien documentados previamente, responden a olas de distantes terremotos centros volcánicos, todos han tenido incrementos en la sismisidad; por tanto, el declive sísmico en el volcán tuvo un efecto real sobre el terremoto Denalie de Mt. Wrangell que ocasionó

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un controvertido rompecabezas. Usando muchos métodos independientes para medir la velocidad sísmica en el volcán desde antes hasta después del principal choque y aplicando rigurosas pruebas estadísticas, concluimos que los cambios sísmicos fueron un efecto real del terremoto Denalie.
Sugerimos también que la despresurización del sistema del volcán tuvo efecto por la descompresión (drásticos cambios estáticos) o por la creación de nuevos caminos fue producto de un movimiento brusco (acentuada dinámica). Hasta el momento no podemos diferenciar entre estas dos posibilidades.

***Palabras clave*:** volcanes de Alaska, terremoto de la falla Delani, Mt. Wrangell, decrecimiento de la sismisidad.

INTRODUCTION

The interactions between earthquakes and volcanoes have been a subject of scientific interest for a long time and, in recent years, a topic of research thanks to the increasing number of reports and the improvement of precise instrumentation. The link between tectonic and volcanic activity is better understood by considering the scales of time and space: at very long times (hundreds to millions of years) and large distances (tens of thousands of kilometers) the connection is explained under the theory of plate tectonics and one only needs to glance at a map of global tectonics to see that the distributions of earthquakes and volcanoes are not the result of mere coincidence.

In the very short time scale of minutes to days and distances of kilometers to tens of kilometers, there is documentation of eruptions following the shaking from an earthquake, such interactions are exemplified by the 1975 summit eruption of Kilauea volcano an hour and a half after the 7.5 Kalapana earthquake (Tilling et al. 1976) and the eruption of Cordon Caulle volcano two days following the magnitude 9.5 1960 Chile earthquake (Gerlach et al. 1988; GVN, 2004). In the intermediate-term scale of weeks to months or years, new evidence and research points towards a positive link, meaning that a regional earthquake with Magnitude 6.5 or larger can trigger unrest at volcanoes located several hundreds of kilometers away (Hill et al, 2002).

In all instances of earthquake-volcano interactions, all observations indicate that a distant earthquake is able to cause changes in activity at volcanic (and non-volcanic) areas in terms of increases in seismicity or eruptions. The opposite effect, however, has not been clearly observed and the literature, with only a few exceptions (Alvarado, et al. McNutt and Sanchez, 2002; Sanchez and McNutt, Submitted), does not include documentation of activity at volcanoes being inhibited by either the transient or permanent changes induced by a distant earthquake.

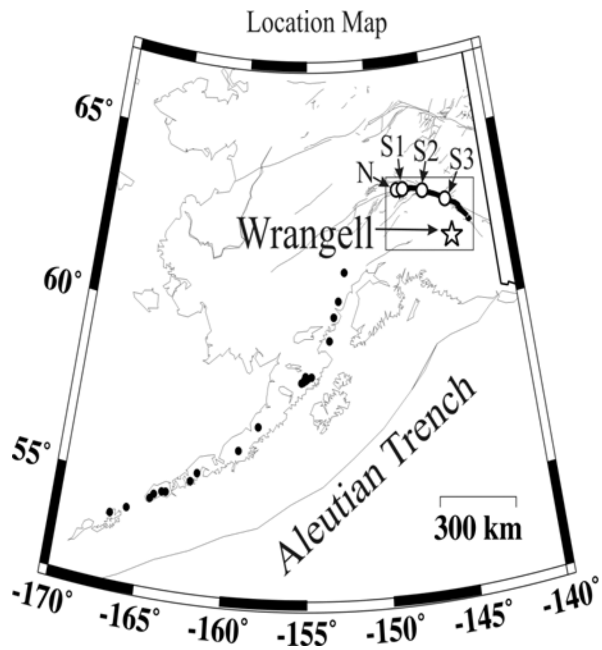


Figure 1. Map of Alaska showing the surface rupture from the DFE (thick solid line) and the epicenters of the October 23, 2002 Nenana Mountain earthquake (N) and the three sub events of the November 3, 2002 Denali earthquake (S1, S2, S3) (Surface rupture from Eberhart-Phillips et. al., 2003, epicenters of N and S1-S3 are from AEIC, 2003); Mt. Wrangell volcano(star); other volcanoes seismically monitored by AVO (dots); and main faults and tectonic structures (gray lines). Rectangle encloses the area of Figure 2.

In November 2002, the Mw 7.9 Denali Fault Earthquake (DFE, Figure 1) triggered short-lived seismicity at volcanic areas located along the rupture direction at distances as far as 3660 km from the epicenter (Eberhart-Phillips et al. 2003) and an apparent intermediate-term drop in the seismicity at Mount Wrangell volcano, located in the perpendicular direction with respect to the rupture and roughly 240 km away from the epicenter. This effect may have lasted for five months.

In this paper we provide evidence, by means of statistical testing, that the decline in seismicity observed at Mount Wrangell volcano was real and can be attributable to the DFE. To do so, we analyze the catalog of located earthquakes at Mount Wrangell, provided by the Alaska Volcano Observatory and test it for the significance of the changes in seismicity following the DFE.

