FIRE and MUD Contents

Precursory Deep Long-Period Earthquakes at Mount Pinatubo: Spatio-Temporal Link to a Basalt Trigger

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ABSTRACT

About 600 deep long-period (DLP) earthquakes occurred beneath Mount Pinatubo in late May and early June 1991. This number is higher than the combined total number of such earthquakes previously reported at all convergent-margin volcanoes worldwide. The DLP earthquakes occurred in two episodes of roughly similar total energy release, from 1700 May 26 to 1210 May 28 and from 2114 May 31 to 1510 June 8. During these same periods, at least 25 hours of very low amplitude DLP tremor also occurred, in 1- to 10-hour-long episodes. The DLP earthquakes exhibit clear P- and S-phases on three-component records. A P-S converted phase, which apparently converts at the base of the dacite pluton at about 14 kilometers in depth, is also observed. DLP earthquake spectra are dominated by very narrow-bandwidth signals with a quality factor of 20 to 50. The dominant frequency for the vast majority of the events, including the 10 largest, is at 2.0 hertz. A few events with dominant frequencies of 3.2-3.3 or 3.6-3.9 hertz were also observed, usually at the beginning of swarms which contain large-amplitude 2.0 hertz events. The Reduced Displacement for the largest event is about 3,100 square centimeters (magnitude ~3.7), making it the largest DLP earthquake ever reported. The events locate from 28 to 35 (possibly 40) kilometers below, and about 6 kilometers northwest of, the summit and apparently shallow with time.

There is a striking temporal correlation between the two episodes of DLP activity and geological and seismological changes at the surface. For example, the onset of DLP seismicity on May 26 was accompanied just 1 hour later by the onset of shallow long-period earthquakes, and within 4 hours by continuous, shallow long-period tremor and a large steam emission. The highest DLP moment release occurred on June 4 and was followed within 3 hours by shallow long-period earthquakes and increased steam emission. Significant DLP activity continued through June 7, accompanied by inflation of the summit and a rapid acceleration in shallow seismicity beneath the summit which gradually shoaled with time. DLP seismicity waned with the emergence of a dome containing inclusions of a very primitive, freshly quenched olivine basalt that had recently arrived (several days to a few weeks prior) from the deep crust. The spatio-temporal development of the DLP seismicity and its temporal correlation with subsequent surficial activity, especially the emergence of the primitive basalt, is taken as evidence that the DLP seismicity is the elastic manifestation of the injection of deep-seated basaltic fluids into the base of the magma chamber. As such, the DLP seismicity provides the first direct evidence for the location and timing of such injections prior to a major eruption and strongly supports the notion that these basalt injections triggered the eruptive sequence at Mount Pinatubo. If basaltic injections trigger all eruptions of such size, early recognition and quantification of large-scale DLP seismicity may provide one of the best tools for predicting the timing and size of such destructive eruptions.

Note to readers: Figures open in separate windows. To return to the text, close the figure's window or bring the text window to the front.

INTRODUCTION

Long-Period (LP) earthquakes and LP tremor have been observed at many volcanoes (for example, Koyanagi and others, 1987). The vast majority of LP events reported in the literature originated at depths of less than 3 km.

Such shallow LP events have preceded eruptions at many volcanoes (for example, Chouet and others, 1994). LP events deeper than 10 km are less commonly observed. Such earthquakes have been noted in California beneath Long Valley caldera (Hill and Pitt, 1992) and the nearby Mono cones (Mitch Pitt, USGS, oral commun., 1994), Mt. Lassen and Medicine Lake volcanoes (Steve Walter, 1988, 1991), and the Clear Lake volcanic complex (Steve Walter, USGS, oral commun., 1994). Other deep LP earthquakes occur regularly under Kilauea volcano, Hawaii (Koyanagi and others, 1987), many have been observed under Mount Spurr, Alaska (Power and Jolly, 1994), one at Izu-Ooshima, Japan (Ukawa and Ohtake, 1987), and at several other active volcanoes and low-velocity in Japan (Hasegawa and others, 1991).

Table 1 lists some basic parameters of deep long-period (DLP) events. Note that, apart from Mount Pinatubo, fewer than 400 DLP earthquakes have ever been observed at all convergent-margin volcanoes combined. At Kilauea, however, about 540 DLP "tremor episodes" occurred during 1962-83 alone (Shaw and Chouet, 1991), with each episode estimated to contain 3 to 300 DLP earthquakes (Shaw and Chouet, 1989). Aki and Koyanagi (1981) reported statistics on 200 of those episodes and showed that the LP nature of the signal is governed by the source process, not by path or receiver effects. Koyanagi and others (1987) showed that both shallow LP earthquakes and shallow LP tremor at Kilauea have identical spectral properties and that both phenomena must, therefore, have a similar source mechanism. Aki and Koyanagi (1981) showed that the process responsible for the DLP events beneath Kilauea has been "a generally steady process which does not seem to be significantly affected by major eruptions and large earthquakes" in the vicinity over at least 17 years.

Table 1. Reported deep long-period earthquakes worldwide.

[Under "Location," the data for "Northern Honshu, Japan" are for the combined data for nine volcanoes and low-velocity zones. Under "Quantity," "reported" indicates the total number of DLP earthquakes counted, including those too small to locate. For "Kilauea," these numbers are taken from the number and duration of tremor episodes given in Shaw and Chouet (1989), where they assumed that tremor there was produced by the sustained occurrence of LP earthquakes at the rate of three per min. "Time Span" is for the whole period of time for which records have been scanned, rather than the duration of a particular episode or active period, for example. M_{MAX} is the estimated amplitude magnitude, except where duration magnitude is noted by M_D , of the largest known DLP earthquake at that location. Note that LP events within the 5-15 km depth range are called "intermediate depth" by Aki and Koyanagi (1981)]

Location	Quantity located/reported	Time span	M_{max}	Depth (km)	F_0 (Hz)	Singly/swarms	Reference
Convergent margin							
Mount Pinatubo	11/400	1991-1991	3.8	28-35(40?)	2-3.8	swarms	This paper.
Long Valley, Calif.	50/?	1989-1992	2.2	10-20	1.5-3	both	Mitch Pitt (USGS, oral commun., 1994).
Mono Cones, Calif.	2/?	1989-1992	1.5	25-35	1-3	singly	Mitch Pitt (USGS, oral commun., 1994).

Mt. Lassen, Calif.	25/50	1984-1992	2.4	13-22	1-3	singly	Walter (1991).
Medicine Lake, Calif.	2/3	1984-1992	2.9M_D	16	1-3	singly	Walter (1991).
Clear Lake, Calif.	15/?	1984-1992	2.7M_D	16-25	2	both	S.R. Walter (USGS, oral commun., 1994).
Mt. Shasta, Calif.	0/0	1984-1992	none	none	none	none	S.R. Walter (USGS, oral commun., 1994).
Mt. Spurr, Alaska	100/250	1991-1992	1.8	15-40	1-3	both	A. Jolly (USGS, oral commun., 1994).
Mt. Redoubt, Alaska	0/0	1989-1992	none	none	none	none	Power and others (1993).
Mt. St. Helens, Wash.	4/4	1980-1992	1.6	11-33	?	singly	A. Qamar (USGS, oral commun., 1994).
Izu-Ooshima, Japan	1/1	1985-1986	2.7	29	1	singly	Ukawa and Ohtake (1987).
Mt. Moriyoshi, Japan	20/20	? - ?	?	27-37	?	?	Hasegawa and others (1991).
Northern Honshu, Japan	?/?	? - ?	2.5	25-40	1.5-3.5	?	Hasegawa and other (1991).
Intraplate							
Kilauea, Hawaii	?/>48,000	1962-1983	3.2?	30-50	2-10	swarms	Shaw and Chouet (1988).
Kilauea, Hawaii	?/>18,000	1962-1981	?	5-15	?	swarms	Shaw and Chouet (1988).
Yellowstone, Wyo.	0/0	1973-1981	none	none	none	none	Mitch Pitt (USGS, oral commun., 1994).

