High-pressure metamorphism and uplift of the Olympic subduction complex

Mark T. Brandon, Arthur R. Calderwood*
Department of Geology and Geophysics, Yale University, P.O. Box 6666, New Haven, Connecticut 06511

ABSTRACT

The discovery of the critical assemblage lawsonite + quartz + calcite indicates that a significant part of the Cenozoic Olympic subduction complex of northwestern Washington State formed by underplating at a depth of about 11 km. The deep structural level exposed in this area is attributed to the presence of a 10-km-high arch in the underlying Juan de Fuca plate. We postulate that this arch was formed when the southern Cordilleran coastline swung westward as a result of middle Miocene to recent extension in the Basin and Range province.

INTRODUCTION

The Cascadia convergent margin, which flanks the western side of Oregon, Washington, and Vancouver Island, marks the subduction boundary between North America and the Juan de Fuca plate. Along much of this margin, the fore-arc region is underlain by a flat to gently dipping sheet of lower Eocene basalt (Crescent, Siletz, and correlative formations; Wells et al., 1984), which represents the basement on which a subsequent fore-arc basin has accumulated (Shouldice, 1971; Snavely et al., 1980; Heller et al., 1987). This relation is well documented in a series of cross sections across the Cascadia margin (presented in Kulm et al., 1984). The Eocene basalt and fore-arc basin strata were collectively called the Peripheral rocks by Tabor and Cady (1978a, 1978b).

In contrast, the Olympic Peninsula (Fig. 1) marks an anomalous segment of the Cascadia margin. Uplift in this area has tilted the Peripheral rocks into a steep, east-plunging anticline. The core of this structure shows that the lithospheric base of the Peripheral rocks is truncated by a major thrust fault, the Peripheral fault (Fig. 1). Lying structurally below this fault is the Olympic subduction complex, which consists of an assemblage of thrust-imbricated marine turbidites and minor pillow basalt (Core rocks of Tabor and Cady, 1978a, 1978b). Fossil ages from the Olympic subduction complex range from Eocene to Miocene; most Eocene ages are from the basalts.

We maintain that the Olympic uplift provides a rare view of the deep structure of the Cascadia convergent margin. The Peripheral rocks represent a relatively coherent structural lid (i.e., rearward-dipping backstop), beneath which the Olympic subduction complex accumulated by subduction underplating. In this paper we examine the metamorphic and uplift history of this subduction complex and postulate a new interpretation for formation of the Olympic uplift.

METAMORPHISM OF THE OLYMPIC SUBDUCTION COMPLEX

Tabor and Cady (1978b) delimited a general metamorphic zonation decreasing in grade from east to west. They based their zonation on the following index minerals: epidote + chlorite, pumpellyite, prehnite + pumpellyite, and laumontite. Their compilation included the work of Stewart (1974) on laumontite-bearing rocks in the western Olympic subduction complex and that of Hawkins (1967) on prehnite-bearing rocks in the Mount Olympus area (MO in Fig. 1). Zones in the remaining areas were delimited using samples collected during their mapping (Tabor and Cady, 1978b) in the central and eastern Olympic Peninsula.

We have examined thin sections from 138 sandstone samples from all parts of the Olympic subduction complex; most, however, are from the central part. Of these samples, 71 were provided by R. W. Tabor and the remaining 67 were collected by us as part of a fission-track dating project. Medium-grained volcaniclastic sandstones provide a sensitive indicator of met-

