ANNUAL TECHNICAL REPORT: 1994-95

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Short Title:	Earthquake Hazard Research in the Pacific Northwest
Effective Date of Grant::	March 1, 1994
Grant Expiration Date:	April 30, 1995
Amount of Grant:	$71,700
Date Report Submitted:	Sept. 18, 1995

Sponsored by the
U.S. Geological Survey
Grant Number 14-08-0001-G1803

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1.Abstracts of Recent Papers

2. Manuscript submitted to BSSA entitled: "The 25 March 1993 Scotts Mills, Oregon, Earthquake and Aftershock Sequence: Spatial Distribution, Focal Mechanism, and the Mount Angel Fault" by G. Thomas and R. Crosson

Summary

This is the annual technical report for USGS grant 14-08-0001-1803, "Earthquake Hazard Investigations in the Pacific Northwest" for the period 3/1/94 to 4/30/95. Our research focuses on research efforts that have direct bearing on understanding earthquake hazards in the Pacific Northwest, including problems relating to large-scale plate interactions and possible large subduction earthquakes on the Cascadia subduction zone. Improvements in our understanding of earthquake hazards requires a better understanding of regional earth structure and tectonics. The research described here is aimed at achieving better total understanding of the structure and tectonics of the Cascadia subduction zone.

A primary source of data for this research is the Pacific Northwest Seismograph Network (PNSN). The PNSN now comprises over 140 telemetered short period seismograph stations in Washington and Oregon, plus 5 broadband seismograph stations in Washington which are tied directly into the PNSN network data acquisition. Data were also obtained and analyzed from a deployment of temporary stations operated during the Scotts Mills, Oregon earthquake sequence of March and April of 1993.

Investigations by our research group during this report period include completion and submission of two manuscripts to the BSSA, preparation of a manuscript on an earthquake-based refraction profile across the Cascade range of central Washington, completion of research on spectral analysis of data obtained in the Scotts Mills mainshock/aftershock sequence, completion of work on multiple reflections from teleseisms, preparation of a special report on the Robinson Point earthquake of 29 January 1995, and initiation of research on the effect of 3-D velocity structure on earthquake locations in the Puget lowland region of western Washington.

Manuscripts

During the period of this research, two manuscripts were submitted for publication, and several others are in advanced states of preparation. The paper by Dewberry and Crosson entitled: "Source scaling and moment estimation for the Pacific Northwest Seismograph Network using S-coda amplitudes" is submitted and in press for the Bulletin of the Seismological Society of America (BSSA). The paper by Thomas, Crosson, Carver and Yelin entitled: "The 25 March 1993 Scotts Mills, Oregon, earthquake and aftershock sequence: Spatial distribution, focal mechanisms, and the Mount Angel fault" has been submitted to BSSA and is in review. A manuscript by Thomas and Crosson entitled: "Spectral analysis of the 1993 Scotts Mills, Oregon, earthquake sequence" has been completed and is in final review prior to submission to Geophysical Research Letters (GRL). A manuscript by Schultz and Crosson entitled "A 2-dimensional P-wave velocity profile across the Cascade range of Washington State using earthquake sources and regional network observations" is nearing completion for submission to Journal of Geophysical Research (JGR). A special report on the M 5.0 Robinson Point earthquake by S. Dewberry entitled: "The Robinson Point Earthquake of 29 January 1995: An unusual deep crustal earthquake in the central Puget lowland" is being prepared for submission to Seismological Research Letters (SRL). This event is of special interest because it may offer insight into the tectonic behavior of the Seattle Fault.

Scotts Mills, Oregon earthquake sequence: M 5.6 mainshock, 3/25/93

In last year's final technical report, we discussed the findings related to analysis of the mainshock-aftershock sequence. Appendix 2 of this report is a preprint of the paper that was submitted to BSSA [Thomas, et al., submitted] on this analysis. In this paper, we concluded that there is strong evidence linking the Mount Angel fault zone with the occurrence of the Scotts Mills earthquake. This relationship must be taken into account in analysis of the earthquake hazards of the northern Willamette Valley region.

Spectral Analysis

In addition to the conventional preliminary analysis of the aftershock and mainshock phase arrival data (Appendix 2), the Scotts Mills earthquake sequence provided a rich set of high quality signals for further analysis of the waveform data. We have undertaken preliminary spectral analysis of these data, and completed a manuscript on these investigations as noted above. The results of this study indicate that there is strong site control of apparent corner frequencies as measured with the temporary aftershock network of wideband instruments. Sites on Tertiary volcanics have apparent corner frequencies 2 to 3 times higher than otherwise identical stations located on Quaternary sediments. We were able to measure true corner frequencies for selected earthquakes using a spectral ratio technique, effectively removing the site effects for several events. Using the spectral ratio technique, a stress drop of 48±29 bars and a fault rupture radius of 2.7±0.5 km are estimated for the M 5.7 mainshock. A paper presenting these results was given at the 1995 SSA meeting [Thomas and Crosson, 1995].

Using the low-frequency spectral levels of aftershock signals, we were also able to determine independent moments for a number of aftershocks. Moment magnitudes based on these measurements were compared with magnitudes based on coda-durations obtained from the PNSN shortperiod stations. Based on this analysis, we conclude that coda-duration magnitudes may underestimate the magnitudes, relative to the Hanks and Kanamori standard moment-magnitude relationship, by up to .5 magnitude unit for earthquakes below magnitude 2.5. This possible distortion is enough to significantly affect b values based on coda-duration magnitudes, causing slight underestimate of b values. Since b values are used for hazard estimation, refinement and possible adjustment for this bias is a priority for future research. This skew of the magnitude- moment relationship is shown on Fig. 1.

