ANNUAL TECHNICAL REPORT 1979

on

Earthquake Monitoring of the Hanford Region, Eastern Washington

Including Quarterly Technical Report 1979-B

Geophysics Program

University of Washington

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I. Introduction and Operations.

The current fiscal year, 1979, has seen the continued operation of the eastern Washington seismic network with few changes. The same seismic stations which were running at this time last year are still operating, and no major developments in the array configuration, data handling or processing were taken. A great deal of preparation has taken place for a major network expansion. Besides the routine processing of eastern Washington seismic data we have reanalyzed the data and produced a uniform catalog for the years 1969-1974. We have also taken delivery on a new computer and have been developing software for it. We ran a rather unsuccessful downhole experiment and have purchased and built the additional equipment needed for a successful one. We have also installed and are obtaining data from a broadband three-component network of stations for studying average crustal structure and attenuation properties. We have completed a refraction study of the north Cascade mountains and the area to the east. Details of these projects are covered in later sections of this report.

Array Expansion. For the past year we have been anticipating and planning for the expansion of the network into the southern Cascades. Last winter we chose five sites along the eastern flank of the Cascades and began the permitting process and ordered phone lines. Three of the sites are directly conneced to phone lines while two are radio telemetered sites whose receivers are located at one of the phone sites. The locally recording station at Goldendale has been put on a phone line for telemetry to Seattle.

Figure I-1 shows all the stations in eastern Washington and the Cascades including these new installations which are marked as solid triangles. During the past several months we have received additional support for six more stations in the south central Cascades. These stations will be used for a special geothermal research project supported by the U.S. Geological survey. We hope to have these stations installed before the winter bad weather closes the high country. The three most easterly of these six stations are shown as dashed triangles in figure I-1. These stations as well as the five just installed must await the software for the online computer before they can be recorded. At that time all stations in Washington will be simultaneously recorded on the same medium. This includes the 21 stations in the western Washington net, the 33 stations of the eastern Washington net, the 6 new east Cascades stations, the 6 southern Cascade geothermal stations as well as the Newport station, for a total of 66 online stations. The initiation of this recording system should greatly improve the data quality for the entire state.

We do have our new offline computer system up and running now and have begun to develop programs for the analysis of some existing data as well as those data which will become available when the online system is operating. Some of the information reported on in this report is the result of programs written and running on the new PDP 11/70 computer. In fact this report was written, edited, and formatted on this machine. Because of the power and availability of this computer we are beginning to develop some very powerful routines for the analysis of seismic data.

Digital Processing of 3-Component Data. Field operation of digital event recorders over the past year have provided several dozen high quality recordings of local earthquakes in the Cascade region and in the Columbia River basalt. Analysis of these records provides some new opportunities for understanding of both source effects and wave propagation effects.

The first task, which is currently underway, involves the development of wave polarization filters to assist in separation and identification of different classes of waveforms present on the seismograms, particularly in the interval between the P and S arrivals. Traditionally this has been accomplished by simple rotation of coordinates into radial and transverse. SH waves are then identified on the transverse component, and P and SV are further separated on the radial component by comparing it with the vertical. Waves which are "up and away" in polarization are identified as P type motion and those which are "up and toward" (with respect to the source in each case) are SV type motion. This type of analysis can be quite useful as a first cut at interpretation of a seismogram, and we are using it to look at the effects of SV-SH splitting due to anisotropy. As more data become available in digital form, it is possible that this simple type of analysis will enable us to address the nature and extent of anisotropic wave propagation. At this point we are developing the computer routines to do this separation on the data in hand.

Other more sophisticated methods of polarization filtering have been proposed in the past. Generally they make use of a frequency decomposition of the signal followed by some type of least square fitting of horizontal and vertical components. Application of these methods has generally been limited to low frequency (less than 1 hz) body waves. The two areas of research responsible for this development have been teleseismic body wave work in connection with the nuclear test discrimination problem, and earthquake prediction.

Since most models of stress dependent seismic velocities depend on dilatancy or some other crack related phenomena, they also predict an anisotropic effect. For this reason, we believe it is important to clarify the basic structural anisotropy that appears to exist in layered volcanic rocks such as those of the Columbia Basin. This is a moderately difficult task, but it is essential to have this information in hand before the possibility of stress induced anisotropy can be evaluated. For this reason, the initial polarization filtering work will be concentrated on SV-SH separation.

Our evaluation of the literature in this field indicates that the available filters are most heavily oriented towards separation of in-phase and out-of-phase motion on radial and vertical components. The obvious application of this process being the separation of compressional waves from Rayleigh waves and SV waves. This process, sometimes referred to as "rectilinear mode filtering" does not seem particularly appropriate for our problem of SV-SH separation. One of the drawbacks is that it is necessary to define a time window for the pulse before the analysis is done. This may be appropriate for teleseismic work, where the frequency content is quite predictable, but it does not appear that it will work well for local earthquake data where there are a variety of frequencies present, and the seismic phases may be separated by only a very small time interval. One of the challenges is to develop a technique that can separate SV and SH when the time interval between them is only a fraction of their predominant period. Any signal decomposition technique designed to accomplish this will of necessity be non-linear.

Another important application of 3-component digital seismic data is in the comparison of recorded data from explosions and earthquakes with theoretically calculated motion (synthetic seismograms). The technical capability to calculate ground motion from an earthquake has increased greatly during the past several years. The experimental capability to measure such motion and compare it with predictions is just now emerging. This work has important application to unravelling the crustal structure in eastern Washington.

Aplications of computer graphics in seismology. The basic information that emerges from the operation of a seismic network is the time-space location of earthquakes and some measure of their energy release. Traditionally this information is displayed on seismicity maps and on cross section plots to show how earthquakes are distributed within a region. Except in a few instances where the interpretation is very clear, the final interpretation of such data is limited by the ability of an observer to assimilate the information and synthesize it into a geologic model. In the case of an aftershock sequence after a large earthquake, the three dimensional distribution often leads to a picture of the buried fault plane responsible for the main shock. This is a particularlrly simple example where the time dependence, and the energy release are not critical for understanding the process. In other situations there may be a relationship between earthquakes in one part of a region with those in another. Physically this makes sense, since each earthquake represents a readjustment in the stress field, it is reasonable to expect that they will be interrelated. To properly evaluate this type of interaction, the time-space-energy function must be displayed in a manner that will assist in the interpretation. Application of computer graphics to this problem is just in its infancy.

During the past several years we have experimented with some simple 3- dimensional display techniques. These methods have focussed on three methods of visually assisting an interpreter, the use of stereographic pairs for 3-D viewing, the use of operator controlled motion of a display to provide visual clues for depth perception, and the use of time lapse views to provide insight into the time rate of change of seismic energy release.

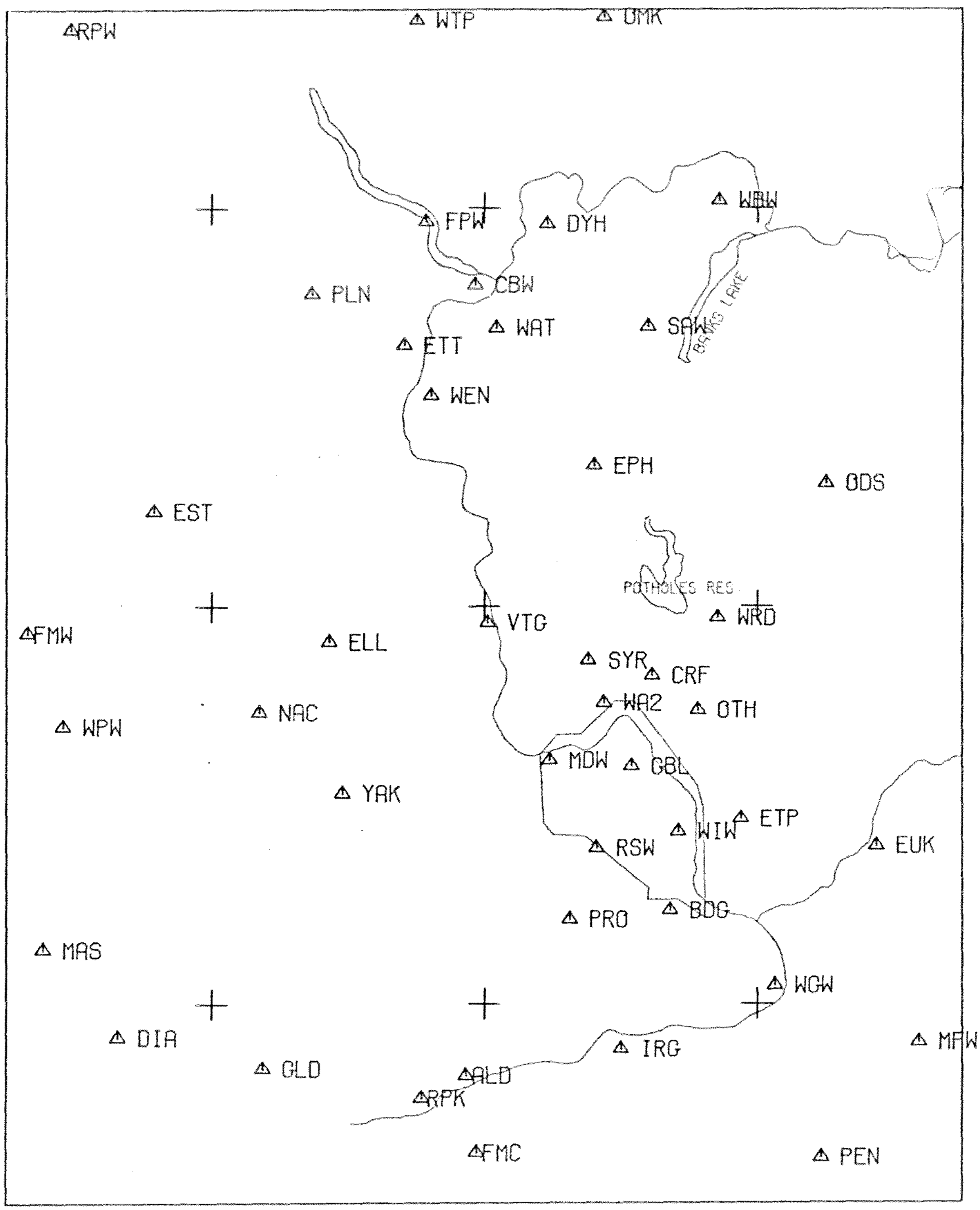
Our current effort in this area is in refining the experimental tools developed over the past several years and producing software for our new computer system so that these techniques can be applied on a more routine basis. We have been adapting some of the programs developed here as well as some developed elsewhere. Notable among the programs obtained outside the University has been the 3-D graphics package developed by Bishop and Foote at Battelle N. W. Laboratories. The basic Fortran programs for stereo projections including the effects of perspective have been converted and installed on our DEC 11/70 system. Work on obtaining and installing the various data bases for mapping is not yet complete.

Downhole Experiment. Unfortunately we have made little progress with our proposed downhole project. We ran a rather unsuccessful preliminary experiment in Febuary and March of this year. A single component old-style deep hole seismometer was installed in DC-3 along with five radio telemetered surface instruments within several kilometers of the wellhead. There were numerous problems with the downhole seismometer, probably because it is quite old and mechanically unstable. There were additional problems with the tape recording system for those periods of time when the downhole instrument appeared to be working. There were no nearby earthquakes during the relatively short period of time when all parts of the system were operational. From this experiment we have gotten a good qualitative feel for the variation in seismic noise level with different depths below the surface. The noise on or near the surface in the area around DC-3 is unusually high. It is probably 12db higher than nearby sites on bedrock such as Gabble Mountain and perhaps as much as 20db higher than at a depth of 1km. These are just rough estimates for use in designing future experiments.

We have recently put in quite a bit of effort trying to find deephole seismometers which are commercially available and not prohibitivley expensive. In April we determined that a three-component downhole instrument made by Geospace is the only instrument that comes close to meeting our requirements. Since this item was not in our original budget there was some delay in obtaining approval for its purchase. The lead time on such an item is now up to 21 weeks. We have ordered a deephole set and a shallowhole set for comparative purposes. These units are suposed to be shipped by the end of July and we hope to start another experiment later in the summer.

The remainder of this report is divided into several sections dealing with various research topics which are completed or at an advanced stage. Chapter III deals with the ongoing study of the temporal distribution of earthquakes in eastern Washington. Chapter IV is the final report on our magnitude calibration study which has developed our current coda length magnitude scale. Chapter V is a complete and thorough introduction to our study of the average crustal structure and attenuation properties for the whole state. This project is nearing the halfway point. The development work is complete, some of the data ~~has~~ have been obtained and the analysis is beginning. The primary effort on this project is by Sheng-Sheang Bor for his PhD dissertation which he hopes to complete in a year and a half. Chapter VI is a part of a PhD dissertation presently being completed by Al Rohay on the velocity structure of the Cascade mountains. The complete catalog of relocated events from the U.S. Geological Survey eastern Washington network is included as an appendix to this report.

There are also four large mapps that are included at this time since they summarize the seismicity of the last ten years. Figure I-2 shows the better located earthquakes during the period 1969 - 1974, that period during which the U. S. Geological Survey operated the network and which the appendix covers as a catalog. Most of the events on this map occur in the central plateau since that is where most of the U.S.G.S. stations were located. (See figure A-1 of the appendix.) Next the catalogs of the U.S.G.S. and the U. of W. are combined in figures I-3, I-4, and I-5. Each of these is a different selection of events from the combined catalog. In Figure I-3 all very well located earthquakes are plotted without regard to their size or depth. A strict rms and epicentral error requirement was used as well as restrictions on number of stations recording the event and the azmuthal gap. This figure represents the best 758 earthquake locations in the catalog. Figure I-4 Takes a slightly more lenient subset of the catalog but only includes events which are less than 6 km in depth while figure I-5 is the complementary deeper events. For interest we also include figure I-6 in the text which shows a map of all the located explosions for the past decade.

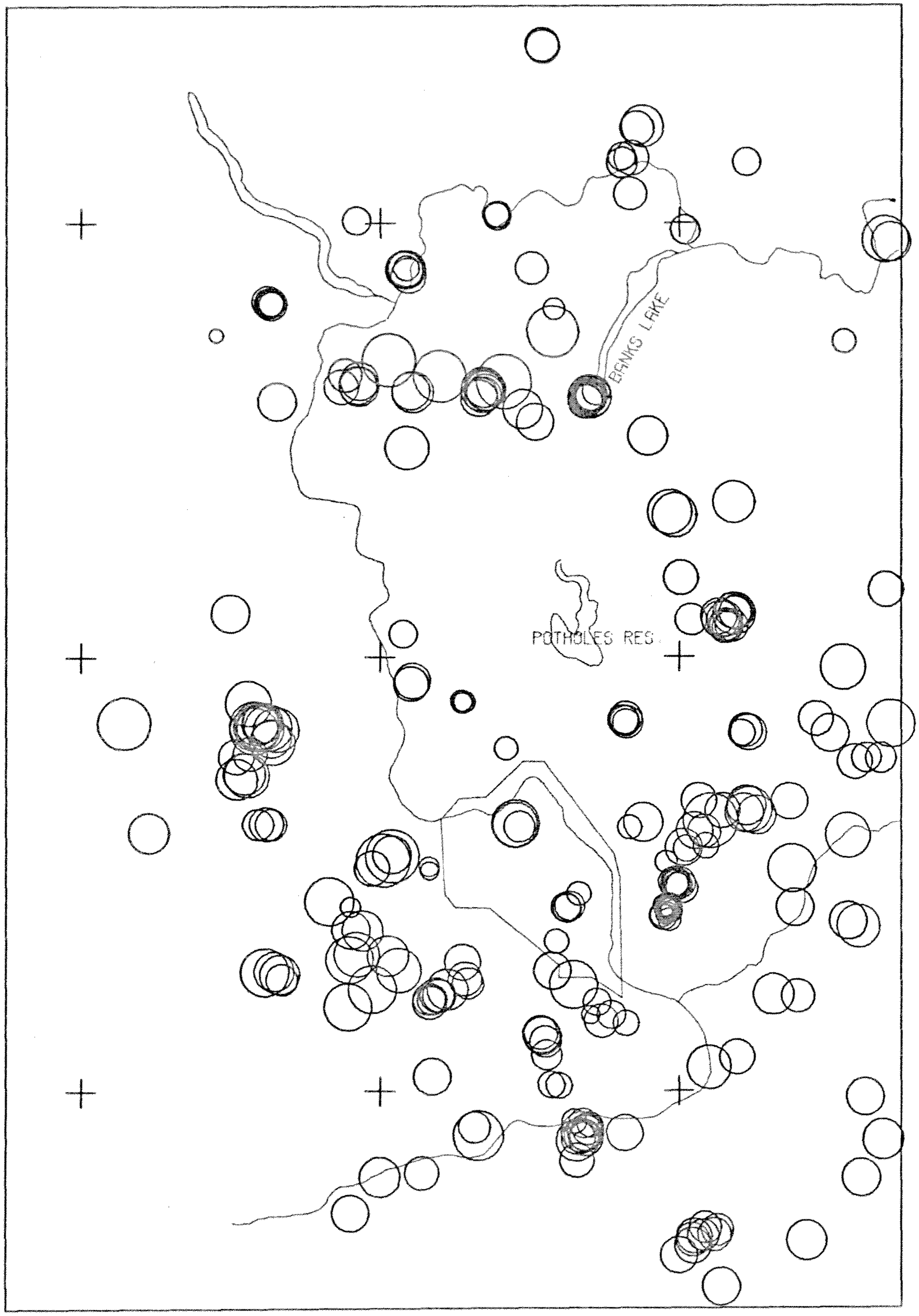


EASTERN WASHINGTON SEIMOGRAPH STATIONS JUNE 1979

CENTER OF MAP IS 47.00 N 120.00 W

MAGNITUDE KEY ○ 0.0 ◯ 1.3 ◯ 2.7 ◯ 4.0

Fig. I-1



EASTERN WASHINGTON EXPLOSIONS 1969 - 1979
CENTER OF MAP IS 47.00 N 119.75 W

MAGNITUDE KEY ∘ 0.0 ◯ 1.3 ◯ 2.7 ◯ 4.0

Fig. I-6

III. Seismicity

This chapter deals with several specific seismicity studies which we have engaged in during the past year as well as reviewing the general earthquake distribution of the past twelve months. Having a complete and uniform catalog from the spring of 1969 to the present has allowed us to look at some statistical patterns over this decade. These studies are specifically related to the temporal distribution of various classes of events and the size distribution of these same classes as revealed in recurrence curves. First we deal with the seismicity pattern of the past year, which has been mostly rather ordinary. Figure III-1 shows the epicenters for the past twelve months. The two major areas of activity are the norh flank of the Saddle Mountains and the Chelan-Entiat area. There have only been 43 earthquakes which fall into the deeper than 6 km class and of these all but 8 were in the area around Lake Chelan. There have been no intense swarms similar to the spring of last year. There have been a few events in the general Ellensburg region indicating this continues to be the only area of the Cascades with any significant activity.