DATA

The Alaska Volcano Observatory (AVO) monitors the seismic activity at Mount Wrangell since July 2000, when two seismograph stations were installed. Locations of earthquakes were possible beginning August 2001, when two additional stations were deployed (Figure "Wrangell area map"). Since then and until the time of this writing (March, 2003) 5,451 events triggered the automatic detection system, but only 508 could be located (Figure 2). The reason for this low detected/located ratio stems from a combination of low signal-to-noise ratio, limited number of seismograph stations and most important of all, the characteristics of the first arrivals of the earthquakes at the different stations. Of all located earthquakes, 86.4% have been identified as low-frequency seismic events (LF). LF events show emergent first arrivals to the seismograph stations, have not distinguishable P- and S-arrivals, and have spectra with dominant frequencies between 1 and 5 Hz (Minakami, 1974). The remaining percentage of seismic events have been identified as high-frequency earthquakes, which may not different in their source mechanism from earthquakes that occur along tectonic areas such as the Denali (Alaska) , San Andreas (California), or North Anatolian faults (Turkey).

Because seismic activity in the Mount Wrangell area is monitored and recorded in continuous and trigger-mode we can use three largely different datasets: digitally filtered waveform data and unfiltered analog seismograms of station WANC (located 3.2 kilometers west of the vent) and the catalog of located earthquakes. We will refer to each of these datasets as pseudohelicorder, helicorder, and located earthquake counts, respectively. The time period covered by our different data sets is September 1, 2002–December 31, 2002 for the pseudohelicorder and helicorder event counts and August 5, 2001–March 18, 2004 for the catalog of located earthquakes.

METHOD

We measure the seismicity rates at Mount Wrangell based on daily event counts out of the three datasets described and the significance of the changes is estimated by applying the z (Habermann, 1987) and β (Mathews and Reasenberg, 1988) statistical tests to the data in both, ordered and randomized time series.

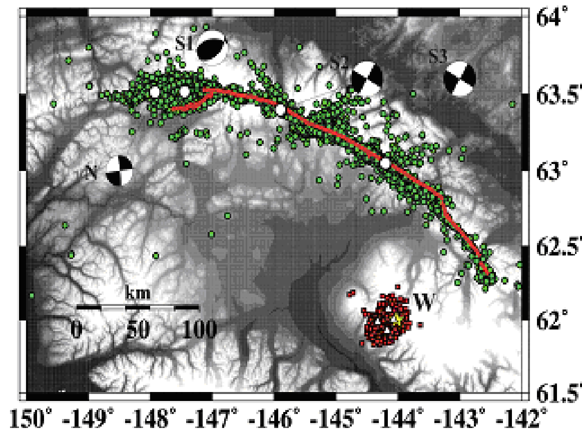


Figure 2. Detail of rectangular area showed in Figure 1. Red thick line: surface rupture from the DFE; fault plane solutions shown for the DFE sequence earthquakes in lower hemisphere projections with compressional areas in black and dilatational areas in white; green circles: Some of the earthquakes located by AEIC between October 23, 2002 and October 31, 2003, with depths above 30 km and Magnitude 3 and above. (4,653 epicenters are shown); red squares: volcanic earthquakes (Lf and HF) at Mt. Wrangell located by AVO between August 5, 2001 and March 18, 2004; white triangles: seismograph stations on Mt. Wrangell. See other conventions as in Figure 1.

We define the seismicity rate changes at the volcano in terms of the rates during the periods before and after the DFE. The z test is a parametric test for estimating the difference between two means, in our case the two means being compared are simply the averages of daily earthquake counts for the period preceding and following the DFE. Z-statistic formula is:

$$z = \frac{(M1 - M2)}{\sqrt{\frac{(S1)^2}{N1} + \frac{(S2)^2}{N2}}} \tag{1}$$

Where M1, M2 mean rates during the two periods (before and after the DFE); S1, S2: standard deviations of the rates; N1, N2: numbers of earthquakes in each time period. The z value computed is interpreted in terms of significance as the number of standard deviations from the mean of a normal distribution (i.e., z = 1.64 represents 90% significance, z = 1.96 1represents 95% significance, and z = 2.57 represents 99% significance) (Habermann, 1987). The sign of the z value indicates the polarity of the change (i.e., $z < 0$ indicates rates increases, $z \approx 0$ indicates no change, and $z > 0$ indicates rate decreases). We also use a moving windows technique to compute the z values. This is done by finding all possible subdivisions of the data in two halves and computing the z value for all pairs of means, allowing us to estimate the significance of a difference in mean seismicity rates at any given point (time) in the curve.

We investigated the sensitivity of the z values versus time curve to the presence of abrupt changes in seismicity rates by computing the z values of several synthetic data sets with varying amounts of change (decreases) in the number of earthquakes per unit time. Our synthetic datasets have the following features: (1) The seismicity rates from before to after the change are different but remain constant within each period; (2) the seismicity rate after the change is some normalized percentage of the seismicity rate before the change; (3) the value of the earthquake count at the middle point of the time series is zero, simulating a useless record the day of the DFE. To somewhat mimic our pseudohelicorder counts data, we created synthetic time series containing one break point and two segments, simulating drops in seismicity of 100% (total drop of seismicity), 95%, 75%, 50%, 25% , and 0% (no change in seismicity). We also investigated the effect of having a seismicity rate that decreases in two steps preceding the main drop. We did this by creating a data set that has two break points and three segments with constant rate each. In this way we simulate a constant rate, and a drop of 25% that precedes a second drop of 50%, after which the seismicity stays at that level. The length of these synthetic data sets is 101 data points.

To mimic our data of located LF seismic events, we created a longer synthetic data set in which there are two break points and three segments. In this case the second segment contains a drop in seismicity of 80% and the third segment represents a recovery of the seismicity rates to background values (rate equal to first segment). The length of this synthetic data set was 487 data points.

Seismicity rates changes, can also be characterized using the -statistic as described Mathews and Reasenberg (1988):

$$\beta = \frac{n_a - E(n_a)}{\sqrt{\text{var}(n_a)}} = \frac{n_a - n_b(t_a/t_b)}{\sqrt{n_b t_a / T}} \qquad (2)$$

Where:

n_a=Number of earthquakes in the region of interest after the perturbing main shock.

t_a=Time period after.

