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I. INTRODUCTION AND OPERATIONS

Introduction

This report covers the operation and research performed for D.O.E and the N.R.C. by the University of Washington Geophysics Program on the seismicity and structure of eastern and southern Washington during the past year. These contract help support parts of the *Washington state regional seismograph network.* There are presently 114 stations in Washington and northern Oregon whose data are telemetered to the University for recording, analysis, and interpretation. The Department of Energy supports the stations on the east flank of the Cascades and eastern Washington. The Nuclear Regulatory Commission partly supports stations in southern Washington and Northern Oregon. Other major parts of the network are supported by the U.S. Geological Survey. Minor amounts of support have been received from the state of Washington, Washington Public Power Supply System, and the National Science Foundation. Section I of this report covers the details of the operation of the network in eastern and southern Washington and northern Oregon.

Details of the past year's seismicity is covered in section II. There was little seismicity of note this past year anywhere in the state in marked contrast to the previous two years. Section III covers some recent advances in our crustal structure studies. Besides summarizing some recent results of investigations of crustal and upper mantle structure of the north Cascade mountains we cover in this section a study using laterally inhomogeneous velocity structures to model the transition between several tectonic provinces. Synthetic seismograms are calculated to compare with observed record sections constructed from the digital recordings of medium sized earthquakes. Section IV summarizes research using a bore-hole seismometer and also describes a technique for recovering velocity, attenuation and fracture porosity from a cross-hole seismic survey run by a subcontractor for

Rockwell Inc. a few years ago. Section V is a preliminary report on a teleseismic P-wave delay study using the digital records of # teleseisms recorded at # stations of the state wide network. This study farther defines the position and velocity contrast of the subducting Jaun de Fuca plate beneath Washington State and extends previous work into northern Oregon.

Two major student research projects have been completed during the past year, both supported by D.O.E. funds. Al Rohay completed a PhD dissertation entitled: *Crust and Mantle Structure of the North Cascade Range, Washington.* The abstract for this work appears at the end of section III and the entire thesis may be obtained from the University as part of this report. Eric Lanning completed a Masters thesis entittled: ---- Its abstract appears at the end of section IV. Both of these works should be considered addenda to this report and may be obtained in their entirety from the Geophysics Program on request.

Network Operations

At the end of September 1982, signals from 108 seismograph stations in Washington and Oregon were being recorded at the University of Washington. Figure I-1 shows stations that have been recording during the past year; the total has sometimes been as high as 114. 13 of these stations are operated by the U. S. Geological Survey and are tapped off at mixer sites; the University is responsible for operation of the remaining stations. Filled in triangles indicate stations intirely supported by D.O.E. or N.R.C. contracts. Those stations are listed in table I-1.

The Eastern Washington regional network consists of 36 stations in Washington east of the Cascade crest, and 2 stations in northeast Oregon. The monitoring sites are not evenly distributed throughout the region. Instead, they center about known areas of elevated seismicity near Hanford and Lake Chelan. A small number of stations located outside these areas enhances the overall state-wide coverage. In late 1981 and continuing in 1982, 6 new stations were installed in northern

Oregon to even out coverage and to provide close-in data for earthquakes in the Portland area, where several intensity VI to VIII earthquakes have occurred in the last century. The new Portland area stations have already recorded several small local events.

The telemetry net operated mainly in a stable manner during the course of the year. Loss of support from WPPSS forced us to shut down the North Cascade stations RPW and LYW, because phone line costs could not be met. LYW will be turned into a radio site, while RPW will be moved to a location several kilometers to the north where a signal can be radioed out of the Skagit Valley. The east flank Cascade stations formerly supported under the same grant continue in operation for the time being, since phone line charges could be met from available funds.

Some breakdowns occurred in late 1981 at about the time that snow rendered many sites inaccessible. As a result, a few stations (TBM, PLN, CBW, and DYH) were inoperative for much of the contract period. Most sites had to be visited at least twice for repairs during the year. The services of the Stanwyck Corporation technician, Mr. Don Hartshorn, proved invaluable in maintaining an overall high percentage of uptime. Given available resources, it would not have been possible for a Seattle-based technician to achieve the same results.

While the network configuration was changed only slightly during the year, some effort was expended in an attempt to upgrade and calibrate the equipment at selected sites. Calibrated S13 seismometers, damped 0.7 critical, were installed at GBL, ETT, and WA2, and one was prepared for MDW but to date has not been installed. The seismometer changes at GBL and ETT were combined with installation of new, low-noise telemetry systems designed by S. T. Morrissey of Saint Louis University. These systems are somewhat narrower in bandwidth than the conventional USGS microearthquake systems, but it is felt that the difference actually is advantageous because the new system has less response only at frequencies higher than about 10 Hz. Given the prevailing attenuation patterns in the northwest, there is little useful data but much telemetry noise in the band between 10 and 35 Hz where the USGS system has its peak response. Therefore, the new systems make a considerable improvement (generally 6 to 12 db) in the signal-to-noise ratio -- headroom that we intend to use to improve the dynamic range of individual stations.

Calibration curves for the ETT system are presented as Figures I-2 and I-3. The GBL system differs slightly in the damping ratio, but the overall response is similar in both frequency and absolute gain level. Figure I-2 gives the combined response of the seismometer- amplifier/VCO system in both millivolts/millimicron and Hz deviation/millimicron. In other words, it is the information placed on the line at each site. Figure I-3 gives the overall system response as seen at the discriminator output at the University, and recorded on the PDP 11/34 digital system. It is given in both millivolts/millimicron and digital counts/millimicron. The response curves for the same stations as recorded on the digital system at Rockwell-Hanford Operations will differ from those as recorded at the University because different discriminators are used.

These curves are preliminary; the seismometer calibration was done at the University but we are relying on the designer's calibration curves for the amplifier/VCO and discriminator components. It would be desirable to run a complete calibration from each site through the entire telemetry system. Such calibrations will be run from the Hanford area sites in the spring of 1983.

The GBL and ETT installations were the first made in an effort to provide at least some calibrated stations throughout the University of Washington network. At present (September 1982), similar systems have been installed at FMW (Mt. Rainier), OBC (North Olympics), SBO (east-central Oregon), and KMO (northwest Oregon). In addition, calibrated L4-C systems are operating at PGO (western Oregon) and STD (Mt. St. Helens). In the coming year, 2 or 3 additional calibrated S13 systems will be installed in the eastern Washington regional net.

In the latter part of 1982 and in 1983, much of the technical staff's attention will be occupied by attempts to further upgrade the eastern Washington network and reduce the high operating costs. 18 of the Morrissey-design amplifier/VCO units have either been ordered or are authorized; such wholesale replacement of the Develco 6202 units seems justified in light of numerous recent station failures that are due to aging components (mainly leaking capacitors). We have been authorized to buy 12 radio pairs; these will be used to cut down on the high phone line charges (estimated yearly cost $36K at present rates). At the same time, there seems to be a possibility that Bonneville Power Administration will provide us with free use of some of their unused microwave circuits at several points in Washington and Oregon. If this comes to pass, it will mean the virtual elimination of long-line charges for the mixed signals coming back to Seattle from bridge points at Wenatchee and Richland. This could mean a savings of 15 to 25% of the annual phone line costs. It would also probably mean some reshuffling of the network, depending on the points where we are granted access to the circuits.

TABLE I-1 D.O.E. - N.R.C SUPPORTED SEISMIC STATIONS

STA	LAT	LONG	TIME	NAME
AUG	45 44.17	121 40.83	10/81	Augsburger Mt.
BDG	46 14.08	119 19.05	7/75-	Badger
CBW	47 48.42	120 01.960	7/75-	Chelan B
CRF	46 49.51	119 23.09	7/75-	Corfu
DAV	47 38.30	118 13.56	7/75-	Davenport
DYH	47 57.63	119 46.16	7/75-	Dyer Hill
ELL	46 54.58	120 34.10	7/79-	Ellensburg
EPH	47 21.13	119 35.77	7/75-	Ephrata
EST	47 14.28	121 12.53	7/79-	Easton
ETP	46 27.89	119 03.54	7/75	Eltopia
ETT	47 39.30	120 17.60	6/77	Entiat
EUK	46 23.75	118 33.72	7/75-	Eureka
FPW	47 58.00	120 12.77	7/75-	Fields Pt.
GBL	46 35.86	119 27.59	7/75-	Gable
GLD	45 50.33	120 48.85	8/77-	Goldendale
JBO	45 27.00	119 51.00	9/82-	Jordan Butte
KMO	45 39.00	123 27.00	9/82-	Kings Mt.
MDW	46 36.80	119 45.65	7/75-	Midway
MFW	45 54.18	118 24.35	7/75-	Milton-Free.
NAC	46 43.98	120 49.47	8/79-	Naches
NEW	48 15.83	117 07.22	/77-	(USGS)
NLO	46 05.30	123 27.00	10/81	Nicolei Mt.
ODS	47 18.40	118 44.70	7/75-	Odessa
OMK	48 28.82	119 33.65	7/75-	Omak
OTH	46 44.34	119 12.99	7/75-	Othelo
PAT	45 52.85	119 45.68	6/81-	Paterson
PEN	45 36.72	118 45.78	7/75-	Pendleton
PGO	45 28.00	122 27.17	6/82-	Gresham, Or.
PHO	45 37.14	122 49.80	4/82-	Portland Hills
PLN	47 47.08	120 37.97	6/77-	Plain
PRO	46 12.76	119 41.15	7/75-	Prosser
RSW	46 23.47	119 35.32	7/75	Rattlesnake
SAW	47 42.10	119 24.06	7/75-	St. Andrews
SBO	45 02.00	120 06.00	9/82-	Squaw Butte
SYR	46 51.78	119 37.07	7/75-	Smyrna
TBM	47 10.17	120 31.00	7/79-	Table Mt.
VTG	46 57.48	119 59.24	7/75-	Vantage
WA2	46 45.40	119 33.76	5/78-	Wahluke2
WAT	47 41.92	119 57.25	11/76-	Waterville
WBW	48 1.07	119 08.23	7/75-	Wilson B
WEN	47 31.77	120 11.65	7/75-	Wenachee
WGW	46 2.68	118 55.96	7/75-	Wallula Gap
WIW	46 25.93	119 17.29	7/75-	Wooded Is.
WPW	46 41.92	121 32.42	4/80-	White Pass
WRD	46 58.19	119 08.60	7/75-	Warden
WTP	48 28.27	120 14.87	8/77-	Winthrop
YAK	46 31.73	120 31.22	7/79-	Yakama

