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Eruptions of Pavlof Volcano, Alaska, and their Possible Modulation by Ocean Load and Tectonic Stresses: Re-evaluation of the Hypothesis Based on New Data from 1984–1998

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Abstract—Thirteen of sixteen magmatic eruptions of Pavlof Volcano in nine of the years from 1973 to 1998 have occurred between September 9 and December 29. Volumes of erupted material range from 0.3 to 16×10^6 m^3 (dense rock equivalent). A significant correlation exists between the eruptions and yearly nontidal variations in sea level and may result from ocean loading. Calculated volume changes beneath the volcano due to ocean loading are from 0.02 to 0.6 times eruption volumes, and it is postulated that the volcano acts as a long-period (several months) volume strainmeter, with lava being preferentially erupted when strain beneath the volcano is compressive. Previous observations of a tilt reversal, and new observations of tectonic activity and eruptions in the spring and summer of 1986, also suggest tectonic modulation of eruptions. The volcano appears to be responsive to small, slow changes in ambient stresses or strains, and these changes may modify or trigger eruptions.

Key words: Volcanoes, eruption, triggering periodic eruptions.

Introduction

Pavlof Volcano (latitude 55°24′N, longitude 161°54′W) is a 2518-m-high stratovolcano which sits on the Alaska Peninsula roughly in the middle of the Shumagin seismic gap (Fig. 1). It is the most persistently active volcano in North America, with over 50 reports of activity and 40 documented eruptions since it was first sighted by Russian explorers in the 1760s. An earlier paper by McNutt and Beavan (1987) discussed the eruptions that occurred from 1973–1984, and modeled the effects of possible modulation by ocean load and tectonic stresses. This paper re-evaluates these effects based on new data from the years 1984–1998.

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Earlier Work

All nine magmatic eruptions at Pavlof from 1973 to 1984 occurred in the fall, between September 9 and November 20 (Fig. 2). Four of them, in four different years, occurred in the same 4-day period between November 11 and 15. Volumes of erupted material range from 0.3 to 16×10^6 m^3 (dense rock equivalent) at an average rate of about 3×10^6 m^3 per year (Table 1). The volumes are estimated from eyewitness reports and photographs for several eruptions; the others are estimated from a locally derived relationship between eruption volume and volcanic tremor duration and amplitude (McNutt and Beavan, 1987). A significant correlation exists between the eruptions and yearly nontidal variations in sea level and may result from ocean loading. Sea levels are corrected for atmospheric pressure (Ingram *et al.*, 1976; Reed and Schumacher, 1981). Calculated volume changes beneath the volcano due to ocean loading are about 10 percent of eruption