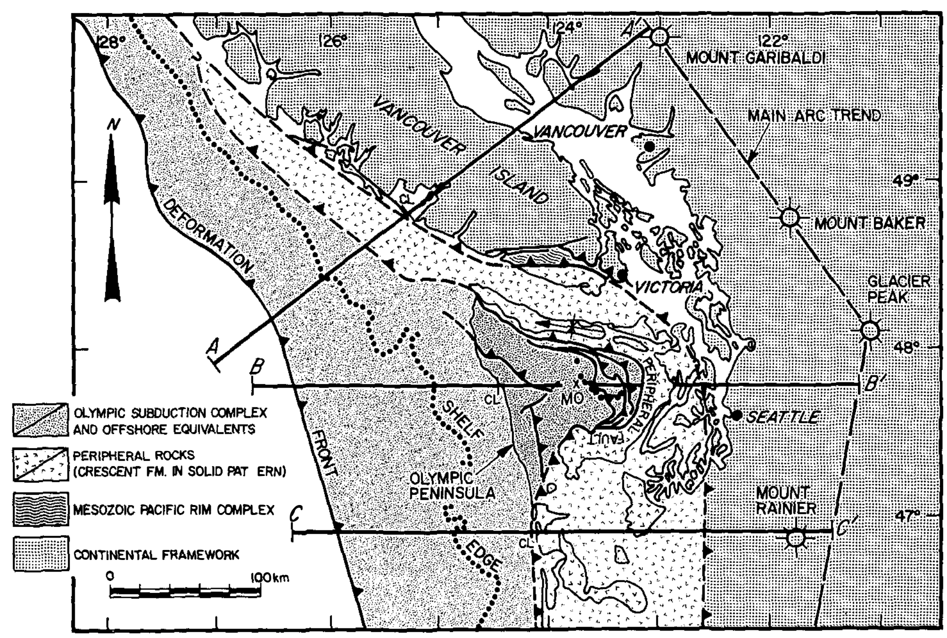

Figure 1. Geologic map of Cascadia margin in western Washington and southern Vancouver Island. Small solid circles indicate lawsonite localities in Olympic subduction complex. MO marks Mount Olympus, highest mountain on Olympic Peninsula. Cross sections are shown in Figure 2; alignment marks labeled CL (coastline) are used there.

*Present address: Department of Geological Sciences, University of British Columbia, 6339 Stores Road, Vancouver, British Columbia V6T 2B4, Canada.

GEOLOGY, v. 18, p. 1252–1255, December 1990

amorphic conditions at very low grades because of reactions involving calcium-aluminum silicate minerals (e.g., volcanic plagioclase, laumontite, prehnite, lawsonite). Most samples are dominated by a nondiagnostic assemblage of calcite + white mica ± chlorite. Lawsonite, prehnite, and laumontite react to form this assemblage when X_{CO_2} is greater than about 4% (Thompson, 1971; Nitsch, 1972). Many samples also showed extensive development of fibrous overgrowths and discontinuous dissolution seams, both of which are associated with the development of a widespread pressure-solution cleavage. The overgrowths consist mainly of quartz and white mica.

Our observations suggest the following modifications of the metamorphic zonation of Tabor and Cady (1978b). Starting with their easternmost zone, epidote + chlorite, we found no metamorphic epidote, but detrital epidote is quite common. Even if metamorphic epidote is present, the assemblage epidote + chlorite has a broad stability range that overlaps completely with the prehnite-pumpellyite facies (Frost, 1980); it is not particularly diagnostic of metamorphic grade. Fission-track dates for sandstones from this zone indicate that maximum temperatures were between 100 ±10 °C and 200 ±50 °C (the blocking temperatures for apatite and zircon, respectively; see Gleadow and Brooks, 1979); apatite is annealed and zircon is not (Brandon et al., 1988b, and unpublished data).

Tabor and Cady (1978b) identified a pumpellyite zone to the west, followed by a prehnite + pumpellyite zone. We agree with their zonation; however, we have found ten samples in these two zones (small solid circles in Fig. 1) that contain a distinctive fibrous intergrowth of lawsonite and pumpellyite. The full metamorphic assemblage is lawsonite + pumpellyite ± prehnite + quartz + white mica ± calcite ± chlorite. The lawsonite + pumpellyite intergrowth appears as fibrous mats that are "pencil-lead" gray in plane light and locally display a patchy first-order orange birefringence (crossed nicols). This particular mineral habit is fairly common in older rocks to the north and northeast of the Olympic Mountains (San Juan Islands and North Cascade Mountains; Glassley et al., 1976; Brandon et al., 1988a; Brandon, 1989). Previous workers in those areas concluded that the mineral intergrowth was composed solely of lawsonite; however, further X-ray diffraction and microprobe analyses indicate that it is a submicroscopic intergrowth of lawsonite and colorless Al-rich pumpellyite. Deer et al. (1986, p. 217) discussed some coarser grained examples of coherent intergrowths of pumpellyite and lawsonite. We propose that the two metamorphic zones in this area should be renamed pumpellyite + lawsonite and prehnite + pumpellyite + lawsonite. Fission-track dates from these zones indicate maximum temperatures ranging from 100 °C to more than 200 °C (apatite is usually annealed, and two out of three zircon samples are annealed; Brandon et al., 1988b, and unpublished data).