$E(n_a)$=Expected number. $E(n_a) = r.t_a$, and

$r = n_b/t_b$ is the *ambient* rate of earthquakes estimated from a background period, t_b.

n_b= Number during the background period.

Note that for a binomial distribution $var(n_a) = Npq$ where $p = t_a/T$, $q = 1 - p = t_b/T$, $T = t_a + t_b$ and $N = n_a + n_b$. Because the observed value of n_a may be anomalous, $E(n_a)$ is used instead, and $N = n_b(1 + t_a/t_b)$.

The sign of β in this case indicates the polarity of the change in seismicity rate (i.e., β < O represents decrease in rates, $\beta \approx 0$ means no change, and $\beta > 0$ represents rates increases). Because there are various methods to compute β that differ in their underlying assumptions and the validity of those assumptions cannot readily be verified (Gomberg et al. 2003) we do not ascribe absolute significance level to the rate variations computed with β statistic and only use to spatially assess the nature of the seismicity rate changes in the area of Mt. Wrangell.

RESULTS

Pseudohelicorder and Helicorder counts

The plots of cumulative daily number of volcanic earthquakes estimated from pseudohelicorder plots and helicorder records are shown in Figure 3. The absolute values in numbers of events are different because of the different criteria used to count earthquakes on each type of record and the different noise levels (usually less noise in pseudohelicorder plots) and sensitivities. The change in slope indicating the change in seismicity rate is more obvious in the pseudohelicorder counts curve. The z values at the date of the DFE are $z = 6.7358$ and $z = 2.5531$ for pseudohelicorder and helicorder counts, respectively, which correspond to confidence levels for the difference in mean rates before and after de DFE above 99.99% and 99% respectively. From the cumulative number and z values plot for pseudohelicorder counts, we note that the decrease in seismicity begins before the DFE, around the time of the Mw 6.7 Nenana Mountain earthquake (October 23, 2002). The helicorder count data although somewhat noisier also shows the change in rates.

We tested the hypothesis that the data showing the suggested decrease in seismicity at mount Wrangell is no different from background noise, represented by many random series. To do this we randomly permute our data sets of pseudohelicorder and helicorder counts, plot the cumulative number of earthquakes versus time and compute the z values for the permuted series. This is done 100 times for each data set being analyzed and we always plot the randomly permuted series along with the original data. We observe less spread in the random series for pseudohelicorder counts; also the data from pseudohelicorder records equals the random noise towards both ends of the time series, but is different towards the middle of the series. The helicorder count data is no different from the random noise.

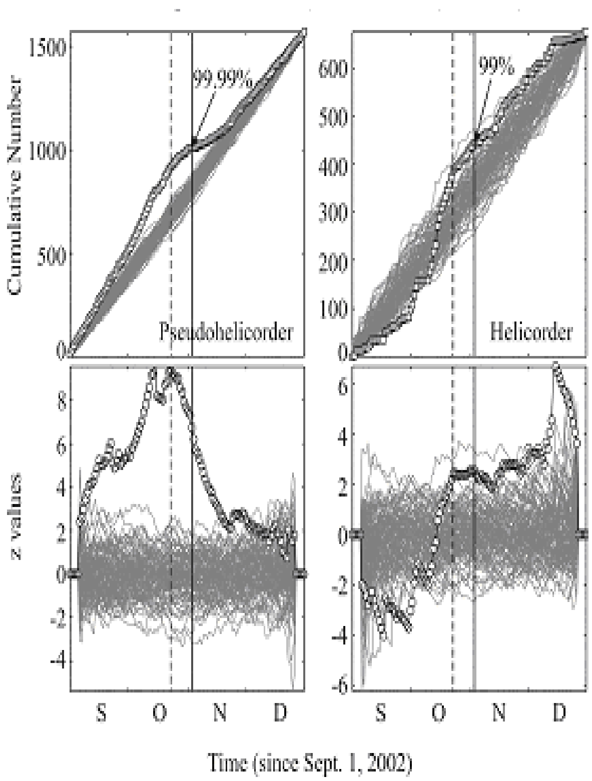
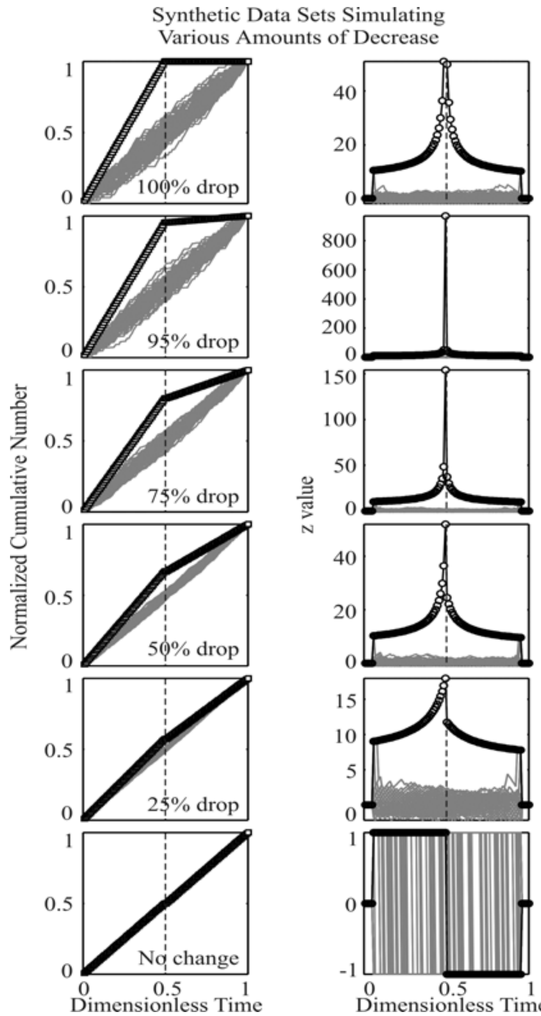