Nowhere have DLP earthquakes been shown to correlate with an impending eruption. At Izu-Ooshima, the only reported DLP earthquake (Ukawa and Ohtake, 1987) was followed about 1 year later by an eruption, but any connection there is tenuous. At Kilauea, DLP activity and eruptions both have been frequent since the early 1960's (Aki and Koyanagi, 1981), but any correlation is unclear. I report the observation of about 400 DLP earthquakes beneath Mount Pinatubo, including the largest DLP earthquake ever observed, which immediately precede, by 1 to 3 weeks, the cataclysmic eruption of the volcano. Peaks in the DLP energy release rates are observed to precede, by 1 h to a few days, major geological and seismological changes near the surface, including the extrusion of an andesite dome containing inclusions of freshly quenched olivine basalt (Pallister and Hoblitt, 1992) that had only very recently (several days to a few weeks prior) arrived from the deep crust.

The evidence is compelling that the DLP earthquakes were produced by the flow of a deep-seated basaltic magma moving upward from near the base of the crust into a magma chamber containing a dacitic residuum and that this magma-mixing led directly to the cataclysmic eruption. I interpret DLP seismicity in terms of a model proposed by Chouet and others (1994), which invokes the choked flow of a mixed-phase magmatic fluid through a rectangular conduit. This model adequately describes the important features observed in DLP waveforms at Mount Pinatubo and can explain other features such as the unusual pattern of DLP energy release over time and the various depth ranges of DLP events at other volcanoes.

THE 22-MIN DLP SEQUENCE OF JUNE 4, 1991

Following the occurrence of a series of steam blasts on the north flank of Mount Pinatubo on April 2, a network of seven telemetered high-gain seismometers was installed on and around the volcano (fig. 1). For a discussion of the network configuration and history, see Harlow and others (this volume) and Lockhart and others (this volume). For the purposes of this study, the network began producing marginally useful records on May 1 and relatively good records on May 10.

Figure 1. The seven high-gain seismometer station locations at Mount Pinatubo prior to the eruption. Contours are 500 and 1,000 m. The original summit elevation was 1,745 m. Map produced using PC-QPLOT (Murray and others, 1993).

From the inception of the network through May 31, data show that a few volcano-tectonic events per hour were originating from a region about 5 km northwest of the summit, while a few volcano-tectonic events per day were originating from another region beneath the summit. On June 4, a 22-min sequence of relatively large-amplitude seismic signals, later determined to be DLP earthquakes, was recorded on the analog record from station PIE. Part of this record is shown in figure 2. The largest swarm of shallow LP earthquakes occurred at this time, ash and steam emissions occurred 1 day later, and 2 days later the rate of shallow volcano-tectonic seismicity suddenly increased beneath the summit. The following day a hybrid andesite dome emerged (for a detailed description of shallow preeruption seismic and geologic activity, see Harlow and others, this volume; Hoblitt, Wolfe, and others, this volume).

Figure 2. Analog Helicorder record from station PIE showing part of the 22-min deep long-period (DLP) earthquake sequence on June 4, 1991. This is the most energetic DLP earthquake sequence recorded at Mount Pinatubo. Note the monochromatic nature of the waveforms and the swarm-like pattern of energy release within the sequence. Other signals shown include a teleseism (T), shallow volcano-tectonic earthquakes (VT), shallow long-period earthquake (SLP), and passing helicopters (H). Tick marks are 1 min apart on each line. The local time (hour) is indicated by "8," "9," "10," etc. Each line is nearly 5 min long.

Two features of the 22-min sequence were initially recognized as unusual, as compared with other seismicity prior to that time: (1) the events appeared monochromatic with a frequency of 2 Hz within the spindle-shaped envelope of individual events and (2) ground displacements were of similar amplitudes at stations near and far from the summit. Regarding the first feature, shallow LP earthquakes and tremor with monochromatic appearance had been noted beginning May 26, but the dominant frequency of this shallow LP activity was 0.8

Hz. A few periods of very low amplitude tremor with dominant frequencies higher than 2 Hz had also been noted, but the signals are broader band and generally correlated with the occurrence of strong rain showers or steam emissions. Ground displacements were of similar amplitudes at station PIE, located 9 km from the peak, and station UBO, located only 1 km from the peak. To the Pinatubo Volcano Observatory Team, which was focussing its attention on shallow seismicity (depths of less than about 10 km), this indicated a deep source for those events. These events were otherwise ignored until well after the cataclysmic eruption.

In late 1991, a search through the collection of about 1,000 digitally recorded waveforms from May 10 through June 12 initially produced two approximately 1-min-long sections of record beginning at 08:59 and 09:08 on June 4, each containing one well-recorded DLP earthquake. The event at 08:59 was later discovered to be the largest DLP event ever reported anywhere in the world. Waveforms are shown for this event at all stations in figure 3A. The other digitally recorded DLP event, shown in figure 3B, contains a third, much smaller DLP earthquake within its coda. Both of the well-recorded events themselves begin, in fact, within the coda of previous DLP events.

Note the nearly simultaneous arrival of the initial wave front at all stations in figure 3, suggesting an approximately planar wave front with nearly vertical incidence. The P-wave first motion is emergent and down at all six stations. Maximum horizontal S-phase amplitudes are observed to be about 5 times larger than maximum P-phase amplitudes. The vertical components of ground motion at the three outermost stations, PIE, BUG, and BUR, which have an angular separation of 25° to 29°, as viewed from the hypocenter, were compared two at a time and found to have maximum correlation coefficients less than 0.4 and thus are uncorrelated. The S-wave arrival is very difficult to pick from the vertical component records but is most clear at GRN and on the horizontal components at PPO. This difficulty owes partly to the near-vertical incidence of the S-phase and partly to the arrival of a converted phase about 1.6 s prior to the S-phase arrival on some records. Figure 4 shows the three-component record and particle motions at PPO for the P- and S-phase arrivals and the converted phase. The converted phase is apparently a vertically arriving shear wave (P to S). The traveltime for the converted phase indicates that the conversion occurred at 13 to 14 km in depth, a location compatible with the base of the dacite pluton, according to Pallister and others (this volume).

Figure 3. Digital seismograms for two events during the 22 min-long sequence on June 4, 1991. Maximum amplitudes in counts are shown for each trace at left. A, This event occurred at 0859 local time. It is the largest deep long-period earthquake ever reported anywhere in the world. Note that the record is offscale (clipped) at stations BUG, GRN, and PIE. Also note the nearly simultaneous arrivals and lack of surface waves at all stations, indicating a deep event from directly beneath the network. B, This event occurred at 0908 and is the largest onscale deep long-period earthquake recorded at Mount Pinatubo. Figure produced using PITSA (Scherbaum and Johnson, 1992).