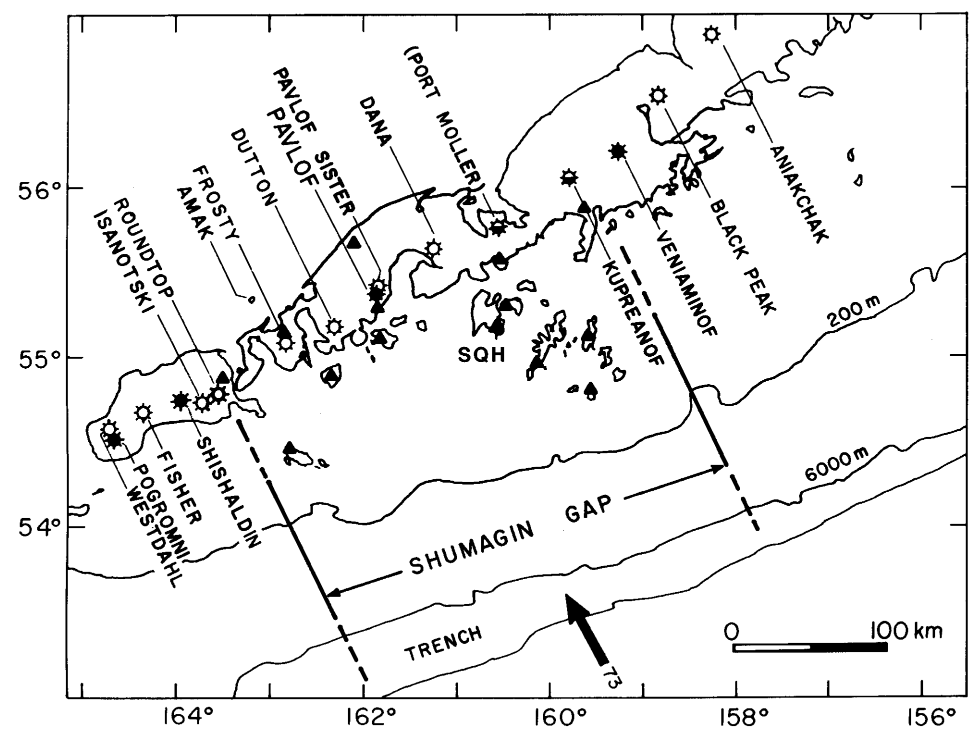

Figure 1
Index map of the study area. Holocene volcanoes are shown as sunbursts; filled symbols indicate eruptions, while half-filled symbols indicate earthquake swarms or geothermal activity. Triangles denote seismic stations. SQH is the levelling line near Squaw Harbor. Shumagin seismic gap boundaries are shown as solid lines for the seismogenic portions of the rupture zone and as dashed lines for nonseismic portions. Arrow shows plate convergence rate in mm/yr.

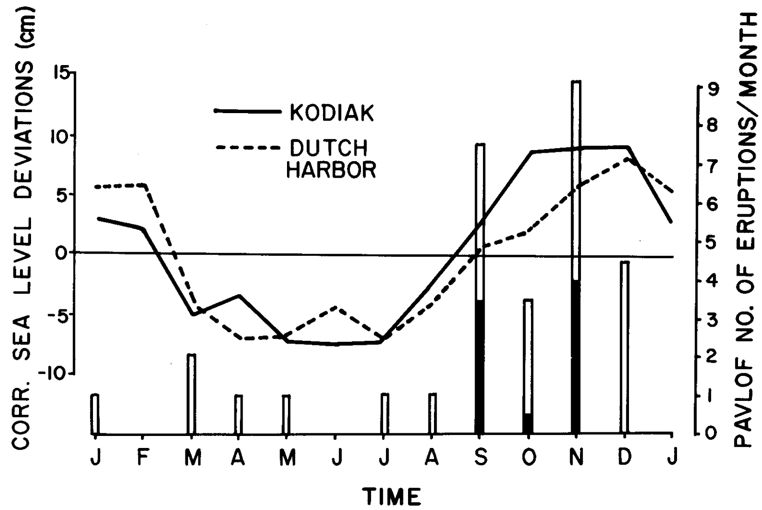

Figure 2
Deviations in monthly mean-corrected (for atmospheric pressure) sea level from long-term mean (1950–1974) and number of eruptions within each month at Pavlof Volcano (1973–1984). Kodiak is 500 km ENE of Pavlof, while Dutch Harbor is 350 km WSW. Histogram bars are solid for magmatic eruptions and open for explosive eruptions. Seismicity for the magmatic eruptions (solid bars) is on average a factor of 10 greater than for the explosive eruptions (open bars). The two smallest magmatic eruptions, September and October 1976, are shown at half the height of the others. Note the correlation between eruptions and increased sea level in fall/winter. (Plot from McNutt and Beavan, 1987.)

volumes, and it is postulated that the volcano acts as a long-period (several month) volume strainmeter, with lava being preferentially erupted when strain beneath the volcano is compressive (McNutt and Beavan, 1987). The amplitude of the sea-level deviation is highest in November, and 4 of the 5 largest volume eruptions from 1973 to 1984 occurred in November. The volcano did not erupt during the period 1978–1980, when tilt, seismic, and sea-level data indicate that deep aseismic slip may have occurred (Beavan *et al.*, 1984). Models of this event predict a volume strain extension beneath the volcano that might have compensated strain from magma injection. These observations indicate that Pavlof Volcano is responsive to small, slow changes in ambient stresses or strains and that these changes may modify or trigger eruptions (McNutt and Beavan, 1987).