Chief Joseph Dam Earthquakes. There have been two sequences of earthquakes which were out of the ordinary during the past year. The first started with a magnitude 4 earthquake on Jan 19, 1979 just south of the Chief Joseph Dam. A number of aftershocks followed this event, the largest being a magnitude 3 event on Jan 21. The activity tapered off considerably over the next few weeks but some residual activity remains at least until April 21 when a magnitude 1.8 was located in the same area The station coverage is adequate to obtain a reasonably weel controlled focal mechanism solution. The first motions for this event are plotted at the top of figure III-2. Since most of the earthquakes in this sequence were moderately deep, (4-8 km) the focal sphere is fairly well sampled. The solution is not much different than some from the south Lake Chelan area.

<u>Walla Walla earthquake</u>. As was reported in last quarter's technical report there was a felt earthquake near Walla Walla on April 8, 1979. We have reanalyzed all available data for this event and determine its epicenter to be at 45 59.7 north 118 26.8 west and its depth to be 4.7 km. The error on the epicentral coordinates is 1.2 km and on the depth coordinate, 1.4 km. An unusual fact about this earthquake is its total lack of associated activity. There were absolutely no foreshocks or aftershocks which could be detected at the nearest station, MFW. The computed magnitude for this event is 4.15 $\pm$.2 . Most events of this size have numerous aftershocks detectable on a nearby sensitive seismograph.

The focal mechanism for this event appears to be a bit out of the ordinary also. Figure III-2 shows the first motion pattern for this event with our interpretation of possible nodal planes. This interpretation is considerably different from ~~than~~ the preliminary one reported last quarter. We are still not fully confident in this interpretation since we had to go to some lengths to get this good ~~of~~ a fit. An alternative interpretation is shown as light lines. The velocity model used to calculate take-off angles was a linearly increasing one starting a 5.1 km/sec at the surface and increasing at 0.05 km/sec/km of depth. We arbitrarily moved the focal depth of this earthquake to 6 km such that the rays leaving the focus horizontally reach the surface at a distance of 48 km. This was necessary to obtain any sort of consistent plot. The maximum principle stress direction for this solution is much more nearly east-west than all other solutions obtained in the plateau. Because of the importance of earthquakes in this region we hope to establish a locally recording station near Walla Walla soon.

<u>Time Distribution Statistics</u>. With the reanalysis of the U.S.G.S. data we now have a uniform catalog with which to examine the temporal statistics of the earthquake activity in eastern Washington. The complete catalog now goes from April 1969 to June 1979 and contains 2587 events of which 345 are known or suspected blasts and 2242 are earthquakes. Some of these events fall outside the boundaries of the eastern Washington network but are still included in the catalog. There are 113 of these leaving 2129 earthquakes of some interest to this project. In the subsequent discussions the analysis does not include data from the second quarter of 1979 since these data were not available at the time of the analysis. This means the period over which the temporal analysis is run is exactly ten years. It should be

remembered that the network array location sensitivity has not been the same over this period.

The earthquakes have been divided into different categories for this analysis because of the different character of the seismicity of different regions. The area just to the south of the southern end of Lake Chelan has been an area of continued activity in which very little swarming is observed. In contrast to this area the shallow events in the central Pasco basin and the Saddle Mountains area show a very marked swarming characteristic. The deep events appear to show little swarming tendency, much like the Chelan area seismicity.

To study the swarm characteristics, events in a 1.4 degree of latitude by 1.6 degree of longitude were selected. These events are plotted in figure III-3 . Only the best located events are plotted in this figure, those with an rms less than 0.2 and estimated epicentral error less than 2.0 km. The very obvious spacial clustering of these events is evident in this figure. We have rather arbitrarily drawn boundaries around what we call individual "swarm areas". These subregions are labeled with a two letter designator for identifying the individual areas. In drawing the boundaries we attempted to include only enough area to encompass all the earthquakes associated in a group. Table III-1 lists these thirteen subregions and gives the important parameters about their location, size, and number of events.

TABLE III-1

Swarm	Latitude south	Latitude north	Longitude east	Longitude west	Area (km^2)	Number of events
West Saddle Mt.	46.820	46.865	119.564	119.797	91.6	46
Frenchman Hills	46.865	46.960	119.514	119.600	71.3	40
Smyrna	46.800	46.855	119.433	119.564	62.9	163
Royal	46.850	46.910	119.317	119.450	69.7	100
Corfu	46.790	46.850	119.331	119.433	53.4	159
Wahluke	46.729	46.780	119.314	119.431	52.1	134
Berg Ranch	46.685	46.728	119.317	119.417	37.6	44
Othello	46.650	46.710	119.183	119.317	70.2	72
Scooteney Res.	46.600	46.680	119.047	119.183	95.0	76
Connell	46.650	46.715	118.850	118.933	47.1	47
Wooded Island	46.390	46.470	119.200	119.333	92.9	179
Eltopia	46.370	46.455	118.964	119.067	76.4	63
Coyote Rapids	46.635	46.730	119.483	119.650	138.5	63
Central plateau	45.800	47.200	118.600	120.200	19560.7	1192

To examine the temporal clustering of events within swarm areas, plots have been made of the number of events per unit time versus time. The ten year period of the catalog is divided into 1000 equal intervals of about 3 1/2 days each and the number of events occurring within a swarm area for each interval is plotted in a histogram manner. These plots are shown in figures III-4, III-5 and III-6 where figure III-4 is for the swarm areas north of the Saddle Mountains going west to east, figure III-5 is for swarm areas just to the south of swarm areas just to the south of the Saddle Mountains and figure III-6 is for other swarm areas like Wooded Island, deep events (which are not swarms) and all events which do not fall into one of the above categories. All events down to magnitude zero are included and for each event larger than magnitude 3 there is a dot placed over the interval in which it occurrs. There are a total of 1699 earthquakes included in these plots. 1192 occur within one of the thirteen selected swarm areas, 300 are shallow but not in a swarm area and 207 are deeper than 6 km and thus are lumped together in their own category.

Note in figure III-6 that indeed the events labeled "not in swarms" and "deep events" display very little temporal clustering. For the deep events there appears to be a weak cluster in the spring of '71 and the summer of '75. The '71 bunch are not much spacially associated though the '75 group are made up of a large felt earthquake in the Horse Heaven Hills and its after shocks. The events which are shallow and fall outside selected swarm areas show almost no clustering. Thus over three quarters of the shallow earthquakes occurring in the last 10 years in this area occur in only 5 percent of the area.

Most of the selected swarm areas have had more than one swarm take place during the past decade. The area called "West Saddle Mt." and "Coyote Rapids" are the only areas without very distinct repeated clustering of events. The most pronounced are "Smyrna," "Royal," "Corfu," and of course "Wooded Island." The swarm areas just to the south of Saddle Mountains (figure III-1) have been surprisingly quiet over the last three years with the exception of "Coyote Rapids" which is isolated on the southwestern end.

The obvious temporal clustering of these earthquakes can also be seen when they are all lumped together as in the bottom part of figure III-7. Many of the individul swarms can be seen as spikes sticking though the background level and in some cases neighboring swarms were active near the same time adding their effect together in this plot. The solid lines represent the number of swarm events per year and is just the sum of the spikes below it in each year. Notice that both 1970 and 1972 had more than 200 swarm events while 1976 had fewer than 20 total. At the top of this same figure a similar plot is made for the earthquakes in the Lake Chelan area. Note that the number of events located before 1975 is far fewer than after illustrating the change in array configuration which took place in the summer of '75. There are no significant spikes representing clusters in these data though there apears to be a very broad hump in activity centered around '77-'78.

<u>Annual Periodicity</u>. Mr. R. E. Isaacson of Rockwell International noticed an apparent annual periodicity in the number of events from the catalog March,1969 through May,1976 (personal communications, 1978). In figure III-8 we illustrate such an effect by plotting the number of events which occurred in a given month over the past ten years. An apparent decrease in activity from winter to summer is evident in this plot though it is not as

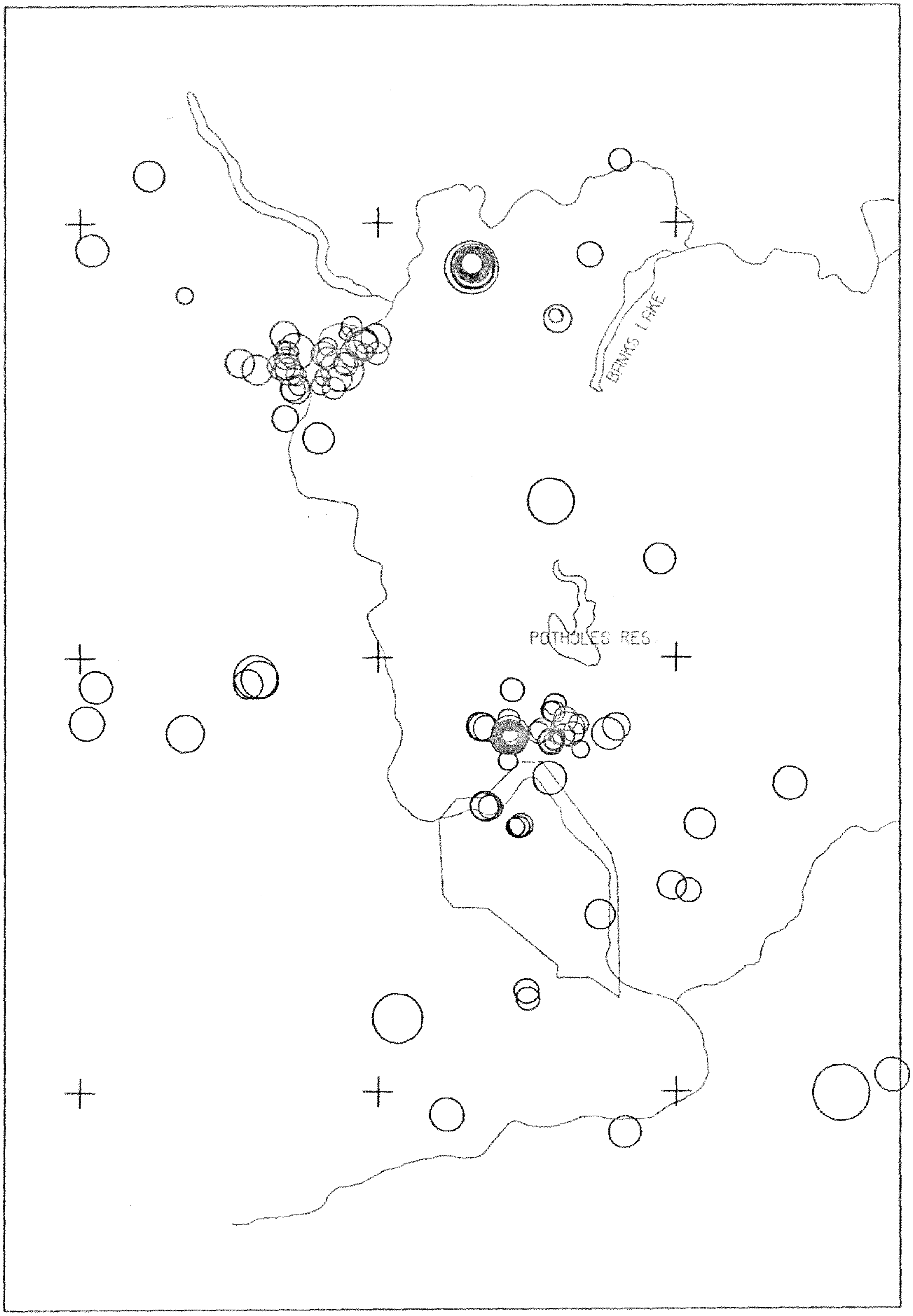
dramatic as that displayed by Mr. Isaacson. The catalog which he used was missing the first half of 1975 which included the Wooded Island swarm. Including this missing section and extending the catalog through the spring of 1979 smooths out the effect somewhat though July does seem to be rather quiet when compared to December or January. This effect may be just an artifact of the spikey nature of the earthquake activity. When events occur in very intense swarms it is possible for a few swarms to strongly effect this type of analysis. We plan further study on this subject because of the obvious implications of an annual cycle on the source for the earthquakes. To investigate this problem, as well as the more general problem of the cause of the shallow earthquakes, we propose to develop a quantitative statistical model for the swarms. In this way the swarms can be considered events whose individual earthquakes are not independent. We would then produce a plot such as shown in figure III-8 for clusters of events rather than individual ones.

Size Distribution Statistics. Using the same catalog as above we next examine the relationship of small to larger earthquakes in several classes. The standard way of approaching this problem is to use recurence curves sometimes called b-value studies. In this analysis the number of earthquakes equal to and larger than a certain magnitude is plotted on a logrithmic scale against that magnitude. With enough data these points will usually define a straight line. The slope of this line is the b-value, the name coming from the following equation:

$$\log N = a - b M$$

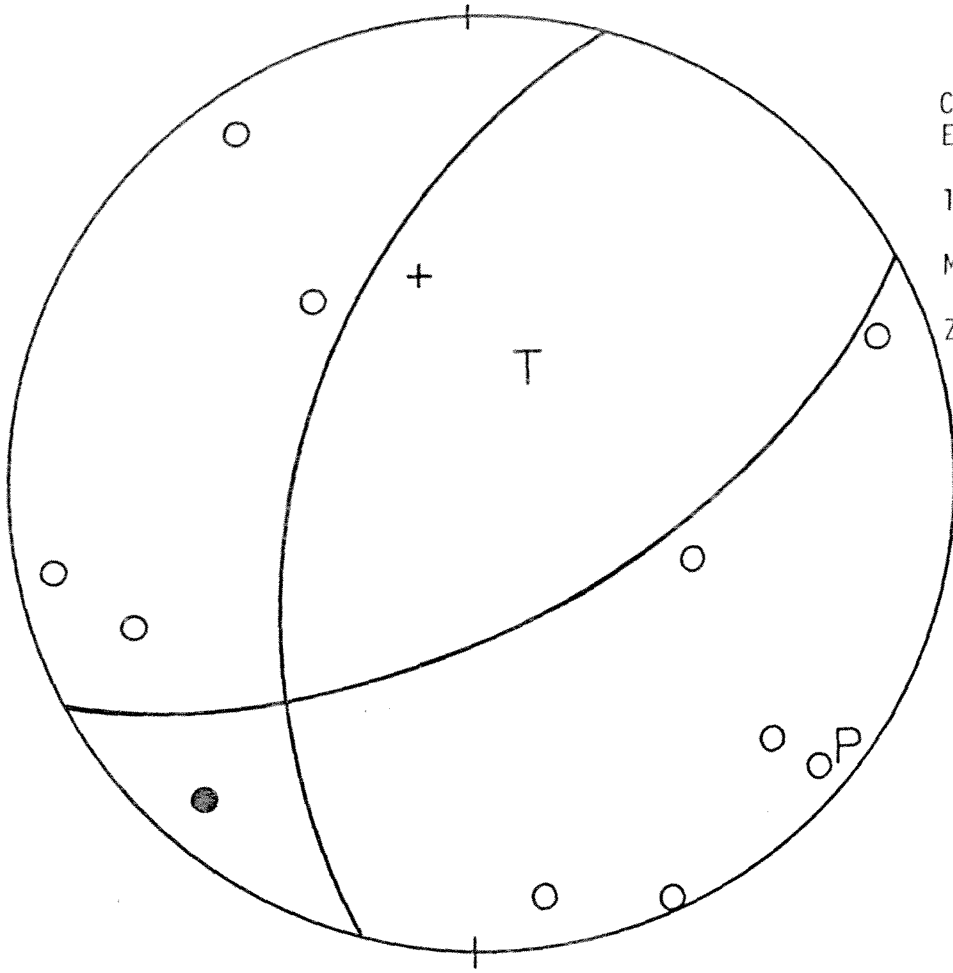
Here M is the magnitude, N is the number of events of this magnitude or greater, "b" is the slope of the curve while "a" is the intercept. Usually "a" has no significance since it is a function of the number of events used which depends on the area covered and the time over which the data ~~is~~ are collected. If the b-value for an earthquake sequence is 1.0, this means there are 10 times more earthquakes of magnitude 2 than there are of magnitude 3 and 10 times more magnitude 1 events. Thus the b-value or slope of the recurrence curve is a measure of the ratio of the small earthquakes to large events. Traditionally it has been observed that tectonic earthquakes on major plate boundaries such as those found in California or western Washington have b-values around .7 to .8. Earthquake sequences on volcanoes or swarm events have b-values larger than 1.0. There may be some relationship between the degree of homogeneity of the medium or stress field and the b-value for earthquakes.

Since there are easily separable classes of events in eastern Washington it is natural to determine the b-values for these different classes. In figure III-9 we plot the recurrence relation for four classes of events. The shallow earthquakes in the central plateau are divided into different time periods since the technique in determining magnitudes changed in the summer of 1971 (see the appendix). Note in figure III-9 that the b-value for both sets of shallow central plateau events is significantly higher than 1.0. Both the events south of Lake Chelan and the deeper events in the central plateau have more typical b-values around .75. This is additional confirmation that the shallow central plateau events fall into a separate class from the deep events. Their higher b-value is typical of swarm earthquakes in other parts of the world. The swarms have relatively more smaller earthquakes than larger ones in comparison to earthquake sequences which follow a more traditional main shock-aftershock sequence.