Figure 4. Particle motions at station PPO. A, 10-s-long record from station PPO showing the locations of the 1-s-long windows, containing the P-, P-S, and S-wave arrivals, used for the particle motions. Maximum amplitudes in counts are shown for each trace at left. B, Upper plot shows radial (Z) versus tangential (E) particle motion, while lower plot shows the particle motion in the horizontal plane, for the P-wave arrival. C, Same as B, for the P-S converted-phase arrival. D, Same as B, for the S-wave arrival. Figure produced using PITSA (Scherbaum and Johnson, 1992).

Figure 5. Epicenters of the two best-recorded and best-located deep long-period earthquakes. These events occurred at 0859 and 0908 June 4 during the 22-min sequence of deep long-period activity. The first event is the larger of the two. The horizontal location error, at the 95% confidence level, is +-2.5 km. The events locate at 33.5 and 33.6 km, respectively, below the elevation of the summit. The vertical location error is +-0.7 km. Contours are 500 and 1,000 m. Figure produced using PC-QPLOT (Murray and others, 1993).

Figure 6. Reduced Displacement (*RD*) and cumulative RD versus time for the 22-min sequence of June 4, shown in part in figure 2. Note the three subgroups (top), with small events occurring first and the largest event occurring at or near the end of each subgroup.

By using P-arrival picks from each of the six vertical component stations and an S pick from the horizontal components at station PPO, I obtained preliminary hypocenter locations for both events at depths of 34 to 35 km beneath the edifice. For final solutions, additional S-arrival picks were incorporated and checked by running the location program with different combinations of S picks and then running the final set of picks used with a 3-dimensional velocity model with stations delays determined from traveltime inversion (for details, see Mori, Eberhart-Phillips, and Harlow, this volume). These additional measures yielded hypocenters with depths of 33.5 and 33.6 km below, and about 5 km northwest of, the pre-June 1991 summit. The relative precision at the 95% confidence level is +-0.7 km vertically and +-2.5 km horizontally. Owing to uncertainties in the seismic velocity model below 20 km in depth, the absolute depth of these hypocenters may actually vary from the calculated values by as much as 3 km. Figure 5 shows the location of these two events.

The 22-min sequence contains 7 of the 9 largest DLP earthquakes (with amplitudes >=12 mm at station PIE) recorded at Mount Pinatubo. The coda duration of the largest DLP earthquake is about 150 s long, as measured from the analog record from station PIE. This duration is similar to the duration of tectonic earthquakes of such amplitude, distance, and depth. The magnitude of this event is estimated to be 3.7, making it the largest DLP earthquake ever reported anywhere in the world. Other unusual aspects of the DLP sequence are: (1) the distinctly nontectonic temporal pattern of energy release and (2) the very narrow-band spectral content, with a dominant frequency of either 2.0, 3.25, or 3.7-3.8 Hz, depending to a large degree on the temporal location of the event within the sequence and (or) magnitude of the event.

The 22-min sequence is composed of 3 temporal clusters of 5, 7, and 10 events, respectively, although there may be a few additional small events hidden within the codas of identified events. The clusters have durations of 4, 5, and 7 min, respectively, with a 1.5-min interval of quiet between the first and second cluster and about 5 min of quiet between the second and third cluster. Figure 6 shows the amplitude of each event as a function of time. Of particular interest is the fact that the largest event is never the first event in a cluster. In fact, the events more or less increase in amplitude within each cluster, with the largest event occurring last within the first two clusters. Note also that the longest interval between events follows the largest event of the entire 22-min sequence. To our knowledge, the only other instance of such a pattern of seismic energy release is for DLP earthquakes at Long Valley caldera, Calif. (Mitch Pitt, USGS, oral commun., 1993). This pattern is certainly contrary to typical mainshock-aftershock patterns commonly observed for tectonic sources, in which the largest event occurs at or very near the onset of the sequence and the number of events decreases logarithmically with time thereafter.

Each of the three clusters within the 22-min sequence begins with small amplitude events characterized by a dominant frequency of 3 to 4 Hz. These events are followed by larger amplitude events characterized by a dominant frequency (F_0) of 2.0 +-0.03 Hz. Amplitude spectra were computed for a 40-s window taken near the beginning of the first cluster, for each station. The spectra were then normalized and stacked. The result is shown in figure 7A. One can see a very narrow-band spectral peak located at 3.27 +-0.03 Hz. Figure 7B shows the spectrum, similarly computed for the largest event of the second cluster (the largest of all 400+ events). The spectral peak for this event is located at 2.03 +-0.05. The quality factor, $Q = F_0/\Delta F$, of the spectral peaks for these two events ranged from 20 to 50, at each station. Figure 7C shows the spectrum for an event near the middle of the third cluster. For this event there are two spectral peaks of similar amplitude located at 2.03 +-0.04 and 3.74 +-0.04, and Q for both peaks ranges from 20 to 70 at each station.

Figure 7. Stack of normalized (vertical) amplitude spectra from all stations and cross-spectra for two of those stations, BUR and PIE, for a 20-s window beginning with the first arrival for three well-recorded deep long-period earthquakes during the 22-min sequence. X-spectrum, cross-spectrum. Spectral units (horizontal axes) are in hertz. All three events have significant energy at 2.03 +-0.04 Hz. A, Small first event at onset of first subgroup; dominant frequency is 3.27 +-0.03 Hz. B, The largest deep long-period earthquake ever recorded, which occurred at 0859, at end of second subgroup (waveforms for this event are shown in fig. 3A). C, Event near middle of third subgroup, at 0908 (waveforms for this event are shown in fig. 3B); this event has a strong secondary peak at 3.74 +-0.04 Hz. Figure produced using PITSA (Scherbaum and Johnson, 1992).

Table 2 lists the individual events identified within the 22-min sequence along with the maximum peak-to-peak amplitude and dominant frequency for each event. Note the great similarity between this sequence at Mount

Pinatubo and observations of deep tremor made in Hawaii by Aki and Koyanagi (1981): "A typical major episode starts with high-frequency ... with moderate amplitude, grows into violent motion with low frequency (1.5-3 Hz) dominating, and becomes gradually weaker ... near the end of the episode."

Table 2. Chronology of the 22-min deep long-period earthquake sequence of June 4, 1991.

[Sequence is grouped into three temporal clusters. "Dominant frequency at station PIE" was estimated from analog records. Where two dominant frequencies were found, the more dominant frequency is listed first. For those events for which digital records exist, "Dominant Frequency of stacked spectra" was calculated for the stack of normalized spectra for the vertical components at all six stations. This same pattern of initially small-amplitude, higher frequency events followed by larger amplitude, lower frequency events is also observed during 10 other DLP earthquake sequences during June 1-6.]

