University of Nevada
Reno

THE 1968 ADEL, OREGON, EARTHQUAKE SWARM

by

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Reno

August, 1976
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ABSTRACT

Historical seismic activity in Oregon is investigated; it is found that earthquake swarms are not atypical of seismic activity in southern Oregon. Swarms tend to occur in regions of heterogeneous geologic structure, or at low stress levels. A swarm of earthquakes occurred during May, June, and July of 1968 near Adel, Oregon; the largest of which had magnitude 5.1 on May 30. The focal mechanism of this shock indicates left-lateral oblique-slip motion on a plane that dips 80° E and strikes N 40° E. The vertical component of motion on this plane is reverse fault motion, which implies a change from extensional to compression stress. 169 events were selected for detailed analysis: epicenters fell within a rectangle about 15 km long (N-S) and 6 km wide; depths averaged about 9 km. A duration-magnitude relationship was established for the tripartite array; the value of b was found to be 1.13.
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CHAPTER 1

SEISMICITY OF OREGON, EARTHQUAKE SWARMS

Beginning in late May of 1968, a swarm of earthquakes occurred near the town of Adel in Southern Oregon, and continued through June and July. Magnitudes of the earthquakes ranged from 5.1 down--many were felt in Adel and nearby towns. Some moderate damage was done by a few of the larger shocks. No surficial expression of fault motion was found; some cracks in pavement and unconsolidated fill were attributed to slumping.

Oregon is found to have had a historically low level of seismicity, mainly centered on the Portland area. Smaller events and swarms predominate in the southern part of the state. The occurrence of swarms is related to low stress and heterogeneous structure.

The Adel Earthquake Swarm

The town of Adel, in the Warner Valley in south-central Oregon, was shaken by a swarm of earthquakes that started on May 26, 1968, and continued through the month of June of that year. The largest earthquake in the swarm had a magnitude of 5.1 (USCGS Preliminary Determination of Epicenters); many shocks in the swarm were felt in Adel and nearby towns, and some did moderate damage. It is because there were several events of approximately the same
magnitude, and no distinctly larger main shock, that the Adel sequence is considered to be a swarm. Most of the large earthquakes occurred during the first ten days of the swarm. After June 5, 1968, only 6 events of magnitude greater than 3.3 are reported in the Preliminary Determination of Epicenters; 20 such earthquakes were reported between May 27 and June 5. Table 1 is a listing of the larger felt events in the swarm.

**Damage.** The most strongly felt shock of the series (magnitude 4.7 on June 3, 1968) damaged nearly every building in and near Adel. Nearly every chimney fell (all were old). Books in the school library were thrown to the east from shelves on the west wall, but not from shelves on the north wall. Groceries were thrown from shelves and glassware broken at the Adel store. Small objects were thrown about in dwellings, more strongly in trailers than in houses. One rock wall of an old storage building collapsed.

**Geologic Effects.** A field investigation (Couch and Johnson, 1968) discovered cracks in the asphalt of State Highway 140 about one mile west of Adel, apparently caused by the slumping of unconsolidated fill about one mile south of Adel. Rocks were dislodged from the fractured and jointed basalt flows of the fault scarp that forms the west wall of Warner Valley. These rockfalls and accompanying dust clouds were seen from Adel all along the west wall of
TABLE 1

Summary of felt events of Adel Swarm
(From United States Earthquakes, 1968)

All Times are PDT

May 26: Observer at Adel, Oreg., reported series of shocks began on May 26 (time not given) and were felt daily to June 11.

May 29: 16:35:59.8. Epicenter 42.3° north, 119.8° west, southern Oregon, W. Magnitude 5.1. Felt at Adel and Lakeview (V), Oregon., and Fort Bidwell and Willow Ranch (east side of Goose Lake, IV), Calif.

May 29: After 21:00. Intensity V at Adel, Oreg.

June 3: 05:27:39.7. Epicenter 42.2° north, 119.8° west, southern Oregon, W. Magnitude 4.7 VI. Felt over approximately 7,000 square miles of southern Oregon and northeastern California. At Adel, Oreg., old chimneys fell or were cracked, and part of an old rock cellar fell and rest of buildings cracked. Ground fissured 2.5 miles northwest of Fort Bidwell, Calif., along Bidwell Creek; a house sustained some foundation cracking and shifting.

June 3: 18:38:29:0. Epicenter 42.3° north, 119.8° west, southern Oregon, W. Magnitude 4.0. Slight damage reported at Adel.


June 4: 02:58:22.8, 20:51:56.8. Epicenter 42.3° north, 119.9° west, southern Oregon, W. Magnitudes 4.2 and 4.7, respectively. Felt at Fort Bidwell, Calif.

June 11: 17:20:56, 17:46:22.4. Epicenter (1) 42.1° north, 120.0° west; (2) 42.1° north, 119.9° west, southern Oregon, W. Magnitudes 3.4 and 4.3, respectively. Felt at Fort Bidwell, Calif.

June 11: Between 21:00 and 22:00. Very light shock felt at Fort Bidwell, Calif.

June 25: 16:43. Press reported shock in Lake County, Oreg., at this time.

June 26: 01:48. Press reported shock in Lake County, Oreg., at this time.
the valley. Slickensides on this fault scarp were examined for signs of recent motion; no apparent displacement was detected.

Couch and Johnson state that Charles Crump reported no change in the activity of Crump geyser, about three miles north of Adel, either during or after the series of larger earthquakes. L. Richards (1968) states that Mr. Crump reported that Crump geyser had been almost dormant before the June 3 shock, when "it came to life and erupted steadily for twelve hours." Couch and Johnson find increased flow in one hot spring to be the only geothermal effect attributable to earthquake activity.

Geology. The steep cliffs that form the east and west walls of Warner Valley are horsts, or upraised blocks, and the valley itself is a graben, or downthrown block. The generally north-south trending horsts and grabens that form the Warner Valley are transected obliquely by many other faults, of predominately northwest-southeast trend. This horst-and-graben structure is typical of the Basin and Range physiographic province. Near vertical slickensides indicative of high-angle faulting are visible along the wall of the valley (Couch and Johnson, 1968).

