Annual Progress Report

on

MICRO-EARTHQUAKE MONITORING OF THE HANFORD REGION

Geophysics Program
University of Washington
Seattle, Washington 98195

March 30, 1976

TABLE OF CONTENTS

LIST OF ILLUSTRATIONS

1. INTRODUCTION

In July, 1975 the University of Washington assumed responsibility for the seismic network in eastern Washington. This network consists of 29 vertical seismograph stations plus three horizontal components. The entire network is broken down into two sub-nets. The Hanford Net consists of 16 stations, located on or near the Hanford Reservation and south of Smyrna, and enables the procurement of epicenter locations of all earthquakes of magnitude 1.0 and greater within the Hanford Reservation. North of this network is the Central Net, which consists of 13 stations, two of which (Vantage and Warden) are closely spaced at the north edge of the Hanford Net. The eleven remaining stations are spaced farther apart, to the north, in order to obtain a regional seismic picture of the entire eastern Washington area (Figure 1).

2. DATA FROM QUARTERLY TECHNICAL REPORTS

Since July,1975 the University of Washington's Geophysics Program has submitted two Quarterly Technical Reports. These reports list all earthquakes that have instrumentally-determined epicenters, including date, shock, time, location, depth, magnitude, and all stations that were used in the epicentral determinations and their arrival times. A third Quarterly Technical Report will be submitted on or before May 1, 1976, and will contain events located in the first three months of 1976.

We have received all films and related information from the U.S. Geological Survey for the first six months of 1975. This data is being scanned, as time permits, and will be submitted at a later date.

3. DATA FROM SPECIFIC STUDIES

We have used a modified velocity model to locate the last half of the 1975 epicenters (Table I). Presently we need more data, especially for deep events greater than 10 km, to improve this model. A detailed report on signal recovery, epicenter determination, and magnitude relationships is included in this report (Section A).

Section A: Data Processing

The thirty-two channels of eastern Washington seismic data are telemetered via four leased telephone lines to the University of Washington Geophysics recording facility. The FM subcarriers are discriminated and demodulated before being recorded on two Geotech develocorders (16 channels each) at a film speed of approximately one inch per second. One reel of film contains data for two days for half the array. One channel from a central station is recorded on a helicorder to allow lab personnel to quickly note large earthquakes without scanning the films. A subcarrier spectrum analyser is used daily to check the incoming data for inoperative stations or communication lines. When more than one station is noted to be down, personnel are sent to the field to make repairs as quickly as possible.

Once or twice a week the films are scanned by a record technician on a film viewer to detect earthquakes. Two logs are kept by this technician, one of all locatable events recorded by more than three stations and a second for smaller events showing on only a few stations. This latter log is called the swarm log since it is made up predominantly of small events showing on only one station and being associated with a known swarm area such as Wooded Island.

Earthquakes which seem to occur within or near the eastern Washington array and which are well recorded on more than three stations are selected for arrival time picking. An enlarged photographic copy of the appropriate stretch of film is made on an Itek viewer-printer. A second is approximately 1.0 cm long on this enlarged record. The record is then "picked" or timed using an acoustical digitizer in conjunction with the Geophysics Program's Varian 620/F computer.

The semi-automated picking process consists of several parts. The process involves an operator interacting with the computer through a teletype, sonic digitizer, and CRT display screen. A typical record is read in the following sequence: (1) The record is aligned on the digitizer pallet and points are given to indicate the time scale and to check the precision of the alignment. (2) The date and time of the event (event identifier) are typed in on the teletype. (3) Individual stations are picked by first indicating the station and phase to be read using a special "menu" template at the bottom of the digitizer pallet, then digitizing one point on the timing line, then several points for the particular arrival indicating arrival time, estimated error, sense of first motion, and maximum amplitude. P and S phases and coda length can be read for up to 50 stations in this way. (4) After all arrivals have been read, the operator tells the computer to try to locate the event. As the interactive program is executing the intermediate results are displayed on the CRT as a map with a changing epicenter, depth, and error estimate. If a solution is found satisfying certain error criteria, the routine is terminated, or the operator can intervene if he feels a low residual solution is not possible. (5) Residuals for each picked arrival are displayed, allowing the operator to easily detect obvious reading errors. Errors can then be corrected by returning to the picking part of the program, and then the event is relocated. (6) When the readings and solution are acceptable to the operator, he can then have all event information listed on the printer and all timing and location information saved on a magnetic tape file. The operator can then proceed with additional earthquakes.