Availability of data

After the mainshock, groups from Denver and Golden (USGS) operated 15 portable digital recorders, including 11 DR200 recorders and 4 Reftek Passcal recorders for a period of 10 days beginning as early as about 18 hours after the mainshock [Carver, 1993]. These instruments all had short period, three component sensors. In addition, a USGS Menlo Park group operated 11 Geos instruments for a period of up to 20 days following the mainshock. Several analog recorders (MEQ's) were also operated by the University of Washington. Other groups, including Oregon State University and the University of Oregon, operated broadband portable instruments for various periods of time in the region.

The portable USGS instruments provide the only close-in digital recording of aftershocks

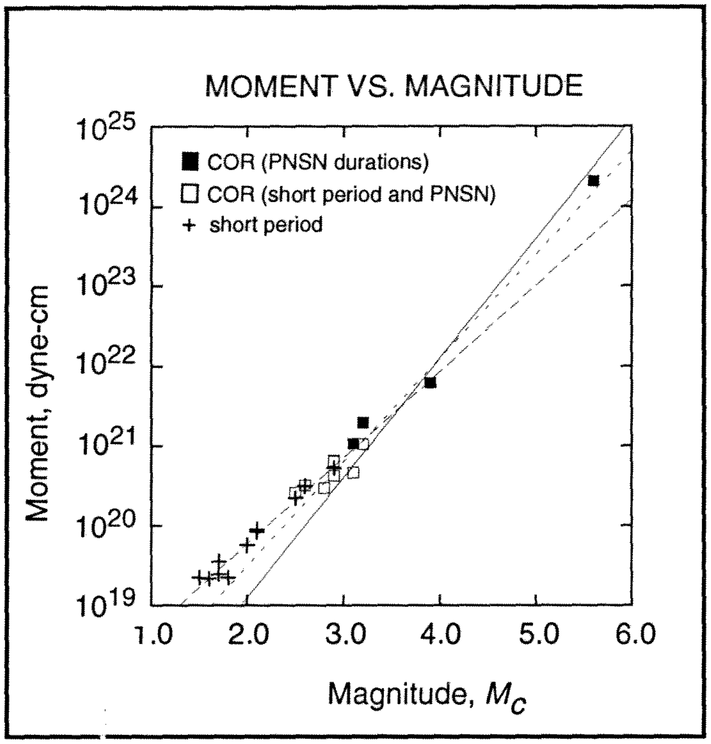

Fig. 1. Moment vs. coda duration magnitudes, Mc, for data from the Scotts Mills earthquake sequence. The solid line is the Hanks and Kanamori [1979] moment-magnitude scale. The dotted line is a linear fit to the broadband data, and the dashed line is a linear fit to the short period data. Filled boxes are broadband data with magnitudes estimated from PNSN durations, open boxes are broadband data with magnitudes estimated from PNSN and short period durations, and plus (+) symbols are short period data with magnitudes estimated from PNSN and short period durations. All magnitudes are from Thomas et al. [1995].

of this earthquake. Our group has assembled and reformatted the available digital data for the Scotts Mills aftershock sequence into a useful high quality data set for further analysis. The current data set of digital trace data is approximately 250 MB in size, and includes about 80 locatable earthquakes (aftershocks) recorded during the deployment of the portable digital stations. The data have all been event-parceled, and the data set includes all auxiliary files of locations, instrument parameters, and other data. This data set is unique in quantity and quality for the Pacific Northwest, and is publicly available, through the UW anonymous FTP site. The data have been acquired by John Nabelek of the Oregon State.

Array analysis of crust and upper mantle structure

We have essentially completed a study analyzing multiple reverberations of P waves from teleseismic waves recorded on the PNSN. Using deconvolution techniques, we originally hoped to be able to map single-bounce multiple reflections from the subducting Juan de Fuca slab over much of the western part of the PNSN. We have found however, that these reflections can be mapped over only a fraction of the network stations, either because of background noise in the short-period range, or due to site-specific scattering. The emphasis in this study has focused on causes of station-to-station variation in the reverberation response of the network stations. It is currently being completed as the Ph.D. thesis topic of Shawn Dewberry.

In conjunction with this investigation, we discovered a new and promising method of deconvolution of multi-station network data using so-called "cepstral" deconvolution. The method allows us to leverage the redundant information in network data to isolate the local receiver response from the common source response generated by the distant earthquake. Two papers on this subject were presented at the 1994 Fall AGU meeting [Crosson and Dewberry, 1994; Dewberry and Crosson, 1994]. Although this discovery was made too late in the current investigation to be fully exploited, it holds promise for several types of network investigation in the future.

Robinson Point earthquake: M 5.0 mainshock on 1/29/95

On 29 January 1995, an unusual M 5.0 earthquake occurred in the central Puget lowland. The event occurred at a depth of approximately 20 km, and occurred close to midway between the cities of Seattle and Tacoma. This earthquake was widely felt throughout the Puget Sound Basin although only minor damage was reported. Fig. 2 shows the epicentral location of the Robinson Point earthquake and a NS cross-section showing the mainshock/aftershock hypocenters.

Although the epicenter of this earthquake was almost coincident with the M 6.5 earthquake in 1965, this was a crustal event in contrast to the 1965 event which was clearly an intraslab event. The Robinson Point sequence includes one M 1.8 foreshock, the main shock, and 25 aftershocks recorded through 14 April, 1995. Aftershock epicenters cluster about the mainshock; the hypocenter distribution shows a steeply dipping distribution with the mainshock at the bottom, near 20 km (after relocation with modified station corrections). P-wave first motions from the mainshock indicate reverse faulting with east-west trending nodal planes. Figure 3 shows a cross-section view of the sequence and the P-wave polarity focal mechanism. The independently determined moment-tensor focal mechanism (by John Nabelek of OSU) was similar.