EASTERN WASHINGTON EARTHQUAKES JULY, 1978 - JUNE, 1979
CENTER OF MAP IS 47.00 N 119.75 W
MAGNITUDE KEY o 0.0 ◯ 1.5 ◯ 3.0 ◯ 4.5

Fig. III-1

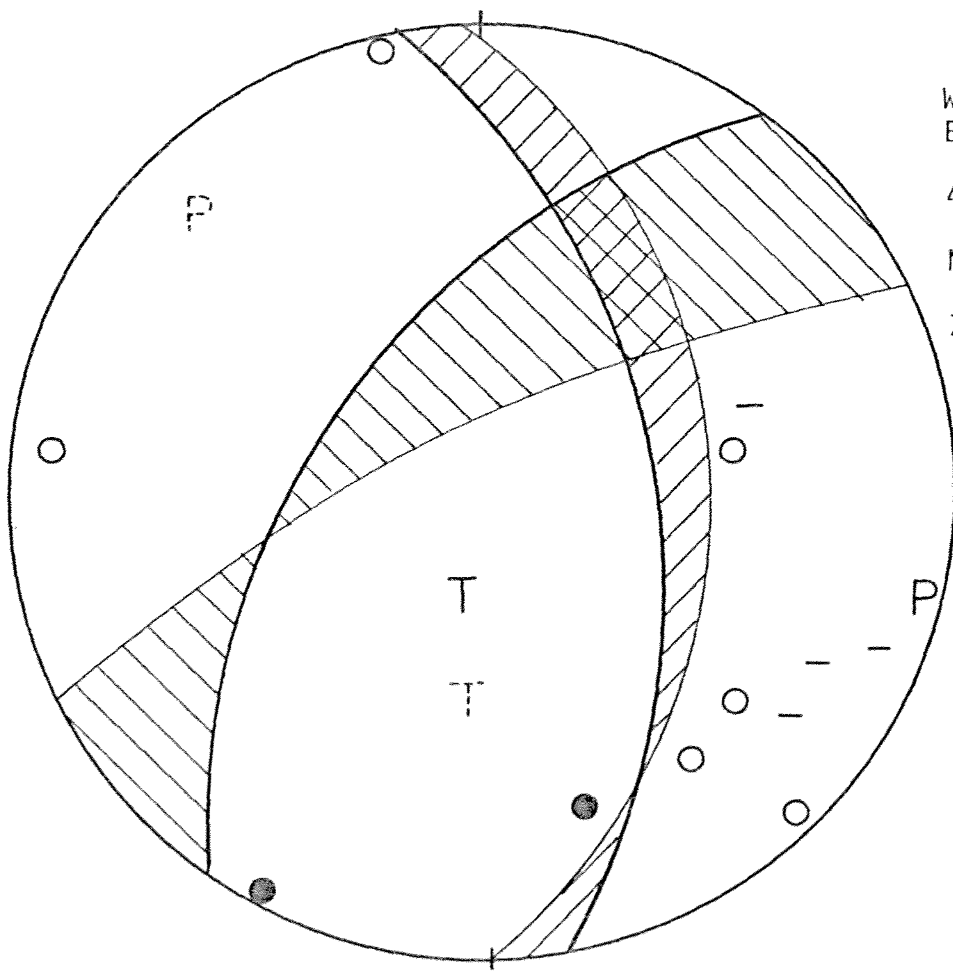


Chief Joseph Dam
Earthquake

1/19/79 1455Z

M = 3.9

Z = 6.7



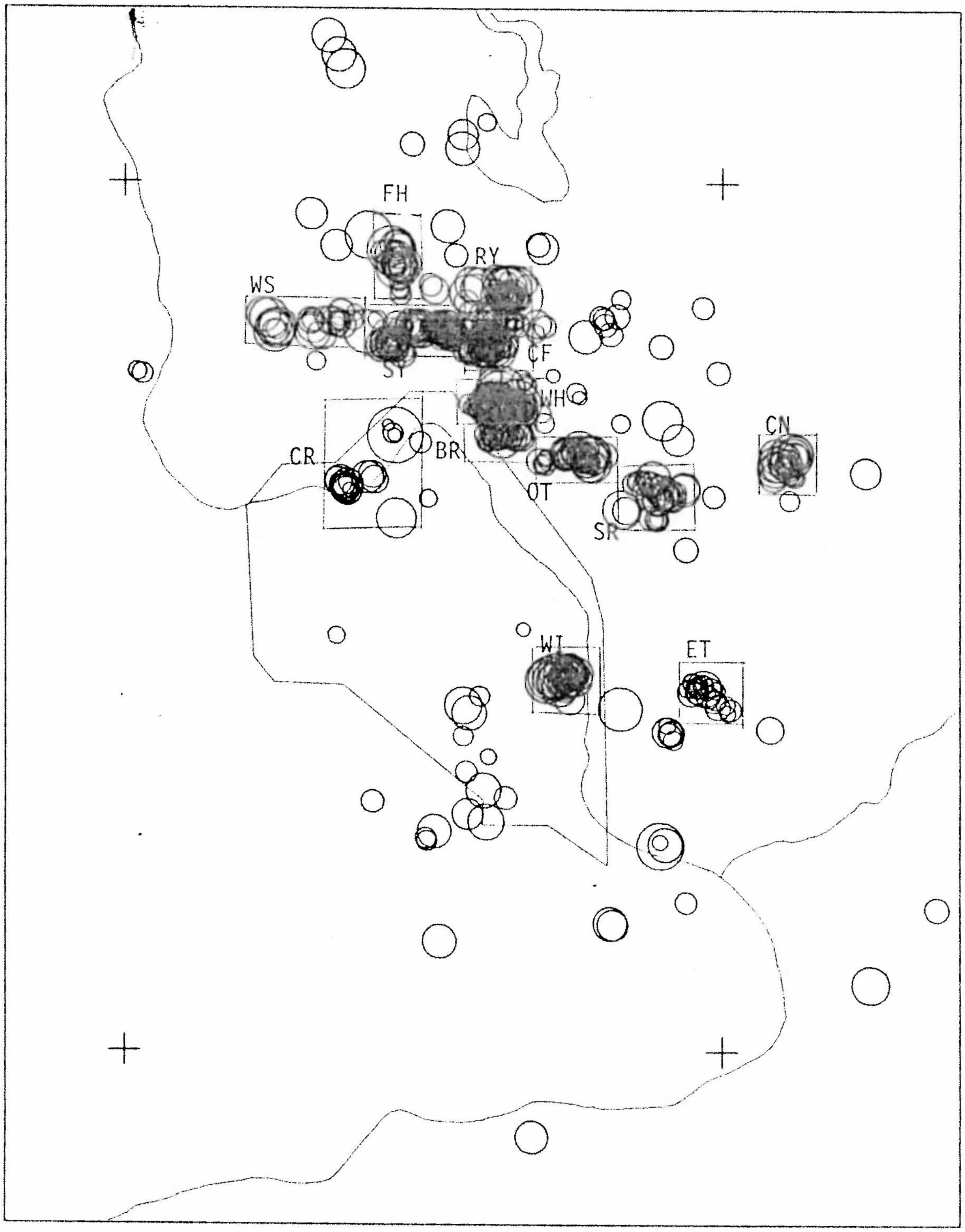
Walla Walla
Earthquake

4/8/79 0729Z

M = 4.1

Z = 4.7

Fig. III-2



WELL LOCATED SHALLOW EARTHQUAKES 1969 - 1979
CENTER OF MAP IS 46.50 N 119.40 W

MAGNITUDE KEY ○ 0.0 ○ 1.5 ○ 3.0 ○ 4.5

Fig. III-3

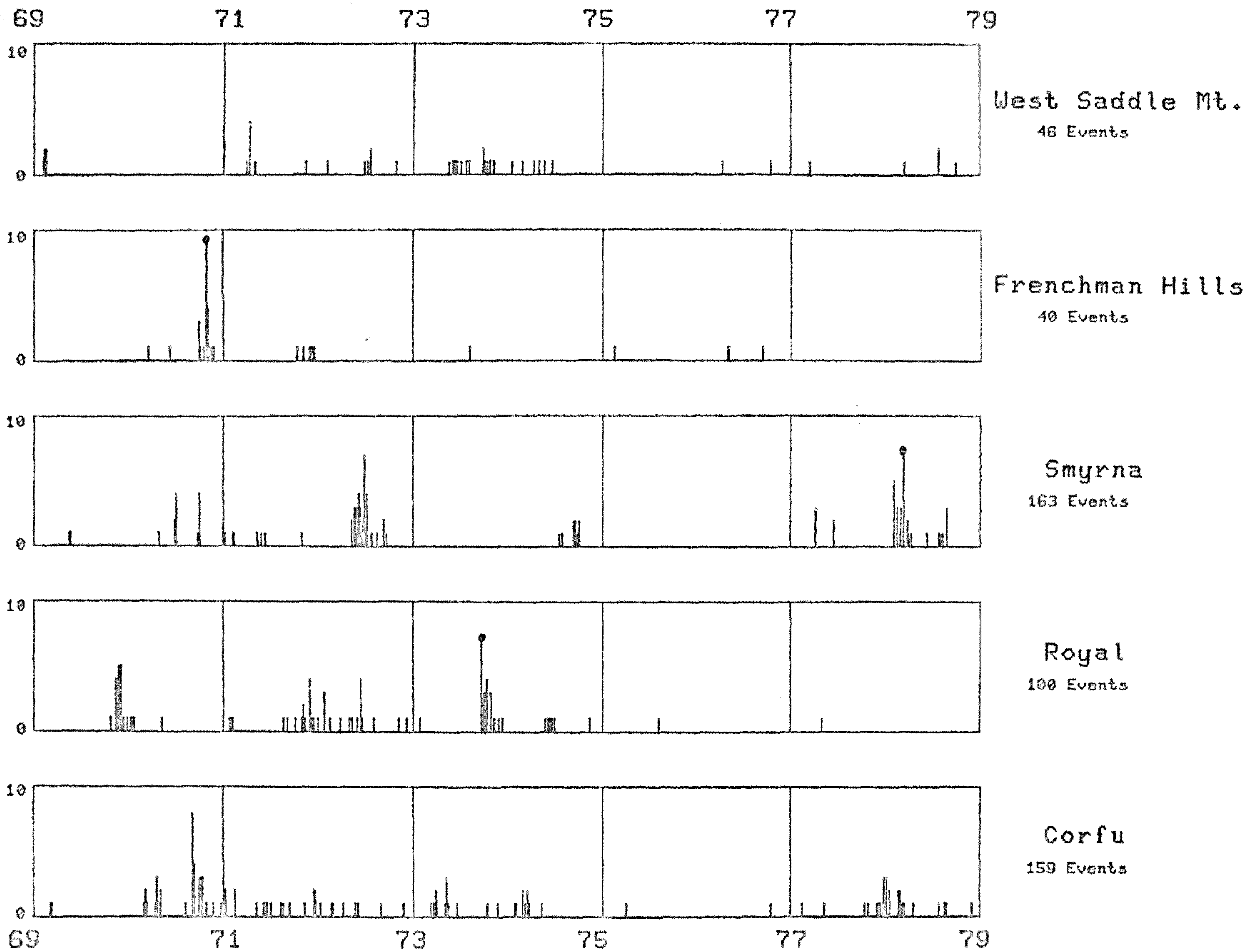

69 71 73 75 77 79
10
0
West Saddle Mt.
46 Events
Frenchman Hills
40 Events
Smyrna
163 Events
Royal
100 Events
Corfu
159 Events

Fig. III-4

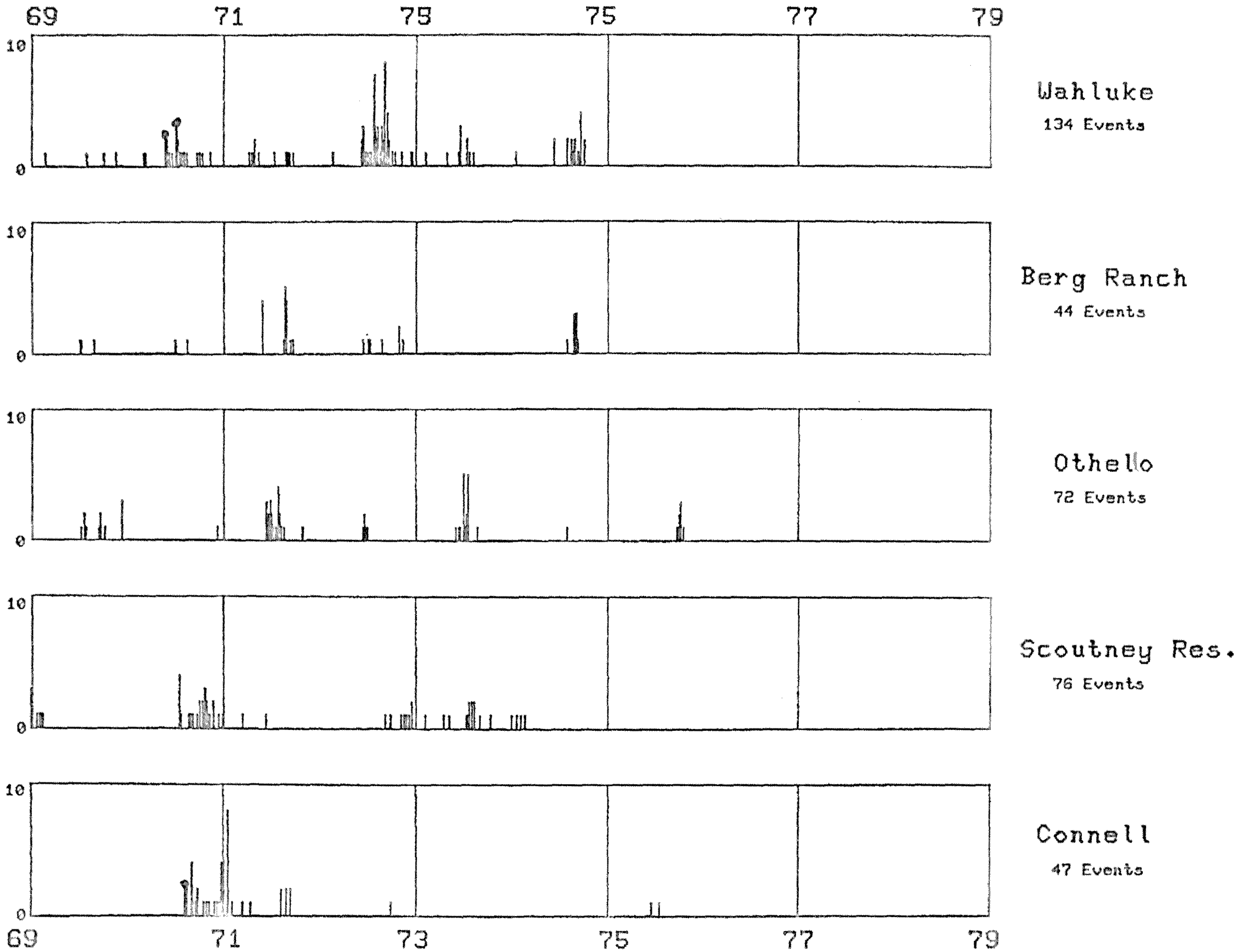

69
71
73
75
77
79
10
0
Wahluke
134 Events
Berg Ranch
44 Events
Othello
72 Events
Scoutney Res.
76 Events
Connell
47 Events