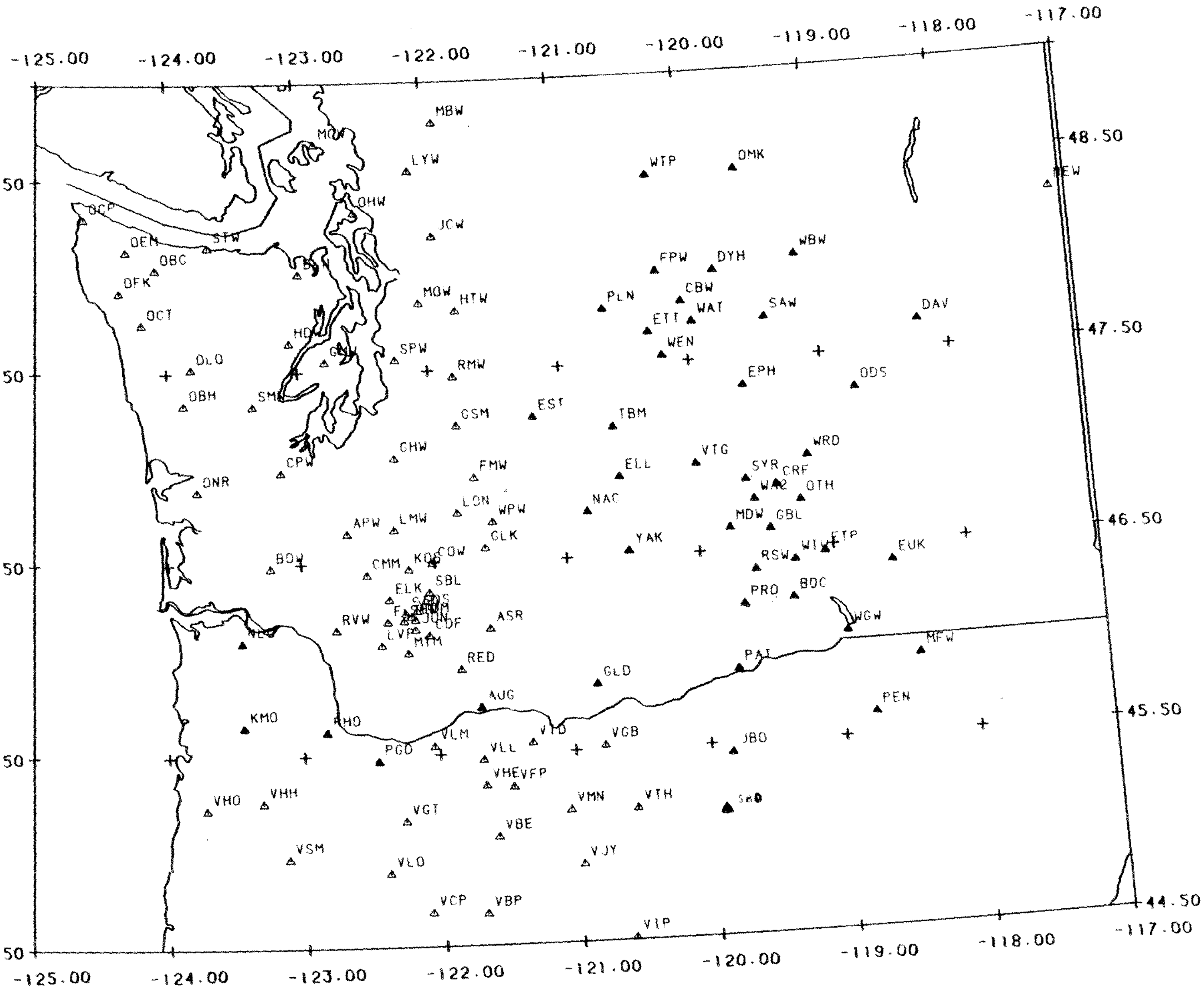

Figure I-1. Currently operating seismograph stations in the Washington regional seismograph network. Filled in triangles indicate stations intirely supported by D.O.E. or N.R.C

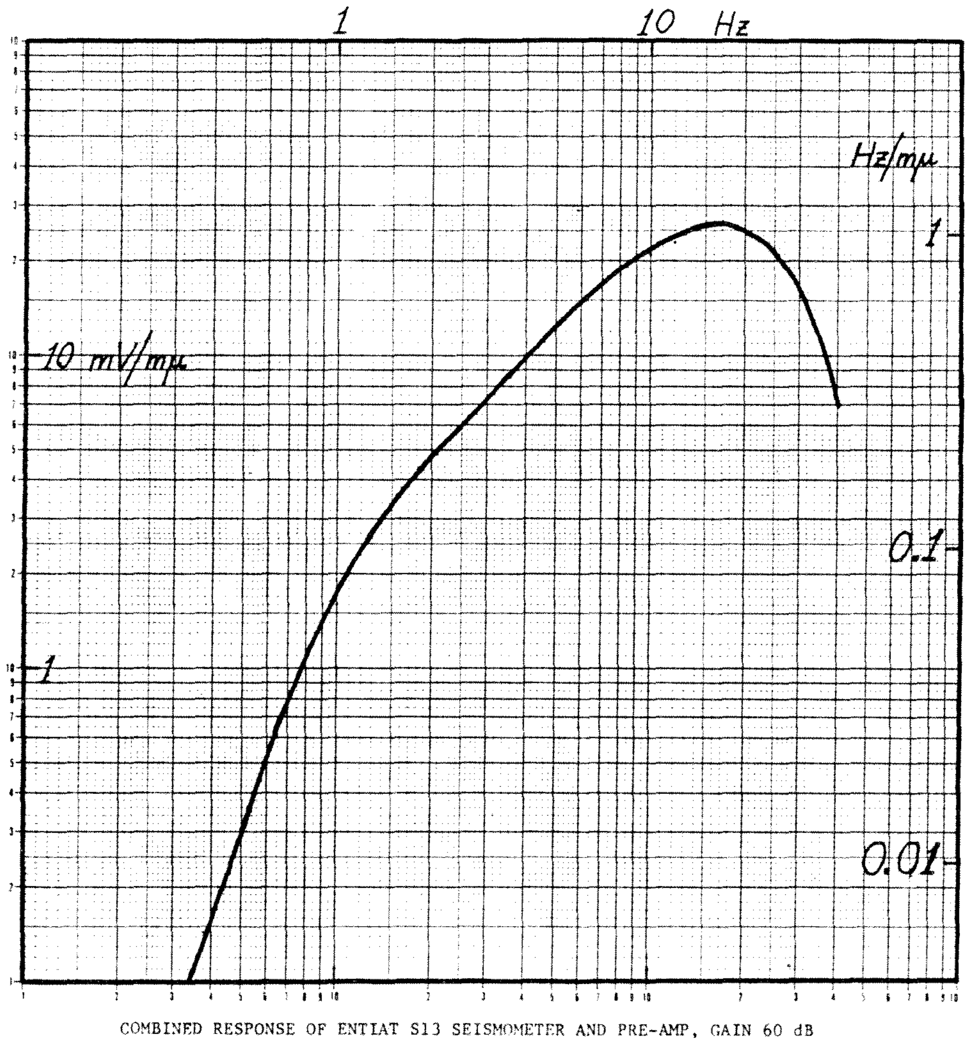

Figure I-2. Magnification curve for calibrated short-period field system.