Time (local)	Amplitude (peak-to-peak) at station PIE (mm)	Dominant frequency at station PIE (Hz)	Dominant frequency of stacked spectra (Hz)
08:51:00	1.4	3.7 ?	-
08:52:13	1.4	3.25, 2.0	3.27, 2.03+-.03
08:52:40	3.8	2.00	-
08:53:30	2.8	2.0	-
08:54:30	3.8	2.0	-
08:56:12	4	3.4-3.8	-
08:56:40	36	2.0	-
08:57:30	23	2.0	-
08:58:12	13	2.0	-
08:58:40	26	2.0	1.98 +-.04
08:59:50	50	2.00	2.00 .03
09:06:40	2	3+	-

09:07:40 2	3.7	-
09:07:55 13	3.83, 1.98	3.78+-.05
09:08:18 6	2.00, 3.75	2.03, 3.74+-.04
09:08:50 6	1.98	2.00+-.04
09:09:12 15	2.0	-
09:10:10 3.3	3.7	-
09:11:30 4	2.0	-
09:12:20 1.2	2.0	-

COMPREHENSIVE SEARCH FOR OTHER DLP SEISMICITY

Once the existence and basic characteristics of DLP earthquakes under Mount Pinatubo, such as a monochromatic 2- to 4-Hz signal with an S-P arrival-time interval (when the S-phase arrival was clear) of 4 to 6 s on the station PIE records and similar ground displacements at all stations, were recognized, an exhaustive search was made for other preeruption DLP events. We began with the simple expedient of overlaying the analog record from station UBO upon the record from station PIE over a light table. Because station UBO, located 1 km from the summit, was operated at 18 dB lower gain (8 times lower magnification at 2 Hz) than station PIE, located 9 km from the peak, signals of similar or larger amplitude on station UBO must have originated at shallow depth near the summit. All other signals were marked as potential DLP events. Many of these signals were obviously related to regional earthquakes, teleseisms, cultural noise, especially helicopters, or transient meteorological conditions near station PIE (wind and rainfall often vary greatly at different locations around the volcano). Such events were discarded from consideration. A type of signal found difficult to distinguish from the DLP earthquake signal was from occasional magnitude ~3 tectonic earthquakes originating at distances of a few hundred kilometers. At this distance, the spectral content of these events is somewhat peaked within the 1- to 3-Hz band, owing to the attenuation of higher frequencies with distance, and the angle of incidence is steep enough to make the S-phase difficult to recognize on vertical component records. Larger regional earthquakes from this distance are readily distinguished by the S-P interval. An additional feature useful for recognizing DLP earthquakes was their unusual tendency to occur in groups in which the largest events occurred between the middle and end of the sequence, whereas tectonic events at regional distances tend to appear either singly or in groups of several with the largest event at or near the beginning of the sequence.

A list of 440 potential DLP earthquakes was eventually compiled for the interval from May 1 through the end of June 11. The earthquakes fell into two temporal groups, the first from 1700 May 26 to 1210 May 28 and the second from 2114 May 31 to 1510 June 8. Most of these events occurred either in large clusters lasting a few hours, with the largest event in a cluster occurring near the middle or end of the cluster, or in small clusters of events with similar amplitudes. For both types of clusters, there usually was at least one event in each cluster with large-enough amplitude on the PIE record that an S-P arrival-time interval of 5-6 s could be observed clearly. I feel confident that at least 400, or 90 percent, of these potential DLP signals are true DLP earthquakes.

Because the identification problems are the greatest among events with small amplitudes, the erroneous inclusion of a few small-amplitude regional events is likely in our list but should have very little effect on our conclusions. For example, the inclusion of 40 such events would inflate the cumulative Reduced Displacement, discussed below, by less than 1 percent. Of the 350 potential DLP earthquakes with amplitudes >=1 mm at station PIE, I believe that at least 330, or 94 percent, of these are in fact DLP earthquakes, and I am confident that virtually all 110 events with amplitudes >=3 mm are DLP earthquakes.

To quantify LP earthquakes, it is convenient to use Reduced Displacement (*RD*) as defined by Aki and Koyanagi (1981). For fixed distance to the source, *RD* is proportional to the displacement, so that the logarithm of *RD* is similar to the magnitude of tectonic earthquakes. The *RD* for the DLP earthquakes and tremor at depths of 34 to 35 km in depth under Mount Pinatubo is given by (see appendix A):

$$RD = \text{Amplitude at PIE x 600 } [\text{cm}^2]$$

where the amplitude is measured in centimeters. *RD* for the observed DLP earthquakes ranges from 30 to 3,100 cm^2 (M ~1.7-3.7), and the cumulative *RD* totals about 68,000 cm^2, roughly equal to a single M 5 earthquake.

In addition to the 400-plus DLP's, there are several long periods of time during which nearly monochromatic signals of about 2 Hz were apparently sustained with amplitude near background-noise levels. Because these long signals are of 2 Hz and because DLP earthquakes are embedded within most of these periods of tremor, these signals were determined to be DLP tremor. These periods of apparently continuous, deep tremor occurred at the following times: 1800 May 26-0400 May 27, 1200-1300 May 27, 1600-1900 June 1, 0200-0300 June 2, 0100-0200 and 1700-1800 June 3, 1000-1600 June 4, and 0600 to 0800 June 7. If we assume, as do Shaw and Chouet (1989) for DLP tremor at Kilauea, that the tremor is composed of three DLP earthquakes per minute, and that each of these DLP earthquakes has an average *RD* of 20 cm^2 (M ~1.5), the additional cumulative *RD* due to observed DLP tremor is about 90,000 cm^2, roughly equivalent to a single earthquake of M 5.2.

The estimated moment release during these DLP earthquakes and during periods of DLP tremor over the interval May 1-June 12, is shown in figure 8. Also shown for the same time interval is the estimated moment release during SLP earthquakes and during volcano-tectonic earthquakes beneath the dome, as well as SO_2 emission.

Figure 8. Moment release, as calculated for double-couple earthquakes (proportional to energy), per 4 h for four types of seismicity and SO_2 emission. A, Deep long-period earthquakes; B, Deep long-period tremor; C, Shallow long-period earthquakes; D, shallow volcano-tectonic earthquakes beneath the dome; E, SO_2 emission in 1,000 metric tons per day. Figure produced using PC-QPLOT (Murray and others, 1993).

Figure 9. S-phase arrivals for seven well-recorded deep long-period earthquakes at the vertical component station GRN. Maximum amplitudes in counts are shown for each trace at left. Date and time of each event is shown at right. Records have been low-pass filtered with a 4-pole Butterworth filter with a high-frequency cutoff at 6 Hz. Vertical line at 3.7 s indicates the P-phase arrival. Note that the S-phases arrive at similar times except for the bottom trace, where it clearly arrives earlier. Figure produced using PITSA (Scherbaum and Johnson, 1992).

Figure 10. Locations of all 10 locatable deep long-period earthquakes and shallow volcanic-tectonic seismicity for the period from May 26 through June 7, 1991. Triangles are seismic stations. Contours are 500 and 1,000 m. A, deep long-period earthquake epicenters. B, Shallow volcano-tectonic earthquake epicenters; the large cluster of shallow volcano-tectonic earthquakes coincides with the location of the new dome first observed on June 7. C [and D], Cross section along A-A' showing both deep long-period earthquakes and shallow volcano-tectonic earthquakes. Magnitudes as in A. D, Depth versus origin time for the deep long-period earthquakes. "?" indicates events for which the depth was estimated from station PIE analog records. Magnitudes as in A. Figure produced using PC-QPLOT (Murray and others, 1993).

Of the 400-plus DLP events identified, all or part of only 18 were recorded digitally. Only 10 of these were recorded well enough to be located. Our inability to record a greater percentage of the DLP earthquakes is due to two main factors: (1) stations located on the flanks of the volcano were operated at low gain settings because of the high levels of seismic noise originating from within or very near the volcanic edifice and (2) the trigger algorithm was set to recognize only shallow high-frequency events. An improved trigger algorithm has since been developed that triggers much more reliably on the low-frequency signals often present at volcanoes (Evans, 1992).