Fig. III-5

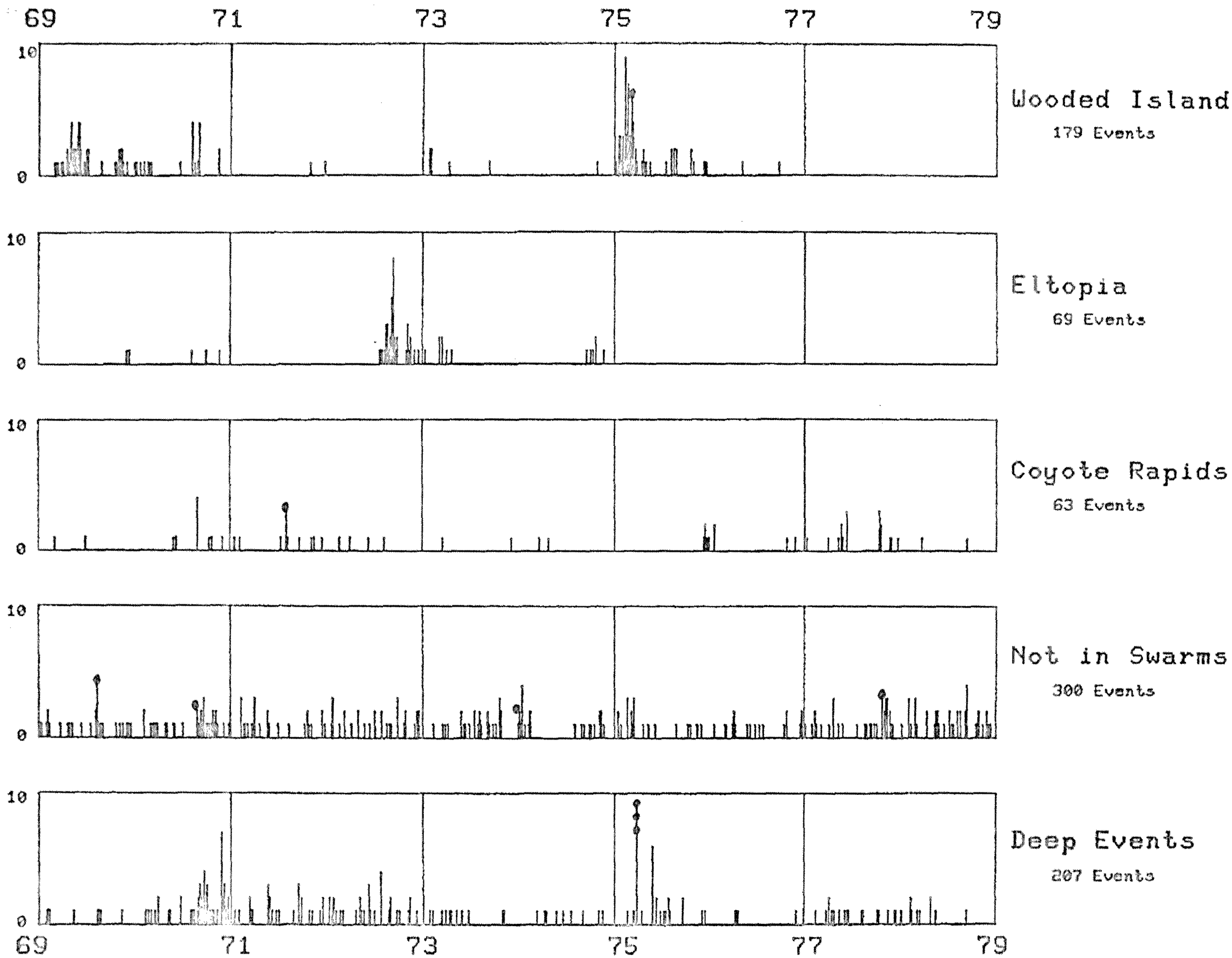
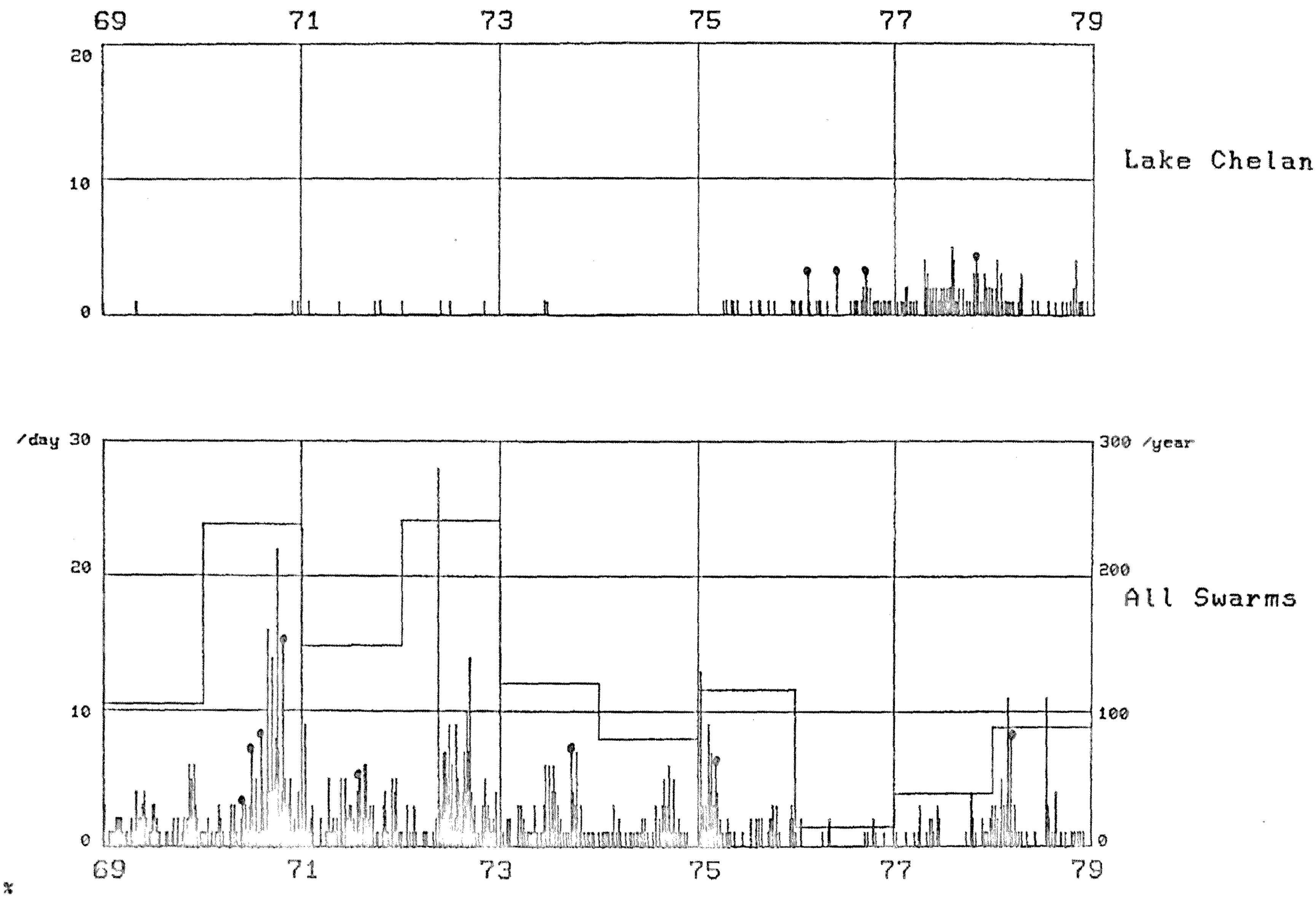

69
71
73
75
77
79
10
0
Wooded Island
179 Events
Eltopia
69 Events
Coyote Rapids
63 Events
Not in Swarms
300 Events
Deep Events
207 Events

Fig. III-6



69
71
73
75
77
79
20
10
0
Lake Chelan
/day 30
20
10
0
300 /year
200
100
0
All Swarms
69
71
73
75
77
79
%

Fig. III-7

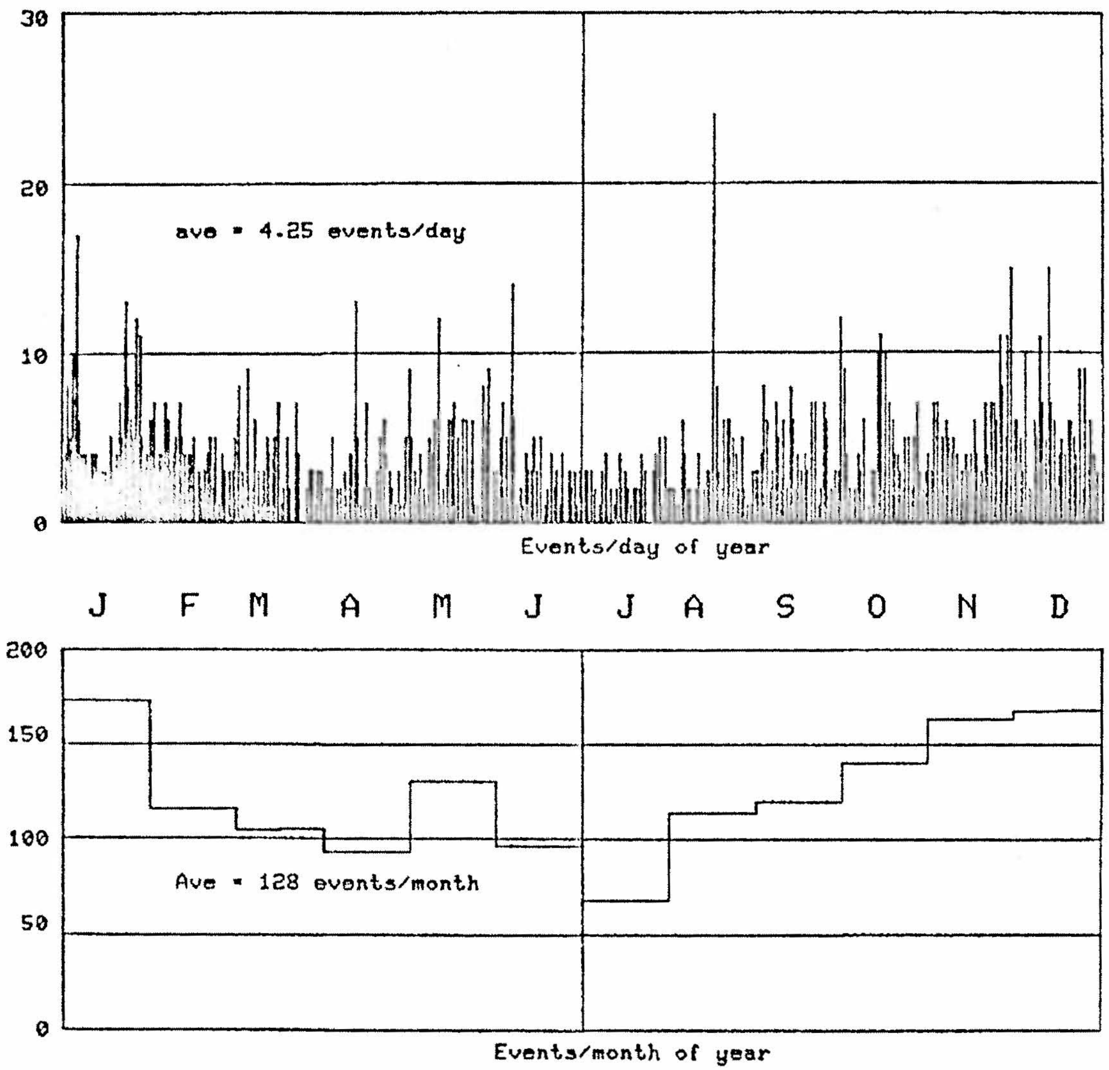

30
20
10
0
ave = 4.25 events/day
Events/day of year
J F M A M J J A S O N D
200
150
100
50
0
Ave = 128 events/month
Events/month of year

Fig. III-8

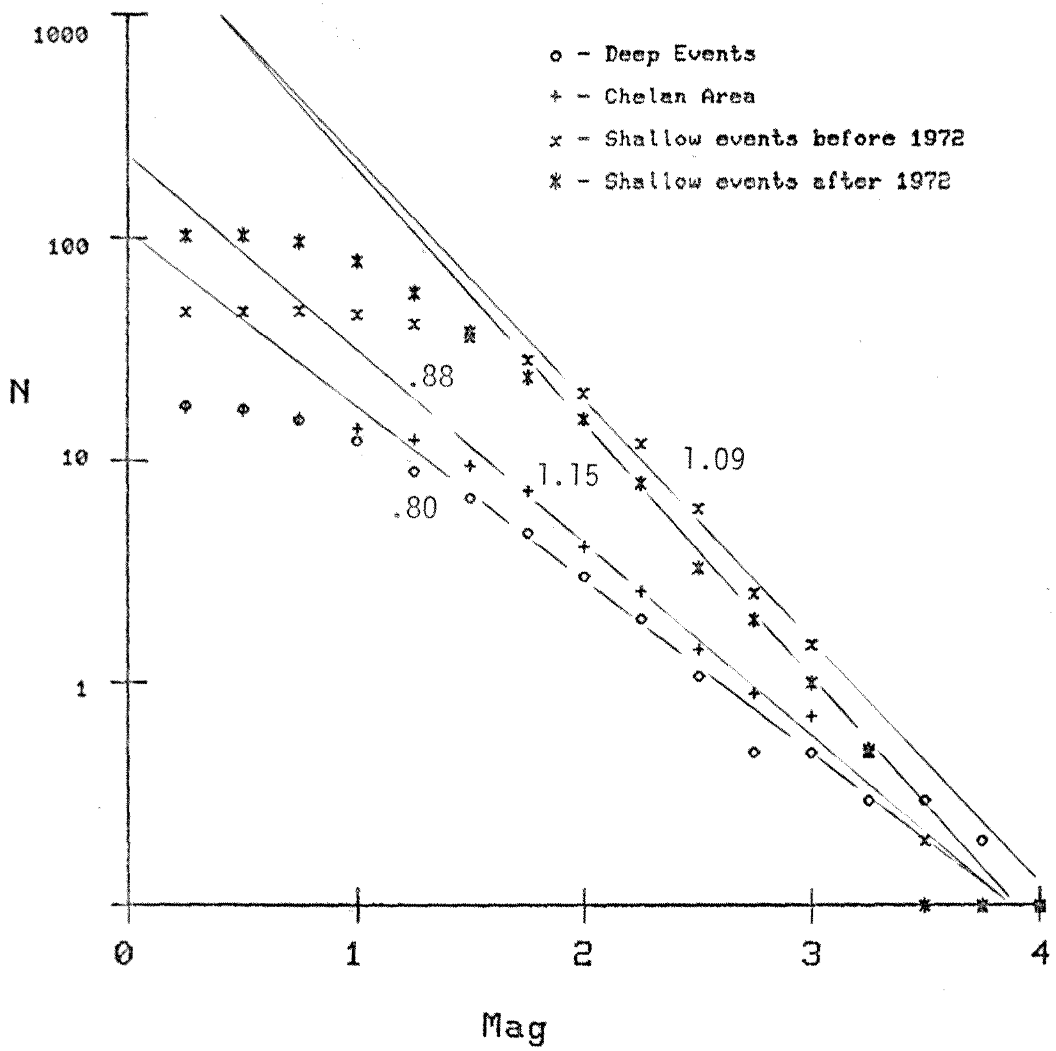

o - Deep Events
+ - Chelan Area
x - Shallow events before 1972
* - Shallow events after 1972
1000
100
10
1
N
.88
1.15
1.09
.80
0
1
2
3
4
Mag

Fig. III-9

WOOD-ANDERSON CODA LENGTH MAGNITUDE STUDY

Introduction

This study was undertaken in order to compare, for earthquakes occurring in eastern Washington, the coda length magnitude formula presently being used with the local Richter magnitudes (M_L) derived from the Wood-Anderson standard torsion seismometers. The presently-used formula for computing coda length magnitude in both eastern and western Washington is

$$m = 2.82\ (\log_{10} C) - 2.46 \quad \text{where} \quad m = \text{coda length magnitude}$$
$$C = \text{coda length (in seconds)}$$

For coda lengths as defined for magnitude determination in eastern Washington, the coda (C) is the time interval, in seconds, from the initial P wave arrival until the trace amplitude returns to twice the noise level observed prior to the initial P wave arrival. In western Washington the coda length formula is the same; however, the length of the coda is defined as extending from the initial P arrival until the trace amplitude is the same as the background amplitude prior to the P wave arrival. What this means is that for the same trace of a recorded event, the coda length used for magnitude determination in western Washington would be somewhat longer than that used if obtaining a magnitude by the eastern Washington method.

A previous study (Malone and Bor, Geophysics Program's Annual Report for 1977) of the Richter M_L and coda length magnitude in eastern Washington concluded that the presently-used formula for western Washington was all right for use in eastern Washington. The report also mentioned that this conclusion was obtained with a limited amount of data.

The present study was accomplished to obtain a direct comparison between the Wood-Anderson standard torsion seismometer Richter M_L and the coda lengths scanned on eastern Washington's Central and Hanford arrays.

Method of Study

The procedure used in this study was to install two horizontal Wood-Anderson seismometers in eastern Washington and to obtain M_L from the local earthquakes they recorded. These instruments were first installed at Entiat in 1977, then at Richland in 1978. Each location was occupied for approximately one year. Table I shows the location and station abreviations for stations where Wood-Anderson instruments recorded the earthquakes used in this study.

Table I. Wood-Anderson Seismograph Station Locations

ENT	47°40'17" N, 120°13'24" W
RIC	46°20'50" N, 119°16'28" W
NEW	48°15'18" N, 117°07'12" W

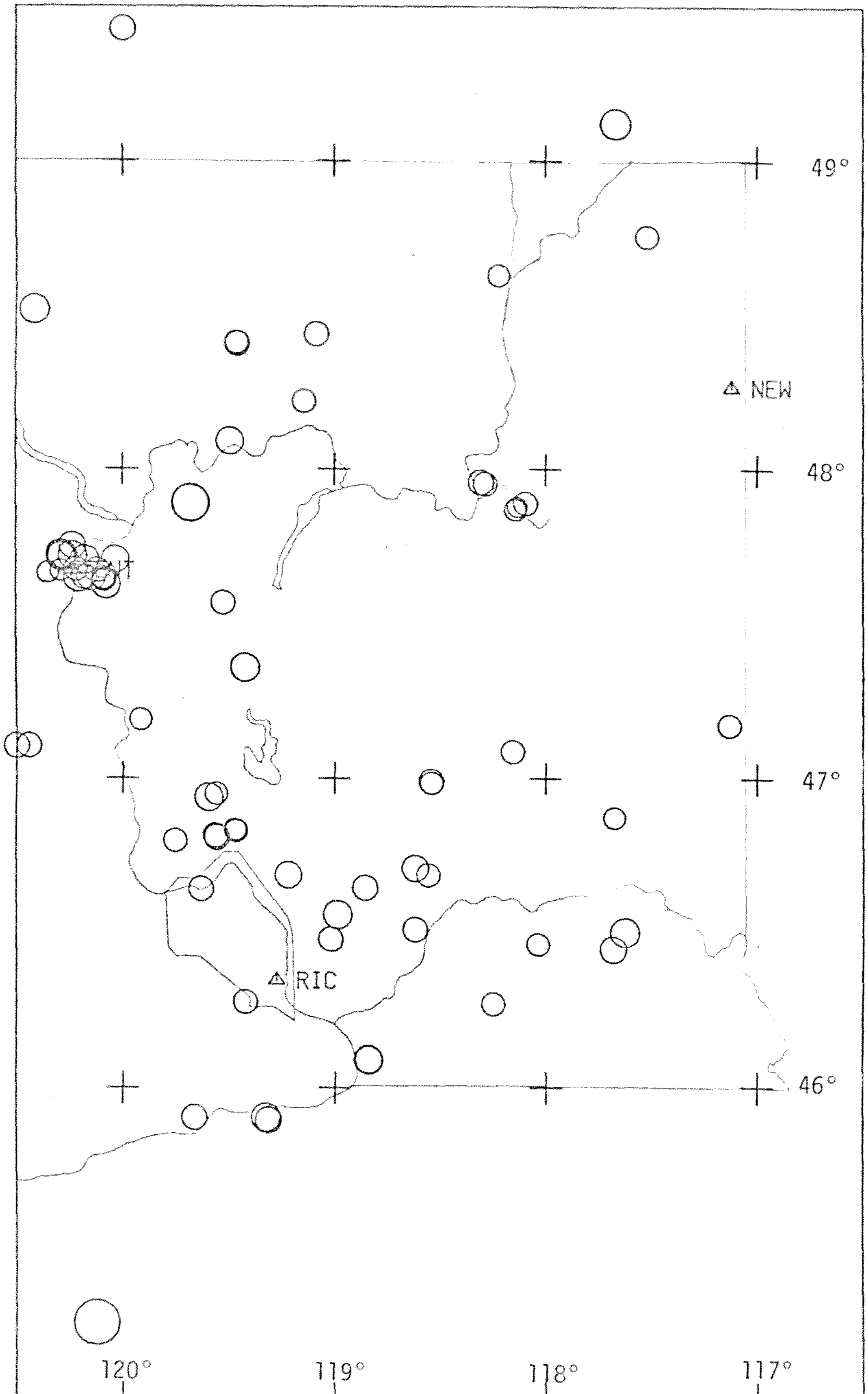
These seismograph systems needed 115 VAC to operate, plus housing from the weather and every-other-day record changes. Because of these requirements, the locations were subjected to cultural noise, thus requiring a maximum magnification of 5600. Because of the relatively few large earthquakes occurring in the area of study, the limited maximum magnification of the instruments, and the limited time, additional M_L data were needed. These data were obtained from the US Geological Survey's Newport Observatory located just north of Spokane. This observatory is in a remote, and therefore culturally quiet, area. The Wood-Anderson magnification is 70K in the winter and 140K in the summer. The events from the Newport Wood-Anderson instruments, used for 1975 through 1978, were those with hypocentral locations obtained from the eastern Washington arrays.