Other volcanoes and earthquakes have shown apparent seasonal periodicities. Etna terminal and flank eruptions from 1323 to 1980 show significant seasonal clustering with a main peak in November and subsidiary peaks in March and May (Casetti *et al.*, 1981). Eruptions of Kilauea from 1832 to 1979 clustered in May (data given in Dzurisin, 1980). In Iceland, Hekla's eruptions from 1104 to 1970 and all known Icelandic eruptions from 1550 to 1978 clustered in May, but not significantly (Thorarinsson and Sigvaldsson, 1972; Gudmundsson and Saemundsson, 1980). Earthquakes in the Torfajokull-Myrdahlsjokull area of Iceland, on the other hand, clustered in the fall, and were interpreted to be triggered by glacial loading (TRYGGVASSON, 1973). Shallow earthquakes at Rabaul caldera from 1967 to 1984 showed an annual periodicity with a peak in December (P. Lowenstein, writt. com., 1985). The formal statistical significance for several of these cases has not been established.

OHTAKE and NAKAHARA (1999, this volume) summarized the observations of earlier workers on seasonal earthquakes, and demonstrate a convincing case of periodicity for great ($M \geq 7.9$) earthquakes along the Nankai and Sagami troughs. The historical earthquakes all occur in the seven months from August to February, a distribution which has only a two percent probability of occurring by chance (OHTAKE and NAKAHARA, 1999, this volume). They interpret seasonal atmospheric pressure variations to be a triggering mechanism, and model the stress changes on the shallow plate interface. Both the seasonal distribution of these earthquakes and the phenomena modeled are remarkably similar to the observations at Pavlof volcano.

New Data

Two new series of eruptions occurred from April 1986 to May 1988 (Fig. 3 and Table 1) and later from September to December 1996 (Table 2). The 1986 eruptions were stronger and of longer duration than any others known (MCNUTT *et al.*,

Table 1

Magmatic eruptions of Pavlof Volcano, 1973 to 1998

Year	Start date	Volume $\times 10^6$ m^3	Comment
1973	Nov. 13	6.4	Photographs
1975	Sept. 13	4.5	
1975	Sept. 23	1.2	Photographs
1976	Sept. 9	0.3	Ground observations
1976	Oct. 18	0.5	
1976	Nov. 10	7.6	Pilot reports
1980	Nov. 11	6.1	Pilot reports, photographs
1981	Sept. 26	10.8	Pilot reports, photographs, ash sample
1983	Nov. 14	12.5	Pilot reports
1986	Apr. 19	–*	Pilot reports
1986	May 28–Sept. 1	16	Unusually long duration; observed by geologists
1986	Nov. 5	2	Pilot reports
1986	Dec. 8	3	Pilot reports
1987	May 23	2	Pilot reports
1987	Oct. 14	1	Pilot reports and ground observations
1996	Sept. 15	7	Pilot reports, photographs

* Volume of April 19, 1986 eruption is included in estimate for May 28 to Sept. 1, 1986 eruption.