Both pulses of DLP seismicity apparently become shallower with time, as estimated from S-P arrival-time intervals from the PIE analog record. Both pulses seem to begin with events near 40 km in depth and are followed by a majority of events located at about 33 to 34 km in depth. Near the end of the second pulse, several events originated at about 28 km in depth. Figure 9 shows the P- and S-phase arrivals for 7 well-located, onscale DLP earthquakes at station GRN, a vertical-component-only station, located relatively far from the summit, for which S-phase arrivals are particularly clear. Note that the S-P arrival-time interval for the last event is noticeably shorter than for those above, indicating a distinctly shallower origin. Figure 10 shows the locations of the digitally well-recorded DLP earthquakes, all of which occurred during the second pulse of DLP seismicity, during June 3-6. At the 95% confidence limits, the horizontal errors are within 3 km, and the vertical errors are within 1.2 km for all of the events shown. Of the 10 hypocenters shown, 8 are statistically indistinguishable from, and include, the 2 well-located events discussed earlier (largest octagons) that occurred within the 22-min sequence. Figure 10D shows the depths of these well-located events versus time.

Amplitude spectra were computed for 11 DLP earthquakes all together, including the three shown in figure 7, for the three outermost digitally recorded stations, BUG, BUR, and PIE. The spectra were normalized and then stacked for each event. Of these events, five are nearly monochromatic at 2.0 Hz, and two are nearly monochromatic at 3.25 Hz. The remaining four events have dominant frequencies of both 2.0 and 3.75 Hz, and have a more broad-band spectral content, similar to the spectra of "hybrid" earthquakes (Lahr and others, 1994), with considerable additional energy over the band from 3 to 6 Hz. The two events located at 28 km in depth are both of this type. Stacked spectra for one each of the first two types of events and two of the hybrid type are shown in figure 11. At least one event of each of these three types locates at 33 to 34 km in depth and about 5 km northwest of the summit. Because the travel path between the source and the receiver is virtually identical for these events, the differing spectral contents must be attributable to the source process rather than path or site effects.

Figure 11. Stack of normalized amplitude spectra from the four 1-Hz vertical component stations farthest from the summit (BUG, BUR, GRN, and PIE) and cross-spectra for BUR and PIE, for four events. A, 1534 June 3. B, 2116 June 5. C, 0557 June 6. D, 1744 June 6. The first three events occurred at 33 to 34 km in depth about 5 km northwest of the summit. The fourth event originated 5 to 6 km shallower. Figure produced using PITSA (Scherbaum and Johnson, 1992).

TEMPORAL LINK WITH NEAR-SURFACE GEOLOGICAL AND SEISMOLOGICAL ACTIVITY

In this section, I briefly describe some of the major geological and seismological changes at or near the surface of the volcano that followed shortly after high rates of DLP energy release. The temporal correlation between the two is so striking that it seems inescapable that the process which produced the DLP activity deep beneath the volcano also directly influenced processes within the uppermost portions of the magma chamber.

The first of the two principal pulses of DLP seismicity began at 1700 May 26 and lasted about 2 days. Figure 12 shows part of the analog record from station PIE for the beginning of this period. The activity began with a 30-min swarm of DLP earthquakes that included 7 events with RD of 90 to 270 cm^2. Another swarm of at least 16 DLP earthquakes began about 8 h later. These 2 swarms appear to initiate and terminate a 10-h period of continuous deep tremor. The first shallow LP earthquake recorded during the month of May (others are seen on records from April 8-10) occurred 1 h after the beginning of the DLP activity and was followed by at least four smaller shallow LP earthquakes within the next hour. About 2 h later, there began a 4-h period of continuous shallow LP tremor, with a dominant frequency of 0.8 Hz, that was superimposed upon the 2-Hz DLP tremor. This period of shallow tremor was accompanied by a major increase in steam emission from vents high on the north flank of the volcano, at a location where the dome later emerged. Measurements of SO_2 gas emissions taken at about 0800 on May 27 showed a threefold increase over the previous measurement taken on May 24 (fig 8E). This was the largest increase in SO_2 emission observed prior to June 10 (see Daag, Tubianosa, and others, this volume).

Figure 12. Helicorder record from May 26, 1991, showing the first deep long-period earthquakes (DLP) and tremor (DLP TREMOR), the first shallow long-period earthquakes (SLP), and tremor (SLP TREMOR). The first DLP earthquake occurred at 1807. The first SLP earthquake occurred at 1907. Other signals on this record are shallow volcano-tectonic earthquakes (VT) from an area 5 km northwest of the summit, regional earthquakes (R), and a passing helicopter (H).

After nearly 3 days of subsequent DLP quiescence, the second pulse of intense DLP activity began in earnest on June 1 and lasted into the morning of June 7, followed by a few additional events through June 8. The highest rate of DLP energy release occurred very early on June 4 and included the 22-min swarm with the largest single DLP (*RD* =3,100 cm^2). The second major episode of SLP earthquakes began about 3 h after the 22-min swarm (see fig. 2), and the steam emissions became much more intense about this time. During a 36-h period from 2100 June 5 through 0900 June 7, high-amplitude DLP's were frequent; this period includes 16 events with *RD* >= 350 cm^2. The largest event during that period, with *RD* of 2,300 cm^2, occurred near the end of the period. Early on June 6, shallow high-frequency volcano-tectonic seismicity increased dramatically beneath the summit and gradually shoaled from 2 km depth toward the surface (Harlow and others, this volume). This shallow seismicity culminated with the emergence of a dome on June 7. The dome was composed primarily of hybrid andesite but contained inclusions of freshly quenched olivine basalt (Pallister and Hoblitt, 1992) which, according to Pallister (USGS, oral commun., 1992), had a probable residence time within the upper portion of the magma chamber of no more than a few days.

By 2030 June 8, DLP earthquakes and tremor apparently ceased, though two small events may have occurred on June 9. One week later, Mount Pinatubo erupted in the largest eruption the world had seen in almost 80 years. Note that recognition of possible DLP seismicity after June 11 is essentially impossible, owing to (1) very high levels of shallow activity from June 12-15, (2) destruction of the network by the cataclysmic eruption on June 15, and (3) the generally 12-dB lower gain of the replacement network installed at the end of June. The 2-week-long period from May 26 through June 8, 1991, contained both the most energetic, and largest concentration, by far of DLP earthquakes ever recorded beneath any plate margin volcano worldwide.

SUMMARY OF OBSERVATIONS OF DLP EVENTS

APPEARANCE OF DLP EARTHQUAKE WAVEFORMS

On three-component records, DLP earthquake waveforms exhibit clear P-, P-S-, and S-phases. Spectra are dominated by very narrow bandwidth low-frequency signals with Q=20-50. The wave field is uncorrelated over the diameter of our network, 16 to 19 km, which represents an angular separation of 25 to 29°. These observations are similar to those made at Kilauea volcano by Aki and Koyanagi (1981) and Redoubt volcano by Chouet and others (1994). DLP coda lengths at Mount Pinatubo are not significantly longer than coda lengths of tectonic events of similar amplitude and distance.