After a number of earthquakes have been picked and preliminary epicenters determined, the output tape is taken to the CDC 6400 computer where cards are punched containing all the read data in a format which is standardized for all our location routines. For the final locations of earthquakes in eastern Washington, we are using a modification of the U.S. Geological Survey program "HYPO71" (Lee and Lahr, 1971; see appendix for write-up on using the University of Washington version, HYPO71-1). This is essentially the same program that has been used by the U.S. Geological Survey for locating eastern Washington earthquakes during the past several years. Minor differences exist in the velocity model we are using and the parameters for magnitude determinations. Both of these aspects of the location routine are presently under development. Catalogs of locations using HYPO71 are produced quarterly and are included in reports to ERDA as well as being included in the state of Washington earthquake catalogs.

Comparisons of locations and magnitudes of earthquakes located by the U.S. Geological Survey (1969-1974) and those by the University of Washington (1975-) should be made with care. We feel that systematic bias in epicenters between the two catalogs should be small (less than 1-2 km). Systematic differences in depth may be slightly larger though probably not greater than 3 km. We feel that systematic differences in magnitude determinations are large--perhaps as much as one magnitude unit. Since the coda length method of determining magnitude (used in both cases) is largely an empirical one and both sets of empirical constants were determined in areas geologically different from eastern Washington, and since the definition of coda length itself is somewhat arbitrary, we feel that neither catalog may contain absolute Richter local magnitudes. To this end we plan additional work to specifically determine the appropriate coda length magnitude empirical constants for the Columbia basin region and to do a simple attenuation study of seismic wave propagation in the area. These studies will be used to modify the magnitude determinations from both catalogs and to give insight into the basic tectonic setting of the region. The studies will include among other things comparing amplitude versus coda length of selected events recorded on calibrated instruments and the installation of a standard Wood-Anderson seismograph in the area.

Section B: Interpretation of Wahluke Seismic Data from Fall 1974 Array

The micro-earthquake seismic data for our array on the Wahluke Slope during the fall of 1974 has been analyzed and a preliminary structural interpretation made. The seismic data was collected in conjunction with the resistivity field test described in Section C, and since there were no obvious variations in resistivity associated with micro-earthquakes, only preliminary hypocenter locations were performed. The array was operated from

October 16 to November 28, during which time 36 locatable events were recorded. The hypocenters are shown plotted in Figure 3. The origin of the coordinate system is one mile east of the Grant-Adams County line on Washington State Highway 24. Hypocenter solutions were determined using a computer program based on an iterative method of minimization of the sum of the square of the residuals between calculated and observed travel times to all stations. Because the earthquakes were located to the southwest of the array, the errors in calculated depth and epicenter are about 0.15 km. As in our previous study of earthquakes on the Columbia Plateau (Malone *et al.*, 1975), the epicenters are scattered with no obvious lineament to suggest a single fault plane. If, however, one plots the depth versus distance north (structural geology suggests thrusting to the north) of the array origin, as shown in Figure 4, one can find two general trends. One trend of hypocenters dips to the south at about 80^{o} and another intersects this trend and dips at about 20^{o} to the south. The 20^{o} dip to the south is consistent with a thrust mechanism. We are continuing to work on this data and hope that refined hypocenter locations along with the addition of marginally located events not included in this report will reveal a clearer picture of the seismicity versus structural geology for the Wahluke Slope.