The age of initiation of faulting is bracketed between late Pliocene or early Pleistocene, continuing to Recent; displacements on the largest faults in southern Oregon are several thousand feet (Donath, 1962). The local
relief in the Warner Valley is about 2400 feet. Taking 2.5 million years as the age of the faulting and 3000 feet as the displacement, an average displacement of $1.2 \times 10^{-3}$ ft/yr ($3.7 \times 10^{-2}$ cm/yr) is determined. If it is assumed that the uplift of the mountain occurred in steps of 10 to 20 feet, an interval of 34,000 to 17,000 years between episodes of uplift is indicated. Using the return-period equation of Ryall and Douglas (1974) for northwestern Nevada, a return period of 14,000 years is to be expected for a M = 7 earthquake within a radius of 30 km of a given site. Figure 1 shows a fault map of the area, with the epicenters of the larger events from the Preliminary Determination of Epicenters plotted.

The valley floor is generally covered with Pleistocene to Recent alluvium, and in some places along the west wall with landslide debris. These overlie Quaternary lacustrine deposits, which in turn overlie Tertiary volcanic rocks, principally basalt. The cliffs that bound the valley are composed of the same Tertiary basalt that underlies it (Walker and Repenning, 1965).

Seismicity of Oregon

Historically, Oregon has had a low level of seismicity, particularly when compared to its neighboring states. The occurrence of earthquakes within the state, however, indicates that it is tectonically active. The record of
Figure 1. Fault map of the Adel area, from Couch and Johnson (1968). Epicenters are from PDE.
seismicity is mainly noninstrumental, based on reports of felt earthquakes. These reports show the north and northwest portions (particularly the Portland area) to have had the greatest numbers of earthquakes—but to some degree this merely reflects the state's population density.

Berg and Baker (1963) have tabulated felt earthquakes in Oregon from 1841 through 1961, based mainly on published accounts of felt earthquakes. Their list therefore consists mainly of events that had Modified Mercalli intensity of III or greater (see Table 1). Magnitudes are estimated only for the larger events, of magnitude greater than about 4.5. Although earthquakes have occurred offshore with magnitudes as high as 6.8, the largest earthquake with an epicenter on land had an estimated magnitude of about 5.5. Figure 2 is plot of the epicenters listed by Berg and Baker, updated to 1970.

As can be seen in Figure 2, most of the felt earthquakes have occurred in the north part of the state, with a large portion (nearly 1/3) in the Portland region. Berg and Baker state that the distribution of felt earthquakes is statistically related to the population distribution. Since the average felt earthquake in Oregon is small (about intensity IV) many earthquakes in sparsely populated regions go unreported, and felt events are assigned to the location where they were reported. Although instrumental data from 1959-1970 shows the occurrence of earthquakes to be more
Felt, 1841-1870

○ - From Berg and Baker (1963)

Instrumental, 1959-1970

× - UCB
+ - ISC
□ - USCGS

Figure 2. Earthquakes in Oregon, 1841-1970.
widespread and evenly distributed, there is still a preponderance of events in the north part of the state.

Although the overall level of seismic activity is lower in southern Oregon than in northern Oregon, it is also less diffuse—the earthquakes occur closer together in both space and time. This spatial and temporal clustering of the southern Oregon events seems to indicate that swarms of earthquakes (earthquake sequences with no prominent "main event") are presently the predominant mechanism for the release of seismic energy there.

The horst-and-graben structure of south central and southeast Oregon is typical of the Basin and Range province. Tocher (1958) found that few earthquakes of magnitude less than 6 exhibit any surface breakage at all, and that nearly all shallow earthquakes of magnitude equal to or greater than 7 do. The presence of structures indicative of large-scale block faulting, then, implies that the formation of the horst-and-graben structure in southern Oregon was associated with earthquakes of magnitude around 7.

Ryall and Douglas (1974) found that the overall level of seismic activity in northern Nevada and southern Oregon is about one-fifth that in western Nevada. Figure 3 presents the results of their search for epicenters in a region including northern Nevada and southern Oregon, from 1860-1969. Figure 4 shows felt earthquakes of intensity V or greater in the western United States (VI or greater in
Figure 3. Earthquakes in northern Nevada and southern Oregon, 1860-1969. From Ryall and Douglas, 1974.
Figure 4. Map of historic (through 1970) earthquakes in the western U. S. of intensity V or greater (VI or greater in California). From Coffman and von Hake (1973).
California; Coffman and von Hake, 1973). These show the low level of seismicity in the Northern Nevada-northeastern California-southern Oregon region in historic time.

In the vicinity of Adel there have occurred four other earthquake sequences, one near Lakeview, one near Paisley, one in Surprise Valley, and a swarm near Denio, Nevada.

The Paisley sequence began April 18, 1906, with an intensity III earthquake. Four events were felt on the next day, the most strongly felt having an intensity of V. Earthquakes continued to be felt in Paisley until April 29 (Townley and Allen, 1939). There is geothermal activity near Paisley; this sequence may also have been a swarm.

The Lakeview sequence started on January 10, 1923, with an earthquake of intensity VI, which was felt as far away as Susanville, California (110 miles). Aftershocks were felt for several days afterward (Townley and Allen, 1939).

The earthquakes in Surprise Valley consisted of a weak foreshock, and two weeks later (on March 2, 1951) a main shock associated with the explosion of a dormant hot spring (Berg and Baker, 1963).

The Denio, Nevada, swarm began in late February of 1973; a number of events were felt during the week of the sequence. The largest event in the swarm had a magnitude of 5.3 (Richins, 1974); some minor damage was associated
Earthquake Swarms

Mogi (1963, 1967) classifies the occurrence of earthquakes into three idealized types: (1) a large earthquake without foreshocks, but frequently followed by numerous aftershocks—most large tectonic earthquakes are of this type; (2) the type of sequence where foreshocks precede the principal earthquake and numerous aftershocks follow it; and (3) earthquake swarms, in which the number and magnitude of the earthquakes increase gradually with time for some duration, and then decrease after some duration. There is no single predominant earthquake in the swarm. Naturally occurring earthquake sequences are in general gradational between these types.