Eighty-six M_L magnitudes were computed, 12 from Entiat, 10 from Richland, and 64 from Newport. This study used 79 earthquakes. A map locating those earthquakes may be seen in Figure 1, and Table II lists the earthquakes where both coda length and M_L were computed.

Table II. Earthquakes Used in This Study

STA	MO	DY	YR	HR	MN	LAT		LON		M/P	M/R	CDA	ST	WAM	DST
NEW	09	18	75	12	19	47	47.8	118	13.9	3.5	3.4	124	10	2.6	100
NEW	10	11	75	00	50	48	26.2	119	05.1	2.3	2.5	058	08	2.0	150
NEW	12	31	75	03	23	46	41.2	119	13.1	2.7	2.7	070	10	2.3	240
NEW	04	13	76	00	47	45	14.6	120	07.3	4.6	4.8	381	09	5.1	410
NEW	05	15	76	13	04	47	42.4	120	02.0	3.0	2.7	067	09	2.4	230
NEW	05	20	76	00	34	47	05.2	118	09.3	2.1	2.3	048	13	1.8	150
NEW	06	15	76	01	01	46	26.9	117	41.0	3.0	2.6	064	10	2.3	205
NEW	07	02	76	01	36	46	59.1	118	32.4	2.7	2.4	053	09	1.7	180
NEW	07	09	76	20	35	47	57.5	118	18.3	2.9	2.9	080	11	2.1	095
NEW	07	10	76	17	54	46	59.4	118	32.5	2.9	2.6	064	08	2.1	180
NEW	07	26	76	20	13	47	57.2	118	17.1	2.4	2.7	071	11	1.9	095
NEW	07	28	76	18	55	47	11.3	119	54.9	2.3	2.2	045	09	1.6	245
NEW	07	30	76	01	53	46	57.0	119	33.5	2.7	2.5	057	07	1.7	235
NEW	08	05	76	01	38	46	27.8	118	02.3	2.8	2.1	040	08	1.7	215
NEW	08	07	76	20	31	46	52.4	117	40.6	2.1	2.0	037	05	1.6	165
NEW	08	17	76	21	44	45	54.3	119	19.5	2.4	2.0	039	10	2.6	310
NEW	08	20	76	00	01	47	05.2	120	56.8	2.3	2.3	049	18	2.5	325
NEW	08	30	76	16	34	47	39.0	120	10.6	3.0	2.7	070	10	2.0	240
NEW	10	31	76	21	39	46	00.7	120	50.2	2.9	3.4	106	06	2.2	380
NEW	11	05	76	01	07	47	10.3	117	07.9	2.4	2.5	059	13	1.8	120
NEW	01	27	77	07	47	46	56.3	119	35.6	3.2	3.1	095	25	2.5	240
ENT	03	07	77	04	45	48	30.9	120	24.8	2.8	2.5	057	14	2.7	095
ENT	03	09	77	10	44	47	39.0	120	12.3	2.5	2.0	039	13	2.8	010
NEW	03	31	77	22	29	45	54.2	119	39.7	3.1	2.4	058	12	2.1	330
ENT	04	08	77	21	53	47	39.1	120	08.6	2.1	2.0	037	12	2.6	010
NEW	04	14	77	19	55	45	07.5	120	52.8	2.8	2.9	077	10	2.6	455
NEW	04	21	77	07	23	49	07.3	117	40.1	3.6	3.1	093	11	2.9	105
NEW	04	28	77	20	33	46	16.2	118	15.0	2.1	2.2	045	11	1.8	240
NEW	05	02	77	02	22	48	45.2	117	31.2	2.1	2.3	049	07	1.8	065
NEW	05	07	77	15	13	47	38.6	120	05.2	2.1	2.1	040	12	1.5	240
ENT	05	07	77	15	13	47	38.6	120	05.2	2.1	2.1	040	12	2.0	015
ENT	05	09	77	23	50	47	43.1	120	14.0	2.0	2.3	050	04	2.6	010
NEW	05	10	77	22	13	46	16.6	119	25.4	2.1	2.2	045	10	1.9	285
ENT	05	12	77	09	48	47	40.3	120	07.1	1.6	1.5	025	04	2.0	010
ENT	05	17	77	04	14	47	39.8	120	20.9	1.5	1.5	025	09	1.5	010
NEW	05	18	77	20	06	46	30.2	117	37.6	2.3	3.2	106	15	2.7	200
NEW	05	23	77	11	49	49	25.6	119	59.8	2.2	2.5	056	07	2.1	250
NEW	05	29	77	12	32	46	41.2	118	33.4	2.1	2.2	047	12	1.8	210
ENT	06	11	77	16	24	47	40.5	120	13.1	1.8	1.6	028	08	1.9	005
ENT	07	13	77	07	15	47	03.6	120	57.2	3.8	3.8	164	10	3.0	040
NEW	07	13	77	07	15	47	03.6	120	57.2	3.8	3.8	164	10	4.0	320
NEW	07	13	77	08	00	47	35.5	121	42.2	2.4	2.3	049	05	2.2	355
NEW	07	13	77	12	49	47	04.2	120	59.2	2.2	2.2	044	13	2.2	325
NEW	07	13	77	14	33	47	03.9	120	59.1	2.4	2.4	052	14	2.6	325
NEW	07	23	77	19	17	46	37.5	120	54.5	2.0	1.9	035	05	2.0	345
NEW	08	29	77	12	40	47	43.1	120	17.1	2.8	2.4	052	13	2.3	250
ENT	08	29	77	12	40	47	43.1	120	17.1	2.8	2.4	052	13	3.0	010
NEW	09	11	77	04	13	46	38.5	119	37.7	2.7	2.3	050	15	2.0	255
ENT	10	09	77	01	16	47	40.2	120	17.3	1.6	1.7	030	10	1.5	005
ENT	10	22	77	14	18	47	42.3	120	10.6	1.7	1.6	028	10	2.6	010
NEW	10	23	77	15	18	47	06.1	120	30.0	2.2	2.6	061	19	2.1	290
NEW	11	08	77	12	11	47	07.8	120	44.0	2.2	2.2	045	22	2.6	305
NEW	11	30	77	23	30	46	47.9	119	45.1	2.0	1.9	036	09	1.8	260
NEW	12	28	77	13	40	47	06.2	120	26.5	2.1	2.2	046	13	2.0	285
NEW	01	22	78	00	52	46	33.6	118	59.1	2.2	2.5	056	09	2.6	275
NEW	01	28	78	13	40	47	06.2	120	26.5	2.1	2.5	058	12	2.0	285
NEW	03	04	78	19	47	46	05.4	118	50.4	2.8	2.4	053	09	2.6	275
RIC	03	04	78	19	47	46	05.4	118	50.4	2.8	2.4	053	09	2.4	045
NEW	03	16	78	22	28	47	34.2	119	31.5	2.3	2.4	052	12	1.9	200
NEW	03	22	78	03	08	48	05.4	119	29.5	3.3	3.2	099	13	2.4	180
NEW	04	06	78	21	41	47	37.6	120	04.6	2.2	2.4	054	10	2.7	235
NEW	04	07	78	21	14	45	53.8	119	18.8	1.7	2.2	043	07	2.2	310
NEW	04	16	78	19	45	47	45.1	120	14.0	3.3	3.0	087	13	2.3	245
NEW	05	01	78	22	58	45	38.7	118	51.2	2.0	2.3	050	07	2.1	225
NEW	05	02	78	21	04	46	30.7	118	37.1	3.0	2.4	053	08	2.0	225
NEW	05	22	78	23	22	48	13.2	119	08.5	1.9	2.1	041	12	1.8	150
NEW	05	25	78	23	29	48	24.6	119	27.7	1.9	1.8	033	09	1.8	175
NEW	05	28	78	02	33	46	48.9	119	33.7	2.0	2.0	038	12	2.0	250
NEW	06	07	78	01	54	45	50.0	119	28.2	2.0	2.2	044	06	1.6	245
NEW	06	07	78	02	50	46	49.8	119	27.9	2.0	2.1	041	05	1.7	245
NEW	06	11	78	02	43	46	50.5	120	57.3	3.1	3.1	097	08	2.1	335
NEW	06	13	78	22	24	48	24.3	119	27.5	1.8	2.1	041	09	2.0	175
NEW	06	27	78	02	18	46	51.4	120	57.6	3.9	4.1	214	23	3.8	335
RIC	06	27	78	02	18	46	51.4	120	57.6	3.9	4.1	214	23	3.5	140
NEW	08	18	78	22	40	47	53.4	118	05.7	1.6	2.2	046	05	1.9	085
NEW	10	10	78	12	04	47	21.5	119	25.2	3.2	3.3	110	17	2.6	205
RIC	10	10	78	12	04	47	21.5	119	25.2	3.2	3.3	110	17	2.6	115
RIC	10	19	78	16	11	46	48.5	119	33.3	2.1	2.3	047	17	2.1	055
NEW	11	08	78	23	35	47	52.2	118	08.6	1.8	2.0	037	06	1.6	090
RIC	11	14	78	22	05	46	28.8	119	01.0	1.7	2.3	047	05	2.0	025
NEW	11	24	78	17	53	46	56.0	120	56.8	2.0	2.3	050	11	2.4	330
NEW	12	11	78	23	34	47	52.7	118	08.1	3.1	2.8	071	12	1.3	090
NEW	12	21	77	00	27	48	37.6	118	13.3	2.0	1.9	036	11	1.6	090
RIC	12	26	78	15	57	46	42.5	118	37.1	2.1	2.1	041	05	2.3	065
NEW	01	19	79	14	55	47	53.4	119	40.8	4.1	4.2	236	13	3.8	200
RIC	01	19	79	14	55	47	53.4	119	40.8	4.1	4.2	236	13	4.0	175

STA = station; MO = month; YR = year; HR = hour; MN = minute; LAT = latitude; LON = longitude, M/P = coda magnitude published; M/R = coda magnitude rescanned; CDA = average coda in seconds; ST = number of stations used to compute rescanned coda magnitude; WAM = Richter local magnitude M_L from Wood-Anderson instruments; DST = distance in km from hypocenter to Wood-Anderson station.



EASTERN WASHINGTON EARTHQUAKES FOR MAGNITUDE CALIBRATION
CENTER OF MAP IS 47.25 N 118.50 W

MAGNITUDE KEY ○ 0.0 ○ 2.7 ○ 5.3 ○ 8.0

Figure 1

All 79 earthquakes used in this study were rescanned in order to obtain coda lengths. This was done to obtain more coda lengths per coda length magnitude computed and also so that one person would be scanning all the data using the same subjective weighting process in deciding each coda length.

After completion of rescanning and recomputing all the coda length magnitudes, it was observed that no significant bias was introduced relative to the original coda length magnitudes. Many coda length magnitudes were changed, although the change was relatively small, and statistically the number of events and the amount of increased magnitude was very close to the number of events and the amount of decrease in magnitude. Thirty-five events were raised an average magnitude of 0.26 and 32 events were lowered an average magnitude of 0.26. Figure 2 shows a plot of the number of events versus the magnitude changes.

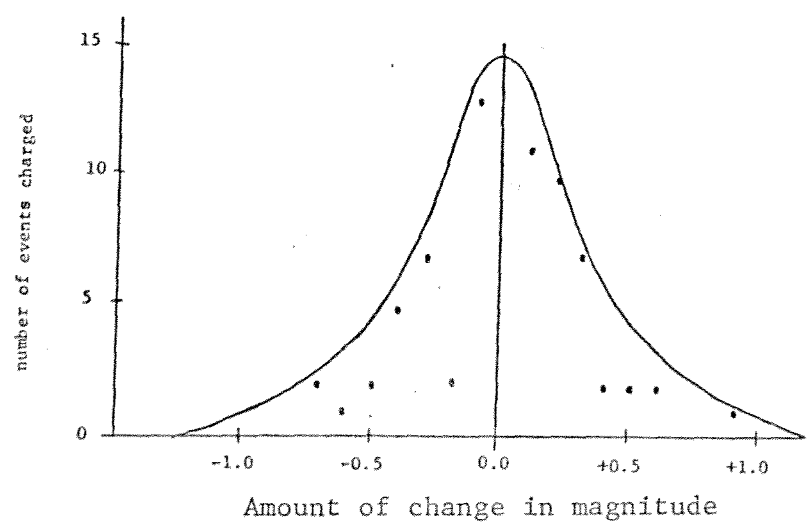


Figure 2. Plotted results of magnitude changes versus amount of change. The hand-drawn bell-shaped curve shows the similarity to a Poisson's distribution of these changes.

Results

All coda length versus M_L magnitude data were plotted and two straight dashed lines were drawn, one using a simple regression of x on y, then another using a simple regression of y on x. A solid line with an average slope of these two dashed lines and passing through the intersection of these dashed regression lines is the resultant solid line, which has the formula $m = 2.75 \pm 0.35 (\log_{10} C) - 2.59$ (see Figure 3).

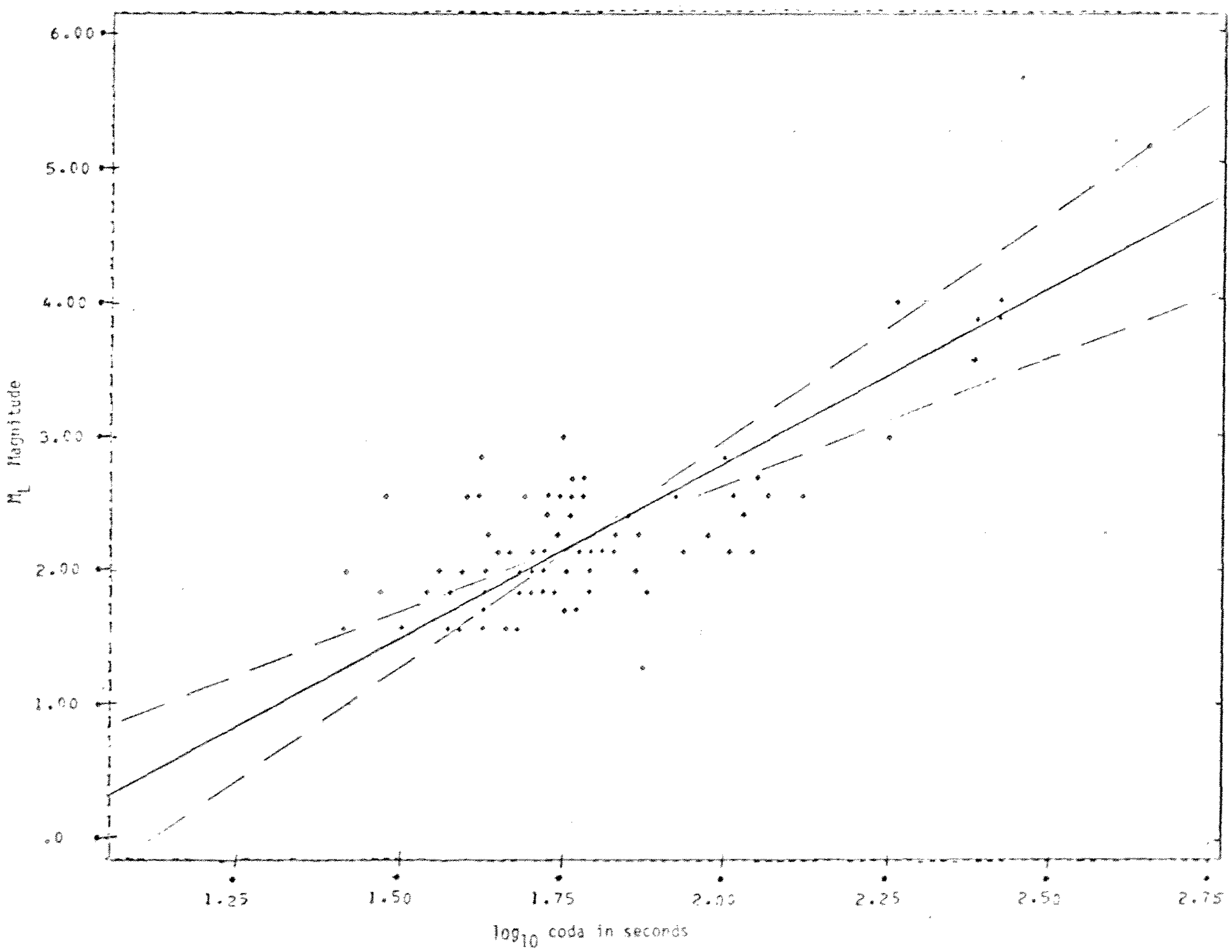


Figure 3. Plot of all data used in this study. Shown as the two dashed regression lines and the resultant solid line giving the formula $m = 2.75 (\log_{10} C) - 2.59$.