Minor damage was reported in Auburn, Tacoma, and Puyallup and shaking was felt throughout western Washington to as far away as Salem, Oregon and Vancouver, British Columbia. The Robinson Point mainshock was the largest earthquake to occur in the Puget-Willamette lowland region of the Pacific Northwest since the M 5.6 Scotts Mills earthquake of 25 March, 1993, and was the first moderate-sized Puget Sound earthquake to be recorded by the three-component, wide dynamic range, broadband instruments recently added to the PNSN. This broadband data provides the only unclipped local records of the mainshock.

The Robinson Point sequence included relatively few aftershocks, none larger than magnitude 2.2. A comparison of the Robinson Point aftershock sequence to aftershock sequences of several other moderate-sized, well-recorded crustal earthquakes (the 1981 Elk Lake [Grant et al., 1984], 1981 Goat Rocks [Zollweg and Crosson, 1981], 1990 Deming [Qamar and Zollweg, 1990], 1993 Scotts Mills [Thomas et al., 1993; Thomas et al., submitted], and 1989 Storm King Mtn.) provides some insight into the nature of deep crustal sources in western Washington. Aftershock sequences associated with the four moderate-sized crustal mainshocks at depths of 15 km or less (1981 Elk Lake, ~7 km; 1981 Goat Rocks, ~3 km; 1990 Deming, < 5 km; and 1993 Scotts Mills ~15 km) differ systematically from sequences following two moderate-sized deeper-crustal mainshocks (1995 Robinson Point, ~20 km; and 1989 Storm King Mountain~18 km). The four shallower mainshocks are characterized by: a significant number of aftershocks

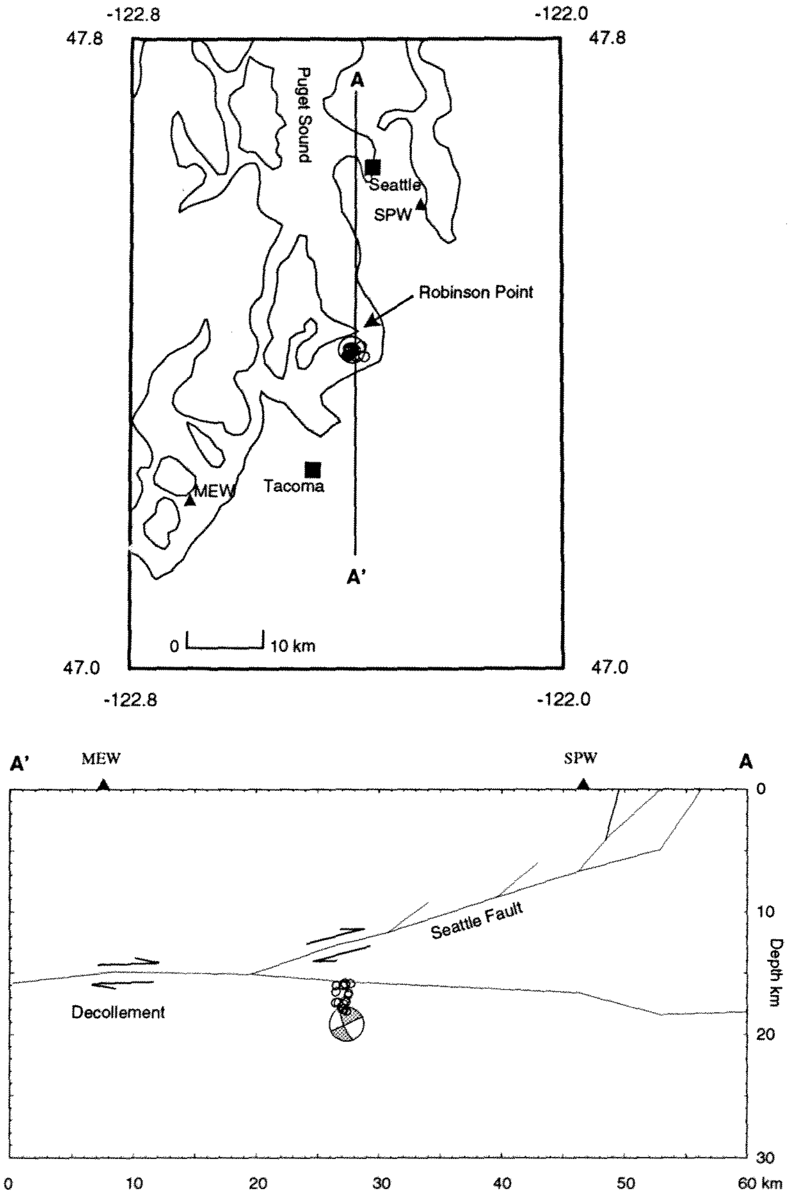

Fig. 2. Map (top) showing location of Robinson Point mainshock and aftershocks, and cross-section (bottom) along NS profile shown in top figure showing hypocenter depths for aftershock in relationship to the structure proposed by Pratt and others [1994]. The focal mechanism quadrant plot is the projection of the best focal mechanism onto the plane of the cross-section, showing the apparent alignment of the south dipping fault plane with the proposed Seattle fault extension at depth.