Laboratory studies of the microfracturing of rock specimens have shown similarities between the behavior of rocks in the laboratory and earthquakes. The three modes of occurrence of earthquakes described above are also found as modes of failure of rock samples. Mogi (1966) finds that the type of failure that will occur is determined by the degree of heterogeneity of the specimen and the uniformity of the applied stress. Type (1) failure occurs in a homogeneous medium with nearly uniform applied stress; type (2) failure occurs when the structure and/or space distribution of the applied stress are not uniform; and type (3) failure occurs in a highly non-uniform medium.
and/or by the application of concentrated stress. These three types of failure correspond to Mogi's three types of earthquake sequence.

In tectonic earthquakes (earthquakes due to the release of regional tectonic strain, as opposed to volcanic earthquakes or earthquakes induced by reservoir loading or nuclear explosions) the stress seems to be applied uniformly. The characteristics of the earthquake sequence are determined by the degree of homogeneity of the rocks at the depth of the earthquakes. At shallow depths, the structural state of the earth's crust may be dependent on the degree of fracturing. Mogi (1966) concludes that main shock-after shock sequences occur in homogeneous regions, that foreshocks occur in moderately fractured regions, and that swarms take place in highly fractured regions (Fig. 5).

In earthquakes associated with volcanic activity, the stresses related to the intrusion of magma into a chamber are likely to be highly concentrated; the structure in such an instance is also likely to be quite non-uniform. Earthquake swarms are frequently associated with areas of current or geologically recent volcanic activity (Richter, 1958).

Scholz (1968a), in his study of the microfracture of rocks, found the homogeneity of the rocks tested to have little effect on their reaction to stress. He found instead that it was the state of stress that influenced
Figure 5. The three types of earthquake sequence and their relation to structure. From Mogi, 1966.
their behavior. Swarmlike activity, with many small events, occurs at low stress; and larger events, with a higher proportion of large events, occurs at higher stresses.

Richins (1974) and others (Sykes, 1970; Ward and Bjornsson, 1971) have noted an association between earthquake swarms and areas of geothermal activity. Two possible mechanisms have been presented: high fluid pressure in the pore space of the rock in the presence of hydrothermal activity, and weakening of the rock due to corrosion by hydrothermal fluids.

That seismic activity is influenced by pore pressure is well-documented; in Denver and Rangely, Colorado, the occurrence of earthquakes has been shown to be related to the pressure of fluids injected through bore holes (Healy et al., 1968, Raleigh et al. 1975). A mechanism proposed by Hubbert and Rubey (1959) provides an explanation: in a rock under stress and containing fluid in its pore spaces, the total stress $S$ is resolved into two components: $S = s + p$, where $s$ is the effective stress on the rock and $p$ is the pore pressure. $p$ is a hydrostatic pressure; faulting or other nonhydrostatic deformation of the rock must be in response only to the effective stress $s$. A high value of pore pressure, $p$, thus has the effect of reducing the effective stress, $s$, required to exceed the strength of the rock and initiate seismic activity.
It is also possible that swarms occur because the
crustal rocks of the region are not sufficiently strong to
withstand large stresses—in which case stress would be
relieved by numerous small earthquakes. Scholz (1968b)
point out that stress corrosion is an important mechanism
in the fracturing of rock, and that the process is
accelerated at high temperature.

Both of the above mechanisms are probably active
in the weakening of rock and the production of earthquakes
in the areas of geothermal activity.
CHAPTER 2

INSTRUMENTAL STUDY OF THE ADEL SWARM

Shortly after the beginning of the Adel swarm on May 26, 1968, the Seismological Laboratory of the University of Nevada at Reno (UNR) installed instruments in the Adel area to monitor the activity. This chapter discusses the instrumentation, the data acquired, and the results of the data analysis. A first motion plot was made to determine the mechanism of faulting, hypocenters were determined, magnitudes were measured and recurrence curves drawn, and the rate of die-off of activity was analyzed.

UNR Study of the Swarm

On June 5, 1968, a field crew from the UNR Seismological Laboratory was in Adel. The center of activity was located by setting up temporary stations at various sites in the Adel area and monitoring activity. Distances from these sites to the active region were approximated using S-P times, and instruments were then installed to monitor the activity.

Instrumentation. The instrumentation consisted of: (1) a tripartite array of vertical seismometers, wired to a trailer where their signals were recorded on magnetic tape, along with WWVB time code and the signals of a three-component seismometer station located near the trailer; (2) a pair of portable Wood-Anderson seismographs;
(3) two seismic event recorders (SER's); and (4) a seismometer recording on a helicorder, first at the Adel school and later at the church near Adel.

The system recorded at the trailer consisted of four Electrotech EV-17 vertical and two EV-17H horizontal seismometers, wire-lined to the trailer. One vertical and the two horizontal-component seismometers were operated at the same position near the trailer to form a single three-component station. The trailer contained a tape recorder, electronics to process the signals, and a WWVB time code receiver. The seven signals (six seismic and one time) were recorded on the slow-speed tape recorder, as described by Ryall et al. (1968). The geometry of the tripartite array is shown in Figure 6. This system was operated continuously from June 6 to July 25, 1968. No useful data was acquired from June 23 to 25 due to broken wires; there were several other periods of 10-15 hours during which one or another of the traces was dead, so that no earthquakes could be located.

The portable Wood-Anderson station consisted of two horizontal torsion seismometers, recording on 35mm photographic film. These records are analyzed on a Surveyor's Service Company film viewer, which then gives standard Wood-Anderson Magnification and response. The Wood-Anderson station was operated June 25, 26 and 27.