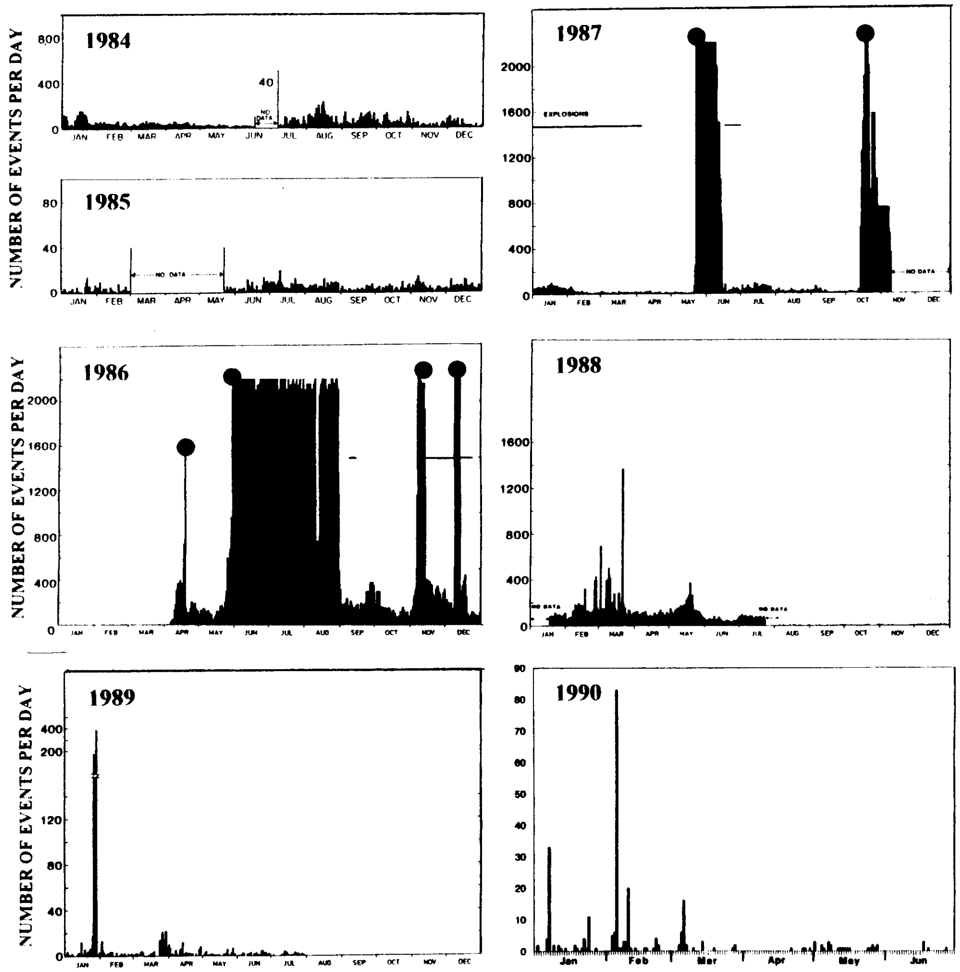

Figure 3
Seismicity plots for Pavlof Volcano for the time period 1984 to 1990. Plots show the number of B-type and explosion events per day. Periods of volcanic tremor are normalized to number of events using a previously published relationship (McNutt, 1987). Round dots indicate eruptions accompanied by lava fountains; horizontal bars above the data indicate time periods of explosions. Note the different vertical scales for mid-1984 through 1985, 1989, and 1990. The 1989 data also show a scale break for the swarm in late January. Horizontal (time) scales are the same for all years except 1990.

1991). They produced a possible pyroclastic flow for the first time, and modified the structure of the vent area. They also occurred from April through August, thus changing the temporal pattern of activity that had persisted from 1973–1984 (Fig. 4). Note that although the May 28 to September 1, 1986 portion of the eruption is a single long-lasting eruption, it contributes to the three months June, July and August in Figure 4 (the minor contributions from only three days in May and one

Table 2

Pavlof Volcano 1996 eruptions

Start date	End date	Activity, based on seismicity, satellite images, and pilot reports*
Sep. 15	Dec. 4	Continuous Strombolian eruption with stronger pulses: Oct. 18 strong volcanic tremor; poor weather Nov. 4 strong volcanic tremor; ash column to 25,000 ft a.s.1.; ash plume 100 miles long Nov. 22 strong volcanic tremor; steam plume between 20,000 and 30,000 ft a.s.1.
Dec. 10	Dec. 14	Strong volcanic tremor; poor weather; one observation of ash column of 15,000 ft a.s.1.
Dec. 26	Dec. 29	Strong volcanic tremor; poor weather; observations of hot spot (satellite image), lava flow, and ash plume several tens of miles long

* Observations are reported in the original units of ft and miles.

day in September are down-weighted). The 1996 eruptions returned to the earlier pattern of fall eruptions.