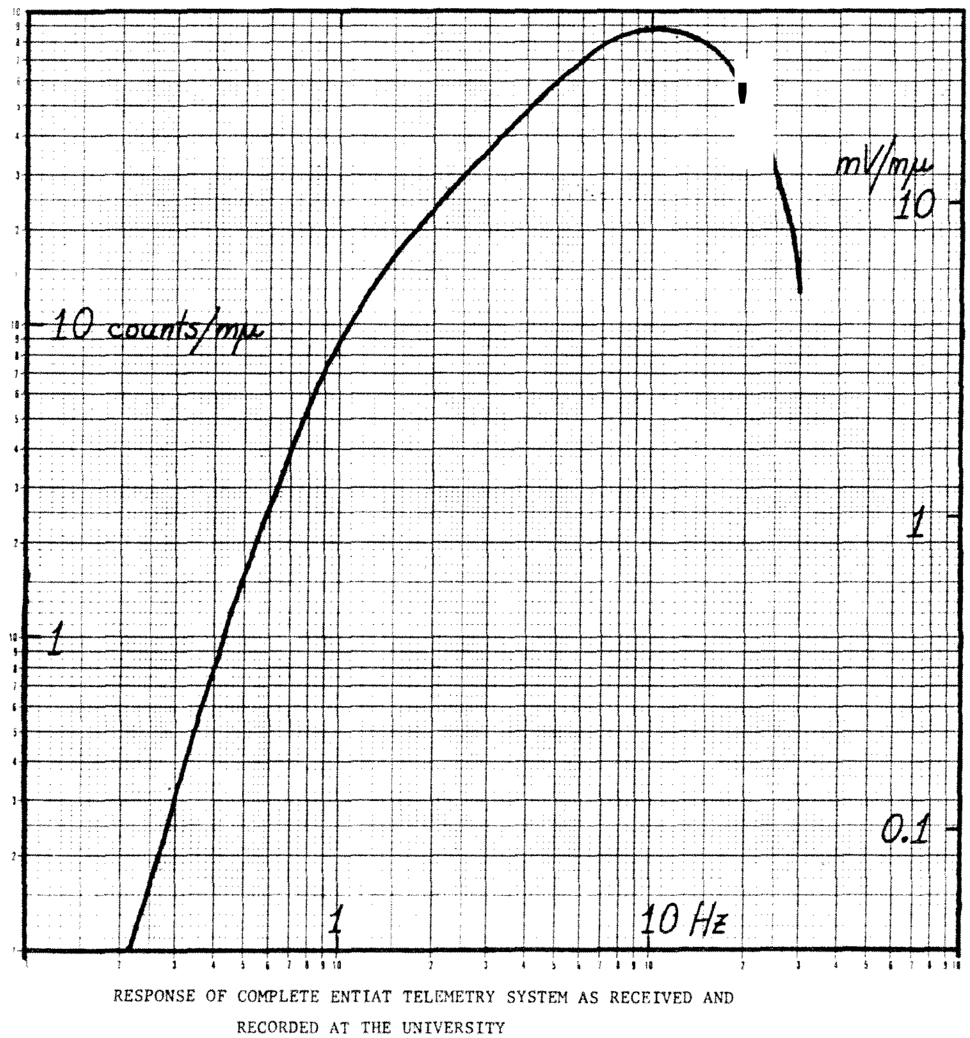

Figure I-3. Magnification curve for complete calibrated short-period system into online computer.

II. SEISMICITY 1981 - 1982

Introduction

During the period 1 July 1981 through 30 June 1982, the level of seismicity in Washington state east of Puget Sound has been low compared with the previous year and half. During the past year no significant earthquake has occured in Washington State or Northern Oregon. The level of seismic activity appears to be returning to the level of seismicity observed prior to the eruption of Mt. St. Helens.

Data

The data base is complete for the time period discussed in this report. The digital method of recording and processing the seismic data on- and off-line has kept up with activity levels during this period. Activity levels in the state of Washington this period have been low relative to the high levels of activity experienced in 1980 and the first half of 1981. During the past year, only 12 events of magnitude 3 or greater have been recorded in the state of Washington. The maximum magnitude earthquake was a M=4.1 which occured on 1 March 1982. This is the largest aftershock of the Elk Lake earthquake(M=5.5, 14 February, 1981) to date and occured at nearly the same location as the main shock. This earthquake was a felt event and was assigned a maximum intensity of V by NEIS. A magnitude 3.7 earthquake occured in the Puget Sound on 12 November 1981. This was the second largest magnitude earthquake to occur in the last year and was felt over an area of approximately 1000 square kilometers in northwestern Washington.

Eastern Washington and Northern Oregon

During the period 1 July 1981 through 30 June 1982, 632 events which occured in Eastern Washington and Northern Oregon were processed. Of these 221 were known or probable blasts; the majority occured at the Coulee Dam, Dry Falls Dam, and Ice Harbor Dam, and in southwestern Washington. Figure II-1 and II-2 show the known and probable blasts in Eastern Washington and Northern Oregon, respectively. The remaining 411 events were earthquakes, of which 60 were hand-picked because the on-line computer system did not record them for one reason or another. Figure II-3 and II-4 show the epicenters of earthquakes in Eastern Washington and Northern Oregon, respectively. Appendix I contains the event catalog for this period. It may show changes from the preliminary catalogs published in the quarterly technical reports because errors have been found and corrected in the interim.

Two clusters of earthquake activity are evident in Eastern Washington. The western most concentration of activity occurred in the aftershock region of the Goat Rocks earthquake. The largest magnitude earthquake in Eastern Washington in the past year was a magnitude 3.0 which occured on 23 January 1982 in this region. This earthquake was not felt due to its location in one of the least-populated areas of the state. The second cluster of activity is in the area southwest of Lake Chelan which has been typically active in the past. Two felt events, occurring in this area, had magnitudes of 2.5 and 2.4. In addition to the above mentioned clusters, diffuse activity occured throughout the Pasco Basin. There was a concentration of activity north of the Saddle Mountains but no significant patterns were observed. Figure II-5 shows the focal mechanism plot for a M=2.3 which occured in the Pasco Basin at a depth of 17 kilometers. While not a well constrained solution, it does indicate a thrust mechanism which is in agreement with mechanisms found for other events in this region.

Two felt events occured in southwestern Washington. A M=2.4 was felt in Camas and a M=2.0 was felt by a few people in Chalatchie. A magnitude 2.5 earthquake on 21 July 1981 in southwestern Washington was recorded on sufficient number of stations with clear first motions to allow construction of a first motion plot. This is shown in figure II-5. The north-south plane is well constrained, however the east-west plane is not. The mechanism is strike-slip with a variable component of dip-slip, depending on the orientation of the east-west plane. This mechanism is consistent with focal mechanisms found for other events in the southern Cascades (Elk Lake and Goat Rocks earthquakes). A concentration of epicenters west of Mt. St. Helens is noticeable; many of these may be blasts identified as earthquakes as this area has experience numerous blasts in the last year.

The appendix to this report is a catalog of the located events between July 1, 1981 and June 30, 1982. The locations reported in this catalog have been determined using a location routine obtained from Dr. Bob Herrmann at St. Louis University and extensively modified and tested here at the University of Washington. Azimuthal weighting is used and obviously bad readings are automatically thrown out. There is a special depth adjustment algorithm for events with poorly controlled shallow depths.

Most of the columns in the appendix are self explanatory. Times are in coordinated universal time (PST + 8hr). The * sometimes following the depth means that the depth has been fixed. **$** and # mean that the maximum number of iterations has been exceeded without meeting convergence tests and both this and the depth has been fixed. Events flagged with these symbols may be very poorly located even if the quality factors are good. NS/NP is the number of stations and the number of phases used in the location determination, and the model code matches the models given in table III-1. The *types* listed in table II-1 are as follows:

X- Known explosion

P- Probable explosion (based on seismogram character)

F- Earthquake reported to have been felt

H- Hand picked event from film records (Computer recording not available)

This catalog is a subset of the state-wide catalog which now has 12,602 entries from Jan 1, 1980 to the present. This includes some 4,041 regional and teleseismic events which are not located or processed other than saving the trace data. This complete catalog is kept nearly up to date including the addition of new events as they are analyzed and the corrections to older events as they are made.