Section C: Wooded Island Seismicity During the Summer of 1975

Based on preliminary locations furnished by Mitch Pitt of the U.S. Geological Survey (U.S.G.C.), the University of Washington Portable Seismic Array was deployed in the vicinity of Wooded Island on the Columbia River beginning June 20, 1975. The array configuration was changed at several times during the operating period. Preliminary location of micro-earthquakes within a few days of their occurrence allowed us to modify the array configuration to optimize locatability of future events, i.e., at least one station directly over the greatest concentration of hypocenters for depth control and other stations at varying distances and azimuths for epicentral control. Preliminary attempts at focal mechanisms also allowed us to move stations to try to get better coverage of the focal sphere. Several factors limited this endeavor. First, construction noise from Washington Public Power Supply System Reactor No. 2 limited us in distance and hence take-off angle to the north and west. Second, topography inhibited our telemetry capabilities to the east. Third, no earthquake of magnitude greater than $M_L = 1.6$ occurred during operation of the array and hence attenuation of seismic energy require small hypocentral distances.

The array is shown in Figure 5. Figure 6 shows during what periods particular stations were in operation. Equipment failures due to vandalism, high temperatures, and failure of portable power caused a loss of about 25% of the data. An insert on Figure 6 shows the percent of coverage obtained. More meaningful statistics on the error in hypocenter location are obtained for every station added after the fourth, which is the minimum number of stations required for leaving all hypocenter coordinates free in the location procedure. We had 5 or 6 stations in operation at least two-thirds of the time.

Over 365 events were detected by our array, but only 72 were large enough to be located. The located events ranged in magnitude from -0.8 to +1.6. Hypocenter solutions were determined using a computer program based on an iterative method of minimization of the sum of the square of the residuals between calculated and observed travel times to all stations. The velocity model used in the hypocenter location procedure was derived using stratigraphic information from several deep-core holes drilled in conjunction with the basalt stratigraphy study being conducted by Atlantic Richfield Hanford Company (ARHCO) for the Energy Research and Development Administration (ERDA) and from a sonic transit log for the 3406 meter deep Rattlesnake Hills No. 1 well, 30 km to the west of Wooded Island. Using an unconformity in the TiO_2 content of the basalt as described by Siems et al. (1975), we transcribed the velocity data directly from the Rattlesnake Hills No. 1 well to core holes DH3 and DB1. DH3 is 670 meters deep and 4 km south of the center of the array. DB1 is 300 meters deep and is the same well that served as the northern transmitter electrode in our resistivity array.

Figure 7 shows the velocity model used in the hypocenter locations. The sedimentary layer in the area has an average P-wave velocity of 2.12 km/sec as determined by site investigations for Washington Public Power Supply System Reactors No. 1 and No. 4, e kilometers to the northwest. The depth histogram for the located events (Figure 7) has a broad, but definite peak centered at about 0.9 kilometers. The activity diminishes to just a few events above 0.4 and below 1.4 kilometers. The lower limit correlates very well with the top of the low velocity layer. This low velocity zone from about 1.5 to 2.3 km depth corresponds to very thin, highly altered, and broken basalt flows as determined from lithologic data from several deep-core wells. The higher velocities above this correspond to much thicker and more competent flows which can probably store the strain energy released in the observed micro-earthquakes. The apparent upper limit of seismic activity at 0.4 km may be an artifact of the computer program use for hypocenter locations. If in the iterative procedure, the correction vector called for placing the next guess hypocenter above the surface of the ground, the depth was set equal to 0.5 km for that guess. Subsequent iterations may not have been able to converge on hypocenters with depths much shallower than this artificial limit. Lithologic data suggest that the basalt flows above 0.5 km, are in fact competent and not much different from those below 1.5 km depth. The sediment depth on top of the basalt is about 0.1 km. It is, therefore, likely that these shallow events do in fact locate nearer to the surface. In our ongoing work, refined location of these events will include an attempt to circumvent this apparently program-induced upper limit.

From the plot of epicenters (Figure 5a) there appears to be no single fault plane on which these earthquakes occur if one examines all located events. However, a selection of events which were located using 5 or more stations and which had an average sum squared residual per station of less than 0.01 sec, revealed a slightly different picture. A plot of these more accurately located epicenters shows that the activity is more concentrated and that there is a tendency for the epicenters to align parallel and on strike with a possible lineament seen as a contrast in vegetation.