Figure 3. Seismicity rates and z values estimated from earthquake counts on Pseudohelicorder and helicorder records from station WANC on Mt. Wrangell., between September 1, 2002 and December 31, 2002. Top: Cumulative number of volcanic earthquakes versus time; bottom: plot of z values versus time. Squares and circles represent the cumulative sum and z values curves, respectively, for actual data; gray lines represent the result of 100 random permutations of the data in each plot. Vertical dashed and solid lines mark the dates of the Nenana Mountain earthquake (10/23/2002) and the Denali fault earthquake (11/3/2002), respectively. Percentages indicate the confidence levels at which the two means, before and after the DFE, are different as evaluated by the z-test.

Analysis of Synthetic Counts Data- Pseudohelicorder counts.

Figure 4 shows several cumulative numbers versus time and *z* values versus time plots for six synthetic data sets all representing a sharp decrease in seismicity, with varying amounts of change. The results show that the spread in the randomly permuted counts increases as the amount of change in seismicity rate increases and so does the separation between observations and random permutations. This means that the stronger the rate change, the more we should be able to see it above random noise. Also we see that even in the case of small rate changes the z values versus time curve is sharply peaked *near* the time of the change. Careful examination of the *z* values curves reveals that exactly one data point before the time where the rate change starts, the peak occurred. This is



not

The *Figure 4.* Cumulative sum and z values plots for six synthetic data sets with one break point in the middle and varying amounts of rate decrease. A vertical dashed line marks the time of the change, in which a value of zero in the counts has been assigned, simulating a useless record. The top plot of z values, for a drop in seismicity of 100%, has been cut off at z≈50 because the peak value is at infinity. Other conventions as in Figure 3.

caused by the number of data points being either even or odd, or by the presence of a zero count at the time of the change, nor by the position of the time of change in the data set. The effect is expected because the z values curve is tracking the evolution of the cumulative sum in time. Let *n* be the total number of data points and n_i be the position of the breakpoint.

The cumulative curve will be constant until exactly n_i-1 and then drops to a different rate. That is where the two segments being compared are the most different.

An additional interesting result from our testing is the observation that the steepness of the rise in z values depends on the strength of the rate change.

The *z* values curve is sharply peaked for large changes in rate and rises more gradually for small rate changes (Figure 5).

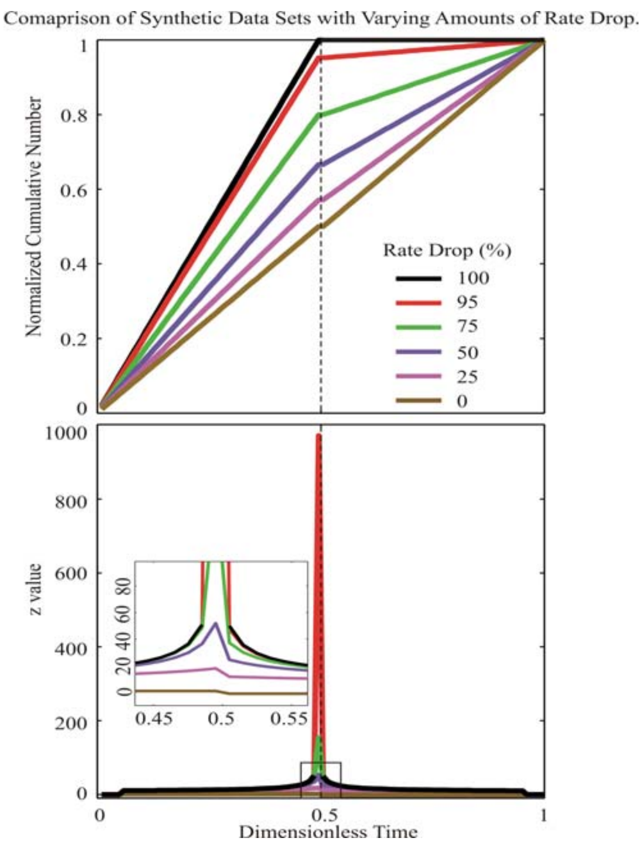


Figure 5. The data sets used in Figure 4, plotted at the same scale to allow comparison of the z values curve. The inset is an enhanced view of the rectangular at the base of the z values plot. See other conventions as in figure 4.

The analysis of a synthetic data set with two break points (Figure 6) shows that when a minor decline in rate precedes a larger one, the *z* values curve is double-peaked, but surprisingly, the first peak has a larger *z* value, regardless of the length of the two segments preceding the main break point. As the length of the second segment shortens, the peaks come together into one. Similar effects are also present in the pseudohelicorder count data, with the two largest peaks in *z* values occurring several days before the date of the DFE. The first peak in z values occurs on October 15, 2002, and it is sensing a zero count value (unusable record) on October 16, after which the counts drop for a few days. We tested the influence of this zero value by replacing it with a value similar to its neighbors and the double peak persisted, although with slightly decreased significance.