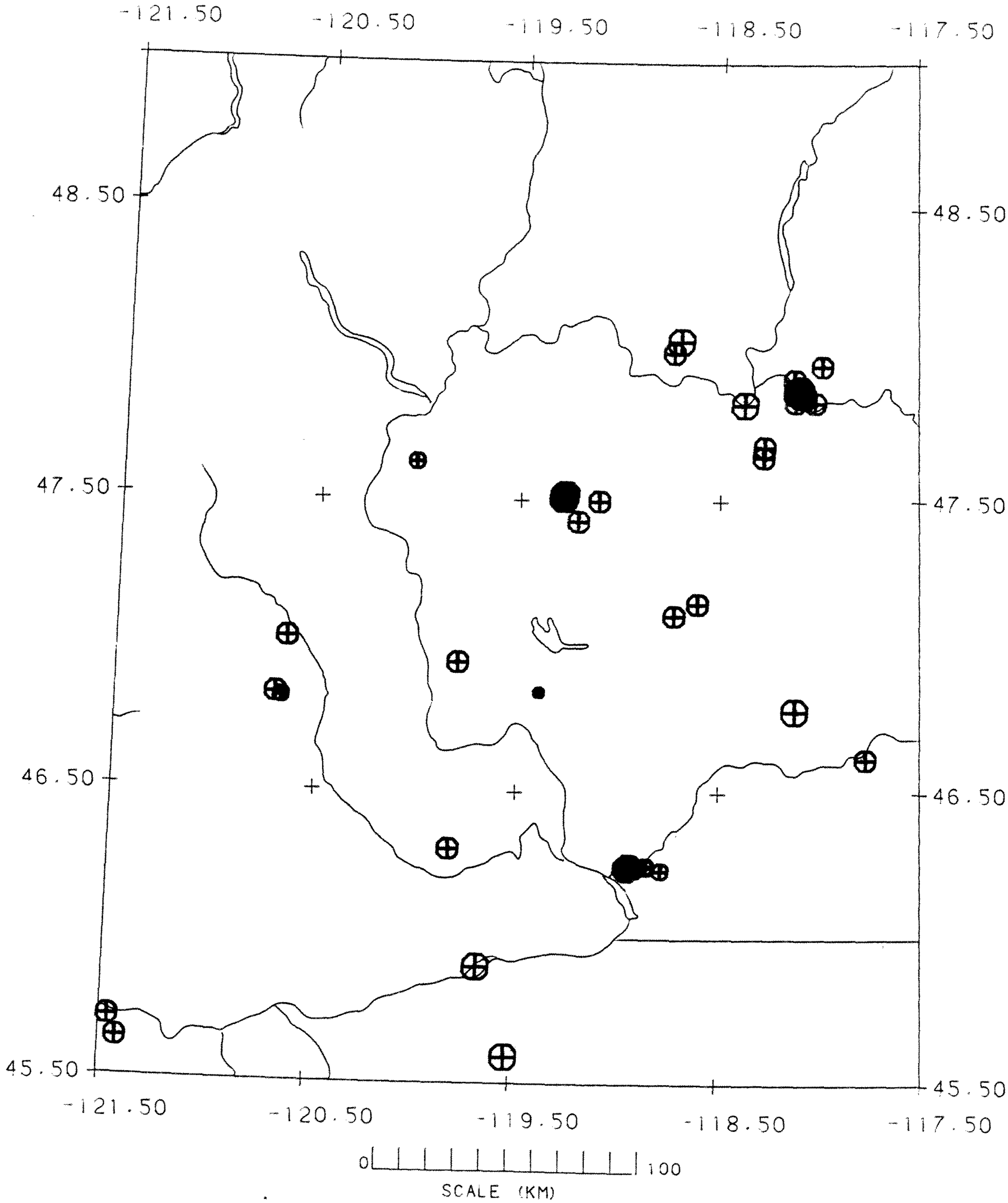

Figure II-1. Eastern Washington known and probable explosions 1 July 1981 - 30 June 1982

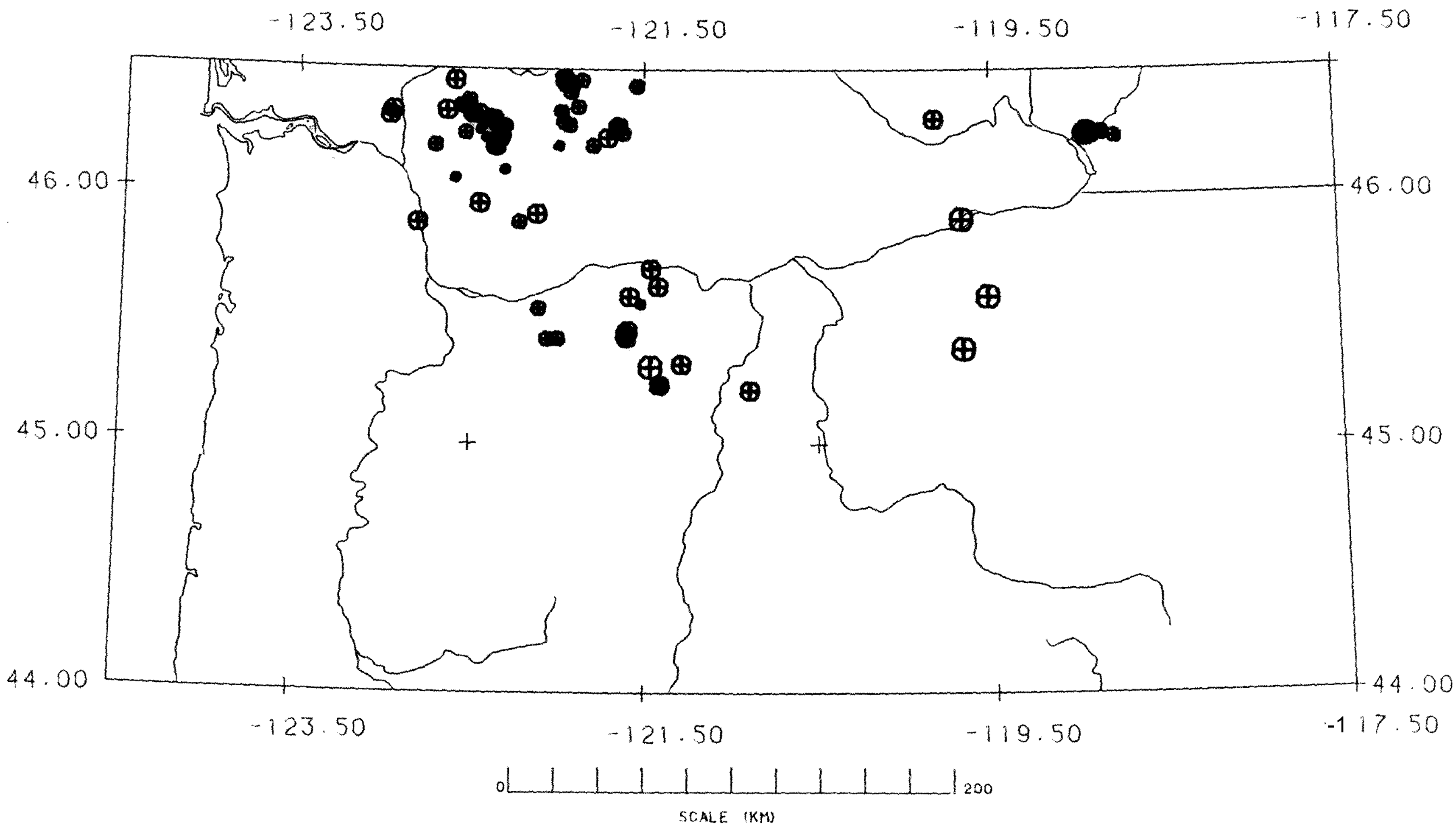

Figure II-2. Southern Washington - Northern Oregon known and probable explosions 1 July 1981 - 30 June 1982