Another structure in the area which may be related to the seismic activity is a gentle anticlinal fold in the basalt. The axis of this fold is believed to strike N 70^{o}W or about 50^{o} west of the postulated lineation. This fold is presumably an extension of one of the linear ridges which exist to the northwest and die out as they come on to the Hanford Reservation. Regional structural geology suggests that these folds are currently active structures. The very shallow nature of the Wooded Island Seismicity could be interpreted as tension failure near the free boundary associated with compressive folding at depth.

Our previous work on the Eltopia Swarm (Malone _et al._, 1975) concludes that a variety of focal mechanisms can coexist within a swarm and both structures discussed above may be active and interrelated.

Ongoing work includes refinement of the hypocenter locations using HYPO71 and station corrections developed on the basis of station elevation and variations in thickness and stratigraphy of the overlying sediments. Construction of focal mechanisms based on preliminary locations was unsuccessful, but we hope to have better results using the refined hypocenter locations. Examination of other aerial photos is also planned in conjunction with further field examination. Work has already begun on extracting events recorded at Wooded

Island station of the Hanford Network from film records. This process involves picking times of micro-earthquakes which arrive at Wooded Island station first and assuming they are located in the same focal volume as those located during our array operation. Times to the nearest minute, coda length, and polarity of first motion are recorded for each event. We hope to extend the reading of the film records to sometime before the April 22, 1974, event located by the U. S. Geologicual Survey and the apparent beginning of activity for this swarm.

Resistivity Monitoring.

The shallow nature of the earthquake swarm activity means that one can get much closer to the hypocenter than with tectonic earthquakes. In a swarm such as Wahluke or Wooded Island, events with hypocentral distances less than 1 km are fairly easy to plan on. This means that many reported earthquake precursers such as resistivity variations should be observable for much smaller events. During this support period we have deployed an active D. C. resistivity monitoring array. Although the data has not yet been fully analyzed, one event at the Wooded Island site shows an apparent resistivity anomaly.

Our measuring technique involves two transmitter electrodes fed by a constant current power supply which periodically reverses polarity and two receiver electrodes connected to a differential amplifier. The amplified and filtered output is digitized and recorded in series form on a single channel of the FM tape recorder which also records the seismic signals. The apparent resistivity is the ratio of the received voltage to transmitted current. In principle this allows us to detect resistance variations as small as one part in 10^4 (60 db dynamic range). Two other channels on the tape also record the transmitted current and the received voltage with a dynamic range of somewhat less than 40 db.

The initial deployment of the resistivity apparatus was on the Wahluke Slope with a Wenner type array with 100-meter spacing. Since the earthquakes had hypocenters several hundred meters from the array, no resistivity variations were seen as might be expected. The array was then changed to a dipole-dipole configuration with two 300-meter dipoles separated by 1 km. We immediately ran into serious difficulty with very large 60 Hz signals (several volts) induced by a 115 kilovolt power transmission line at the site. Re-design of the receiver was required. By the time this was accomplished, the Wahluke activity had died, and we shifted out attention to Wooded Island.

The Wooded Island seismic array was run for about two weeks to provide epicentral data for location of the resistivity array. Figure 8 shows a map of the site. The first preliminary locations showed the center of activity to be on the west bank of the river. The recording trailer which was initially located at a source of commercial power on the east bank was moved to the west bank and transmitter (T) and receiver (R) dipoles indicated by the dotted lines in Figure 8 were installed. The northern transmitter electrode is a 300-meter, deep well penetrating into the basalt. The southern transmitter electrode is a large, buried steel plate soaked in a salt solution. The receiver electrodes are copper-sulfate porous pots. We again had a 115 kilovolt power very close to the array with 2.5 volts of 60 Hz on the receiver electrodes. The new receiver was able to detect the 15 mv received square wave with no difficulty.

The magnitude 0.86 earthquake plotted on Figure 8 would appear to be ideal for seeing some sort of resistivity fluctuation. We have not yet decoded the digital resistivity data. However, we found no anomalous change on the direct recording of the received signal. Thus any effect must be less than about 5%.