(70-600 events with M > 0.0) and an exponential decay in event frequency with time. In contrast, the deeper mainshocks show relatively few aftershocks (< 30 events) and no significant decay in occurrence with time immediately following the mainshock. In addition, both of the deeper mainshock sequences lack aftershocks larger than magnitude 2.4.

The close spatial relation of the Robinson Point sequence to the Seattle Fault (SF) is also of great interest. Gravity data, geologic evidence, and seismic reflection data indicate that the

Seattle Fault is an east-west trending blind thrust fault dipping to the south. The surface manifestation of this fault is approximately 30 km to the north-northwest of the mainshock epicenter as shown on Fig. 2. Geologic observations suggest that a reverse slip earthquake with 7 m of sudden uplift of a marine terrace at Restoration Point occurred on the Seattle fault about 1100 yr ago [Bucknam et al., 1992].

The cross-section in Figure 2 shows the projected position of the Seattle Fault at depth according to Pratt et al. [1994]. The horizontal line is the intersection of the decollement surface proposed by Pratt et al. to lie at about 17 km depth beneath the entire Puget lowland. In this model, the Seattle fault, which is nearly vertical at the surface, becomes a south-dipping thrust fault at depth, joining the decollement surface in the vicinity of the Robinson Point earthquake hypocenter. Although the cross section of Fig. 2 shows the sequence occurring beneath the decollement and not on the Seattle fault (or its depth extension), there are significant uncertainties about the depth of the decollement surface and the dip and position of the Seattle Fault at depth. In part, this uncertainty stems from our lack of knowledge of an accurate 3-D velocity structure in the central Puget lowland.

In general, the suite of deeper crustal earthquakes in the Puget lowland lies beneath the decollement surface proposed by Pratt et. al. Whether or not the Robinson Point earthquake sequence occurred on the Seattle Fault remains unresolved at present. However, its depth and focal mechanism suggest that it may be related to the Seattle fault or to the decollement system. A better understanding of this event, and other deep earthquakes within the Puget lowland is important for understanding and testing the new models, and ultimately for our understanding of earthquake hazards in this region. In order to study these earthquakes, we need a better model of the 3-dimensional seismic velocity structure of the Puget lowland and vicinity. Because of the potential special significance of this earthquake, we are currently preparing a special report on the Robinson Point earthquake for submission to Seismological Research Letters or to BSSA, short notes [Dewberry, in prep.].

3-D velocity structure of the Puget lowland

In papers presented at the 1994 Fall AGU meeting [Pratt et al, 1994; Stephenson and Pratt, 1994], Pratt and others presented a new tectonic and structural interpretation of the central Puget lowland. This model, referred to here as the PJPS model, is based primarily on newly available industry seismic reflection data obtained in Puget Sound, in addition to gravity, magnetic, surface geology, and inference from other regions. In the PJPS model, the Puget lowland lies on a north-directed thrust sheet of 10 to 17 km thickness, which is cut by thrust faults and deformed by fault-bend and fault-propagation folds. The thrust sheet is ultimately driven by oblique motion on the Cascadia subduction interface, although the details of this mechanism remain obscure.

The Seattle Fault is of particular interest and concern due to its proximity to the major population centers of western Washington. Although this fault could be inferred from the gravity based interpretation done by Danes et al. [1965], recent investigations have produced considerably more information about its structure and potential for earthquake generation. Evidence from Restoration Pt. in the central Puget lowland, and elsewhere around Puget Sound suggests that the Seattle Fault may have generated a large earthquake in the immediate vicinity of Seattle about 1000 ybp [Yount, 1992; Johnson, et al., 1994; Bucknam et al, 1992; Atwater and Moore, 1992]. More than 3 meters of displacement may have occurred during that event [Bucknam et al, 1992]. The SF is an important structural element of the PJPS model.

Understanding the SF, its structure and driving mechanism is therefore of major importance in assessing earthquake hazard in western Washington.

In the PJPS model, the Puget lowland thrust sheet is bounded by strike-slip faulting on the northeast and east (S. Whidbey Isl. fault, Puget fault) , and possibly the west (Hood Canal fault on the east side of the Olympic Mts.). Within the thrust sheet, there are several sets of down-dropped and uplifted blocks, that correspond in general to known sedimentary basins with relative gravity lows and arches with relative gravity highs. The Seattle basin, north of the SF, is the most pronounced of the basins. The SF is a (blind) thrust fault dipping 17° to 25° to the south, extending from near the surface to about 15 km depth where it intersects the decollement. The basal unit within the thrust sheet is the Crescent formation, basaltic accreted terrain of Eocene age that is exposed in upturned and faulted sections along the east margin of the Olympic Mts. The thrust sheet model is consistent in a general way with focal mechanisms of central Puget lowland crustal earthquakes. Many of these earthquakes exhibit thrust mechanisms on north dipping or south dipping fault planes that are consistent with N-S compressions (Ma et al., in press).

The PJPS model, whether it is correct in detail or not, makes it clear that the Puget lowland is underlain by complex geology that presents extreme lateral seismic velocity contrasts from the Crescent volcanic terrane with typical P wave velocities of 6.5, to deep sedimentary basins with P wave velocities as low as 2-3 km/sec. It is clear that we must consider the effects of 3-dimensional structure in locating and studying earthquakes. Questions have arisen about the influence of 3-D structure on producing the scatter of earthquake hypocenter locations based on 1-D velocity structure. An important component of this research project, begun during this report period, is the investigation of 3-D velocity structure and its influence on earthquake locations and other types of analysis. Here we briefly describe the results of our initial studies.