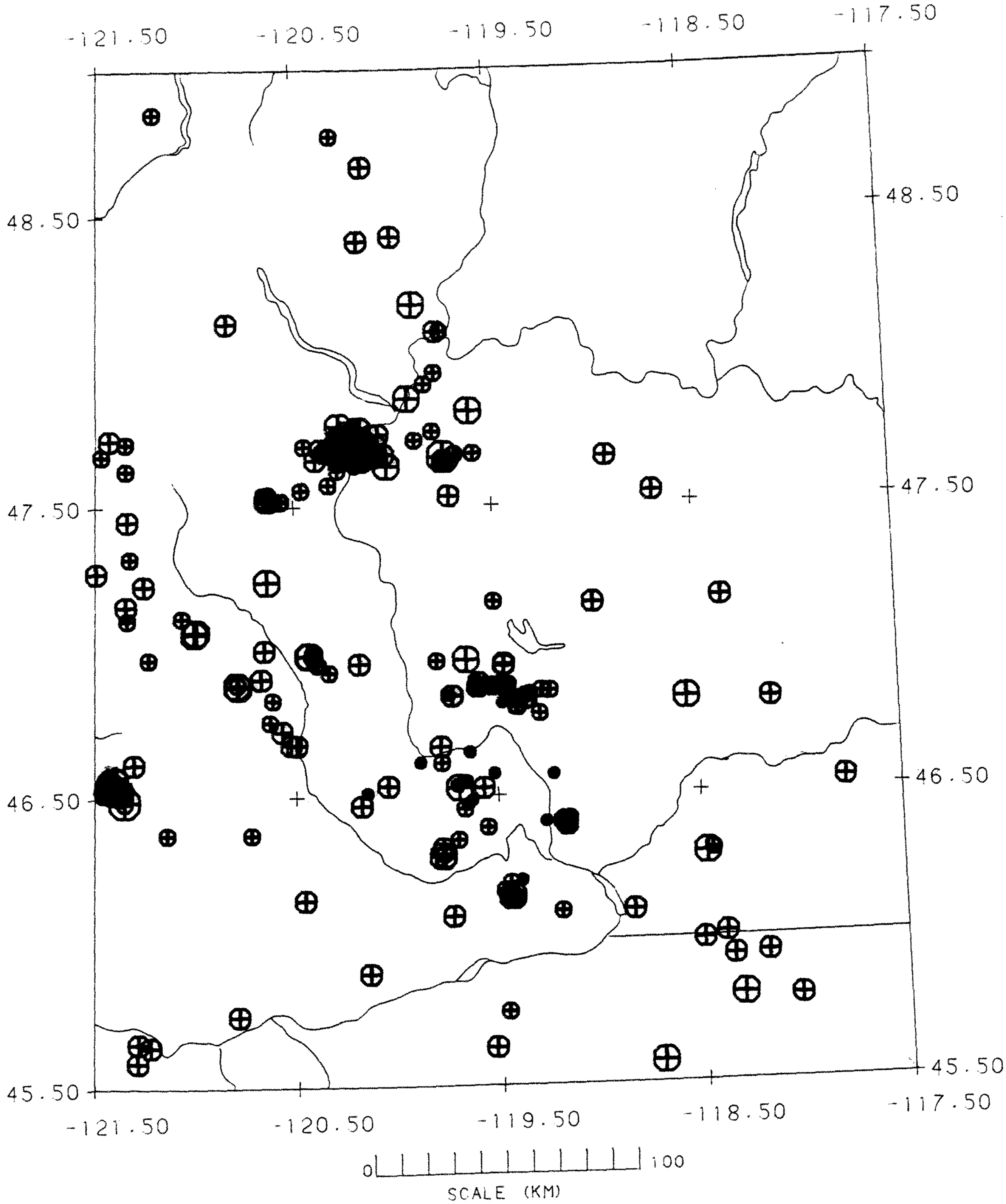

Figure II-3. Eastern Washington Earthquakes 1 July 1981 - 30 June 1982

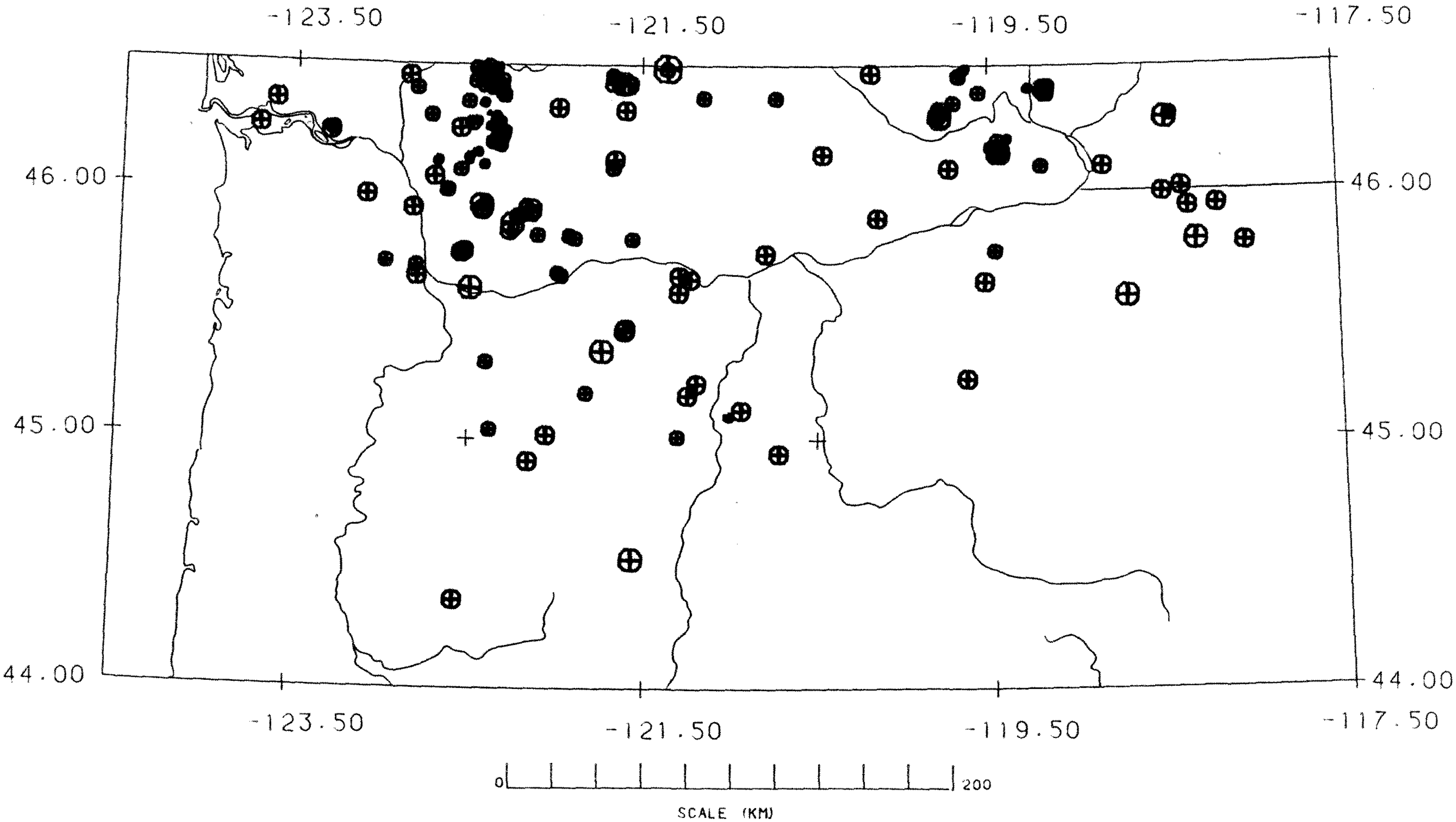

Figure II-4. Southern Washington - Northern Oregon Earthquakes 1 July 1981 - 30 June 1982

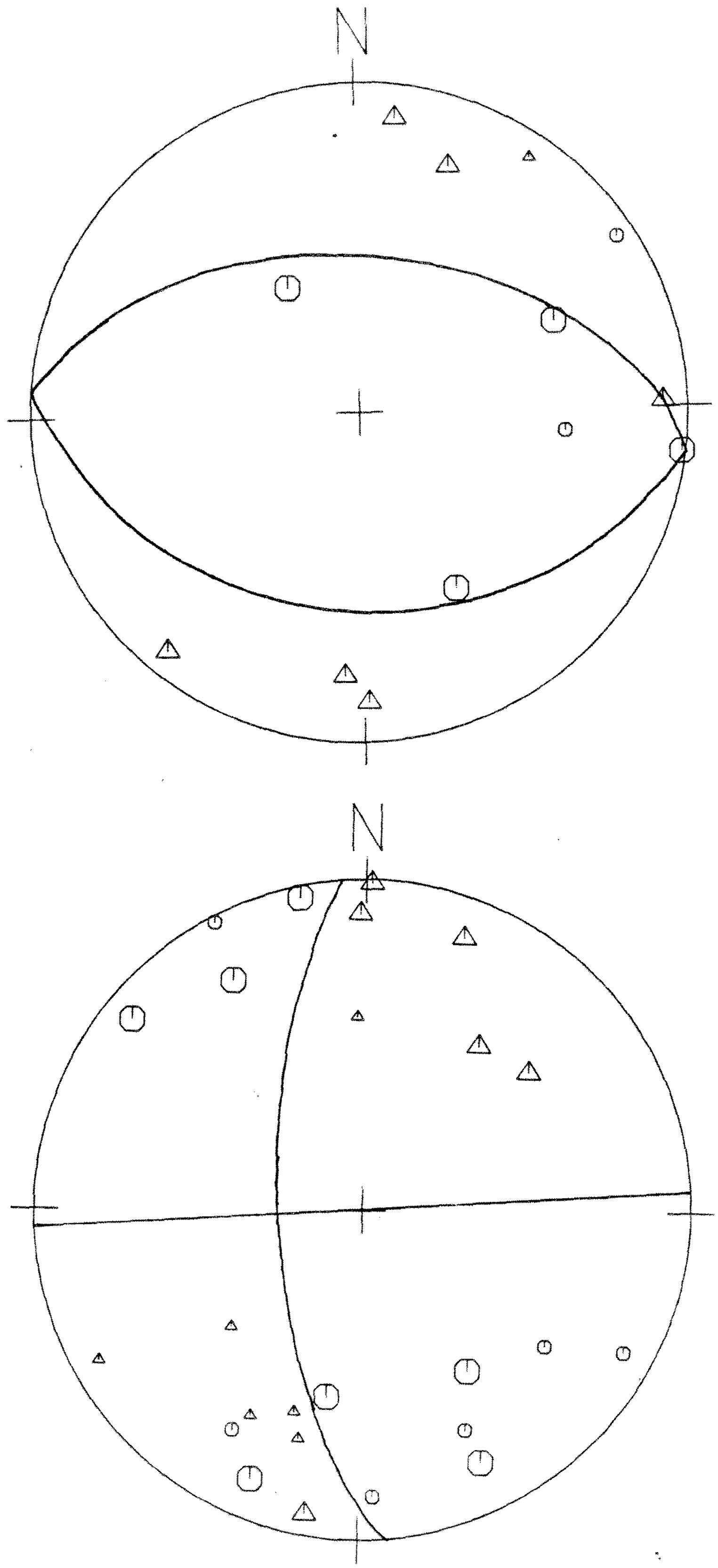

Figure II-5. Focal mechanism solutions for: Top- Sept 23, 1981 at 1628Z in central columbia basin, depth= 17.4 km, mag= 2.3; Bottom- Jul 21, 1981 at 2210 in south west Washington, depth 9.6 km, mag= 2.5.