The array and recording trailer was initially powered by a motor generator provided and maintained by ARHCO (Atlantic Richfield Hanford Co.). This caused considerable difficulty because of the cycling of the primary load: the airconditioner in the trailer. After one disaster where the trailer voltage rose to 160 VAC for several hours and damaged some of our equipment, we decided to move the recording trailer back to commercial power on the east side of the river despite the logistical problems resulting from having to drive 20 miles south to Pasco in order to cross the river. The transmitter dipole was enlarged somewhat with the southern electrode now being another well 90 meters deep which penetrates a few meters into the basalt. The circuit resistance of this transmitter is only 13 ohms. It draws 2 amps at 26 volts opening the future possibility of running the transmitter with more reliable power sources, such as batteries or a thermoelectric generator.

The new receiver was also enlarged and signal strength with the large array was similar to the earlier configuration. The only serious interference at the receiver site was a rancher's electric fence.

A signal proportional to the tranmitted current was sent by radio across the river. Unfortunately the basic time base for the system remained at the transmitter, and, therefore, we were unable to operate the digitizer at the receiver end. During most of the time the received signal was recorded directly with about 40 db dynamic range. This means that our detectability for resistivity changes was no better than 1% or 2%.

During the time that all configurations of the resistivity array operated, we recorded 18 locatable events. Ten of these are "within" the array meaning their hypocenter lies within one transmitter-receiver spacing of the center of the array. Five of the events have $M_L > 0.35$. With only one exception, none show any evidence of a resistivity anomaly.

The anomalous event is the magnitude 0.48 earthquake plotted on Figure 8. This event is anomalous for a variety of reasons. First, its epicenter is west of most of the activity and is only a few hundred meters from one of the transmitter wells. Second, the frequency of the seismic signals is noticeably lower than the other events in the swarm. Third, it is extremely shallow. If the calculated depth were reliable, it would be in the sediment. We have been unable to force the location deeper, but we are hesitant to make any firm conclusion yet as to the true depth. Additional data from the Wooded Island station of the Hanford Net which lies only a few hundred meters west of the epicenter will help. Finally, the event appears to have both a pre-seismic and a co-seismic resistivity fluctuation.

The resistivity data is shown together with one seismic channel in Figure 9. During the time period the transmitted current remained constant within the recorder resoltuion. Therefore, the apparent resistivity is proportional to the amplitude of received square wave. This gives one a measure of the resistivity at each transition of the received voltage. The transitions during this time occur every 30 seconds. Prior to the event, the record is very quiet. The very small short-period noise is almost all due to the local electric fence. There is also some telluric noise with a period close to a minute. The transition immediately prior to the event is 20% less than the preceding transitions. This is most easily seen by sighting along the record. All the transitions for several hours previous are within 5% of each other. One is tempted to conclude that 10 seconds prior to the event the resistivity was 20% below normal. About three seconds after the first seismic arrival, the signal appears to return to its normal level. This is accompanied by high frequency noise which may be seismic, or, judging by similar spikes at other times, may be either telluric or from the electric fence. The first transition after the event is slightly, but probably not significantly, larger than normal.

There are several unlucky circumstances which could produce the observation without a change in ground resistivity. One is a "box car" telluric signal which turns on precisely at the transition prior to the event and turns off 10 seconds later. A second is a drop in receiver amplifier gain in the period between the second transition prior to the event and the transition following the event. A third possibility is an overload of the input amplifier with a signal outside the filter passband. Normally, the gain is set so that with the filter switch out, the amplifier output is well below the maximum of 10 volts. However, an unusually strong signal could saturate the input stage effectively reducing the gain for the desired signal. Unfortunately, it is conceivable that the rancher's electric fence, which produces short bursts of 60 Hz energy might be able to produce the requisite interference, although the normal level after an active R-C filter before the amplifier input is hundredths of volts.