In our initial investigations of the relationship between earthquakes in the central Puget lowland and the PJPS model, we chose a population of 1057 previously well-located crustal earthquakes in the central Puget lowland. We wished to examine the accuracy of hypocenter locations, particularly depth, in relationship to the PJPS model. These earthquakes all have digital records from the PNSN, and were initially chosen using the following criteria: M > 0.5, Gap < 100°, Depth 2 to 35 km, P phases > 4, S phases > 0, Time period 1980 - present, Qual. BB or better.

To accurately compare hypocenter locations with the structural model, we created a digital representation of major features of the PJPS model using formats compatible with Xmap8 (Lees, 1995) (Xmap8 is an interactive X-window based application for viewing, manipulating, and printing a variety of earthquake and structure information on an interactive basis). Using Xmap8 and our digital representation of the major fault surfaces of the PJPS model, we are able to plot maps and cross sections of these structures combined with earthquake hypocenters, velocity model information, station locations, and other important data.

Fig. 3 is an Xmap8 image of epicenters of the selected events in the target region of our investigation. This figure also shows most of the stations used in the relocation experiments, cross-section locations, and the surface intersection of the main branch of the SF in the central part of the figure.

The locations of all earthquakes in Fig. 3 were initially determined using a conventional earthquake location program (LQUAKE) which we have used and refined over a period of years. LQUAKE is based on Geiger's iterative non-linear least squares method, and uses a 1-D, piecewise constant, velocity model for both P and S wave first-arrival ray-tracing. The locations

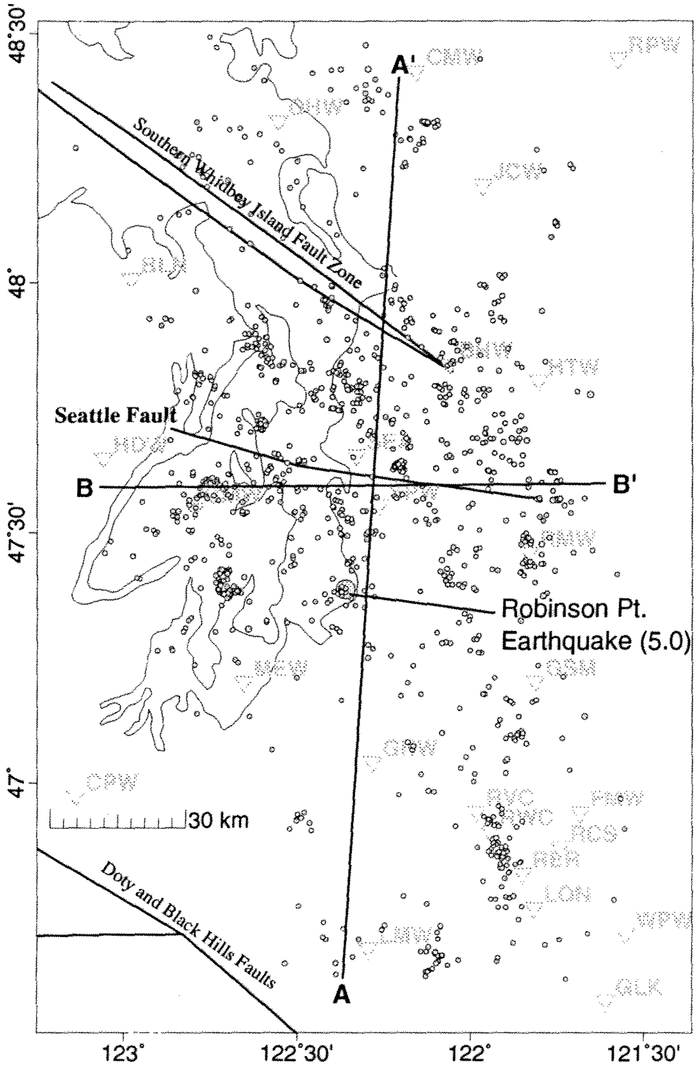

Fig. 3. XMAP8 [Lees, 1994] epicenter map for the central Puget lowland region, showing principal thrust faults from the Pratt et al. [1994] model. Two cross-section lines are shown as A-A' and B-B' which are referred to in later figures. Corridor widths for these cross-section projections are about 12 km. Faults of the Pratt et al. model are entered as physical surfaces, and the intersection of these surfaces with the cross-section profiles are shown on the cross-section figures.

determined by LQUAKE have been compared extensively to other location programs such as HYPO71, HYPOELLIPSE, and FASTHYPO; all produce similar solutions. Hypocenters were initially determined with the standard model (PS2) used for routine location of earthquakes in the Puget lowland (Crosson, 1976). Most of the hypocenters from the initial location calculations lie between 15 and 30 km depth. Although there is some localized clustering of epicenters, there is no obvious spatial relationship between epicenters and the projections of known or inferred faults.

Fig. 4 shows 12 km wide corridors (cross-sections) along A-A' and B-B' on Fig. 3. Section A-A' shows the SF in cross-section, the Seattle basin, and the decollement surface from the PJPS model. Section B-B' intersects the SF (actually 2 branches) along its strike direction. Owing to the corridor widths, these projections do not show exactly the same earthquakes.