III. STRUCTURAL STUDIES

The velocity structure for various parts of Washington have been determined by a number of investigators over the past few years. Many of these studies were supported by the *Eastern Washington earthquake monitoring project.* In the past year a major study of the crust and upper mantle structure of the North Cascades range was completed by Alan Rohay as his doctorial dissertation. The abstract for this dissertation appears at the end of this section and the complete dissertation is included as an addendum to this report. This work not only derived a fairly detailed crustal model for the North Cascades but also compared this model to those of surrounding provinces and suggested improvements to these models. It carried the velocity model into the upper mantle and defined a dipping high velocity slab interpreted to be the subducting Juan de Fuca plate.

The Rohay study as well as most of the others listed in table III-1 used first arrival picks, read from film records, as the primary data source. Only first arrival times were used to invert for the velocity structure. With the advent of digital recording it is now possible to use more than the first arrival.

In last year's annual technical report, samples of seismic record sections from blasts and earthquakes were shown. These sections, produced from the digital data archive, showed more detail than could be interpreted by first arrival techniques alone. No additional interpretation was attempted at the time. During the past year we have generated additional record sections from other sources and different combinations of stations to study primarily the Cascade Range and the transition into eastern Washington. We have also developed the procedure to generate theoretical travel time curves using a two dimensional laterally inhomogeneous velocity structure and synthetic seismograms from this model. These travel-time curves can be compared to those obtained by a visual examination of the observed record sections, and the synthetic seismograms can be compared to the recorded data. This technique can allow us to refine and improve the velocity models currently in use and may allow us to detect rapid lateral velocity variations which may be of great tectonic significance. In this report we summarize the 26 record sections obtained thus far, compare 8 of these sections with calculated travel-time curves and synthetic seismograms determined using velocity models previously determined, and then illustrate how one of these models can be modified to improve the comparison of the theoretical with the observed.

Existing Velocity Models

There are six one dimensional velocity models which are currently in use for locating earthquakes and which were determined by specific structural investigations. A summary of these structures for separate tectonic provinces is given in table III-1.

Figure III-1 is a map showing the boundaries for the tectonic provinces whose velocity models are listed in the table. The respective models are used for locating earthquakes within each province. Because our hypocenter location routine (and all others commonly in use) is designed to use only one velocity structure we must choose one and only one model for locating an event. Obviously there are problems for those events near province boundaries or earthquakes large enough to be well recorded in provinces other than their own. Our solution to the first problem is to assign an earthquake to a province in which most of the nearest stations lie. The solution to the second problem is down weight stations far enough away to be in another province and to apply station corrections appropriate to that model for stations in a different province.

The boundaries shown on figure III-1 are somewhat arbitrary, often having little data to establish their exact position, not to mention the nature of the transition across the boundary. The models chosen for each province are, by necessity, an average of various determinations within that province. A time-term method

TABLE III-1 VELOCITY MODELS

Model and Source	velocity	depth(km)
Puget Sound (P1)	5.4	0.0
Crosson (1976)	6.6	10.0
	6.7	16.0
	6.9	22.0
	7.1	32.0
	7.2	35.0
	6.9	38.0
	7.8	41.0
Eastern Washington-	3.70	0.0
Columbia Basin (E1)	4.70	0.8
Malone (1977)	5.15	1.2
	6.05	7.5
	7.20	19.0
	8.00	28.0
North-Eastern	5.10	0.0
Washington (N1)	6.05	0.5
Malone (1977)	7.2	19.0
	8.0	24.5
North-	5.1	0.0
Cascades (C1)	6.0	1.0
Rohay (1982)	6.6	10.0
	6.8	18.0
	7.1	25.0
	7.9	35.0
Oregon Cascades (O1)	3.0	0.0
Leaver (1982)	4.7	1.3
	6.0	3.4
	6.3	7.6
	6.5	11.0
	7.0	31.0
	7.7	44.0
St. Helens Area (S1)	4.8	0.0
Combination of	5.0	1.0
O1 and C1 models	6.0	3.0
	6.4	10.
	7.0	30.
	7.8	42.

was used to establish at least parts of the C1 (Rohay, 1982), N1, E1 (Malone, 1977) models, a simultaneous inversion for velocity and earthquake hypocenters was used for the P1 model (Crosson, 1976) and a detailed, many channel, refraction survey was used to establish the O1 model (Leaver, 1982). This later technique used ray tracing through a laterally inhomogeneous medium to generate theoretical travel time curves and synthetic seismograms. The use of synthetic seismograms can add a great deal of information to a crustal structure study because not only the time of arrival of individual phases but the amplitude and waveform as well can be calculated and compared to the observed data. Thus far their is no way to directly invert the seismograms for structure, but seismograms generated synthetically from a given source and structure can be compared to the observed ones to check the validity of the model. This technique is used in this section to check the accuracy of the models used in Washington and to examine some details of the transitions from one model to another, particularly from the Cascade (C1) model to the eastern Washington models (N1 and E1).

Observed Profiles

Using earthquakes as seismic sources for refraction lines has both advantages and disadvantages. The advantages are that they are free, big enough to record well at even great distances, and are usually good generators of impulsive P and S waves. The disadvantages are that one does not know when or where they are going to occur and can not plan ones station distribution to take best advantage of them. There exact location and origin time can only be determined precisly by knowing the velocity structure and that is the purpose of the experiment. We can minimize these disadvantages because we have many stations operating over a large area all the time and thus we "catch" the earthquakes when ever and where ever they occur. By using stations near the epicenter to locate the event and a velocity model determined independently from controlled explosions and of sufficient accuracy for the shallow crust, we can then used the arrivals from this earthquake to study the velocity structure of the deeper crust at some distance from the event.

Figure III-2 is a map showing 9 medium large earthquakes which have been used as sources for 26 refraction profiles put together from the data of the regional seismic network. Most of these events have well determined hypocenters using 14 to 41 phases from stations within 60km and they are all less than 10km deep. Table III-2 gives the details of each of these events. 26 record sections have been constructed using these events as sources and various combinations of network stations. An example of such a record section is shown in figure III-3 for the Cle Elum earthquake of 18 February, 1981 and stations distributed to the southwest.

TABLE III-2 REFRACTION LINE SOURCES

DATE	TIME	LAT	LON	DEPTH	MAG	MODEL	LOCATION
Sep 19, 1980	22:53	47 54.49	121 51.37	5.79	3.8	P1	Sultan
Nov 19, 1980	21:35	46 57.24	119 28.67	0.50	3.3	E1	Smyrna
Feb 2, 1981	1:23	46 15.28	121 0.10	3.46	4.0	C1	Toppenish
Feb 14, 1981	21:27	46 20.63	122 14.16	8.33	3.8	S1	Elk Lake
Feb 18, 1981	6:09	47 12.49	120 54.84	6.94	4.2	C1	Cle Elum
Mar 15, 1981	7:23	47 59.20	121 30.25	4.81	3.6	C1	Granite Falls
May 28, 1981	8:55	46 31.97	121 25.03	3.74	4.6	C1	Goat Rocks
Jun 23, 1981	0:05	48 50.86	122 9.61	3.02	3.4	C1	Mount Baker
Feb 18, 1982	3:27	47 39.65	119 44.92	6.75	2.8	N1	Waterville

The first arrival at most of the stations between about 60km and 160km usually defines a major crustal velocity of between 6.0 and 6.8 km/sec. This velocity, determined by fitting a line by eye through the best arrivals is indicated near each line in figure III-3 which shows the general direction and length overwhich this velocity was observed. Note that the major crustal velocities going north-south in the Cascade range are between 6.55 and 6.7 km/sec, while those in eastern Washington are significantly lower, between 6.1 and 6.4. There is considerable variation for paths which cross the transition between the Cascades and eastern Washington from a low of 6.0 km/sec for the Goat Rocks earthquake to the northeast and a high of 6.79 km/sec for the Smyrna earthquake and paths directly to the west.

It is not possible to use these data directly to solve for crustal structure at this time. One problem is that we do not know the depth of most of these earthquakes precisely enough to use the intercept times of the velocity determinations to establish refractor depth. We are also not sure if we are seeing the same crustal layer in each of these cases. Careful examination of each of these lines may provide useful information on the details of the velocity variations of the main crustal refractor but we choose to explore a different method of approach at this time.

Comparison of Theoretical and Observed Travel Times

The best eight refraction profiles have been selected for which to compute theoretical travel time curves and synthetic seismograms. These lines use the larger, better recorded earthquakes with the most uniform station distribution over a narrow azimuth. We have selected laterally varying velocity structures which are roughly linear transitions between the one dimensional models given in table III-1. In each case the model is four layers with possible dipping interfaces and lateral velocity variations over a half space. Each layer has a slight increasing velocity gradient within it (0.1 km/sec increase between top and bottom of a layer). This causes rays entering a layer at a shallow angle to be bent upward. These "turning" rays are used to approximate head waves which are not generated in our ray trace program.

The ray trace routine uses zeroth order asymptotic ray theory to trace rays through laterally varying structures. Its output consists of a plot of the rays shot through the velocity model, a travel time plot, and files describing the model and rays. The use of asymptotic ray theory allows both horizontal and vertical velocity gradients and permits the boundaries to be almost any shape. Some of the disadvantages of the technique are that head waves must be approximated by turning rays and that each type of arrival at a given distance (eg. PP, Pg) must be explicitly specified.