Two statistical tests were performed on the data from 1973 to 1998 (Table 1) to determine the significance of the temporal distribution of eruption onsets. To test for uniform versus von Mises (or other one-humped alternatives) the Rayleigh test (Mardia, 1972) was conducted. For number of cases $n = 16$, the regression statistic

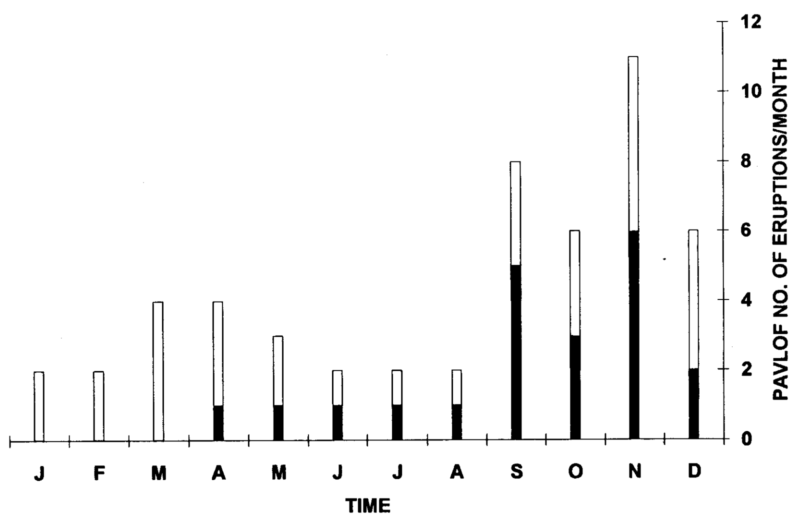

Figure 4

Histogram of the number of eruptions within each month of Pavlof Volcano from 1973–1998. Histogram bars are solid for magmatic eruptions and open for explosive eruptions. Here each eruption is plotted at the same height regardless of volume erupted. Note that despite the increase in the number of eruptions from January to August (compare with Fig. 2), the eruptions are still concentrated in the fall/winter months from September to November.