VARYING DOMINANT FREQUENCIES BETWEEN DLP EARTHQUAKES

Most DLP events were found to have a dominant frequency of 2.0 Hz. Some sequences that contained large amplitude events, however, were observed to begin with small amplitude events with dominant frequencies of 3.2-3.3 or 3.7-3.8 Hz. At least 11 sequences from June 1 through June 6 began in this manner and were followed by larger 2-Hz events within 1 min. This observation is very similar to that made by Aki and Koyanagi (1981). A few isolated events were also observed with a 2-Hz dominant frequency, but these also contained strong additional energy in the 3- to 6-Hz range and appear similar to the "hybrid" events of Lahr and others (1994).

DEPTH AND RESTRICTED VOLUME OF HYPOCENTERS

Of 11 locatable events, which span the interval June 3-6 and which include events with the four types of spectral signatures described previously, 9 originated from within a small source volume located about 33 to 34 km beneath, and about 5 km northwest of, the summit near the base of the crust. These events originated within a maximum volume defined by the associated 95% error ellipses: an oblate spheroid about 5 km across horizontally and about 2 km in vertical extent. The actual source volume for the events may be much smaller. The two remaining locatable events (the two that occurred last) locate at about 5.5 km shallower than the previous nine events. S-P arrival-time intervals from the PIE analog records indicate that, within each of the two pulses of DLP seismicity, the earliest events originated at depths as great as 40 km and the events generally became shallower with time.

ENERGY RELEASE PATTERN

Reduced displacements of the 400-plus observed DLP earthquakes ranged from 30 to 3,100 cm^2 (M ~1.7-3.7). The 25 h of DLP tremor had an average reduced displacement of 20 cm^2. Within well-recorded DLP earthquake swarms, amplitudes are observed to generally increase with time, with the largest event occurring at or near the end of the swarm. When the largest event occurs before the end, it is generally followed by the longest inter-event interval within that swarm. To our knowledge, the only other instance of such a pattern of seismic energy release is for DLP earthquakes at Long Valley caldera, California (Mitch Pitt, USGS, oral. commun., 1993).

CORRELATION OF DLP ACTIVITY WITH NEAR SURFACE PHENOMENA

The initial burst of DLP activity on May 26 was followed 1 h later by the onset of shallow (depth <2 km) long-period earthquakes. Shallow long-period tremor and large steam emissions followed within 2 h, and the subsequent measurement of SO_2 showed a threefold increase over previous levels for the month to that date. The most intense DLP activity, on June 4, was followed within 3 h by shallow long-period earthquakes and increased steam emissions. DLP seismicity continued at a relatively high rate for 3 more days and was accompanied by inflation of the summit and a dramatic increase in shallow volcano-tectonic seismicity which shoaled with time. The DLP seismicity apparently tapered off after the emergence of the andesite dome on June 7.

A MODEL FOR DLP EARTHQUAKES

In this section the characteristics of DLP earthquakes and tremor observed above are interpreted in terms of a model proposed recently by Chouet and others (1994). The model is shown to be compatible with both the range of DLP event depths observed at Mount Pinatubo and the variety of generally wider depth ranges observed at other volcanoes. In the section following this, the model is used to interpret the seismological and geological events leading up to the cataclysmic eruption of Mount Pinatubo.

The observations summarized in the section above place strong constraints upon possible models:

1. The narrow-bandwidth, long-period characteristic cannot be ascribed to either a path or receiver effect but must be an effect of the source process.

2. Over a 1-week interval, from June 1-7, this source repeatedly produced DLP sequences that began with events having dominant frequencies of 3.2-3.8 Hz followed by larger events having dominant frequencies of 2.0 Hz; this pattern implies a repetitive, nondestructive source process of fixed-scale length.

3. This source process seems to be confined to a relatively small volume, less than a few kilometers in diameter, located near the base of the crust and almost directly beneath the volcano's summit; this small source volume appears to rise with time.

4. The energy release pattern within swarms, wherein DLP earthquakes generally increase in amplitude and the largest is followed by the longest period of quiescence, implies a process in which the timing is governed by a feature akin to a pressure relief valve.

5. Most importantly, this process near the base of the crust immediately preceded major geological and seismological effects at the surface, more than 30 km above, within 1 h to a few days, and therefore implies a causal relationship.

The combination of these observations provides compelling evidence that the DLP earthquakes and tremor were produced by flow, near the base of the crust, of basaltic fluids upward into the magma chamber. I believe that remnants of these fluids were preserved as inclusions of basalt within the andesite dome that emerged on June 7. These basalt inclusions have characteristics compatible with a deep-crustal origin and a probable residence time within the upper magma chamber of days to a few weeks at most (J.S. Pallister, USGS, oral commun., 1992).

In Chouet and others (1994), a mixed-phase magmatic fluid is forcefully injected through a crack and, upon reaching choked-flow conditions, excites the crack to resonance. In this model, the choked-flow condition is met when the flow speed exceeds Mach 1 locally. Such condition can be achieved at flow speeds near a constriction in the conduit if the magmatic fluid contains a free gas phase. Constriction can greatly increase flow speeds locally, while the compressibility of a mixed-phase fluid can greatly reduce the sound speed of an otherwise gas-free fluid (for example, Kieffer, 1977). Constrictions would be plentiful in a magma-transport structure consisting of magma-filled tensile cracks as proposed by Hill (1977) and Shaw and Chouet (1991).

Note that the choked-flow condition may be met only when the void fraction lies within a narrow range: (1) enough bubbles must have formed to lower the sound speed of the fluid to flow velocity, yet (2) the volume percent of gas bubbles must remain small enough that the magma remains somewhat incompressible, else choked flow ceases and (or) the capacity of the fluid to sustain resonance is destroyed through acoustic absorption. Though the required range of the void fraction for sound speed minima, and therefore choked flow, in basaltic magma is unknown, it almost certainly depends on pressure and possibly bubble radius (Kieffer, 1977). The void fraction for basaltic fluid at most times and places along the path of ascent undoubtedly lies outside this range and implies that basaltic fluid mostly rises aseismically.

The model requires that a free gas phase exist, at least temporarily, at our DLP source depth of 28 to 35 (possibly 40) km below sea level. To be a reasonable candidate, this gas must have a very low solubility and exist in relatively large quantity within basaltic magmas. CO_2 is the leading candidate for such a gas because it has by far the lowest solubility and exists in the greatest quantity among volatiles commonly present in basaltic magmas (Bottinga and Javoy, 1990). Gerlach and Graber (1985) showed that CO_2 is present in large quantities (0.3 to 0.6 wt%) in basaltic magmas from Kilauea and "is transported predominantly as a fluid saturating the parental melt from depths of ~40 km." Stolper and Holloway (1988) showed that basaltic CO_2 probably begins to exsolve in the 30- to 50-km depth range. Evidence for the existence of CO_2 as a separate gas phase at this depth range is provided by Willshire and Kirby (1989) and Kirby (1990), who found that xenoliths, which originated at upper mantle depths, showed clear evidence of healed fractures which once contained CO_2 gases.

DLP earthquakes have been reported to occur over certain depth ranges at some volcanoes but very different depth ranges at other volcanoes (see table 1). In particular, at Long Valley caldera, Mount Lassen, and Clear Lake, DLP earthquakes are reported from 10 to at most 25 km in depth while at Kilauea Volcano, most DLP events occur between 30 and 50 km in depth. Although, as noted above, depths of DLP earthquakes can be very difficult to estimate without particle motion studies using three-component data, the variety of depth ranges is most likely real. Because the production of DLP seismicity depends critically on the existence of a small volume percent of a free-gas phase, which depends on the solubility of that gas, which in turn depends on magma composition and on pressure, the variety of depth ranges of DLP events should reflect differences in magma composition between volcanoes and different pressure-temperature profiles along the deep magma conduits beneath different types of volcanoes. Significantly different depth ranges of DLP events should be expected from contrasting tectonic regimes such at Hawaii, where oceanic crust overlies a mantle plume, versus northern California, where thick continental crust overlies a subducting slab, owing to corresponding differences in magma composition. For example, the magma at Kilauea is likely more basaltic, with a higher CO_2 content, than that of plate margin volcanoes, so CO_2 saturation should occur at a greater average depth there.