The program evolved from the Cerveny RAYTRACE program written in 1977. Since then it has been extensively modified by various members of Walter Mooney's group at the USGS(McMechan & Mooney, 1980). The only changes that have been made at the UW have been to machine dependent input-output code plus the correction of several small errors. Thus far the raytrace program has been used at the UW to model refraction data along the Washington continental margin (Taber, 1981; McClain, 1981) and to model teleseismic data throughout Washington State (Michaelson, section V of this report).

Besides tracing rays through a complex velocity medium this program can calculate amplitudes for each ray by computing the geometric spreading of the ray tube and the transmission and reflection coefficients of the ray at each boundary. Finally, a synthetic seismogram can be generated using a separate program by constructing a time wavelet of a linear combination of a unit impulse and a Hilbert transform, summing over all arrivals at a given distance and convolving the result with an appropriate source function. An output file from the first program is used as the input to the second which calculates and plots the synthetic seismogram. The process is fairly complicated and requires a number of long computer runs for each record section since one must find all the appropriate ray paths by trial and error and explicitly designate the path they take. This means that many different ray paths must be tested to assure that a proper waveform is generated.

Almost four hours of PDP-11/70 CPU time were used to test and generate the comparisons shown in figures III-4 to III-11. Each of these figures shows in part *A* the velocity model and rays comming from the earthquake source at the left hand side. These rays travel various least time paths before emerging at the surface. In part *B* of the figures, a reduced theoretical travel time curve is shown (reducing velocity of 7.0 km/sec) along with our most reliable picks of first arrivals. Crosses on the travel time curves indicate the arrival at that point from a turning ray (head wave), and pluses indicate the arrival of a reflected ray, usually one which has been supercritically reflected and thus has not lost energy to deeper layers. For simplicities sake and to save time we have not included rays which are internally reflected at less than the critical angle even though the program is well set up to handle these. Such waves arrive later in the seismogram and are usually smaller when compared to those arrivals we have encluded. We also have not included multiple reflections from the free surface. Such rays can contribute significant energy in the early part of the seismograms but are very sensitive to source depth. Our uncertainty in source depth is large enough for most events that we choose to leave out these rays all together. Part *C* of each figure is the observed record section and part *D* is the synthetic record section plotted at the same time and distance scale. Synthetic seismograms are calculated every 20km starting at 50km from the source.

Discussion

Obviously we are not trying to duplicate the observed seismograms wiggle for wiggle but rather, as a first cut, try to see if the gross features of our travel time curves and synthetic seismograms are similar to the observed. In some cases the agreement is good while in others there are obvious discrepancies.

Figure III-4 and II-5 show a roughly reversed refraction line running north-south down the center of the Cascades from Mount St. Helens in the south to about Glacier Peak in the north. The model used to generate the theoretical travel time curves and synthetic seismograms is a combination of the Oregon Cascade model (O1), and the north Cascade model (C1). Uniformly sloping interfaces and smoothly varying lateral velocities go from the one dimensional structures at each end. Note that in both cases the fit of first arrivals is poor in the range of about 100 to 220 km. The observed first arrivals are up to 0.8 sec. earlier than the computed ones. The comparison for the southward directed section (Figure III-5) is better than that for the northward directed one. In particular the arrivals through the St. Helens area (stations COW - LVP) are on the average only a few tenths of a second early and the secondary arrival at these stations matches well with the predicted secondaries in the synthetic record section. For the northward directed section (figure III-4) there is considerable discrepancy for stations RMW and HTW. Both the observed first arrivals and secondary phases are quite early compared to the synthetics. At a distance of 275 km (MBW) the comparison of the first three observed arrivals and the synthetics shows that the crustal thickness part of of the models is pretty good. There is obviously a problem with the mid-crustal part of our model. Unfortunately there is no single change to the model which would bring all the observations in line with the theoretical ones. An increase in velocity for the mid-crust might help, or a shallower 7.0 km/sec layer. Also a step in one of the mid-crustal boundaries might produce a better fit or a combination of all three of these.

While the one dimensional models for the north and south Cascades do not differ from each other very much, the eastern Washington models are quite different. Figure III-6 shows a profile from the north central Cascades into northeastern Washington. The crust thins from 41 km under the Cascades to only 26 km 330 km to the east. As a first approximation we model this as a uniform smooth transition. It is somewhat surprising that the fit between the observed and theoretical for this profile is much better than that for the previous two. The relative amplitudes of the first arrivals do not match the synthetics well but this type of comparison is not valid since the observed data are from uncalibrated stations and are plotted at an arbitrary gain factor to make the maximum parts of the the various traces roughly equal amplitude. Note that the secondary arrivals on several stations, match fairly well though in some cases there is an obvious time shift of part of a second.

The above profile is normal to the strike of the Cascade-eastern Washington transition. Figure III-7 crosses this transition at an angle from the Goat Rocks earthquake foreshock to north eastern Washington. The first arrival data compares well for the crustal phases though the observed mantle refractions and reflections are too early indicating our model crust is, on the average, too thick. This earthquake was large enough that most stations out to at least 200 km were at least partly saturated making comparisons of secondary arrivals at the near stations difficult. Only at stations ODS, WBW, and DAV can any use be made of the later arrivals. The uniform transition from central Cascades to north eastern Washington seems to fit these data well though the average crust may be a few kilometers thinner through this area.

Figure III-8 shows a profile which is almost identical to that of the previous figure; but the source is 70 km farther to the southwest and at a depth of 8.3 km rather than 3.8 km. The observed first arrival picks shown in part B are from the Elk Lake Main shock which had virtually the same hypocenter as the aftershock 15 hours later which was used to plot the record sections of part C. The model has a 44 km crust at the Elk Lake end and thins to only 28 km at the northeastern end. In this case the observed shallow crustal arrivals and mantle refractions fit the theoretical travel time curves quite well. On the other hand the observed mid-crustal first arrivals (between 100 and 200 km) are later than the theoretical travel times would predict. The synthetic record section of part D indicates that the first arrivals in this region should be quite weak. It may be possible that the our picks have missed the true first arrival but this is doubtful since the main shock is a magnitude 5.5 earthquake which saturates virtually all stations in the state within the first few cycles. Our model might be adjusted to virtually eliminate the early arrivals at the medium distance by a shadowing effect. It is also possible that the earthquake radiation pattern is a minimum for rays leaving the source at the angle required for this travel path.

For modeling the transition from the Cascades into south central Washington, the Columbia basin, we break the section into three parts. Because of the thick sequence of low velocity basalts in the central basin, we start the transition of the upper crust farther to the east than we do the lower crust. For figure III-9 the crust-mantle boundary dips similar to the above model for north-east Washington from a depth of 44 km to 28 km. The top layer, however, remains thin (3.4 km) for 120 km before dipping to a depth of 11 km under the central basin. The observed seismograms for this profile fit the model remarkably well though the two stations YAK and ELL appear to be quite late. Like the previous example, this may be a case of the true least time path having an amplitude too small to be observed. There is a complicated pattern of secondary arrivals and some first arrival discrepancies over small distance ranges in the central basin. Since the stations for all of these profiles are really not in a line but rather are just within a narrow azimuth range it is not unreasonable to expect there to be some discrepancies such as observed here. Particularly in the central basin where the relief on upper and mid-crustal layers may be large, a variety of problems plague the interpretation. As a first cut our model seems to fit the general character of the seismograms quite well, though there is obvious room for improvement.

Figure III-9 and III-10 show profiles into the central Columbia basin from the Goat Rocks earthquake and the Granite Falls earthquake respectively. Both models are similar to the above Elk Lake to central basin model though the transition has been adjusted because of their different distances. For both of these figures the observed first arrivals in the central basin are quite early while the arrivals at intermediate distances seem pretty close to the theoretical ones. Secondary arrivals for both of these events are fairly clear but do not match the synthetic ones that well. We suspect that our model should have a transition to an E1 type model sooner and perhaps a more complicated mid-crustal transition. It seems the eastern Washington type crust may extend into the Cascades farther than we have modeled it here.

Thus far we have taken localized velocity models determined for separate regions and combined them into a transitional model and compared theoretical travel time curves and synthetic seismograms with observed refraction lines generated by earthquakes recorded on the regional network. This technique has shown that the transition between models can be modeled as a smooth one in most cases. Some modification to the transitions has been suggested in a few cases. We are beginning to experiment with more complicated models for a few of the cases where the data is the best. This work has not proceeded far enough to report on in detail at this time. We plan additional numerical experiments to not only work out some of the problems observed thus far but also to try modeling the upper crustal layers in the central Columbia Basin using both earthquakes and quary explosions as sources.

Dissertation Abstract of A. C. Rohay (1982)

Crust and Upper Mantle Structure of the

North Cascade Range, Washington

Teleseismic P-wave residuals recorded by a three-hundred kilometer long east-west network of seismic stations in northern Washington indicate a major high velocity anomaly beneath the north Cascades Range. Relative arrival time differences of up to two seconds are observed for events from eastern azimuths. There are much smaller travel time differences from the western azimuths. This pattern of residuals is compelling evidence for an eastward dipping high velocity slab beneath the north Cascades. The teleseismic residuals are modeled with a 40 km thick slab dipping 55 degrees to the east with a velocity of 8.5 km/sec. The velocity contrast is assumed to occur between 50 and 180 km depths within a 7.7 km/sec asthenosphere. The dip direction is variable, being N70E in northern Washington, rotating to a more easterly direction further south.