The SER's each consist of a seismometer and WWVB receiver, signal-processing electronics, a tape recorder, and
Figure 6. Geometry of the tripartite array.
an endless tape loop. The seismic and time signals are recorded on the tape loop, and the signal is continuously scanned. When the amplitude of the amplified incoming seismic signal exceeds a preset threshold level, transfer of the delayed signal from the tape loop to the library tape is initiated. After the tape loop passes the reproduce head, the incoming signal at the record head erases the previous data and records current data. The amount of delay introduced into the signal is typically 70 seconds, sufficient time to ensure recording the relatively small amplitudes at the beginning of the seismic signal. Minimum transfer time is 60 seconds to guarantee at least one full frame of WWVB time code. Thus, the library tape contains only the seismic information and time code for successive events (Hunt, 1969). SER's were operated at sites A and B (see Figure 7) on June 14 and 15, and 27 kilometers north of Adel at Plush, Oregon, on June 25 (when the tripartite array was down).

The tapes from the tripartite array were played out continuously on a Siemens chart recorder for the period from June 18 to July 24. The playout is low speed (7.2 cm/min), and therefore not useful for timing events to be located. It is of sufficient quality that the earthquakes could be counted and their durations accurately measured for subsequent comparison to magnitude. 169 events were selected for location; these were played out at high speed (75 cm/min)
Figure 7. Locations of the tripartite array, Wood-Anderson seismographs, and SER's.
on a Honeywell visicorder. The precision of timing on the high-speed playouts is ± 0.01 second; due to reading errors the accuracy of the arrival times is ± 0.03 sec.

The SER tapes were played out at high speed (210 cm/min) on a Brush recorder; the timing precision on these is also ± 0.01 sec. The tripartite array was down during most of the time that the SER's were operated; there were eight events recorded on both the tripartite array and the SER's at sites A and B.

Problems with the Data. There were several unusual problems associated with the data for this sequence, all associated with the fact that the data had been stored for nearly eight years and had been partially analyzed by each of several investigators. Just finding much of it was a major problem; for instance, the slow-speed playout of the first nine days of the sequence was never found. Three different sets of co-ordinates for the stations of the tripartite array had been determined by three different investigators; the original field maps were eventually found and the correct co-ordinates taken from them. The Wood-Anderson films were poorly labelled; the date was marked but not the start or stop times. Timing for the films had to be established by comparison of the time of large events to the times of large events on the slow-speed playouts from the trailer system. Several of the SER playouts were identified by time and recording site, but
not by date. Dates were determined using the same method as for the Wood-Anderson films. Finally, there existed several sets of event locations, determined using different velocity models and different location techniques. These were ultimately ignored.

Focal Mechanism

In order to determine the mechanism of faulting, the direction of first motion from some of the larger shocks was determined at a number of stations. Of the large shocks in the sequence, only one (May 30, 1968, 0036; M = 5.1) registered clearly at a sufficient number of stations to give a well-determined focal mechanism. The station distribution near Adel is sparse; only one station, KFO, was close enough to record Pg as a first arrival. All other first motions were either Pn or teleseismic P; most of the Pn first motions were read from the original seismograms, and the P readings were taken from the Bulletin of the International Seismological Center. Table 2 lists the stations used in determining the focal mechanism and their first motions. The other large events had fewer clearly defined first motions, but those that could be clearly read are consistent with the first motions presented for this shock.

The first motions were plotted on a lower-hemisphere equal area projection, and fault and auxiliary planes fitted
<table>
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<th>STATION</th>
<th>CODE</th>
<th>DISTANCE, DEG.</th>
<th>AZIMUTH</th>
<th>FIRST MOTION</th>
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<td>1.39</td>
<td>271.7</td>
<td>D</td>
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<tr>
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<td>UNN</td>
<td>2.21</td>
<td>143.6</td>
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<td>215.2</td>
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to them as shown in Figure 8. The north-striking plane was assumed to be the fault plane due to its alignment with the tectonic features of the Warner Valley. This plane is nearly vertical (dip 80° E) and strikes N4° E. The indicated motion on this plane is oblique slip, with about equal components of left-lateral strike-slip and reverse fault motion. The axis of maximum compression (P) has direction S60° E and plunges 13°; the axis of relative tension (T) has direction S65° and plunges 35°. These axes are similar in direction but opposite in sense to those found for the Denio, Nevada, swarm by Richins (1974) and for the rest of the Basin and Range Province. There the tension axis is nearly east-west and nearly horizontal (plunging slightly east) and the compression axis has a steeper plunge to the southwest.

The orientation of the fault plane from the focal mechanism is in accord with the tectonic features of the area; the mechanism of faulting is not. The usual mode of faulting in the Basin and Range is oblique-slip, with the vertical component of motion being normal faulting—-which implies extensional tectonics. This focal mechanism indicates oblique slip, but with reverse faulting as the vertical component. This indicates that the valley block is rising with respect to the mountain block, under the influence of compressional tectonics. That different stresses act here than in the rest of the Basin and Range...
Figure 8. Lower hemisphere plot of first motions for the earthquake of May 30, 0036. Focal mechanism and stress axes shown.
Province indicates that the area is no longer under the same tectonic influence that produced the area's horst-and-graben structure.

Although the compressional tectonics indicated by the focal mechanism are anomalous with respect to the Basin and Range morphology of the area, some recent investigation of plate tectonic processes indicate that a zone of compression should exist to the north or east of the intersection of the San Andreas and Mendocino faults. McKenzie and Parker (1967) and Atwater (1970) found that the right-lateral motion on both faults requires that crust be consumed (at a subduction zone, with its associated compressional tectonics) somewhere to the north of the Mendocino-San Andreas junction. Atwater considers two models of plate motion and finds that the direction of the compression should be either north-northeast or east-north-east. The direction of compression indicated by the focal mechanism is east-southeast.

**Rate of Occurrence**

To examine the temporal behavior of the Adel sequence, the numbers of earthquakes in twelve-hour intervals were counted on the slow-speed playout, for the entire period from June 18 to July 24. These are plotted in Figure 9.