INTERPRETATION AND IMPLICATIONS

Several effects are predicted as a result of our model: (1) a pressure transient produced by the choked flow itself, (2) a hydraulic pressure wave produced by the addition of a fluid volume at the base of the magma column, and (3) advective overpressure produced by gas bubbles rising in the column.

The pressure transient, ΔP, is related to the behavior of the shock wave produced by the choking, that is, the flow velocity exceeding Mach 1. ΔP is given, according to this model, by the formula derived in appendix 2. The largest DLP earthquake during the initial 1 hr of DLP activity occurred about 6 min after the onset of the activity. *RD* for this event is 240 +-50 cm^2, and, therefore, ΔP is about 80 +-17 bar. This pressure transient travels at acoustic velocity and should have a duration on the order of the coda length of the DLP earthquake.

Hydraulic pressure originating from the addition of a fluid volume at the base of the magma column should propel a corresponding volume of magma upward into the base of the magma chamber and thereby increase the pressure in the chamber proportionately. This effect is long term, but because the volume of the magma column is tiny compared with the volume of the magma chamber and the average flow velocity is fairly slow, the effect is negligible over days to weeks and may be impossible to detect.

Bubbles within the rising magma column will also increase the pressure by means of advective overpressure. Steinberg and others (1989) have pointed out that, within a closed system with constant volume and temperature, a single perfect gas bubble rising by distance Δh through an incompressible fluid of density ρ will increase the pressure everywhere within the system by $\Delta P = \rho g \Delta h$, where g is the force of gravity. Under these assumptions, and using ρ=3.2 g/cm^3, and g=980 cm/s^2, a bubble rising 1 km would increase the pressure everywhere along the interconnected conduit system, including the upper portion of the magma chamber, by 330 bar. Allowing for compressibility of the fluid owing to multiple bubbles, this value is reduced by one-third to 11 bar per kilometer of rise. If the bubbles are entrained within the basaltic magma, then they will rise at, and therefore the pressure will increase with, the flow velocity. At the average flow rate of about 120 m/h (see below), the pressure within the system should increase at an average rate of only 11 bar/h. If the magma column is not strictly continuous, but, rather, comprises a plexus of magma-filled tensile cracks offset by faults, as proposed by Hill (1977) and Shaw and Chouet (1991), the pressure increase may take some time to propagate upward through the system.

One or more of the above effects may account for the observation that, on May 26, the first DLP seismicity recorded since May 1 was followed 1 h later by the first shallow (depth <4 km) LP earthquakes recorded since May 1. This same effect was also observed about 3 h after the 22-min sequence of DLP earthquakes on June 4. In the case of the May 26 events, communication between the DLP source region and the near surface was remarkably rapid, travelling at an average velocity of 8-12 m/s over 30-37 km vertically.

Finally I speculate that the batch of basaltic magma that produced the first DLP earthquakes on May 26 may have been the same batch of basaltic magma that reached the surface as inclusions in the dome that emerged on June 7. If so, the average ascent rate of that batch, assuming a starting depth of 35-40 km to the surface over 11.5 days, is 111-126 m/h, or nearly 3 km/day. A similar ascent rate, over a greater depth range, has been estimated for Hawaii (Decker, 1987). At this rate, the batch of basaltic fluid that produced the May 26 DLP seismicity should have reached the base of the dacite within the magma chamber at about 14 km in depth (Pallister and others, this volume) by about June 2. After a few days of magma-mixing there, the mixed-magma product extruded as a dome on June 7. By the above reasoning, the second batch would have begun reaching the base of the dacite by about June 8 and would have continued to arrive there for the following week. It may have been this second batch of basalt which pushed the system to the plinian stage. Because little hard-rock evidence of basalt was found in the products of the June 12-15 eruptions, other than from remnants of the dome itself, however, this second batch may have led to the final eruptive stages principally by contributing pressure, heat, and volatiles.

To summarize the events leading up to the cataclysmic eruption, as interpreted in terms of this model, I propose that (1) the system was leaky and relatively unpressurized prior to May 26; (2) on May 26, the first batch of basaltic magma began migrating upward from the base of the crust, producing the first DLP activity; this briefly pressurized the system enough to produce minor SLP activity, which, in turn, opened the system further and released enough volatiles to reduce the pressure gradient and stop further SLP activity temporarily; (3) during June 1-4, this first batch of basaltic magma reached the base of the dacite pluton as the second batch of basaltic magma began migrating upward from the base of the crust; the system was still somewhat leaky at this time, as evidenced by the fumarolic activity and lack of shallow LP activity prior to this time; (4) by June 4, pressure within the magma chamber had built sufficiently for the renewal of minor SLP seismicity and, by June 7, had built sufficiently to produce the emergence of a dome; (5) by June 11, the second batch of basaltic magma reached the base of the dacite pluton, dome growth had sealed the system shut, the volatiles became trapped, and pressure began to build at shallow depths below the dome, as manifested by a reemergence of SLP activity at that time, and led to the first plinian eruption, on June 12; (6) after the dome was partially destroyed, during the first plinian eruption, we begin to observe large SLP earthquakes that marked the onset of rapid acceleration of the pressure buildup toward the cataclysmic eruption on June 15 (Harlow and others, this volume).

The DLP seismicity that occurred during the 3 weeks leading to the eruption of Mount Pinatubo is the greatest ever reported, both in terms of number of events and energy released, for any plate margin volcano and was immediately followed by the largest eruption in the world in almost 80 years. If such large eruptions are triggered primarily by basalt injections such as occurred at Mount Pinatubo, marked by DLP seismicity, early recognition and quantification of large-scale DLP seismicity may provide one of the best tools for predicting the timing and size of such destructive eruptions.

The reader is cautioned that DLP events are usually difficult to record, owing to masking from surficial noise at the volcano. DLP events at Mount Pinatubo were, in general, very poorly recorded by stations nearest the summit. DLP events may have occurred at other volcanoes during the last few decades and gone undetected. In order to monitor DLP seismicity effectively, care should be taken to install several high-gain stations at quiet sites at least 10 to 15 km from the volcano summit.

CONCLUSIONS

1. About 400 deep long-period (DLP) earthquakes and 25 h of DLP tremor occurred beneath Mount Pinatubo, in two pulses, during May 26-28 and May 31-June 8. This DLP seismicity is greater than previously reported beneath all convergent margin volcanoes worldwide. Though two orders of magnitude more DLP earthquakes have been recorded at Kilauea, Hawaii, over the last 30 years, DLP seismicity apparently released more seismic energy at Mount Pinatubo during the 2-week-long period from May 26 to June 8, 1991, than at Kilauea during the entire 30 years.

2. Early events in each pulse may have originated as deep as 40 km. Eight of the locatable DLP earthquakes originated at depths of 33 to 34 km beneath a point about 6 km northwest of the summit. Two later locatable events originated at about 28 km in depth beneath the same area. Reduced Displacement for the largest event is about 3,100 cm^2, an amount that makes it the largest deep long-period earthquake yet reported.