The teleseismic analysis is supported by a study of refraction data using the time term method which indicates that the Pn velocity is 7.8-7.9 km/sec, and that the crust is 40 km thick in the north Cascades, thinning to the east to 30 km. The crust has an upper 6.1 km/sec layer above tentatively identified intermediate

layers of 6.4 and 7.0 km/sec at 11 and 22-29 km depths, respectively.

An analysis of teleseismic residuals at stations on the western flank of the Cascades indicates that a change in the geometry of the slab occurs at 47.5 degrees north latitude. The change in the position of the travel time advances indicate an offset in the subducted slab to the west of the Garibaldi-Baker-Glacier Peak alignment. The slab is also interpreted to be much shorter in the southern regions. The position of the offset is correlated with changes in the deep seismicity and volcanic history of the region.

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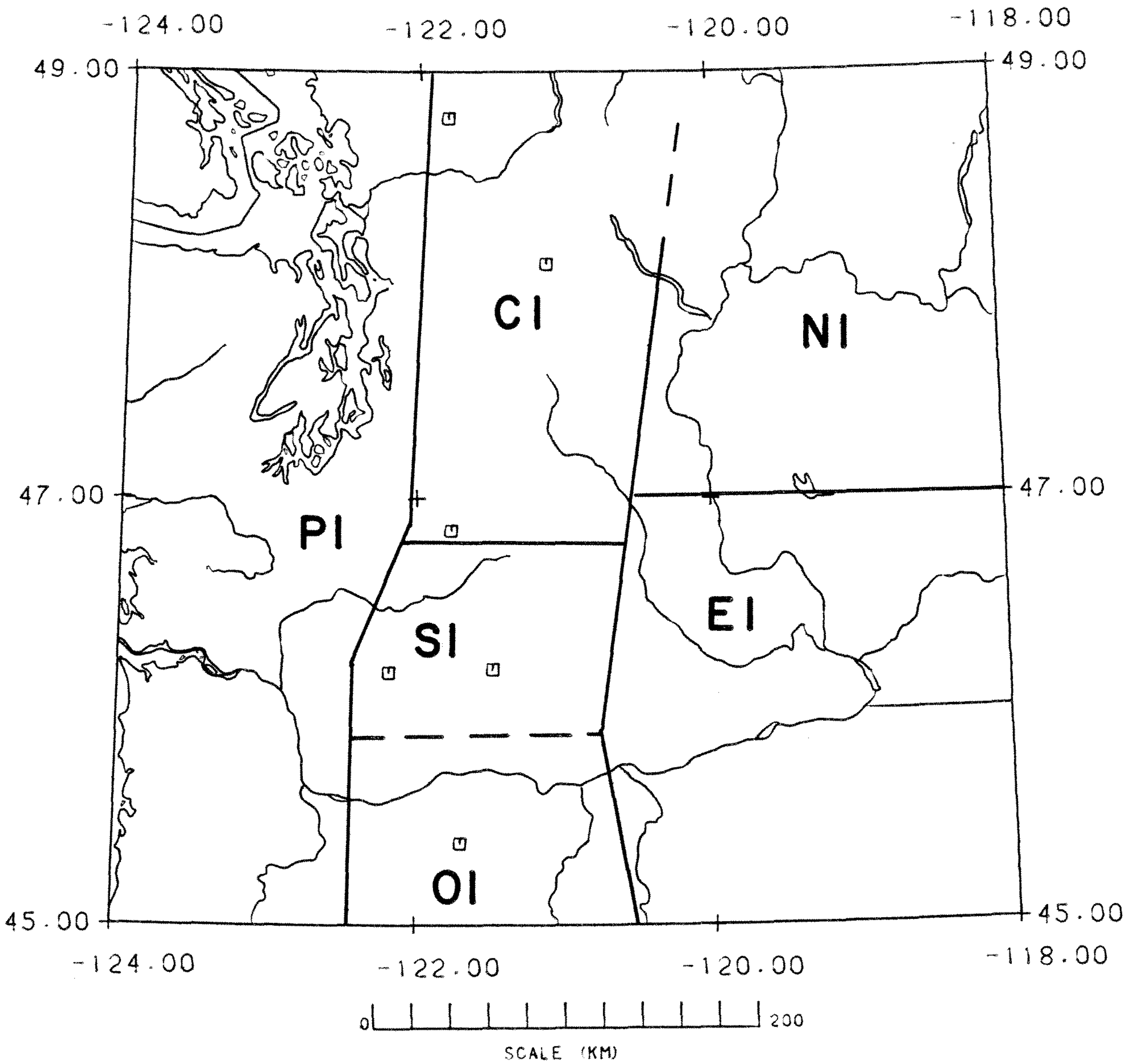

Figure III-1. Velocity model boundary map showing approximate location of transition between significantly different velocity models. The area marked *S1* is the Mount St. Helens area for which we use a combination of the C1 and O1 models. The specific model parameters for each area are found in table III-1.

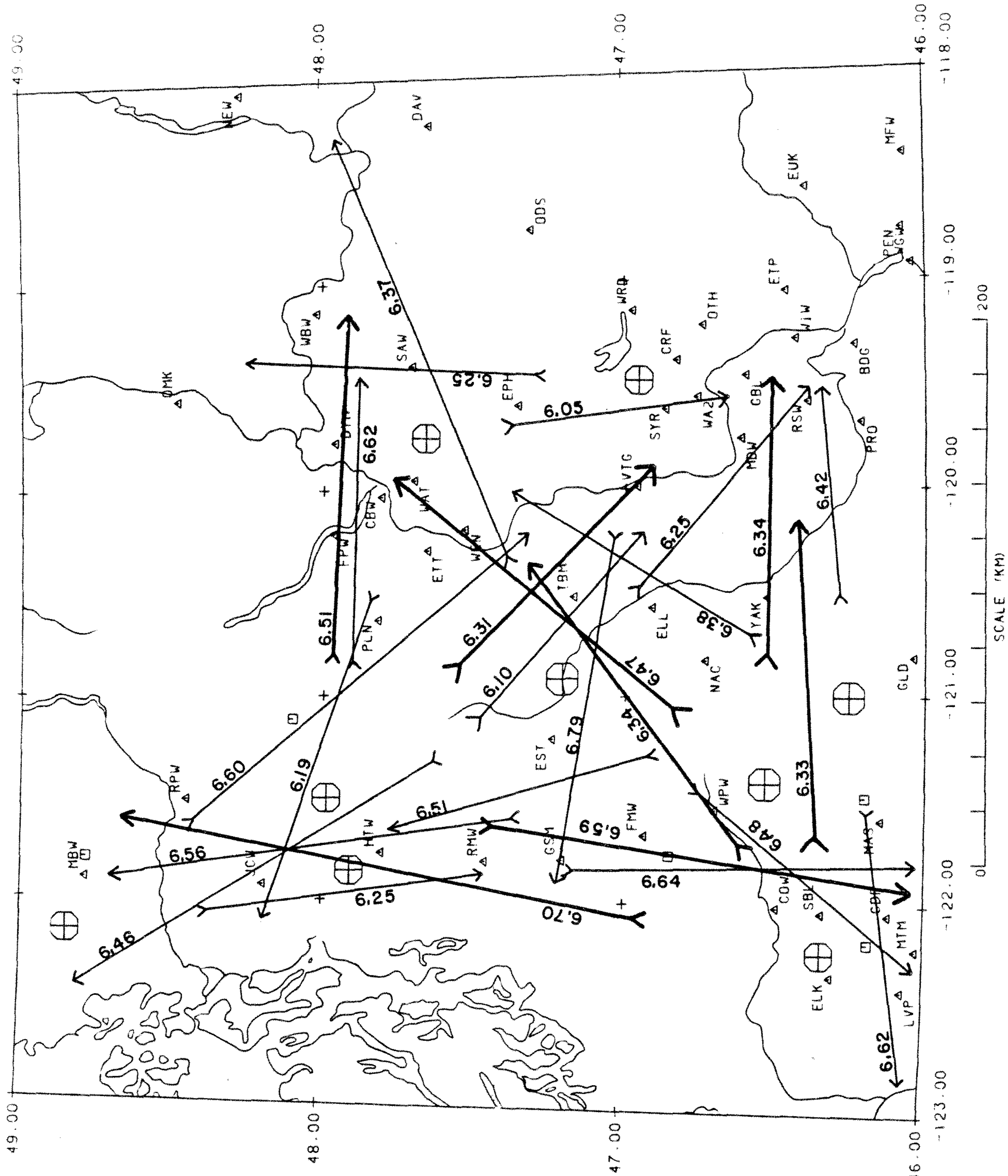

Figure III-2 Refraction profile map showing the various mid-crustal refraction lines put together using earthquake sources and network stations. Average major crustal velocity is indicated near each line. The heavier lines are those for which theoretical travel time curves and synthetic seismograms have been computed.