From Fig. 4 we can see clearly that most of the "well located" hypocenters in the Puget lowland lie significantly deeper than the proposed decollement. Furthermore, the scatter of hypocenter depths makes the association of hypocenters with the decollement problematical, although there appears to be some tendency of the hypocenter depths to mirror the decollement profile. Of course, it is important to recognize that the decollement itself is indirectly inferred

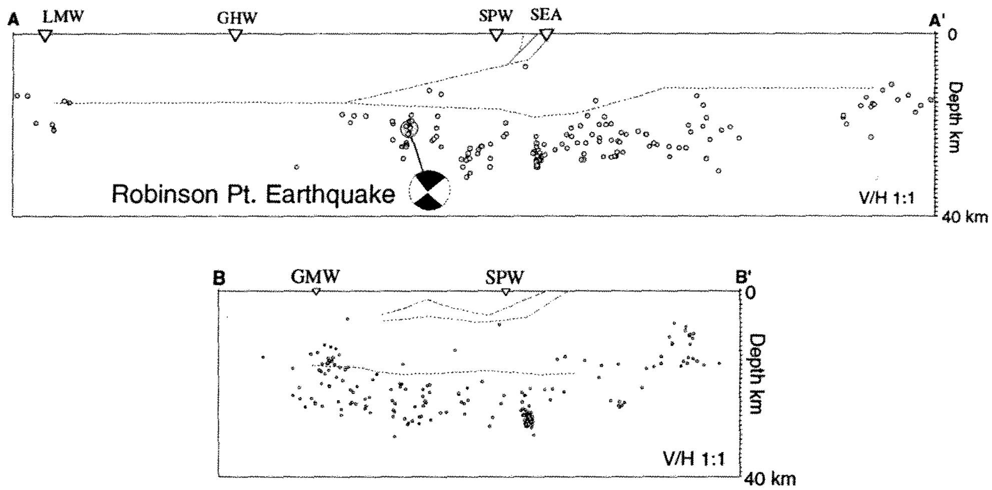

Fig. 4. Cross-sections A-A' (top) and B-B' (bottom) from Fig. 3, showing the hypocenters for test data set using the PS2 velocity model in relationship to the principal fault surfaces of Pratt et al. [1994] model for central Puget lowland.

from structure in the upper 10 km, and is not directly imaged in the seismic reflection interpretations. Two clusters of earthquakes that appear on these two cross-sections, near the Robinson Pt. hypocenter and near stations SPW and SEA, appear to have vertical extension and do not correlate with possible thrust faults.

If these earthquakes are indeed related to the decollement, we must ask if the observed depth scatter results from lack of depth resolution. The same concern can be raised about the vertical clusters - are they real or due to inadequate depth control? Is the overall pattern of earthquakes at mid-crustal depths in the Puget lowland related to deformation at the decollement, and thus to variations in the thickness of the thrust sheet? Can the decollement interface be refined with earthquake data? To answer these questions, we need to establish unequivocally the depth accuracy of hypocenters determined from PNSN observations.

As a first test of hypocenter locations, we computed the changes in locations of our test data set due solely to refinement of station corrections. To make valid comparisons of locations, we used a fixed set of 30 stations within the rectangle 46.75° to 48.5° N, and 121.0° to 124.0° W to relocate all earthquakes. We refined station corrections by several cycles of (a) relocation and computing residual statistics for all locations, and (b) adding the median residuals back into the station corrections. Large residual outliers and stations with poor residual statistics were checked and adjusted individually. The result of these adjustments in station corrections were only minor (typically less than 1 km) changes in hypocenter depth, and virtually no change in event epicenters.

As noted above, the PJPS model and a number of other lines of evidence (e.g., Miller et al., 1995; Pullen et al., 1994) indicate that the velocity structure of the crust in the vicinity of the Puget lowland is quite 3-dimensional in nature. Therefore, we ultimately need to compute earthquake locations based on true 3-D structure. A central problem in using 3-D velocity models is the difficulty in accurately computing travel times from source to receiver. Ray tracing algorithms may have difficulty in computing first arrival times, since solutions found by ray tracing cannot be guaranteed to be first arrivals. To circumvent this difficulty, we have implemented a version of the finite difference (FD) earthquake location procedure of Nelson and Vidale (1990). This procedure, which we refer to as FDL (Finite Difference Location), uses the 3-D finite difference travel time calculation method of Vidale (1990), as modified by J. Hole (pers. comm.).

The FDL earthquake location procedure has several significant advantages over conventional non-linear least squares methods, including: (a) being able to correctly characterize arbitrary 3-D structure for first arrival phases, (b) the procedure is non-iterative, (c) the procedure allows circumvention of the local minimum problem, (d) it is possible to compute confidence intervals correctly based on the full non-linear minimization problem, and (e) the use of alternative minimization criteria can be done without additional computation. There are also disadvantages of the method, such as: (a) intensive numerical work is required to compute travel times from a large number of discrete points in the model to every station (this needs to only be done once for each station in any given location run; thus efficiency is improved by processing many events in one processing run), (b) storage requirements are considerable, (c) only first arrivals are calculated, although they may not be observed, (d) accuracy may be limited in certain parts of the model depending on the structure of the FD grid and other factors.

In spite of the added computation burden, the advantages of FDL are very significant for the earthquake location problem. We have extended the method to include calculation of confidence regions based on single and two-sided F-tests, and we are planning to implement further modifications to enhance the accuracy of the FD travel time calculations. Future enhancements of the FD travel time calculations may be possible, using finite bandwidth schemes, to minimize the problem of computing observed arrivals that have amplitudes too low to observe. These issues are the subject of future research as more experience is gained with this method. The FDL method can also lead directly to full 3-D tomographic inversion of the velocity structure in which both hypocenter and velocity model parameters are determined simultaneously (one form of this inversion was done by Ammon and Vidale, 1990).