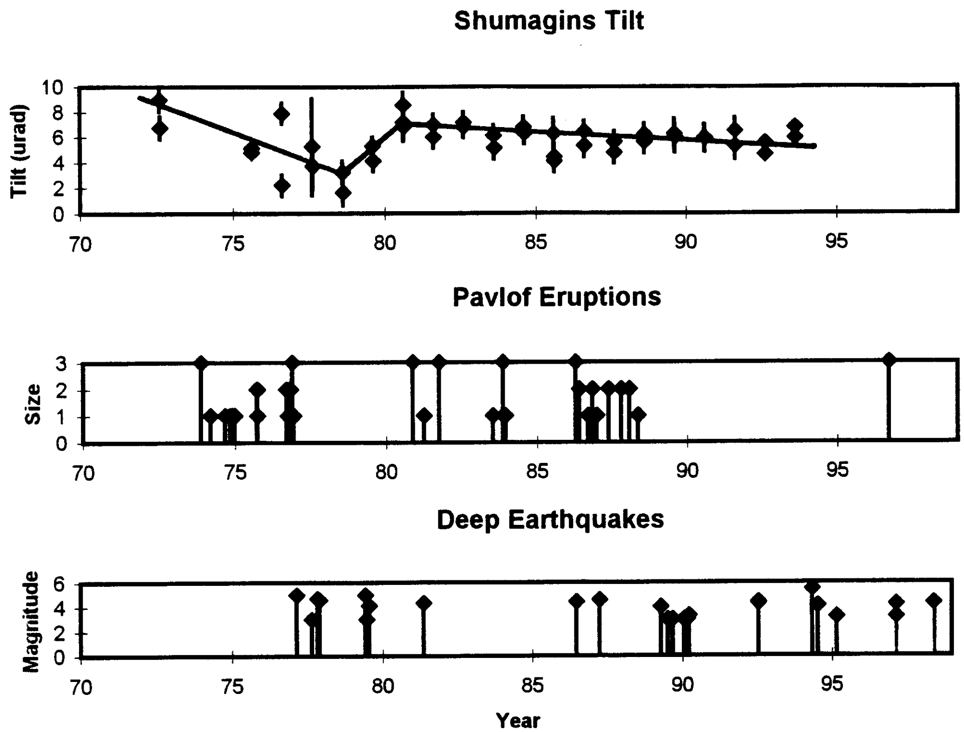

Figure 5
Tilt (microradians), Pavlof eruptions (VEI; Volcanic Explosivity Index), and deep earthquakes (>160 km) in the Shumagin region for the period 1970–1998. An apparent tilt reversal took plate on three level lines between 1978 and 1980. The tilt change is significant at the 98% level on line SQH, which is plotted here (see Fig. 1 for location). The line drawn through the data is our interpretation of the tilt history in the inner Shumagin Islands. During the tilt reversal, the seismicity rate for microearthquakes was higher (not shown; see McNutt and Beavan, 1987), Pavlof Volcano was not erupting and exhibited very low seismicity, and most of the teleseismically recorded deep earthquakes from 1970–1985 occurred in a cluster NNE of Pavlof (see Fig. 6). The estimated magnitude for complete recording is 4.5; quakes plotted as $M = 3$ (smallest events on plot) had no magnitude in the PDE catalog. New tilt data are from Beavan (1994).

$\overline{R} = 0.5602$, thus the probability or p value is approximately 0.005; we can reject a uniform distribution at the 99.5 percent confidence level. A second test conducted was Kuiper's test (Mardia, 1972) of uniformity versus nonuniformity. For $n = 16$, $V_n = 0.5692$, yielding a p value of approximately 0.0004, or a 99.96 percent probability that the distribution is nonuniform. Therefore, both tests indicate that the Pavlof eruption onsets are highly non-random. They appear to be clustered in the fall, as is evident from Figures 2 and 4.

New tilt data from 1984 to 1993 were compiled (Beavan, 1994) and are plotted in Figure 5. These data show continuous slow tilt down towards the trench for site SQH in the inner Shumagins (Fig. 1). No evidence is found for another tilt reversal similar to the one that occurred from 1978 to 1980, however a slight change is noted for the last data point, which occurred just after the May 13, 1993 $M = 6.9$ Shumagin earthquake (see below). New data on deep (>160 km) moderate earthquakes as reported in PDE were also compiled. The deep events all occurred spatially in the region NNE of Pavlof (Fig. 6). While most of the deep events occurred during times of no eruptions, two of the larger events in 1986 and 1987 were associated with eruptions. Also, the cluster of deep events from 1977 to 1979 occurred just before and during a tilt reversal, yet no apparent tilt reversal occurred in 1989 or 1990 during a similar cluster of deep earthquakes.

The new data suggest that the tectonic conditions in the Shumagin region changed in 1986 then returned some time later. It is also worth noting that from 1988 to 1996 Pavlof was in the longest period of repose (8 yr 4 mo) since seismic monitoring began in 1973. Strictly speaking, seismic monitoring by the Lamont-