The figure shows a fairly uniform decrease of
Figure 9. Temporal behavior of the Adel sequence and hyperbolic decay curve, $N(t) = 40,738t^{-1.39}$
activity with time, upon which are superimposed two peaks of high activity on June 22 and 30. The rate of activity at the beginning is probably the end of another peak of activity, or it may be representative of the actual activity level before the beginning of the data sample.

In order to compare the rate of decay of activity of the Adel swarm to other sequences, a curve of the form \(kt^{-b}\) representing hyperbolic decay was fit by least squares for \(k\) and \(b\) to the twelve-hour counts. The equation of this curve is

\[
N(t) = 4.07 \times 10^4 \times t^{-1.38}
\]

where \(N(t)\) is the number of events per hour and \(t\) is the number of hours after the start of the sequence (taken as 0000 on June 3 for the purposes of this calculation).

Figure 9 shows that equation (1) fits the counts fairly well for the period covered by the data. If equation (1) is extrapolated back to the beginning of the sequence, however, it gives an unreasonably high rate of activity (more than 40,000 events in the first hour). The coefficient of \(4.07 \times 10^4\) in equation (1) is thus larger than is reasonable, and most likely an artifact of the time period covered by the data, or the fact that this was a swarm, and therefore may have a different mode of decay than aftershock sequences. The value of \(k\) for the Boxcar aftershocks was 372 (Ryall and Savage, 1969), and was 61 for the aftershocks of the 1966 Truckee, California
earthquake (Ryall et al., 1968).

The computed exponent of $t$ ($-1.38$) for this swarm indicates a rate of decay of activity very similar to the decay rate found for the Denio swarm by Richins (1974). Both are indicative of a faster die-off of activity than is generally found for aftershock sequences. A value of $-0.64$ was found for aftershocks of the Truckee, California, earthquake; the 1966 Parkfield-Cholame earthquake had a decay constant of $-0.96$ (Ryall and Savage, 1969). Mogi (1962) found values of $-0.96$ to $-1.36$ for Japanese earthquake sequences.

If the data extended back in time to the beginning of the sequence, it is possible that the high levels of activity shown in Figure 9 would be seen to be peaks of activity superimposed on a flatter (i.e., more slowly decaying, with a smaller exponent of $t$) curve than represented by equation (1), or that the activity exhibited some mode of decay other than hyperbolic (e.g., similar to Mogi's type 3 sequence shown in Figure 5, or that found by Lee et al. (1971) for the Danville swarm, shown in Figure 10).

Locations

To locate the 169 events selected for detailed analysis, the high speed playout from the tripartite array and SER's were timed to a precision of 0.01 second.
Figure 10. Temporal behavior of the Danville, California swarm, 1970. From Lee et al., 1976.
Station Corrections. Station corrections for the stations in the tripartite array were determined by the following method: the eight events recorded in common by the array and the SER's were located using only the SER's at sites A and B and the center station of the array, ADC. These locations were presumed to be more accurate than locations obtained from the tripartite array alone, since the SER-A, SER-B, ADC array is larger than the tripartite array by an order of magnitude. From these locations, the arrival times at stations ADN, ADS, and ADW were computed and compared to the observed arrival times at those stations to obtain average arrival-time corrections for those stations. The corrections computed for ADN, ADS, and ADW are 0.017, 0.004, and 0.078 seconds respectively. After incorporation of the corrections into the location routine, locations of the eight events using the tripartite array alone corresponded quite closely to their locations determined by using the larger array.

Velocity Model. The velocity model used in locating the earthquakes was determined by W. Savage for the Adel area during his investigation of the sequence; it consists of an 0.5 km thick layer with P-velocity of 3.3 km/sec, overlying a 1.5 km thick layer with P-velocity of 4.7 km/sec, on top of a half-space with P-velocity of 5.9 km/sec. The focal depths and distances to the array made it unnecessary to model the crust-mantle interface.
The vp/vs ratio used is 1.73, corresponding to a Poisson's ratio of 0.25, and giving S-velocities of 1.91, 2.71, and 3.41 km/sec.

**Location Routine.** The computer program used to locate the events with the tripartite array was written by W. Savage and incorporates a subroutine for determining travel times from HYPOLAYR, written by J. Eaton of the U. S. Geological Survey. This program LOC, calculates origin times from S-P readings at all stations for which there are both S and P data. The origin times thus determined are averaged to give the origin time used by the program for locating the earthquake. Using the station with the shortest travel time for P (after applying station corrections), the program makes a first estimate of the hypocenter. Arrival times at each station from the trial hypocenter are calculated through an earth model of horizontal layers of constant velocity. The residuals at each station and partial derivatives of travel time with respect to the coordinates of the hypocenter are calculated. The trial hypocenter is then adjusted to minimize the residuals, until an iteration limit is reached or until criteria of sufficiently small residuals or correction to the hypocenter are reached as described by Richins (1974).

**Converted Phase.** The program's feature of determining an origin time from each set of S and P readings
turned out to be very useful in identifying the direct S-phase. When attempting the locations that were to lead to the station corrections, there was found to be a disparity between the origin times determined from S-P times at sites A and B (which were consistent with each other) and origin times determined from S-P time at ADC. A re-examination of the records from the tripartite array showed a large-amplitude phase arriving later than the one originally picked as S. Considering this later phase to be S gave origin times consistent with the SER's; the S-times on all the records were then re-picked. Figure 11 shows a typical record from the tripartite array, with P and S identified. Also identified is the intermediate phase, which is probably a conversion of S to P at a layer interface, or maybe an S-wave critically refracted on some layer interface below the focus.

The records from Adel are different in appearance from the records of other sequences made on this same system. Records of the Adel events have a generally lower-frequency appearance and a longer and less rapidly decaying coda (see Figure 12). These differences are at least partly due to recording-site effects. The seismic detectors at Adel were in an alluvial (and probably multi-layered) valley, whereas the other records came from detectors placed on or near rock outcrops.

Hypocenters. A map of the epicenters calculated by
Figure 11. Typical Adel seismogram, showing P, S, and converted phase X.