For the FDL procedure, we used a model grid spacing of approximately 1.3 km in x,y, and z, over the latitude, longitude, and depth ranges: 47.75° to 48.5°N, 121.5° to 124.0°W, -10 to 52 km depth. The total number of nodes in the model grid is 1.42 million, and the total number of stations within the grid space is 36 (although no single event has more than 30 observations for any one event). We ran a series of comparison tests of the FD travel time calculations against conventional ray-tracing using a 1-D model. In these tests, we found a maximum travel time difference of .15 sec at a few places near the edges of our grid or near strong velocity gradients. Over most of the model, the time differences were only a few hundredths of a second. We attribute larger differences to the inability of the FD scheme to exactly simulate the piecewise constant velocity model used in the ray tracing calculations.

To speed the grid search required for earthquake locations, we use a coarse grid initially to eliminate uninteresting parts of the location space, followed by a fine grid for location refinement near the minima of the objective function. This procedure results in a large gain in speed without sacrificing accuracy. The hypocenter problem is generally well enough behaved that there seems to be little danger of missing important global, or even local, minima with this multi-resolution approach.

Since in the FDL method, the objective function (RMS weighted residual) is computed directly at every point in the FD grid, we can establish correct statistical tests to map out confidence regions in 3-dimensional space. In fact, such confidence regions can be calculated and mapped in lieu of discrete locations if desired. Although more costly than conventional location methods, FDL lends itself to processing large batches of earthquakes due to the fact that the travel time tables are computed once, stored in virtual memory, and accessed as infrequently as possible. An enormous advantage of the FDL method is that it correctly handles arbitrary velocity models. We believe that it is the most practical method for computing hypocenter locations for complex 3-D structure models.

To test our initial implementation of FDL, we compared FDL locations to LQUAKE locations discussed previously using identical station corrections for both sets of calculations. The results of this comparison are shown in a cross section in Fig. 5 (cross section is comparable

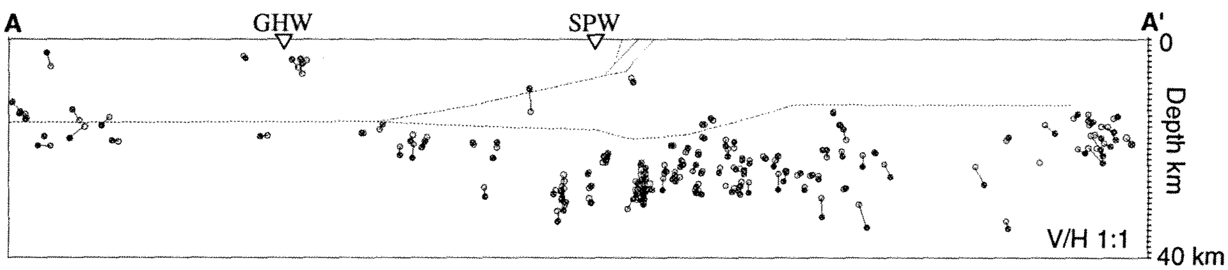

Fig. 5. Cross-sections showing the changes in location resulting from using our adaptation of the 3-D FDL location procedure of Nelson and Vidale [1990]. For this test, we used an approximation to the PS2 (1-D) velocity model. Only small event depth migration is observed; we interpret this to result from slight differences between the FD representation of our 1-D model, and the layered model used with the conventional location method. For this example, closed circles are the original LQUAKE locations, and open circles are the FDL locations.

to Fig. 4; closed circles are the LQUAKE locations, and open circles are the FDL locations). For this test, we used a 3-D FD representation of the PS2 1-D velocity model for the Puget lowland. Although a small number of hypocenter depths have decreased by as much as 5 km, most hypocenters have moved less than 1 km in this experiment, indicating that the FDL method is working correctly, and that the locations based on the PS2 model are quite stable. We attribute the minor changes in hypocenter locations to be mostly the result of differences in the FD representation of the velocity model compared to the piecewise constant representation used by LQUAKE. For each model, station corrections have been adjusted empirically as described above. We conclude that the FD and conventional location schemes give the same results when the models are comparable.

Owing to the fact that overestimating near-surface velocity typically results in computed hypocenter depths that are greater than their true depths, there is concern that our existing "standard" 1-D velocity model may produce hypocenters that are significantly deeper than their true depths. To test this possibility, we investigated the effect of lowering the P and S wave velocity for depths less than 9 km. The original PS2 velocity model reflects average crustal velocity over a larger region than the Puget lowland, and therefore does not correctly model the substantial velocity reduction in the sedimentary rocks of the central basin. The 1-D joint model/hypocenter inversion procedure upon which PS2 is based, has little resolution within the upper 10 km due to a preponderance of hypocenters in the 15-30 km depth range and relatively large station spacing.

For the station and event distribution of the Puget lowland, in 1-D inversion many raypaths travel nearly vertically in the upper 10 km beneath stations, and therefore there is a direct tradeoff between near surface velocity and station corrections. Station corrections were included in the original inversion (Crosson, 1976), and no-doubt had the effect of trading off with near-surface velocity. For example, station SPW had an original station correction of about 1.2 sec from the velocity inversion. SPW is one of the few stations situated within or near the Seattle basin. To a limited extent, station corrections can account for near surface lateral velocity heterogeneity.