This formula, if used instead of the presently-used one, would lower the published magnitudes by about 0.3. There are, however, some considerations to be made before deciding to use this new formula. One is that there are only eight data points above an M_L magnitude 2.9. The relatively few points carry a good deal of weight in arriving at the slope of the new line. Another consideration is that the rescanned codas for these larger-magnitude events increased in all but three cases (two of the three had no change and one was reduced). This rescanning, therefore, actually tended to decrease the slope of the new formula more than if the original published coda length were used. A final consideration is that the new formula's slope has a possible error of $\pm$ 0.35, while the presently-used formula is within that error. See Figure 4 for a comparison of coda magnitudes.

Recommendations

While it appears that the presently-used formula is giving slightly higher magnitudes than the Richter M_L, more data from events greater than M_L = 3.0 are needed in order to calculate a better formula. Obtaining these data will take several years. One way to reduce the time interval is to carefully establish another permanent Wood-Anderson seismograph station of high magnification at a centralized location in eastern Washington. The presently-used formula is close to the new formula, however, and if the difference in magnitude of a few tenths is not significant to any present or anticipated investigation in the near future, the expense of a new Wood-Anderson station does not seem justified.

My recommendation is to continue gathering M_L data from Newport on all earthquakes over magnitude 2.9 until there are enough data points to reduce the error of a new formula.

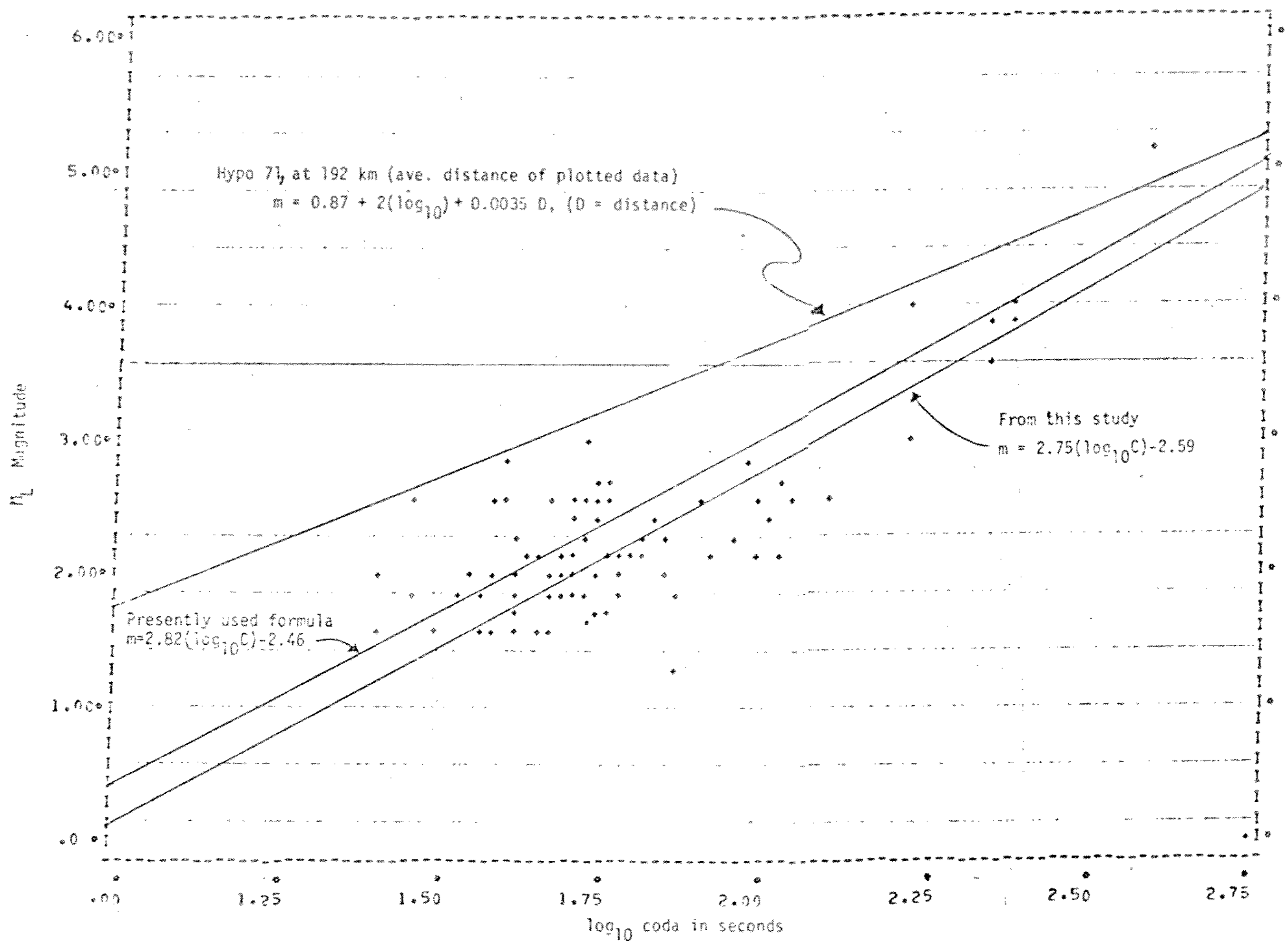


Figure 4. Graphic representation of Hypo 71 formula derived for California earthquakes, the presently-used formula, and the new formula obtained from this study.

V. Surface Wave velocity and Attenuation Study

Introduction. Because of its geographical and geological setting in the Pacific Northwest, the state of Washington has many intriguing tectonic problems. Questions regarding its western margin along the continental-oceanic boundary that are frequently asked are: Is the Juan de Fuca Plate still subducting? What is the regional stress field? What is the tectonic model that will explain the geologic and geophysical observations? The most interesting questins pertaining to the eastern part of the state relate to the origin of the Columbia River Plateau and its relation to the surrounding geologic regions. There is a great deal of published literature about these problems that either discuss observational evidence or models to interprete existing data. However, among all these published work a detailed study of the upper mantle structure , especially a reliably determined lithospheric thickness, is lacking. Since a clear picture of the physical conditions in the upper mantle will provide valuable insight toward a better understanding of the on going tectonic processes, it seems necessary to propose a research project to determine this vital information.

This report contains two main sections. In the first section, a general review of the studied area is given. The proposed method, its feasibility, resolution power and anticipated difficulties is discussed in section two, along with the specific progress made so far.

Geophysical Overview of the Washington State.

Topography and Geology. Physiographically Washington State can be divided into five regions. Starting at the Pacific, a major mountain range exists along the coast, which is loosely referred to in Oregon and Washington as the Coast Range. These mountains rise to their greatest elevation on the Olympic Peninsula in the Olympic Mountains. At the north end of the Olympic Peninsula these mountains drop abruptly into the Strait of Juan de Fuca. Inland from the coastal mountains lies a lowland, called the Puget Willamette Depression after the Puget Sound lowland in Washington and the Willamette River Valley in Oregon. East of the lowland rises the Cascade range, which extends from northern California into southern British Columbia. The Columbia River Plateau is east of the Cascade Range and north of the Blue Mountain Uplift of central Oregon. North of the Columbia River Plateau is a region with mountains and high plateau which extends continuously from the Interior Plains of Montana and Alberta to the Pacific Ocean. This region is referred to as the Okanogan Highland in Washington.

West of the Cascades lies the Puget Sound Basin which is thickly covered by quaternary glacial deposits in the center and is bounded by tertiary and older rocks on the Cascade and Olympic mountains. On the southeastern side of the high Cascades is the Columbia River Plateau consisting of a thick sequence of flood basalts, whereas Mesozoic-Tertiary granitic and metamorphic terrain comprises most of the Okanogan Highland.

Compressional wave velocity structure. Since 1957 there have been many published studies of crustal seismic velocity structure in the Pacific Northwest region. All investigations except for one concern the compressional wave velocity structure. Table 1 is a list of the various P wave velocity models in Washington and surrounding areas. Because of the ambiguity of the problem and the resolving power limitations by the methods and data used, the nature of the data and analyzing method is given in Table 1. Some features generally agreed upon are: (1) The crust in the Puget Sound lowland and Southwestern B.C. near Vancouver Island is quite thick (about 41 to 51 km) with relatively low Pn velocity(about 7.7 to 7.75 km/sec). (2) From Olympia to west central Oregon, the thickness of the crust is relatively thin(16 to 20 km) with low to normal Pn velocity (7.67 to 8.0 km/sec). (3) To reconcile refraction data, the crust beneath the Columbia River Plateau must either be anomalously thin, or unusually mafic, or both. (4) Beneath southeast British Columbia and the Okanogon Highland, the crust is of intermediate thickness (31 to 35 km) with normal Pn velocity (7.9 to 8.0 km/sec).

Seismicity. Most earthquakes in the state occur either in central Puget Sound or along the western flank of the Cascades. Epicenters are diffusely spread through these regions. The average depth of all located events is between depths of 20 and 30 km. (Crosson, 1972 and 1977). Composite focal mechanisms suggest a north-south compression which may be related in some complicated way to the regional stress field associated with the movement of the American Plate with respect to the Pacific Plate. In eastern Washington an area just to the south of Lake Chelan is quite active and isolated swarms of earthquakes occur in the central Columbia Basin. The focal depths of the majority of these earthquakes are shallow (less than 6 km deep). About ten per cent of the central basin earthquakes occur at greater depths, which apparently are unrelated in space or time to the shallow swarms. Results of composite focal mechanism indicate the axis of maximum compression is oriented roughly in a north-south direction (Malone et al., 1975).

Seismic intensity data. General features that can be identified from published isoseismal maps are the complexities in the western side of the Cascades and an asymmetric pattern that is skewed toward the east. The former reflects the complexities of the local geological conditions in the Puget Sound lowland, whereas the latter is a consequence of the regional difference in the crustal attenuation factor (Malone and Bor, 1979).

Gravity data. A compilation of gravity anomaly data (free-air at sea, Bouguer on land) (e.g. Riddihough, 1979; WPPSS Gravity Map, 1977; Woollard and Rose, 1963) reveals interesting features that may bear evidence for the regional tectonic processes (e.g. Riddihough and Hyndman,1976; Riddihough,1979). General features of the gravity map include: (1) A linear negative free-air anomaly (about -25 to -75 mgal) located about 50 to 150 km offshore; (2) A parallel, linear positive Bouguer anomaly (about 25 to 50 mgal) at about 50 km onshore, except for a region of gravity low (about -75 mgal) centered on the Olympic Mountains; (3) Steep gradients and large amplitude negative anomalies (-70 to -110 mgal) in the Puget Sound lowland; (4) A broad Bouguer maximum (-60 mgal) near the Hanford area in the Columbia River Plateau; (5) A tongue of maximum negative anomaly (-110 to -160 mgal) extending from the northeastern Okanogan Highland to the southwestern part of the Cascades.

<u>Heat flow data</u>. Although the measured data points are sparse in this area, they still represent regional characteristics of heat flow. Heat flow values are below normal in the Puget Sound region (0.83 to 1.25 HFU) (Sass et al., 1971) and high in eastern Washington (1.3 to 1.6 HFU) (Roy et al., 1968; Blackwell, 1969 and 1971). The values in the Columbia River Plateau are less than those measured in the Basin and Range and the Northern Rocky Mountains provinces (about 2.0 HFU). This suggests a difference of the crust and upper mantle underneath these two regions.

<u>Tectonic implications</u>. To summarize the hypotheses and models that have been proposed to explain the tectonic setting of Washington State only the major references will be mentioned. According to Atwater's (1970) model, in which magnetic anomaly patterns were used to outline the history of plate interactions at the western margin of North America, the Pacific margin of North America has evolved from a continuous subduction zone in the early Tertiary to the present state of alternating subduction

and transform interactions. Crosson (1972) studied local earthquakes and crustal structure in the Puget Sound area, and found no distinct Benioff zone underneath this region. He concluded that crustal consumption may have ceased. By using pattern recognition techniques, Riddihough and Hyndman (1976) investigated a large set of geological and geophysical observations and concluded that, although the pattern is complex and changing, convergence has continued to the present day south of 50 N. Riddihough (1979) tried to reconcile the gravity and seismic data near Vancouver Island and concluded they were consistent with the active subduction hypothesis. From comparison of geodetic leveling and seismic data between western Washington and Shikoku, Japan, Ando and Balazs (1979) concluded that aseismic subduction is taking place beneath western Washington.

There are three primary interpretations concerning the problem of the origin of the Columbia River Plateau. Scholz et al. (1971) proposed a back-arc spreading model, in which thermal energy originating near the top of a subducting plate mobilizes the overlying mantle, producing a thermal diapir, extensional strain, and partial melting, all culminating in the eruption of basalt. Other recent models rely on interaction of the North American lithosphere with primary upwellings in the mantle. Morgan (1972) suggested primary mantle upwellings in the form of deep mantle plumes to account for a principal driving force for plate tectonics. Other authors have postulated current tectonic models in this area based on the "hot spot" hypothesis. The third opinion is introduced by Simpson and Cox (1977). Based on paleomagnetic data they found that the Oregon Coast Range has rotated clockwise relative to North America. It is this rotation which provides extensional strain and room for the passive upwelling of basaltic magma.

Proposed Study

The previous discussion delineates the diverse and distinct physiographic, geologic, and geophysical characteristics of Washington state. This study will focus on determining the average crustal seismic velocity models and upper mantle structures east and west of the Cascades. Since shear wave propagation is more sensitive to changes in physical conditions in the medium, major questions raised concerning this study are: (1) What is the vertical distribution of S wave velocity in the crust and upper mantle? -- are there positive or negative gradients (i.e. low velocity layer) with depth. (2) How pronounced are the lateral variations in S wave velocity in this region? (3) Does the structure of S wave velocity closely correspond to that of P wave velocity? Or how does Poisson's ratio vary with depth? (4) How does the attenuation (Q) vary with depth as well as with horizontal distances? (5) Is there any correlation between the Q structure, P wave and S wave velocity structure? If any, can they be used to define the lithospheric thickness in this area? (6) Can we relate the focal depth distribution of earthquakes to the variation of the elastic and inelastic properties with depth? (7) Based upon the answers to the above questions, what implications can be made about the tectonic history of this area?

<u>Proposed approach</u>. The determination of shear wave velocity structure by body wave data depends upon accurately determined S wave arrivals. The S wave is often difficult to pick accurately because it is a later arrival on the seismogram and is usually obscured by earlier phases. This situation is even worse if the structure or source is complicated. If we had many 3-component stations to record seismic signals the onset of shear wave energy would be easier to discern. Since the cost is not economically feasible for a large array the proposed approach is to use three long period seismograph stations for a surface wave study. These stations would be deployed along a great circle which, hopefully, will be aligned with several major earthquake source regions in order to collect localized Rayleigh and Love wave phase and group velocity data as well as amplitude information. The collected data will then be used to construct the regional average S wave velocity and Q structure.

In order to resolve upper mantle structure and still obtain a meaningful regional, average crustal structure, the array dimension should be comparable to the maximum desired penenetration depth. Three station sites have been selected according to the above requirement These three stations along with the WWSSN station at Longmire form two tripartite arrays. The advantage of using the Longmire station is that the tripartite method can be applied to obtain structures within the array eliminating the need for the source to be exactly aligned with the receivers. Hopefully, the advantage will shorten the time for data acquisition, while the other three temporary stations can be used to collect data for the attenuation study by the two-station method.

The station coordinates are listed in Table #.

The greatest difficulty in this study is the installation of the instrument and data acquisitiion in the field. The anticipated difficulties in analyzing the data are those typical of surface wave studies in general, such as non-least time arrivals, higher mode contaminations, large uncertainties in amplitude measurements, non-uniqueness of the inversion processes. These difficulties can be reduced to a minimum if proper methods are carefully applied.

Current Progress. Figure 1 is a shematic work chart for this study. Lab tests of instruments have been conducted since September 1978. These include: (1) Modify digital event recorders for long period data acquisition. (2) Repair and test three 3-component long period stations in the lab. (3) Design and build external circuit components for the long period seismographs. (4) Calibration of the whole system in the lab.

In January and February 1979, stations were installed at Mt. Constitution and Liberty and have been in operation since. A temporary station in the lab at the University was also set up in March and has been operating intermittently since. In August this station will be moved to a tentative site at Walla Walla in eastern Washington. Already five large earthquakes have been recorded, all of which were either from Alaska or Mexico and, fortunately, are on the same great circle as the stations. The performance of the digital event recorders has been unsatisfactory because of the nature of the triggering mechanism and general mechanical failures and a lack of recording capacity. Fast tape speeds are not designed for long period data acquisition. Consequently, it has been arranged to borrow 3 complete analog tape recorders from the US Geological Survey. This equipment has been recently tested and installed at the MCW and LIB stations. The next phase of the field program will be mainly using the analog tape recorders to collect data, which hopefully will record events with greater epicentral distances.