On balance, we think the observation is real, although one hesitates to make a firm conclusion without at least a second example. Some comments on the possible origin of the resistivity anomaly are warranted. First, the variation is probably not a result of dilatancy and fluid diffusion. The precursor time scale is too short. The scaling law used for larger earthquakes (Scholz, et al., 1973) predicts a 45-minute precursor. Further data analysis is required to see if such a long precursor is observed. Using pumping test data on the sediments overlying the basalt, we find that the consolidation time scale in the most permeable cases could be as low as 100 seconds. However, most of the sediments and the basalt have much longer pore fluid diffusion times. The most likely cause for a resistivity change, then, is strain associated with pre-seismic slip. The exact means of coupling the strain to the resistivity, however, is not clear. One possibility is that compression acts to increase the saturation of a partially saturated medium. Another possibility is that the strain changes the contact resistance at the wells used as the transmitter electrodes. Although the power source regulates the total current flowing into the ground very accurately, the relative proportion of current flowing out at different depths can change. Normally the wells are line sources. Suddenly changing them to point sources would reduce the received signal in the same manner as a volumetric resistivity increase. The fact that the event hypocenter is extremely close to one of the transmitter electrodes makes this hypothesis plausible.

On-going work consists of two basic types: We are continuing to examine data already collected. This principally means digitizing and synchronously detecting the resistivity signal with the transmitted current. This will allow detection of variations with time scales much longer than a minute. Secondly, we are redeploying the resistivity array on a more permanent basis in order to maximize our chances of detecting further events. As mentioned earlier, the power requirements of the transmitter are so low that we can power it with a thermo-electric generator. Several suitable units have been loaned to us by the U. S. Geological Survey. We have also obtained a 14-channel, low power, 30-day tape recorder which can also operate from a T-E generator. The resistivity array will operate entirely independently of any seismic work and will be correlated on a long-term basis with the Wooded Island station of the Hanford Net. Funding for this continuing field operation is being sought from the Energy Resource Development Administration (ERDA).

The Wahluke Slope and Wooded Island studies were partially funded by ERDA and the U.S. Geological Survey.

4. GENERAL EARTHQUAKE ACTIVITY

Over the last half of 1975, there have been several areas of obvious seismic activity. We have experienced earthquake swarms at Midway, Othello, Emery, and Wooded Island. There have been clusters and possible trends of seismic activity near River Homes and just south of Chelan. The significance of these relatively more active areas is not yet clear. Because of the regional type of spacing of the Central Net there may be more small earthquakes at south Chelan and River Homes than we can actually locate. A possible future course of action is to place the Geophysics Program's portable array in one or both of these areas to better monitor this activity. We hope to re-scan the Hanford Net records prior to July, 1975 to see if we can recognize any past seismic activity from these northern areas. See Figure 2 for a map of the seismicity of eastern Washington for the last six months of 1975.

The U.S. Geological Survey's Report on the Hanford Seismicity between 1970 and 1975 is expected to be in print shortly. When this report becomes available to us, it will help considerably in planning special studies and regional direction for future studies of the Hanford area. Until this report is available, a critical evaluation of the seismicity of the Hanford area will not be possible.

REFERENCES

Lee, W.H.K. and Lahr, J. C., 1975, HYPO71 (revised): A computer program for determining hypocenter, magnitude, and first motion pattern of local earthquakes, U.S. Geol. Survey Open-File Rept. 75-311.

Malone, S. D., Rothe, G. H., and Smith, S. W., 1975, Details of micro-earthquake swarms in the Columbia Basin, Washington, Bull. Seis. Soc. Amer. 65, 855-864.

Pitt, A. M., 1971, Micro-earthquake activity in the vicinity of Wooded Island, Hanford Region, Washington, U.S. Geol. Survey Open-File Rept.

Scholz, C. H., Syyes, L. R., and Aggarwal, Y. P., 1973, The physical basis for earthquake prediction, Science 181, 803-810.

Siems, B. A., Bush, J. H., and Crosby, J. W., 1974, TiO_2 and geophysical logging criteria for Yakima basalt correlation, Columbia Plateau, Bull. Geol. Soc. Amer. 85, 1061-1068.