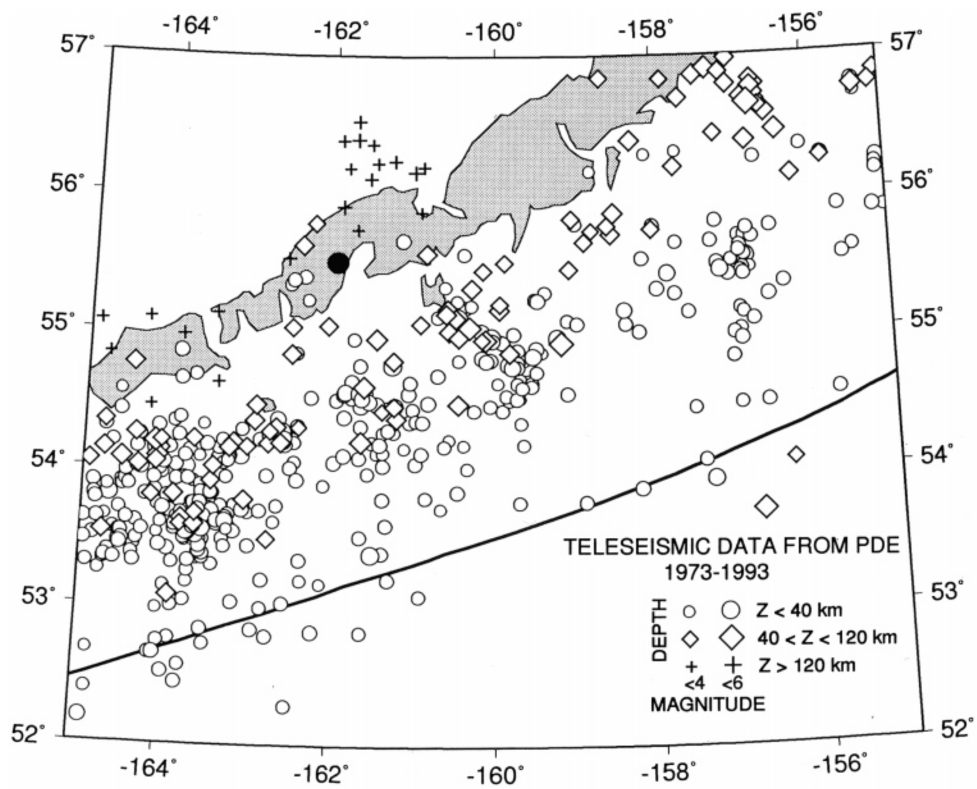

Figure 6
Epicenter map of teleseismic data from PDE for the Shumagin region. All data from 1973–1993 are shown. Note that the only cluster of deep events is located in the back arc region NNE of Pavlof Volcano (solid dot). This spatial pattern was persistent from 1994 to 1998 as well. A time series for these events is displayed in Figure 5.

Doherty Earth Observatory ended in summer 1990. However, stations operated by the University of Alaska Fairbanks at nearby Dutton volcano (35 km SW) remained in operation, but did not record any eruptions at Pavlof from 1988 until fall 1996. Further, there were no pilot reports or any other reports of eruptive activity from 1988 to 1996. In July 1996, Alaska Volcano Observatory installed a new seismic network at Pavlof, which recorded an eruption sequence from September 15 to December 29, 1996 (Table 2; McNutt, 1997). It is salient that the 1996 eruptions again occurred in the fall, similar to the activity from 1973 to 1984.

Discussion

The original work (McNutt and Beavan, 1987) suggested that Pavlof erupted in the fall because eruptions were triggered by compressional stresses or strains due to ocean loading. For modeling, a feeder zone was defined beneath the volcano. This was 40 by 40 km in area extent (average volcano spacing is 40 km), and extended from the base of the crust (30 km) to the top of the Wadati-Benioff zone (100 km). The true magma generation and storage complex probably includes structures with many different rheologies and geometries, which could have a substantial effect on the distribution of stresses or strains from sources such as ocean loading (McNutt and Beavan, 1987). Using the same feeder zone, a 3.8-year period of no eruptions from 1978 to 1980 coincided with an episode of aseismic slip that produced extensional strains at depth. Both effects demonstrated that the volcano was very sensitive and responsive to small changes in stress or strain. It is, however, necessary to re-evaluate the new data in terms of both yearly and longer time scales.