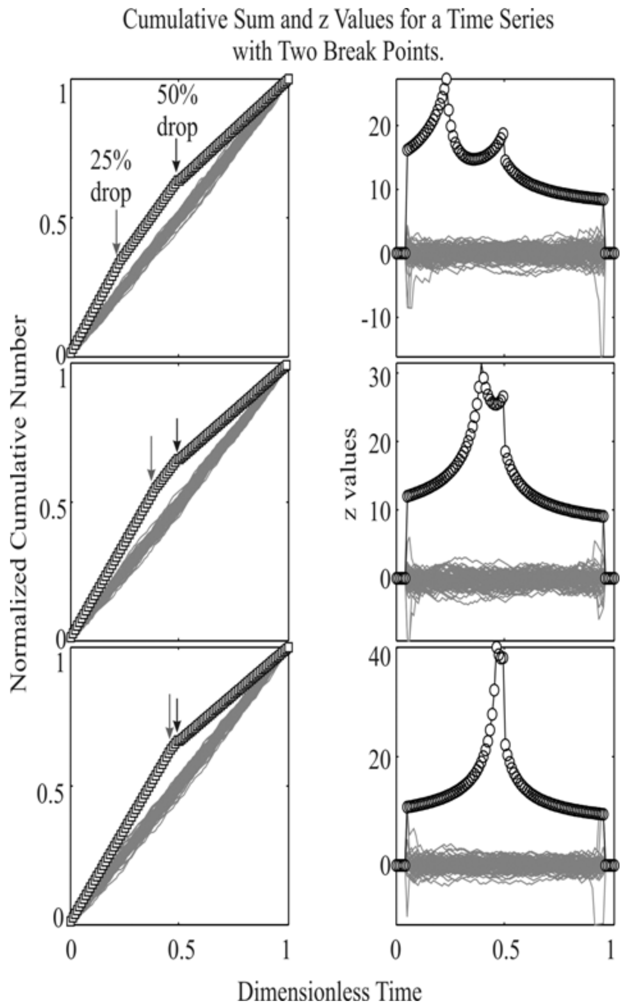


Figure 6. Cumulative number and z values plots for synthetic data with two break points representing an initial drop in rate by 25% followed after sometime by a larger drop of 50%. Vertical arrows mark the time of the drops. The start time of the first (and smaller) drop varies, while the start time of the larger drop is constant. Other conventions as in Figure 3.

The second peak in z values occurs on October 23, 2002 and it is sensing a general decrease in number of daily counts (mostly good days) leading to the strongest drop on November 3. We suggest that the relatively gradual increase in z values leading towards the peak on October 23 reflects the magnitude of the seismicity drop following the DFE, although we cannot reject the possibility that the decrease in seismicity actually started near October 23, 2002.

Analysis of Synthetic Counts Data-Located Earthquake Counts.

When we use a larger data set of 487 data points (equal to the number of located LF seismic events) simulating a time series with a final segment in which the rates recover to the background values, we observed that the *z* values curve has two peaks, the first one representing the drop in seismicity and a second one, with negative *z* value, indicating the recovery (increase) to background rates (Figure 7).

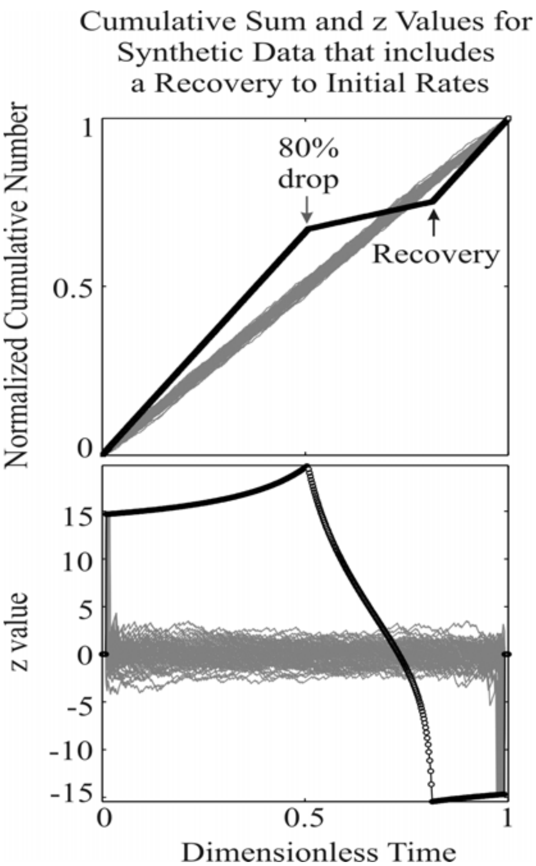


Figure 7. z values and cumulative sum versus time for a synthetic data. This simulates a recovery to pre-perturbation rates following a drop in seismicity by 80%. Vertical gray and black arrows mark the start times of the drop in rate and recovery, respectively. Other conventions as in Figure

Located Volcanic Earthquakes

We examined the catalog of located earthquakes in the Mount Wrangell area from August 2001 through March 2004. In Figure 8 we show a time-depth plot of earthquakes at Mount Wrangell. Notable changes in seismicity rate include the apparent increase associated with the introduction of EARTHWORM acquisition system (Johnson and Others, 1995; Dixon et. al., 2003); the decrease following the DFE, and two other subsequent apparent decreases (July – September 2003 and December 2003) associated with temporary telemetry problems and data disruptions.

We used the computer program ZMAP (Wiemer, 2001) which is suited for analysis of located earthquakes and includes routines to estimate changes in seismicity rates. The catalog includes 508 events with magnitude from -0.2 to 2.7. An automatic estimate of the magnitude of completeness indicates the catalog to be complete for magnitudes $\geq$ 0.9, in agreement for the value reported by Dixon et al. (2003).

To avoid artifacts introduced by artificial rate changes, we select a subset of earthquakes between early March 2002 and late June 2003 (dotted box in Figure 8), also to ensure the completeness of the samples used, we select earthquakes with magnitudes $\geq$ 0.9 for subsequent analysis.