3. DLP earthquake waveforms exhibit clear P- and S-phases. Spectra are dominated by very narrow bandwidth signals with Q=20-50. The dominant frequency for most of the events, including the 10 largest, is 2.0 Hz. A few of the smaller events have dominant frequencies of 3.25 and 3.75 Hz.

4. The first DLP earthquake recorded by the seismograph network, on May 26, was followed 1 h later by the first shallow (less than 3 km deep) long-period earthquakes and 2 h later by a shallow tremor and large steam

emission. The high rate of DLP seismicity during June 4-7 was accompanied by inflation of the summit area, a rapid increase in shallow seismicity beneath the summit, and the eventual extrusion of a dome containing inclusions of a very primitive, freshly quenched, olivine basalt only recently arrived (days to a few weeks at most) from the deep crust.

5. The DLP earthquakes and tremor were likely produced by the forceful injection of mixed-phase basaltic fluids upward through cracks from near the base of the crust into the upper magma chamber, which contained a dacitic residuum. The pressure, heat, and volatiles injected into the dacite pluton by the first pulse of basaltic magma led to magma mixing and the dome extrusion on June 7, which sealed the system. The almost continual arrival of additional basaltic magma at the base of the pluton pressurized the system and led to the destruction of the dome on June 12. Depressurization, degassing, and vesiculation of magma at the top of the reservoir, following the vent-clearing eruptions, together with pressure and volatiles from the arrival of yet more basaltic magma, led to the paroxysmal dacitic eruptions of June 15.

ACKNOWLEDGMENTS

I wish to acknowledge the many members of the Philippine Institute of Volcanology and Seismology (PHIVOLCS) and the Volcano Crisis Assistance Team (VCAT) of the U.S. Geological Survey who labored day and night under very stressful and dangerous conditions to install and maintain the seismograph network and the data collection apparatus. From PHIVOLCS, I especially wish to thank Ed Laguerta, Gemme Ambubuyog, S. Marcial, and A. Melosantos. From VCAT, I especially wish to thank Dave Harlow, John Power, John Ewert, Andy Lockhart, and Tom Murray. Discussions with Bernard Chouet concerning the choked-flow model and its implications were critical. This manuscript was greatly improved by reviews from Motoo Ukawa, Bernard Chouet, Chris Newhall, Dave Hill, and Manoling Ramos.

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APPENDIX 1. CALCULATION OF REDUCED DISPLACEMENT

The formula for Reduced Displacement, which is the RMS displacement corrected for geometrical spreading and instrument response, is given by formula 2 of Aki and Koyanagi (1981):

$$RD = Ar/(2M\sqrt{2})$$

where A is the peak-to-peak amplitude in centimeters of the largest body wave phase, usually the shear wave, averaged over several cycles, r is the hypocentral distance (in centimeters), and M is the instrument magnification to ground displacement at the dominant frequency.

Shear waves from the DLP events beneath Mount Pinatubo arrived at our digitally recording stations within 15° of vertical. At the stations, the particle motion of these shear waves would have been confined essentially to the horizontal plane so that true shear-wave amplitudes would be best measured off horizontal-component recordings. Unfortunately, all but one of the digitally recording seismograph stations, including PIE, the helicorder record upon which the comprehensive catalog was based and the amplitudes taken, contained only vertical-component sensors. Worse yet, the three-component station was operated at the lowest magnification of any station in the network and recorded only a few of the largest DLP events. To find out by what factor vertical-component recordings underestimated true shear-wave amplitude, I compared vertical- and horizontal-component recordings of several DLP earthquakes at our three-component station. I found that shear-wave amplitudes recorded by the horizontal components (A_H) averaged 3 times the amplitudes recorded by the vertical component (A_V). Therefore, for estimates of A, I use A_V measured from the station PIE and multiply that value by this factor of 3.

To calculate the RD of DLP earthquakes or tremor from A_V for body waves at station PIE, I take (1) the depth to be 35 +-1 km; (2) the epicenter to be the average of the best located events, at 15°10'N. latitude, 120°19.2'E. longitude, yielding $r = 36.7$ km = 36.7×10^5 cm; and (3) the magnification at 2 Hz of $M = 6{,}250$. Note that Lockhart and others (this volume) estimate $M = 5{,}000$ +-20%, assuming a sensor sensitivity of 1V/cm/s and neglecting site amplification. Probable site amplification at station PIE, however, causes calculations of RD to average about 25% too high compared with estimates of RD based on data from other stations, so the higher value for M will be used. This will lead to more conservative (lower) estimates of RD.

For DLP events: $RD_{\mathrm{PIE}} = 3A_V r/(2M\sqrt{2}) = A_V \times 600$ +-20% [cm^2]

For the largest DLP earthquake in the catalog, recorded at 0859.5 June 4, A_V was estimated to be 5 +-1 cm at station PIE, so RD = 3,100 +-700 cm^2.

For the largest DLP on May 26, RD = 240 +-50 cm^2.

APPENDIX 2. CALCULATION OF EXCESS PRESSURE

The pressure fluctuation, ΔP_F, produced by DLP earthquakes over the volume of the source is related to RD by formula 12 of Chouet and others (1987):

$$RD = \Delta P_F\,(1/2\pi)(1/Q_s)^{1/2}(V/(2f\alpha\rho b))^{1/2}$$

where Q_s is the quality factor corresponding to radiation loss, V is the volume of the conduit, f is the dominant frequency, α is the P-wave velocity, ρ is the density, and b is the bulk modulus of the basaltic fluid in the crack. This formula relates RD, calculated from the maximum P-wave amplitude (hereafter referred to as RD_P), to ΔP_F. However, because I have calculated RD from the maximum horizontal component of the S-wave (hereafter referred to as RD_S), ΔP_F based upon RDS will be too large by some constant factor. By calculating both RD_P and RD_S for the five most well-recorded events, I find that factor to be 4, that is, $RD_P = RD_S/4$. Thus, in order to use the above formula to calculate ΔP_F, the values for RD, as calculated according to appendix 1, must be divided by 4.

I assume, as Chouet and others (1987) did for Hawaiian DLP tremor, that the quantity $(1/Q_s)^{1/2}$) is on the order of 0.1, α = 8 km/s, and ρ = 3 g/cm^3. I assume a small (~1 vol%) bubble content, although the bubbles will be nearly incompressible at 35 km depth. I assume, for the magmatic fluid, that b is ~10^{10} dyne/cm^2 and we shall use V, as estimated by Shaw and Chouet (1991) for deep fractures beneath Kilauea, to be $V = 2 \times 10^{10}$ cm^3. The dominant frequency for the vast majority of the DLP events, including all of the largest events, is f = 2 Hz. Inserting these values and rearranging gives ΔP_F =138,000 RD_P =34,500 RD_S. Following this reasoning, the smallest DLP events in our catalog produced pressure fluctuations of about 10 bar, while the largest event, (with an RD_S ~3,100 cm^2) produced a pressure fluctuation of about 1 kbar.

Note that the ambient pressure, due to the overburden, at the top of the magma chamber at about 6 km below the surface (Mori, Eberhart-Phillips, and Harlow, this volume) is about 2 kbar and at the DLP earthquake source depth of 34 km is probably about 11 Kbar.

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Last updated 06.11.99