The new data include both additional fall eruptions and several in the spring and summer. However, statistical tests show that the distribution is still highly non-random. Thus, the ocean loading hypothesis still appears to be valid, although other mechanical or tectonic effects may also be occurring. It is speculated that compressional strain may have occurred in 1986 at depth, forcing magma to erupt prematurely starting in April. While no direct data exist to confirm this hypothesis, it is noted that widely separated segments of the Alaska/Aleutian arc were tectonically active during a 6-week period in spring 1986: Mount Augustine erupted 670 km to the east of Pavlof from 27 March–28 April (Kienle *et al.*, 1986) Pavlof erupted 14–26 April (McNutt *et al.*, 1991); and a great earthquake of $M_W = 8.0$ occurred in the Andreanof Islands region 1110 km to the west of Pavlof on 7 May 1986 (Kisslinger, 1988). There is a remote possibility that an arc-wide tectonic strain pulse may have triggered all three of these phenomena.

There is no evidence of another tilt reversal, indicative of aseismic slip (Fig. 5), so it is not possible to evaluate or compare long-term behavior directly. However, the largest earthquake in the Shumagin region in 45 years occurred on May 13, 1993 (Lu *et al.*, 1994; Abers *et al.*, 1995). This $M = 6.9$ event occurred on the main thrust zone, had a 600 km^2 rupture area and 1 m of slip (Lu *et al.*, 1994). The sense of motion for the 1993 earthquake is the same as that for the 1978–1980 aseismic slip event, leading to a net volume strain extension beneath the volcano. Levelling data from summer 1993, after the earthquake, show a slight change in polarity (Fig. 5), consistent with the occurrence of the earthquake (Beavan, 1994). The occurrence of the 1993 earthquake may thus have contributed to the long repose of 8 yr 4 mo from 1988 to 1996. In addition, the deep earthquakes continued to occur mainly in clusters during times when the volcano was not erupting (Fig. 5). This suggests a mechanical connection between deep slab processes and shallow eruptive processes. Most earthquakes at depths greater than 160 km in the Shumagins had focal mechanisms indicating downdip compression (Reyners and Coles, 1982; Hauksson *et al.*, 1984). All available first motion data were examined for the newer deep earthquakes, and all showed downdip compression.

Although the suggested modulating mechanisms, ocean loading, aseismic slip and related tectonic activity, may have the correct polarities and magnitudes to produce the observed effects, it must be kept in mind that correlation does not necessarily imply cause and effect. Therefore, other mechanisms need to be considered. The annual sea-level variations are adding a small amount of strain, which may be in phase with other larger strains produced by processes for which we lack observations. For example, there may be annual modulations in stresses and strains within the solid earth caused by astronomical variations, or rotational changes caused by atmospheric angular momentum variations (e.g., Lowrie, 1997, pp. 54–55). Such effects may be needed to explain similar periodicities of eruptions and earthquake swarms in places as widely separated as Alaska (Pavlof), Hawaii (Kilauea), Italy (Etna), and Papua New Guinea (Rabaul).

The apparent sensitivity of Pavlof to strains and its location in the Shumagin seismic gap suggests that its eruptive activity will change systematically if a large earthquake occurs in the gap. Specifically, based on the above observations, it is forecast that the volcano will be relatively more active before the earthquake and less active afterward, possibly for many years. Other studies have shown that volcanoes tend to erupt more often around the time of great earthquakes (e.g., Carr, 1977; Kimura, 1978), and, in general, volcanoes appear to be affected by, or sensitive to, regional stresses. A recent example is the increase in earthquake activity at Long Valley, California, and elsewhere in the western U.S., triggered by transient stresses generated by the 28 June 1992 Landers, California earthquake (Hill *et al.*, 1993). Thus, seismic recording at volcanoes offers the opportunity to use them as "barometers" to monitor stress conditions prior to, during, and after large earthquakes and other significant tectonic events.

Conclusions

The new computations clearly show that Pavlof's eruptions are non-random; the likelihood of the distribution occurring by chance alone is less than 0.5 percent. New data include a departure from the fall pattern in 1986, then a return later and continuing in 1996. The available data on sea-level variations, aseismic and seismic slip have the correct polarities in terms of compressional and extensional strains, and reasonable magnitudes to modulate the volcanic eruptions. Therefore, all the data are consistent with the hypothesis that Pavlof is responsive to small, slow changes in ambient strains, and these changes modify or trigger eruptions.

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