The other decreases in seismicity rates observed in the catalog of are unrelated to large earthquakes along the Denali fault because the most recent event of note was the Mw 5.7 earthquake on 22 October, 1996 (Ratchkovski et al. 2003), but rather these decreases are caused by other factors such as temporary low signal-to-noise ratio or stations outages. We note that the decrease in seismicity at Wrangell following the DFE was unrelated to "artificial" causes because the same analyst located all the volcanic earthquakes and the stations were operational during the time of the decrease.

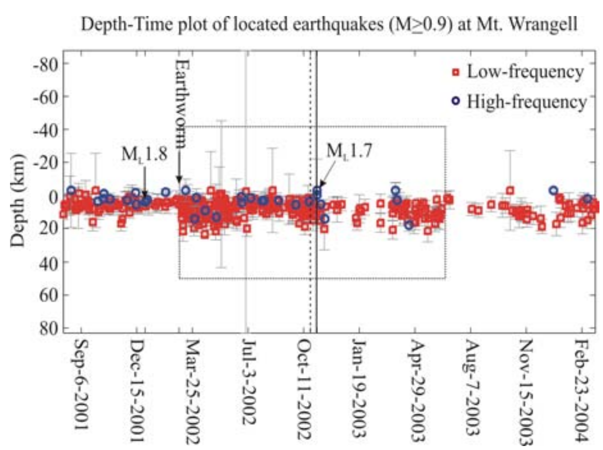


Figure 8. Comparison Depth-time plot for located earthquakes between August 5, 2001 and March 18, 2004 at Mt. Wrangell. Red squares and blue circles with gray error bars mark LF and HF earthquakes, respectively. Only earthquakes with magnitude $\geq$ 0.9 are shown. Arrows indicated the two largest HF earthquakes and the time of change from IASPEI to EARTHWORM acquisition systems. A dotted rectangle encloses the most homogeneous part of the catalog, between March 1, 2002 and June 30, 2003 that is used for all analysis of located earthquakes. Other conventions as in Figure 3.

The catalog of located earthquakes at Wrangell from August 5, 2001 to March 18, 2004, includes at least three decreases of noticeable duration. The only decrease in seismicity at Mt. Wrangell following a large earthquake along the Denali fault is the one that followed the DFE. The other two decreases are artificial (data disruptions caused by temporary telemetry problems) and are not preceded by large earthquakes in the area or by moderate-sized (with magnitude 5.0 or larger) aftershocks of the DFE (AEIC, 2003). Thus although the seismicity rates at Mt. Wrangell fluctuate, the coincidence of the rate decrease onset time with the DFE sequence suggest a causative relation and not the result of chance.

To test the hypothesis that the decrease following the DFE, as estimated from located earthquakes, may not be different from random noise, we follow our previous procedure of randomly permuting our data sets of located earthquakes counts, plot the cumulative number of earthquakes versus time and compute the z values for the permuted series. In this case we repeat the permutation 100 times and show the results along with the original data in Figure 9. We do the analysis for LF and VT earthquakes separately. For LF events the cumulative number versus time plots shows several changes in slope that represent decreases in seismicity. The decrease associated with the DFE sequence, however, represents the breakpoint with the most significance, because the peak in z value happens near the time of the DFE and nowhere else. We found that for LF events the cumulative number and z values curves are different from their respective permuted series. We note that data of located LF seismic events does not show a negative peak in the z values curve that we observed by creating a synthetic data set in which there was a recovery towards background rates.

For HF earthquakes the ordered data is no different from the randomly permuted series. The catalogs of located LF and VT earthquakes show that the decrease in seismicity initiated after the DFE, this is because some earthquakes could be located during the immediate few days following the DFE.

Spatial Mapping of the Amount of Decrease Using β-statistic

To examine spatially the decrease in seismicity in the Mt. Wrangell area we divide the are of interest in a grid with nodes separated 3 km (average of horizontal location errors) and compute the β-statistic using equation (2) for the 30 nearest earthquakes to each node. The two periods being compared to compute the β-statistic are: March 1, 2002 – November 3, 2002 and November 3, 2002 – June 30, 2003. Note that the two time periods are of roughly equal duration (eight months each). The map of the β-statistic for the Mt. Wrangell area is shown in Figure 10 and it shows a general decrease in the seismicity rates everywhere on the volcano. A small area located SW of the vent experienced relatively little to no change, but this is a region with sparse seismicity. The catalog used is rather small, but the seismicity is distributed so that