Fig. 6 shows a modified composite 1-D velocity structure that we tested for earthquake locations. The P wave velocity structure of PS2 is shown for comparison as a dashed line in the

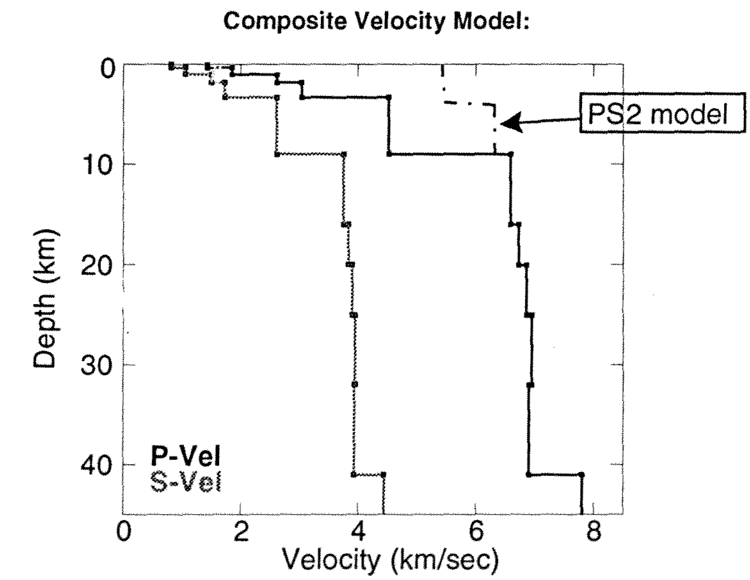

Fig. 6. Composite 1-D velocity model used in relocation tests to check the effect of pronounced low-velocity in the upper 10 km. The P velocity in the upper 9 km of the crust is based on reflection interpretation stacking velocities by Pratt (pers. comm.). Below 9 km depth, the PS2 model is used.

upper 9 km. The upper 9 km velocity structure is based on stacking velocities interpreted from the seismic reflection analysis by Pratt et al. (1994), and obtained from Pratt (pers comm). S wave velocities were assigned using a regional average Poisson's ratio. Deeper than 9 km, we used the PS2 velocity structure, so that the strongest velocity contrast in the model occurs at 9 km depth.

After readjusting station corrections for the modified 1-D velocity model, we relocated the test data set again using LQUAKE. The results of this relocation are shown on Fig. 7, with open circles as the original locations, and filled circles as the relocated hypocenters. Note that Fig. 7

includes only a subpopulation of all of the located events along a corridor that is near but slightly west of line A-A' of Fig. 2.

Fig. 7 illustrates some surprising but important results. Although most of the hypocenter depths changed only a small amount (1 km or less), a sub-population of about 20% of all hypocenters have now moved to depths shallower than 10 km. The subpopulation along corridor A-A' shown in Fig.7 that has migrated to shallow depth appears to be somewhat larger than the 20% represented by the entire suite of test events.

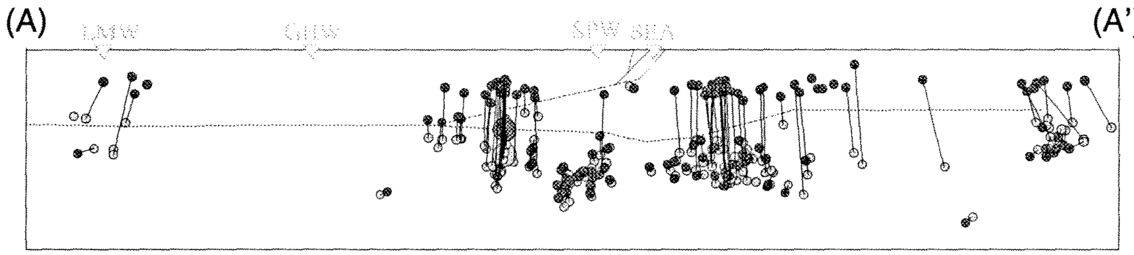

Fig. 7. Cross section slightly west of section A-A' of Fig. 2, showing migration of hypocenters from original PS2 locations to locations using composite 1-D velocity model described in the text. A sub-population of hypocenters migrate to a depth near 9 km depth where there is a strong velocity contrast (as shown in Fig. 6). This effect is due to a double minimum in the objective function resulting from inadequate S arrival control and takeoff angles that are dominantly in the upper hemisphere. We believe that the shallow depths are an artifact, and do not reflect the true locations.

The reason for this behavior is illuminating - the strong velocity contrast in this region produces a situation in which two minima in the objective function, near 8 km depth and near 20 km depth, exist. Some of the hypocenters have migrated to the shallow minimum while others remain at the deep minimum. We believe that this effect is an artifact of the strong velocity contrast at 9 km depth, and is not necessarily a real property of the earth. If there is a velocity contrast near 9 km depth, it may not be on the average as great as we have modeled. On the other hand, this exercise demonstrates that it is possible to have true double minima for particular combinations of earth model and data distribution.

The strong velocity increase near 9 km depth produces a situation where locations immediately above the interface result in all stations receiving refracted rays with high horizontal apparent velocities which are controlled by the mid-crustal rock. If the location were shallower, the cross-over to some direct rays at nearby stations drives up the residual variance, constraining the depth. The residual saddle between 9 and 20 km depth is more difficult to understand and is the subject of further investigation. We are quite certain that this result is both possible and real however, since two totally independent location procedures (LQUAKE and FDL) reflect the presence of the double minimum. Station and observation distributions must be critical in such cases. Owing to the fairly large station spacing in the Puget lowland (we estimate 30-35 km average spacing), and the paucity of high quality S wave observations, depth resolution problems may be more prevalent that we have recognized. Further research into the causes and liklihood of such double location minima is needed to resolve these earthquake depths with confidence. Future research will also focus on the location effects of realistic 3-D velocity models for the Puget lowland.

Publications funded under this grant

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Abstracts

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Acknowledgments

Work under this grant could not be carried out without the data collected by the PNSN; we recognize the dedicated contributions of the electronics technicians, data analysts, and other staff who contribute to the operation of the network.

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