Table VI-1. Compressional wave velocity structures in Washington and surrounding areas

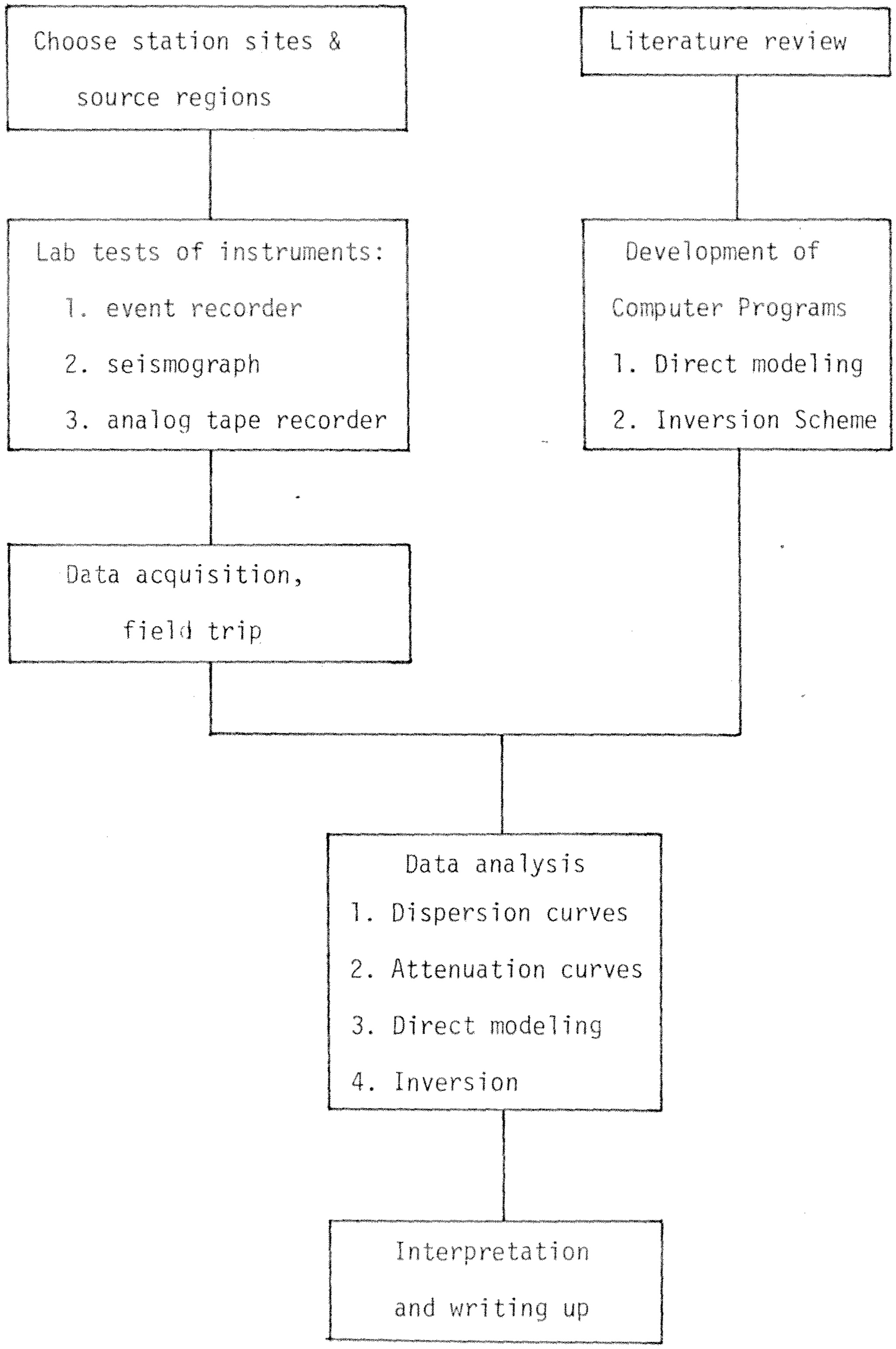
Area	Sedimentary layer (Upper layer)		Intermediate		Moho		Method	Reference
	V_p (km/sec)	h (km)	V_p	h	V_p	h		
Wash. & Oregon (W of Cascades)	5.48	--	6.61	--	7.67	30(?)	Body wave local earthquakes	Dehlinger *et al.*, 1965, 1968
Wash. & Oregon (W of Cascades)	5.5	5	6.1-6.6	35	7.67	35-	Surface waves	Chiburis, 1966
Wash. & Oregon (Olympia to W Central Oregon)	4.4-5.5	10	6.6-7.4	16	8.0	16-	Refraction	Berg *et al.*, 1968
SE B.C. to W Wash.	lack		6.30	32	7.91-8.06	32-	Refraction	Johnson & Couch, 1970
Puget Sound	5.36	10	6.61-7.16	41	7.75	41-	Local earthquakes	Crosson, 1972, 1976 McCollen & Crosson, 1975
West B. C. (Vancouver Is.)	3-6.4	6	6.8	51	7.7	51-	Refraction	White *et al.*, 1965, 1968
Wash. & Oregon (E of Cascades)	5.53	--	6.60	--	7.96	35-40	Body wave local earthquakes	Dehlinger *et al.*, 1965, 1968
Wash. & Oregon (E of Cascades)	5.60	5	6.10-6.60	40	7.96	40-	Surface waves	Chiburis, 1966
Columbia Basin			6.2	5?-18	7.9	18-	Refraction	Hill, 1972

Table VI-1, Cont.

Area	Sedimentary layer (Upper layer) V_p (km/sec)	Sedimentary layer (Upper layer) h (km)	Intermediate V_p	Intermediate h	Moho V_p	Moho h	Method	Reference
Columbia Basin	3.7-5.15	1.2	6.05-7.2	28.0	8.0	28-	Local earthquakes and blasts	Malone, 1977
Northeast Wash. (Okanogan High-land)			6.0-6.6	34	7.9	34-	Refraction	Hill, 1972
Northeast Wash.	5.1	0.5	6.05-7.2	24.5	8.0	24.5-	Local earthquakes and blasts	Malone, 1977
Southeast B. C.			6.1-6.4	31	7.8	31-	Refraction	White et al., 1965
Snake River Plain	5.2	10	6.7	45	7.9	45-	Refraction	Hill & Pakiser, 1966

Table VI-2. Station coordinates of the array

Station Name	Code	Latitude (N)	Longitude (W)	Elevation (m)
Mt. Constitution	MCW	48°40'47"	122°49'56"	693
Liberty	LIB	47°16'48"	120°38'53"	975
Walla Walla	WLA	46°04'48"	118°23'24"	~
Longmire	LON	46°45'00"	121°48'36"	854
Seattle	SEA	47°39'18"	122°18'30"	30



Choose station sites & source regions
Literature review
Lab tests of instruments:
1. event recorder
2. seismograph
3. analog tape recorder
Development of Computer Programs
1. Direct modeling
2. Inversion Scheme
Data acquisition, field trip
Data analysis
1. Dispersion curves
2. Attenuation curves
3. Direct modeling
4. Inversion
Interpretation and writing up

Fig. VI-1.

VI. A Refraction Study of the North Cascades and Northeastern Washington

This study outlines an interpretation of arrival times from blasts and earthquakes to indicate average crustal velocities, layering, and thickness in the Cascade Range and northeastern Washington. A previous refraction study in the north Cascades (Johnson and Couch, 1970) showed a crustal thickness of about 30 km with an average crustal velocity of 6.1 km/sec. The data analyzed in this study indicate an average crustal velocity of 6.4 km/sec, with a crustal thickness of nearly 40 km in the north central Cascades. To the east of the north Cascades the average crustal velocity may be slightly lower. The crustal thickness in northeastern Washington is about 30 km (Hill, 1972).

Crustal Arrivals. A number of quarry explosions have been observed and timed for various paths through and across the Cascade Mountains (Figure VI-1). Both crustal arrivals and, farther out, mantle arrivals are used to study the structure and thickness of the crust. Arrival times corresponding to an apparent velocity of 6.5 ± .2 km/sec were picked from the first observed motions for blasts at Princeton Canada. These were recorded along the western margin of the Cascades as far south as Mt. Rainier (Figure VI-1). Arrival times observed along the eastern edge of the Cascade Range and in northeastern Washington can be fit by a velocity of 6.3 ± .2 km/sec. The arrival times for both regions are plotted in Figure VI-2. The combined data set is fit by

$$T = 1.3 \pm 0.7 + X/(6.4 \pm 0.1 \text{ km/sec})$$

The Centralia Mine blasts were recorded from the southern Puget Sound region across the Cascades and into northeastern Washington (Figure VI-1). The observed first motions correspond to velocities less than 6.5 km/sec. The velocity computed using just the distant stations in northeastern Washington is only 6.1 $\pm$ .14 km/sec (distance range 180-320 km). Using the stations in the Cascades and some of the western Washington stations (Chou and Crosson, 1978), we find a travel time equation of

$$T = 2.5 \pm .17 + X/(6.3 \pm 0.05 \text{ km/sec})$$

These arrival times could be produced by a 6.3-6.4 km/sec layer at depths between 10 and 20 km. This velocity is somewhat low to be considered as a refraction at the Conrad Discontinuity (velocities in this lower crustal mafic layer are usually 6.5 - 7.0 km/sec).

Extensive refraction data in British Columbia has been summarized and re-analyzed by Berry and Forsyth (1975)., who suggest that the large amplitude seismic waves observed after the Moho refraction Pn are due to a guided wave in the crust. This multiply- reflected and refracted energy, ($P_M P$), propagates with an apparent velocity which approximates the average velocity of the crust (hence the notation $\overline{P}$). The velocity that is observed in British Columbia is 6.4 km/sec. A refraction study in northeastern Washington interprets large amplitude emersive arrivals ($\bar{P}$) with an apparent velocity of 6.2 km/sec as such a guided wave (Hill, 1972).

The least-squares fits to the Princeton and Centralia data indicate that the arrival times have an r.m.s. residual of 0.5 sec, while the calculated standard error of the data is less than about 0.2 sec. These data have been grouped into restricted azimuthal ranges, but cannot be considered linear profiles. The large scatter in the data may reflect both lateral velocity variation and uncertainty in the picking of this complex wave train.

Although the standard errors of the velocity are large, there is a trend toward lower apparent velocities in northeastern Washington as compared to those of the Cascades. (Because of different crustal velocities in the Cascades and in Puget Sound, the arrivals in the Cascades from the Centralia Mine may be biased. The apparent velocity predicted by the Puget Sound model (Crosson, 1976) is 6.6 km/sec in the distance range corresponding to the stations in the Cascades (Crosson, 1978).)

Layered Velocity Models of the Crust. A separate time term study of Mount Baker determined a velocity of 6.1 km/sec in the north Cascades. A velocity of 6.1 km/sec is also determined from a time term study of northeastern Washington by Malone (1977). The same velocity is found by Johnson and Couch (1970) in the north Cascades, and by Unger (1972) near Mt. Rainier. The arrival times for the Canyon Lake blast and the Columbia Cement blasts near Mt. Baker fit this velocity out to 170 km. All these data indicate a depth to this horizon of less than 2 km.

Distant arrival times observed in the north Cascades from the Centralia Mine fit an apparent velocity of 7.0 ± 0.1 km/sec (distance range 190-260 km). The intercept time is calculated as 4.2 ± .5 sec, implying an average depth to this horizon of 25 ± 3 km. These arrivals may be refractions (P_I) from the top of an intermediate velocity lower crust (ie. Conrad discontinuity). The depth was calculated assuming an upper crustal velocity of 6.1 km/sec. This estimate may be biased by the differences in upper crustal (0-20 km) velocity structure, although lower crustal velocities of the two regions are comparable. (In this distance range, apparent velocities of 7.8 km/sec are predicted by the Puget Sound model. The observed arrivals may be secondary arrivals.) Secondary arrivals from the blasts near Yakima (Figure VI-1) are fit by a velocity of 6.8 km/sec with an intercept time of 3.5 sec. The depth to this horizon is 24 km. This event's origin time was computed from an earthquake location routine (see below). Johnson and Couch (1970) tentatively identified secondary arrivals from the Greenbush Lake shots (Figure VI-1) with an apparent velocity of 6.9 km/ sec. The intercept time of 3.1 sec implies a depth to this horizon of 20 km.

The arrival times of impulsive secondary arrivals for an earthquake near Lilliwat, British Columbia and from first arrivals from another earthquake near Nachez, Washington (Figure VI-1) fit a velocity of 6.9 km/sec in the north Cascades region (epicenter-receiver distances 180-290 km). Apparent velocities of regional earthquakes recorded at Mt. Rainier are 6.5-6.8 km/sec.

Refracted arrivals in northeastern Washington are observed to fit a velocity of 7.2 km/sec for the Midnight Mine blasts (Malone, 1977). Secondary arrival times observed from the Greenbush Lake explosion indicate a velocity of 6.6 km/sec (Hill, 1972). The depths determined by these two studies were 19 and 22 km, respectively.

To summarize the results so far, the Cascade Range from the Canadian border to Mt. Rainier can be modeled by a 6.1 km/sec upper crust which is 20-24 km thick. A lower crust of velocity 6.5 to 6.8 km/sec is indicated beneath Mt. Rainier and 6.8 km/sec in the north Cascades. The velocities in the lower crust of northeastern Washington are comparable to those observed in the Cascades.

<u>Mantle Headwave Arrivals</u>. The crustal thickness of the Cascades and northeastern Washington can be determined using the observations of Moho refractions, Pn, and the crustal velocity models determined above. Johnson and Couch (1970) obtained ten Pn arrival times in the north Cascades from shots at Greenbush Lake in British Columbia, and obtained velocities of 7.91 ± 0.15 and 8.06 ± 0.06 km/sec along two linear profiles. The values predicted by these two equations are not significantly different in the range 200 to 400 km. The combined data set is fit by the equation

$$T = 6.86 \pm .39 + X/(8.03 \pm 0.05 \text{ km/sec})$$

The average crustal thickness, determined using a crustal velocity of 6.4 km/sec, is 37 ± 2 km. The crustal thickness beneath Greenbush Lake is about 36 km (Berry and Forsyth, 1975).

The Columbia Cement blasts produced three repeatable observations of Pn along the western margin of the Cascades as far south as Mt. Rainier. The travel time equation calculated is

$$T = 7.4 \pm 0.9 + X/(7.9 \pm 0.3 \text{ km/sec})$$

(In this equation the standard errors are those predicted by the standard deviations of the data, about 0.2 sec. The fit to the three points indicates an r.m.s. misfit of only 0.05 sec.) When the average crustal velocity of 6.4 km/sec is used, the intercept time gives an average crustal thickness of 40 ± 5 km along the western margin of the Cascades.

A blasting operation near Yakima was used to obtain five Pn arrivals (Figure VI-3) in the north Cascades. A Pn velocity of 7.8 ± .2 km/sec is determined. No origin time was measured, but the origin time computed by an earthquake location process indicates an intercept time of 6.8 ± 0.7 sec. The blast was well located near its known site and the nearest station was 17 km distant. An average crustal velocity of 6.4 km/sec gives a crustal thickness of 37 ± 3 km. The crustal thickness near Yakima is probable less than this value, due to its proximity to the Columbia Plateau. Crustal thickness there is estimated as 25-30 km (Hill, 1972, Malone, 1977). A thicker crust is implied for the north Cascades .

Two observations of Pn in the north Cascades were observed for two blasts from the Merrit Mine (Figure VI-1). The intercept times calculated, using an assumed Pn velocity of 7.9 ± 0.1 km/sec, are 6.7 ± 0.4 sec. The average crustal thickness is interpreted to be 37 ± 4 km. (The crustal thickness beneath Merrit is about 33 km (Berry and Forsyth, 1975). A crustal thickness of 40 ± 4 km is implied in the north Cascades.) Impulsive Pn arrivals were obtained from an earthquake near Lilliwat, British Columbia and are fit by a velocity of 7.95 km/sec.

Refracted and reflected arrivals from the Moho in northeastern Washington are observed (Malone, 1977) for blasts at Grand Coulee Dam and the Midnight Mine (Figure VI-1). Fitting these arrivals (distance range 70-180 km) to a straight line yields a velocity of 8.00 ± .06 km/sec and intercept time of 5.5 ± .1 sec. These values may be biased, because within distances of about 90 km the arrivals must be reflections. An average crustal velocity of 6.2 km/sec is used to compute an average depth to the Moho of 27 ± 1 km. The reflected arrival at 70 km gives a depth of 25 km. A slightly lower mantle velocity is indicated because this fit includes reflected arrivals.

A Pn velocity of 7.9 km/sec is assumed in the interpretation of refraction data analyzed by Hill (1972) from the Greenbush Lake explosions. However, the data from northeastern Washington and from south of the Columbia Plateau fit a velocity of 8.0 km/sec. The apparent velocity in northeastern Washington is higher than 8.0, which is interpreted as indicating a shoaling of the Moho from 34 km near the British Columbia-Washington border to about 29 km at the northern edge of the Columbia Basalts (roughly near the Grand Coulee blasts). Mantle velocities beneath the Columbia Basalts are 8.1 km/sec, as determined by McCollom and Crosson (1975).

Summary and Discussion. The crustal models determined in this study are compared to models of adjoining regions in Figure VI-4. The velocity of the upper crust in southern British Columbia, northeastern Washington, and in the Cascades as far south as Mt. Rainier is 6.1 km/sec.

The surficial geology near Mt. Rainier evidences widespread Tertiary andesitic flows and Tertiary grainitic intrusions. These intrusions are exposed irregularly into the north Cascades where Paleozoic and Mezozoic meta-sedimentary rocks have been intruded. In northeastern Washington, Mezozoic granitic intrusions predominate over Paleozoic and Mezozoic meta-sediments. The 6.1 km/sec velocity can be associated with either granites or meta-sediments. In the Puget Sound region, Tertiary volcanic and sedimentary rocks are associated with the 5.4 km/sec upper crust (Crosson, 1976). A sharp transition from the Puget Sound to the Cascades upper crustal structure is required.

The middle crust of the Puget Sound, Cascades and northeastern Washington areas appears to be continuous, although this higher velocity material is present at shallower depths in

Puget Sound model (Crosson, 1976). The velocity in the lower crust has been attributed to the presence of granulite facies rocks (Crosson, 1976).

The mean crustal velocity determined from crustal guided waves is 6.4 km/sec in British Columbia and the north Cascades. It is possibly lower (6.2-6.3 km/sec) in northeastern Washington.