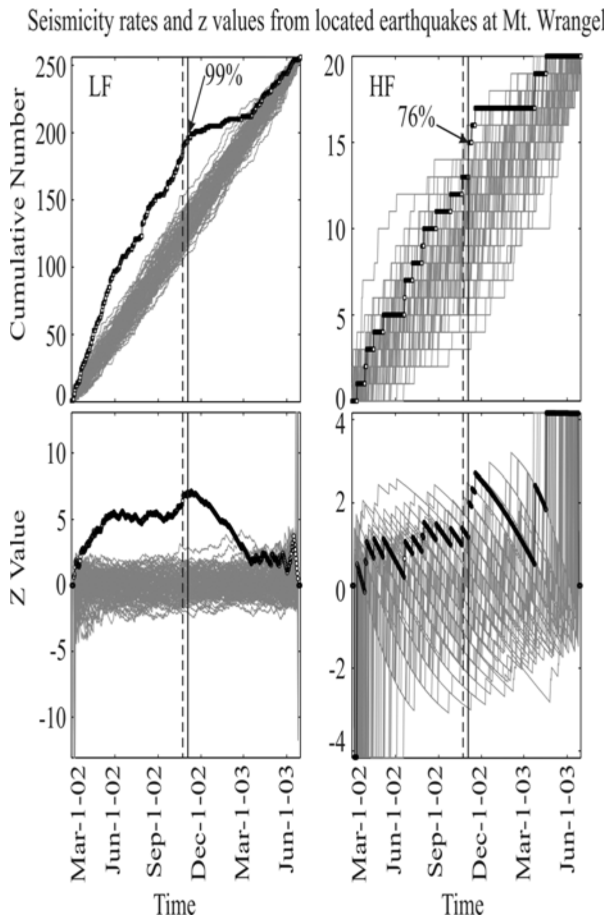


Figure 9. Cumulative sum and z values versus time plot for earthquake counts based on located events between March 1, 2002 and June 30, 2003 (dotted rectangle in Figure 8). Left: plots for LF seismic events. Right: Plots for HF earthquakes. Conventions as in Figure 3.

the radius of circles used to enclose the 30 nearest earthquakes to each node varied between < 2 km for areas around and west of the vent to 10 km around the edges of the region mapped.

DISCUSSION

We examined the rate of occurrence of volcanic earthquakes at Mt. Wrangell from before to after the DFE, at different time scales and using different data sources, to find the amount and significance of an observed decrease in volcanic seismicity. We think we can reliably measure this decrease because Mt.

Wrangell is an area with high rates of background seismicity, and such places are optimal for observation of seismicity declines (Toda and Stein, 2002). The decrease that we see in the seismicity at Mt. Wrangell following the DFE is not the decay from an aftershock sequence of a large earthquake rupturing along the Denali fault. The seismicity along the Denali fault is not collocated with the seismicity at Mt. Wrangell (Figure 2) and because we carefully checked all seismograms at station WANC on Mt. Wrangell, we could easily identify non-local earthquakes and discard them from our counts (see for example Figure 3 of Sánchez and McNutt, submitted).

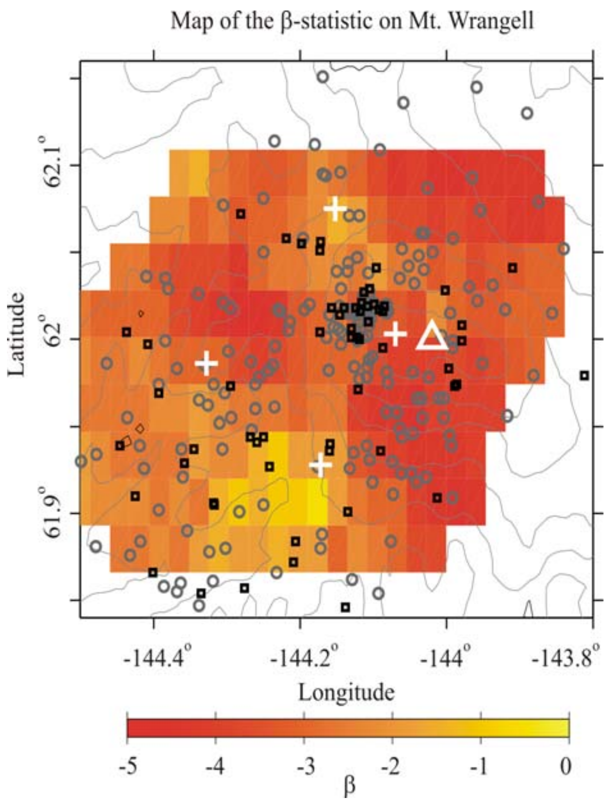


Figure 10. Map of the Mt. Wrangell area showing the seismicity rate changes, as evaluated by the β-statistic. The two time periods being compared are: Pre-DFE: March 1, 2002 to November 3, 2002; post-DFE: November 3, 2002 to June 30, 2003. Triangle: vent. Crosses: Seismograph stations (WANC is the station immediately west of the vent). Gray circles and black squares mark the epicentral locations during the pre-DFE and Post-DFE periods, respectively. Gray lines: Topographic contours.

We used rigorous statistical testing of the earthquake data and found the rates of local volcanic earthquakes (LF and HF together) to have decreased by 40% when the pseudohelicorder record data is used in the time scale of two months before to two months after the DFE (Figure 3). In some areas around the volcano the seismicity may have decreased by as much as 80% when the located earthquake data from eight months before to eight months after the DFE is used (Figure 9). Although visual inspection of the data suggests that the rates of both LF and HF earthquakes decreased (Figures 8 and 9), it is statistically difficult to prove for HF earthquakes because of their low rate of occurrence. From the depth-time plot and the analysis of cumulative number of located earthquakes we found that the decrease lasted for five months after which the seismicity seems to recover to pre-DFE levels.

We investigated the question of why the *z* values curve does not seem to peak exactly on November 3, 2002 (the time of the DFE), but in most cases, before that time. By computing the *z* values on six synthetic data sets with varying amounts of rate decrease (Figures 4 and 5), we found that the *z* values curve peaks exactly one data point before the drop in seismicity begins, which is expected because the rate before the change is constant exactly until that data point. An obvious conclusion of this test is that if the rates of seismicity are not perfectly constant within each period, we should expect to find variations in the behavior of the *z* values curve and the presence of multiple peaks of varying significance, which is what we see in our data (Figure 3). We also found that the z values curve is highly sensitive to sharp rate changes, regardless of the amount of change.

We also tested the effect, in the *z* values curve, of having a decrease in seismicity in two steps before the main perturbation and the effect of a seismicity rate that fully recovers sometime after the perturbation. The motivation for these two tests is the fact that our pseudohelicorder count data shows an apparent decrease in rates beginning sometime before the DFE (Figure 3) and because in the catalog of located earthquakes it *appears* as if the seismicity recovers to pre-DFE levels (Figures 8 and 9).