Figure 12. Seismograms from other sequences, recorded on the same system. From Ryall and Savage, 1969.
the computer routine is presented in Figure 13. The zone containing most of the activity is approximately rectangular, about 15 kilometers long (north-south) and 6 kilometers wide. The entire zone of activity is somewhat greater in area than the above mentioned rectangle; witness the few events to the south and to the east of the array. Although the 169 located events constitute only a small fraction of the total number of events in the sequence, they were chosen arbitrarily and are probably representative of the range of activity. No earthquake was located under the center of the valley (to the east of the area shown).

There was no evidence of migration of activity during the sequence, although the average depth of earthquakes late in the sequence was somewhat greater than during the first part of the sequence.

Figure 14 shows cross-sections along lines A-A' and B-B' of Figure 13, showing the depth distribution of the events located. The dashed line on the figure has the dip of a nodal plane of the focal mechanism in Figure 8. Figure 15 presents a histogram of the depths of the Adel earthquakes, with similar histograms of other sequences for comparison. The focal depths of the earthquakes in the Adel swarm are seen to be similar to those of other tectonic earthquake sequences, and deeper than those of sequences induced by reservoir loading or nuclear explosions.
Figure 13. Map of epicenters of the Adel sequence.
Figure 14a. Projection of hypocenters of Adel earthquakes onto a north-south plane (A-A' in Figure 13). No vertical exaggeration.
Figure 14b. Projection of hypocenters of Adel earthquakes onto east-west plane B-B' in Figure 13). No vertical exaggeration. Dashed line corresponds to the N-S plane of the focal mechanism (Figure 8).
Figure 15. Histogram showing percentages of Adel earthquakes vs. depth.
It should be noted that the hypocenters (particularly the depths) of the earthquakes are dependent on the velocity model used in the analysis; the accuracy of the velocity model determines the accuracy of the hypocenters. Changing the velocity model affects all events similarly; i.e., changing the velocity model has the effect of changing the position of the entire sequence. The absolute locations of the earthquakes, therefore, may only be accurate to about ±1 kilometer, while the relative locations of the epicenters are consistent to a few hundreds of meters. The relative depths are not as well constrained; the accuracy of the depth determination depends more strongly on the accuracy of the velocity model and S-P interval.

Figures 13 and 14 suggest that the sequence is associated with the fault whose scarp forms the western boundary of the Warner Valley but do not conclusively place the hypocenters on that fault. The fault plane indicated by the focal mechanism is consistent in orientation with this fault, but the direction of motion (oblique slip with a component of reverse faulting) inferred from the focal mechanism is inconsistent with the normal fault motion indicated by the slickensides on the scarp. Other investigators of swarm activity have also found a volume distribution of seismic activity for swarms (Lee et al. (1971), for the Danville, California, swarm, The Party for Seismographic Observation of Matsushiro Earthquakes (1967),
for the Matsushiro, Japan, swarm, and Richins (1974) for the Denio, Nevada, swarm), as opposed to the generally planar distribution of seismic activity in aftershock sequences.

Magnitudes, b-value

Seismograms written by the Wood-Andersons (the standard instrument for determining magnitude) provided an excellent opportunity to check some of the methods that have previously been used to determine magnitude and draw recurrence curves, and to calibrate the field system for determining magnitudes.

Wood-Anderson Magnitudes. The magnitudes of all events recorded on the Wood-Anderson seismographs were determined. The Nordquist nomogram could not be used for these determinations; it assumes a constant focal depth of 16 km (A. Ryall, personal communication) and many of the earthquakes had focal distances of less than 16 km from the station. Focal distance rather than epicentral distance was used for the calculation of magnitude and was determined from S-P time. For an earthquake of given magnitude M, the trace amplitude in millimeters, A, on a Wood-Anderson seismograph is given by

\[
\log A = \log A' - d \log D
\]

(2)

where A is the amplitude an event of magnitude M writes at 100 km and D is the epicentral distance (A. Ryall,
personal communication). To determine $d$, the slope of the line representing the attenuation of trace amplitude with distance, an earthquake of $M = 2$ was assumed and amplitudes taken from the nomogram for distances of 150, 100, 60, and 30 km. For a focal depth of 16 km, these epicentral distances are approximately the same as the corresponding focal distances. A log-log plot of the distances versus the amplitudes was made; a straight line resulted with a slope of $-1.66$. Using $A'$ values for different magnitudes, inserting these values into equation (2) and solving the resulting two equations simultaneously, the following relation was found:

$$M = \log A + 1.66 \log r - 0.32 \quad (3)$$

where $M$ is magnitude, $A$ is maximum trace amplitude in millimeters, and $r$ is focal distance in kilometers. Equation (3) can be applied to focal distances less than 16 km; it was used to compute magnitudes of the earthquakes recorded by the Wood-Andersons.

Magnitude-Duration Relationship. An attempt was made to determine magnitudes from the duration (coda length, or F-P time) of the events. A plot was made of magnitude vs. the logarithm of the duration (LD) as shown in Figure 16; although the data are somewhat scattered a least-squares line (of the form $M = c + a \times LD$) was fitted to the points. Magnitudes for all earthquakes in the sequence were
Figure 16. Plot of Wood-Anderson magnitude vs. duration, showing first- and third-order curves fitted.
determined with this equation, and a magnitude-frequency plot was made. A magnitude-frequency plot (recurrence curve) has the form

\[ \log N = a - bM \]  

(4)

where \( M \) is magnitude, \( N \) is the number of earthquakes equal to or greater than \( M \), and \( a \) and \( b \) are constants. The recurrence curve determined by using the straight line magnitude-duration relation was curved (Fig. 17a), so that no value of \( b \) could be determined. Savage (1976), in his discussion of recurrence curves, notes that curved frequency-magnitude plots often result from determining magnitude as a first power function of \( LD \).

Since the magnitude-duration data in Figure 16 appear to indicate a curved relationship, higher-order polynomials in \( LD \) were fitted to the data by a least squares procedure. A third-degree polynomial was visually selected as giving the best fit to the data in the range of durations from 30 to 125 mm (all durations were measured in millimeters; the conversion factor is 1.2 mm/sec).