The mantle velocity in the Cascades is not well determined. In Puget Sound, mantle velocities are anomalously low, 7.7-7.8 km/sec (McCollom and Crosson, 1975; Crosson, 1976). In British Columbia, mantle velocities are interpreted to be 7.8 km/sec south of 51° N (Berry and Forsyth, 1975). The mantle velocity in the Columbia Plateau area is higher, about 8.1 km/sec (McCollom and Crosson, 1975).

If a mantle velocity of 7.8 km/sec is assumed, the crustal thickness determined using average crustal velocities from the Greenbush Lake explosions is inconsistent with the depth determined from the Columbia Cement blasts. Assuming a velocity of 8.0 km/sec implies a dip of 2° to the northwest for the Yakima blasts. A thinner crust in the south Cascades (35 km) and a thicker crust (45 km) in the North Cascades results.

A mantle velocity of 7.9 km/sec and average crustal velocities are used to compute crustal thicknesses in figure VI-5. Estimates based on layered crustal models are generally in agreement with these values. The crustal thicknesses are plotted for representative positions of the offset points of refraction (offset 45 km). An east-west cross section (Figure VI-6) shows the crustal thickness decreasing gradually from 40 km in the Cascades to 30-35 km in northeastern Washington. The thick crust in the Cascades is consistent with the Puget Sound model (Crosson, 1976), but does not agree with the thickness of 25 km found in southwest British Columbia (Berry and Forsyth, 1975). The crustal thickness is interpreted to be 40 km at a point 45 km north of Mt. Rainier. If a lower mantle velocity is assumed, this depth maybe 5 km shallower, but is certainly not as shallow (25 km) as Unger (1972) suggested.

Mantle velocities are required to change from the low values determined to the north and west of the Cascades to higher values to the east of the Cascades. The crustal thicknesses determined in this study can only be considered accurate to about 3-5 km until better information on the transitional mantle velocity in the Cascades is obtained.

The Bouguer gravity low (Figure VI-7) in the Cascade range indicates a crustal thickness of greater than 40 km, and the average elevation of the range implies a depth of 40 km, assuming the region is in isostatic equilibrium (Johnson and Couch, 1970). The upper crust is interpreted to cause the high gradient of gravity along the western edge of the Cascades. The increase in gravity in the south Cascades and the Columbia Plateau may be due to crustal thinning or higher densities in the crust. The slightly higher gravity in northeastern Washington is interpreted to be caused by a thinner crust.

We are presently exploring techniques to include the regional structural differences obtained by this study in our routine earthquake location procedure. It is particularly difficult to include different velocity models for different stations recording the same event. This situation is typical of the few earthquakes which do occur in the Cascades.

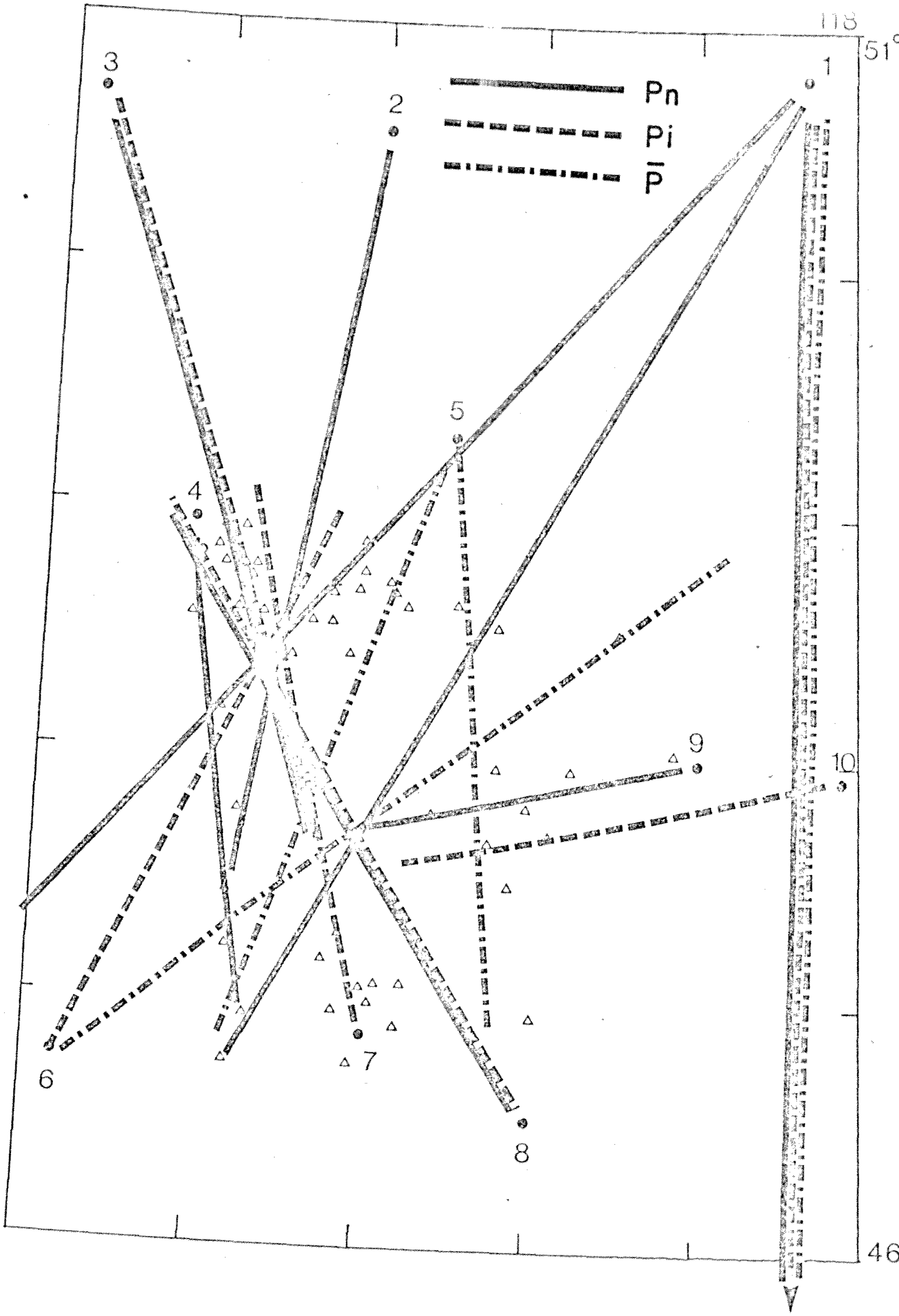


Figure VI-1. Raypaths; 1=Greenbush Lake, 2=Merrit, 3=Lilliwat EQ, 4=Columbia Cement, 5=Princeton, 6=Centralia, 7=Nachez EQ, 8=Yakima, 9=Grand Coulee, 10= Midnite Mine.

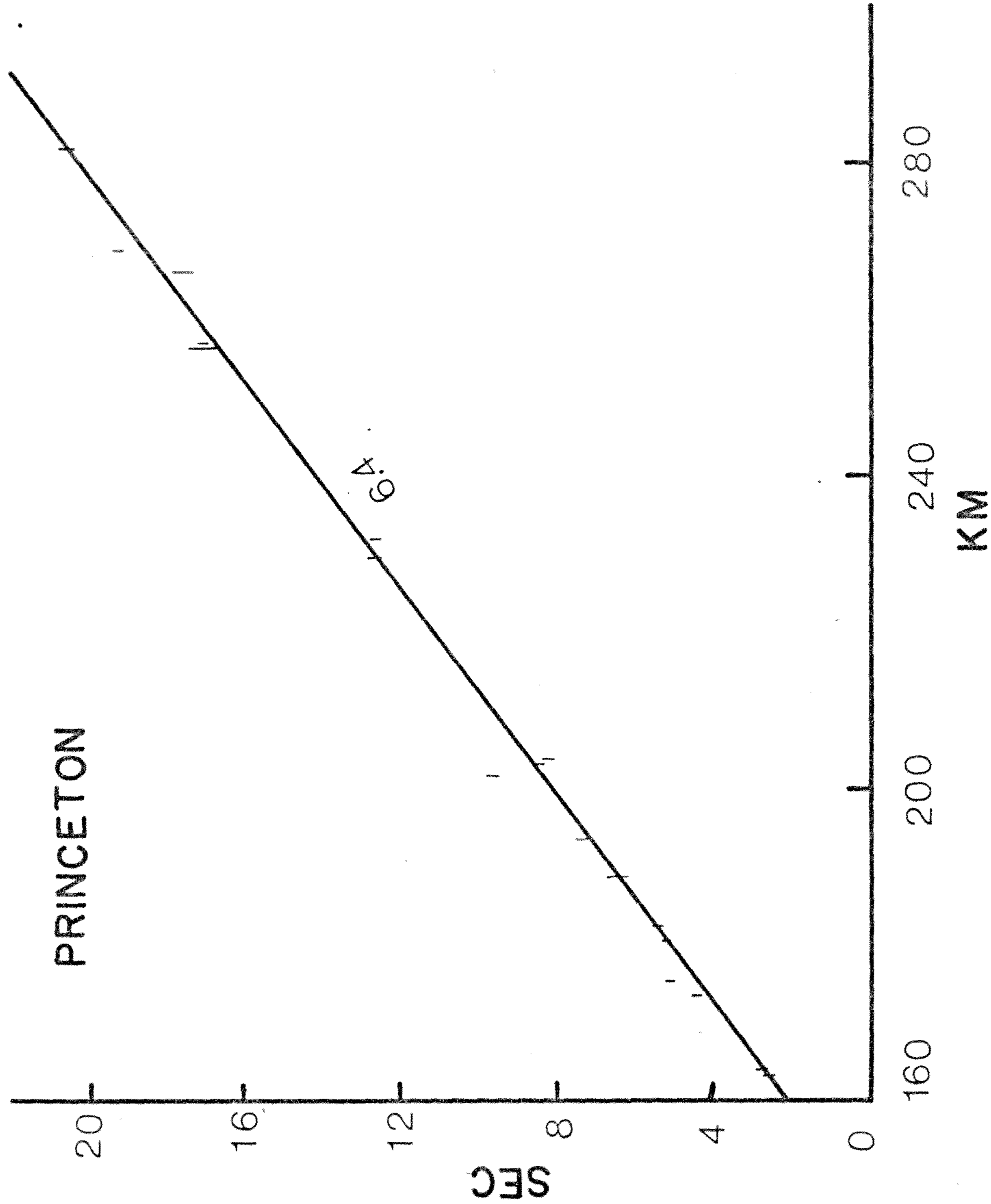


Fig. VI-2

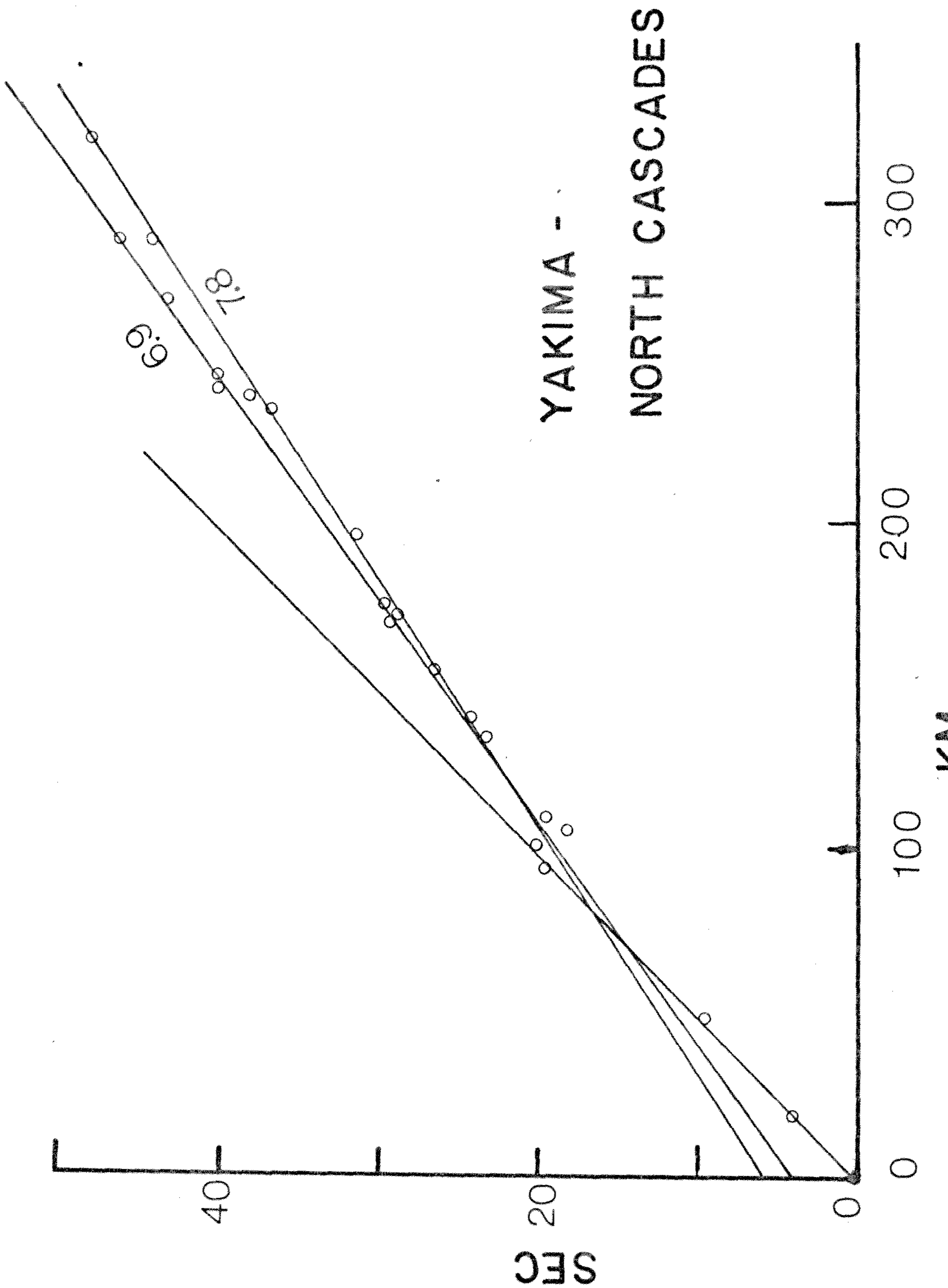
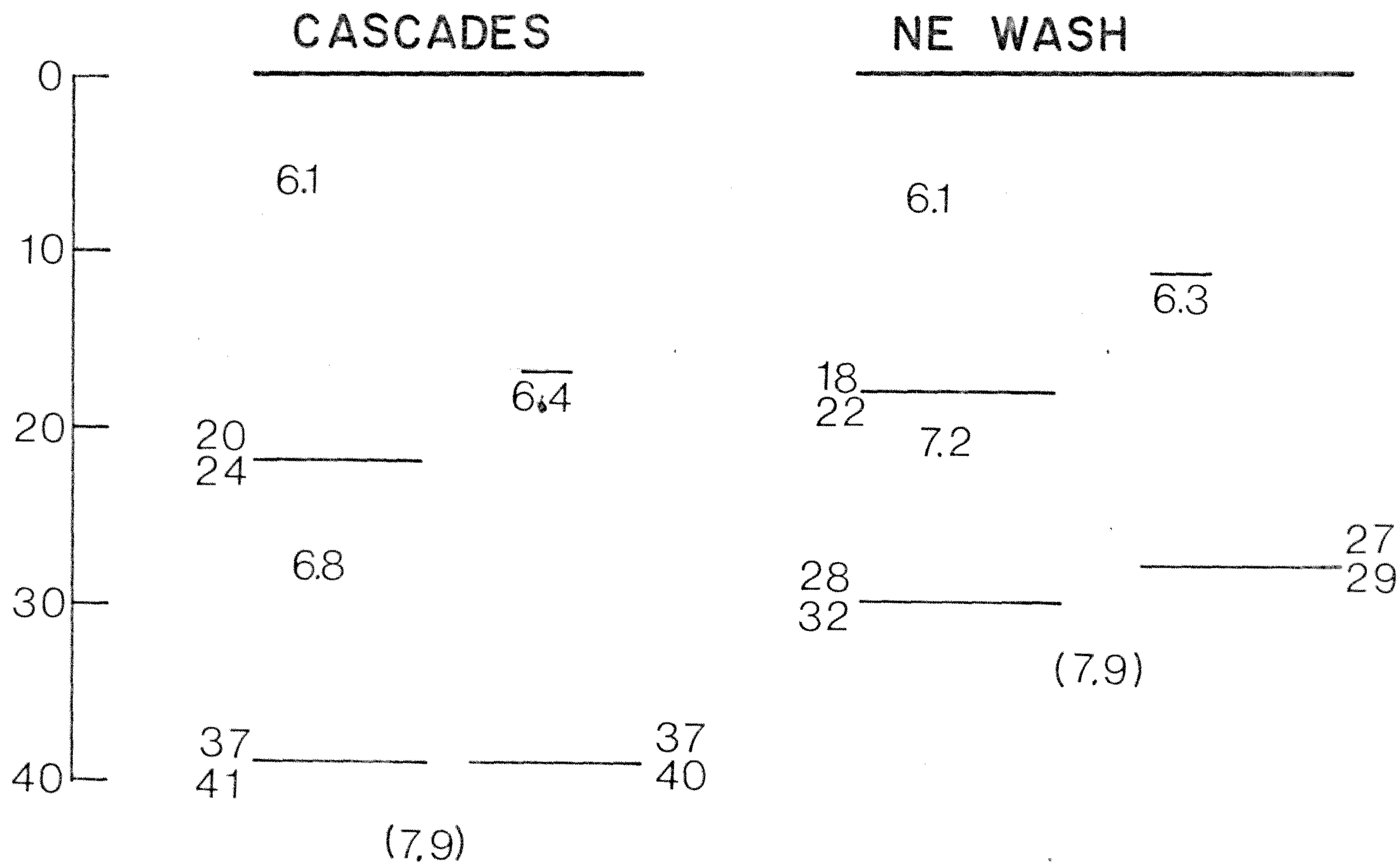


Fig. VI-3


CASCADES
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40
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18
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7.2
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32
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29

Fig. VI-4

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