By simulating a two-step decrease in seismicity preceding the main rate change we observed that the *z* values curve is double-peaked and that the first peak indicates a more significant change (Figure 6). This apparently puzzling result comes about because the *z* values are a combination of the difference in means, the standard deviations of the two samples, and the number of data points in each sample (Equation. 1). Examination of Figure 6 indicates that at the position of the first peak: $M1$-$M2$= 0.4221, $S1$ =0, $S2$=0.1361, and $N2$=77; whereas at the second peak M 1-$M2$=0.3798, $S1$=0.1262, $N1$=50, $S2$=0.07, and $N2$=51. Thus it is apparent that the relative values of the peaks depend mostly on the difference in means and the number of data points on each segment. In the *z* values curve for our pseudohelicorder counts data we see some of the features identified in the curves for synthetic data. The pseudohelicorder counts data shows a *z* values curve with "regional" maxima that happens before the DFE but we also observe that this maximum is double-peaked, indicating that there were some step-wise changes in earthquake rates at the volcano.

From the analysis of the located events at Wrangell, on the other hand, we learned that the seismicity may have dropped after a few days following the DFE. We see this in Figures 8 and 9. By separating LF events from HF events we infer that both types of events may have decreased in rates, but because the low number of HF events available it makes it statistically difficult to prove.

The results of tests with synthetic data suggest that unless the rates of seismicity are fairly uniform before and after the perturbation, it is difficult to use the *z* values to pinpoint *exactly* the time of change. This does not mean the plot is useless, because the *z* values curve does indicate the presence of one or several important changes in rates within the data set under scrutiny, such as in the case of the data presented here.

We also simulated a time series in which an 80% drop in rate is followed by a recovery to the original rates (Figure 7). In this case we observed that the *z* values curve has two peaks, one positive and one negative. We made this synthetic data set based on some features of our located earthquakes counts data in which a drop of about 80% in seismicity is followed by an apparent recovery after five months of lower-than-normal rates (Figure 8 and 9). Because we do not see the negative peak in the *z* values curve of our real data, we infer that the seismicity at Mt. Wrangell did not recover fully to pre-DFE rates, at least until the end time period analyzed (June 30, 2003).

Although seismicity rates at volcanoes vary in the absence of evident external factors, the coincidence of the decrease with the time of the DFE suggests that the drop in seismicity at Wrangell was not the result of chance. The total duration of the catalog of located earthquakes, including those with $M < 0.9$, is 957 days, or roughly 32 months; we see in this time period (Figure 8) three noticeable drops in reported seismicity (one of them following the DFE, the other two are artificial) and the average duration of these three periods of low seismicity is ~ 3 months. Thus there are about 11 3-month-long periods during which we may or may not find a decrease. In the context of a binomial distribution, we have $n = 32/3 = 11$ trials and three successes (decreases). The probability p of success in a given binomial trial is $p=3/11=0.27$. This means that the probability of not finding an anomalously low-seismicity period is $q = 1- p = 0.73$, thus about 73% of the time we find that the seismicity is at background rates.

If we consider a more homogeneous catalog, with magnitudes ≥ 0.9, and only during March 1, 2002-March 18, 2004 (since data acquisition with Earthworm began) the number of available months and 3-month trials get reduced to 25 and 8, respectively, and the probability of finding a decrease rises to $p = 0.36$, making it a more common fact. But this may be a biased test, because we know that the two latest decreases are man-made. In our view, a more balanced test would only include the period March 1, 2002 – June 30, 2003 (dotted box in Figure 8). In this case there are 487 days or roughly 16 months, which means we have only $n = 16/5 \approx 3$ trials. The probability p of finding a decrease is $p = 1/3 = 0.33$, thus $q = 0.69$. This means that about 70% of the time we would find the seismicity to be at background rates.

Because the only large earthquakes along the Denali fault during the time period march 1, 2002 – June 30, 2003, were associated with the DFE sequence and because the decrease in seismicity began rather close in time to the DFE, we infer that it was not a chance occurrence but that the cause was a perturbation of the volcanic plumbing system caused by either the dynamic shaking or the sudden drop in pressure around the volcano (Sánchez and McNutt, submitted).

A spatial mapping of the β-statistic (Figure 10) shows a general decrease in seismicity around Mt. Wrangell with some patches of larger change located south, north and west of the vent and an area of little to no change southwest of the vent. The mapping should we interpreted with caution because the locations of earthquakes at Mt. Wrangell are preliminary.

CONCLUSIONS

A decrease in volcanic seismicity occurred at Mt. Wrangell following the Denali fault earthquake sequence that began with the Mw 6.7 Nenana Mt. Earthquake and ended with the nucleation of the Mw 7.9 Denali earthquake on November 3, 2002. Our data indicates that the volcanic seismicity dropped by as much as 80% during the five months that followed the DFE sequence. Our test with synthetic data showed that unless the seismicity rates are fairly constant during the pre and post-perturbation periods, we could not use the *z* values curve to pinpoint exactly the time of change. The presence of a broad peak, however, indicates a significant change in the rates and that is what we see in the data. Because of the time coincidence of the decrease in the seismicity rate with the time of the DFE sequence we think this is not a change occurrence, but an effect attributable to the DFE sequence. A comparison of our data of located earthquakes at Mt. Wrangell with a synthetic data set reveals that the seismicity at Wrangell during the period November 4, 2002 – June 30, 2003 did not recover to the pre-DFE levels. The mapping of the β-

statistic indicates that the decrease in seismicity occurred around the general Mt. Wrangell area, with some patches showing a stronger change than others.

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