**b-value.** A recurrence curve using the magnitudes from the Wood-Andersons gave a value for \( b \) of 0.97; however, the sample size (49 events) is too small to permit an accurate determination of \( b \). Savage (1976) found that at least 100\( b \) events are required to accurately
Another recurrence curve was plotted, this one using magnitudes determined by the third-order polynomial LD applied to all durations above 30 ms (Figure 17b). The curve is straight between magnitudes 1.9 and 3.6 (where the polynomial is well-determined), and gives a 'b' value of 1.11. This value is considered statistically significant, if based on 650 events.

Ryall and Savage (1970) calculated a 'b'-value for Amel of 0.70, determined by plotting log N vs. the log of the time amplitude in millimeters, from the records from the seismographic array. Their argument that log A is proportional to magnitude because the flat portion of the shear strain's response is the same as the flat portion of Anderson's response is thus shown to be incorrect.

With a well-determined value of b for the magnitude recurrence curve, to 3.0, Ryall suggested the following method for updating the magnitude-duration relationship by low (and possibly higher) magnitudes than those for which it was already established. Earthquake sequences shown linear recurrence curves (Savage, 1970); this being so the magnitudes of events outside the region for which magnitude-duration holds can be assigned according to the next lower or higher line if the two lines are for greater than LD. This leads to a rearrangement of

\[ \text{Cumulative number of events} \]

\[ \text{Magnitude} \]

**Figure 17a.** Recurrence curve found by using first-order magnitude-LD relationship.
evaluate b.

Another recurrence curve was plotted, this one using magnitudes determined by the third-order polynomial in LD applied to all durations above 30 mm (Figure 17b). The curve is straight between magnitudes 1.9 and 3.0 (where the polynomial is well-determined), and gives b-value of 1.13. This value is considered statistically significant, as it is based on 650 events.

Ryall and Savage (1969) calculated a b-value for Adel of 0.70, determined by plotting log N vs. the log of the trace amplitude in millimeters, from the records from the tripartite array. Their argument that log A is proportional to magnitude because the flat portion of the field system's response is the same as the flat portion of a Wood-Anderson's response is thus shown to be incorrect.

With a well-determined value of b for the magnitude range 1.9 to 3.0, A. Ryall suggested the following method for extending the magnitude-duration relationship to lower (and possibly higher) magnitudes than those for which it was already established. Earthquake sequences exhibit a linear recurrence curve (Savage, 1976); this being so the magnitudes of events outside the region for which magnitude-duration holds can be assigned according to the number of events N(LD) with log duration equal to or greater than LD. This amount to a rearrangement of equation (4), to read
The cumulative numbers of events with log duration equal to or greater than a cut-off value of log duration were determined and assigned magnitudes by equation (5). These magnitudes were then plotted against the logarithm of duration as shown in Figure 18. Polynomials in LD were least-squares fit to these points and visually compared to the data for best fit. The polynomials were fitted only to the points with durations between 10 and 125 sec, to stay away from the range of statistical fluctuations due to the small numbers of large events on the high end and the incomplete detection on the low end.

Polynomials of orders 2 through 5 were tried in the range of durations from 10 to 125 sec and behaved essentially identical to each other and to the order polynomial used to fit the data originally. The fourth-order polynomial

$$\log(N/LD) = 1.13 - 3.86 \times 10^{-1} \times LD + 7.43 \times 10^{-2} \times LD^2 - 6.99 \times 10^{-3} \times LD^3 + 3.35 \times 10^{-4} \times LD^4$$

was chosen to calculate magnitudes because of its behavior above and below the range to which it was fitted. It seems reasonable to represent the entire sample.

Equation (5) fits the observations well for the entire sequence of magnitudes between 1.5 and 5.0.

**Figure 17b.** Recurrence curve found by using third-order relationship between magnitude and LD. b-value of 1.13 assigned to the portion of the curve between magnitudes 1.9 and 3.0.
The cumulative numbers of events with log duration equal to or greater than a cut-off value of log duration were determined and assigned magnitudes by equation (5); these magnitudes were then plotted against the logarithm of duration as shown in Figure 18. Polynomials in LD were least-squares fit to these points and visually compared to the data for best fit. The polynomials were fitted only to the points with durations between 10 and 125 mm, to stay away from the range of statistical fluctuations due to the small numbers of large events on the high end and the range of incomplete detection on the low end.

Polynomials of orders 2 through 5 were tried; in the range of durations from 30 to 125 mm all had behavior essentially identical to each other and to third-order polynomials used to fit the data originally. The fourth-order polynomial

\[ M = -0.60(LD)^4 + 3.68(LD)^3 - 7.33(LD)^2 + 6.99(LD) - 1.36 \]  

was chosen to calculate magnitudes because of its behavior above and below the area to which it was fitted; it came closest to representing the entire sample.

Equation (6) fits the observations well for the Adel sequence for magnitudes between 1.25 and 3.5. It remains to be seen if equation (6) can be considered as
Figure 18. Plot of magnitudes (from equation 5) vs. duration for all earthquakes recorded, showing fourth-order curve fitted.
calibrating the system for determining magnitudes of other sequences. The effects of different ranges of distances have yet to be explored, as do the effects of different recording sites (the Adel sites were on alluvium and the records from Adel had a different appearance than most others recorded on this system).

Figure 19 shows the recurrence plotted using all events in the sequence, with their magnitudes calculated by equation (6). The value of b on this plot is also 1.13--it was constrained to that value by the method used to calculate magnitudes.

Richins (1974) found that the recurrence curve for the Denio, Nevada, swarm exhibited an offset at about magnitude 3.5; i.e., the curve continues with the same slope for magnitudes greater than 3.5, but is offset upwards. Figure 17b shows a similar offset at magnitude about 3.0. This offset is not present in Figure 19--the curve-fitting technique constrained all points to lie on the same line.

Temporal changes in b. To determine whether the value of b changed during the sequence, a moving-window calculation of b was made using the method of Utsu (1965). This method gives results closely comparable to a weighted least-squares calculation of b (W. Peppin, personal communication). The window width used in the calculation was 300 events, with a 100-event overlap. The curve was
Figure 19. Recurrence curve for entire sample, magnitudes calculated by eq. 6.
calculated twice, once using magnitudes between 1.8 and 3.2 and again using magnitudes between 1.3 and 2.5 to determine whether the presence of the larger events in the sample was affecting the shape of the curve. The two curves are identical; Figure 20 presents the variation of $b$ with time.
Figure 20. Plot showing change of b with time.

Number of days after start of recording, June 18.
CHAPTER 3

SUMMARY AND DISCUSSION

Oregon has historically exhibited a lower level of seismicity than its neighboring states, particularly Nevada and California. Most of the seismic activity has been in the north part of the state; the largest earthquake to occur in Oregon had magnitude estimated at 5.5. Earthquake swarms, instead of main shock-aftershock sequences, seem to be the predominant mechanism for the release of seismic energy in southern Oregon, particularly in conjunction with geothermal activity.

An earthquake swarm began near Adel, in southern Oregon, on May 26, 1968. This swarm has here been studied in some detail, in an attempt to somewhat illuminate the subject of the southern Oregon seismicity and the occurrence of swarms there.

The epicenters of the events located in this study lie mainly in a north-south trending zone, about 15 km, long by 6 km wide. Focal depths of the events average about 9 km, and are typical of tectonic (rather than induced) seismic activity. These earthquakes seem to be associated with the fault whose scarp forms the western wall of Warner Valley; however, this study cannot conclusively place them on that fault. The distribution of hypocenters may reflect a volume, rather than planar, focal zone.
A first-motion study produced an oblique-slip focal mechanism, composed of approximately equal amounts of left-lateral strike-slip and reverse fault motion on a north-south trending fault. Although the major tectonic features of the area trend north-south, the slickensides and the regional morphology suggest that previous seismicity has been of a normal-faulting nature.

It is this change, in Recent time, from an extensional to a compressive stress system that constitutes the most important result of this study. The 1973 swarm near Denio, Nevada (about 120 km to the southeast of Adel) had an extensional focal mechanism (Richins, 1974); therefore there is a change in the tectonic stress field somewhere between Denio and Adel. There is Recent evidence of extensional tectonism near Adel (Donath, 1962); therefore the change from extensional to compressive stress occurred in the geologically recent past.

Several investigators of plate tectonics have determined that compressive stresses should be seen in Oregon. From a consideration of relative plate velocities, McKenzie and Parker (1967) find that right-lateral strike-slip faulting on both the San Andreas and Mendocino faults requires either creation of crust to the southwest of their junction (which is unlikely) or consumption of crust (i.e., subduction) to the north or east of the junction. Atwater (1970) also finds that subduction (probably beneath the
Cascade mountains) is required for consistency of the plate motion. She considers two models of plate motion, and finds that there should be compression behind the subduction zone in either a north-northeast direction or in a direction slightly north of due east. Although the direction of compression indicated by this focal mechanism is east-south-east, some slightly modified scheme of plate motions may adequately explain the stress system at Adel.

The temporal behavior of the swarm was analyzed: although the decay curve fit the data fairly well, the period studied may not have been representative of the entire swarm. A high initial rate of activity and rapid die-off is indicated.

The presence of Wood-Anderson Seismographs made it possible to acquire a relationship between magnitude and coda length, at least for magnitudes between 1.25 and 3.5. Due to the unusual character of the records from this sequence, it remains to be seen whether this relationship will be applicable to other sequences recorded by the same system.

The value of b for this sequence was determined to be 1.13. Although the validity of comparisons is dependent on the accuracy of the method used to determine b, this value indicates a smaller proportion of large earthquakes than for the western Basin and Range as a whole (for which \( b = 0.81 \), Ryall and Douglas, 1974). Relatively high
b-values are frequently associated with swarm activity (Mogi, 1963, and Ward and Bjornsson, 1971).

Laboratory experiments on the microfracturing of rock have also shown high b-values to be associated with swarmlike activity. Mogi (1963) found that swarm activity (and high b-values) occur in samples of heterogeneous composition. He states that the fractures in such a sample can relieve only a small amount of strain. Hence there will be many small fractures, giving high b-values.

Scholz (1968a) proposed an alternate explanation of the association of high b-values with swarm activity. He found b to depend strongly on the state of stress in a specimen, and that high b-values are associated with low to moderate stress levels.

If the rocks underlying the Adel region are presumed to be highly fractured (perhaps by infiltration and corrosion by the waters of the geothermal field), the region would be expected to respond heterogeneously to stress and, by the Mogi argument, to produce swarm activity with a high b-value. Seismic activity would be initiated at a lower level of stress in a highly-fractured region than in one of more competent rocks, and the Scholz mechanism would also call for swarm activity and its attendant high b-value. The additional assumption of high local pore pressure due to the active geothermal field increases the effective stress, and further reduces the
tectonic stress required to initiate seismic activity. By considering the effect of heterogeneity on the ability to withstand rupture, the Mogi and Scholz findings become identical, at least in their prediction of swarms in a highly fractured region.

Comparison of Figure 20 to Figure 9 (temporal behavior) shows that the large dips in the curve occur at the same times as the peaks in activity. The first dip in the curve contains the two largest events recorded. The shape of the curve is unaffected by deletion of the large events. The value of $b$ is to some extent a measure of average magnitude (Aki, 1965); the larger is the average magnitude, the smaller is $b$. The large events were not included in the sample; possibly the masking of small events by larger ones partially accounts for the dips in the curve.

It can be seen that the changes in $b$ are not premonitory to, but are rather contemporaneous with the occurrence of the large events.

A not unreasonable expectation is that $b$ should increase with time during the sequence, as the stress level and hence average magnitude of the events become smaller. The aftershocks of the 1975 Oroville, California, exhibited a slight decrease in average magnitude with time (Ryall and Van Wormer, 1975). No such decrease in average magnitude (increase in $b$) can be seen during the time period covered by this data.
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