In areas other than the Olympic Mountains, it has been shown that prehnite, as well as lawsonite, can form within the aragonite stability field (Brandon et al., 1988a). For the Olympic samples, we have found only calcite. Aragonite commonly inverts to calcite during uplift of high-pressure metamorphic rocks. However, this retrograde form of calcite is typically very fine grained (Vance, 1968). The calcite in our samples occurs in grains as large as 0.2 mm across and in some cases envelopes smaller grains of prehnite. Thus, we conclude that calcite was a primary phase during high-pressure metamorphism.

The last and most western metamorphic zone is the well-documented laumontite zone of Stewart (1974). Our observations are compatible with Stewart's work, although our samples are limited in this area.

We suggest that the most deeply exhumed part of the Olympic subduction complex coincides with the prehnite + pumpellyite + lawsonite zone in the Mount Olympus area. Lawsonite surely indicates moderately high pressure. Furthermore, prehnite has probably formed by a prograde reaction involving the breakdown of either laumontite (Thompson, 1971) or lawsonite (Brandon et al., 1988a, p. 38). Thus, the occurrence of prehnite suggests temperatures higher than those in adjacent zones. It is interesting that this part of the Olympic Peninsula has the greatest amount of local topographic relief (about 2 km in the vicinity of Mount Olympus), which implies significant recent uplift.

Peak metamorphic conditions in this zone are interpreted to be within the lawsonite + quartz stability field and at pressures slightly above the reaction laumontite = lawsonite + quartz + water because of the adjacent laumontite zone to the west. These constraints indicate temperature and pressure conditions of about 190 °C and 300 MPa (relevant experimental and thermochemical data are summarized in Brandon et al., 1988a, Fig. 28). This pressure corresponds to a depth of 11 km, assuming a mean density of 2700 kg/m^3.

TIMING OF THE OLYMPIC UPLIFT

The fore-arc basin strata in the Peripheral rocks provide a record of the development of the Olympic uplift. Conformable stratigraphic sequences in the northern and southern Olympic Mountains (Tabor and Cady, 1978b; Snavely et al., 1980; Walsh et al., 1987) indicate that the Peripheral rocks remained relatively flat lying and submarine until the end of the middle Miocene. The first indication of unroofing of the Olympic subduction complex is in the Montesano Formation (Tabor and Cady, 1978b; Bigelow, 1987), south of the Olympic uplift. Walsh et al. (1987) cited a late Miocene (12–5 Ma) age for this unit. Marine strata of the Pliocene Quinault Formation unconformably overlie parts of the western Olympic subduction complex. This stratigraphic evidence is compatible with apatite fission-track dates, indicating cooling of the subduction complex at about 7 to 12 Ma (Brandon et al., 1988b, and unpublished data).

LARGE-SCALE STRUCTURE OF THE OLYMPIC UPLIFT

Even though the Olympic subduction complex has been uplifted and exposed, it still resides in its original convergent margin setting. The cross section in Figure 2 shows the structure of the margin in this area. The position of the subducting Juan de Fuca slab is based on earthquake locations from the Washington Regional Seismograph Network catalog, which contains a selected set of well-located events (depth error ±5 km; R. S. Ludwin, 1989, written commun.). Taber and Smith (1985), Crosson and Owens (1987), and others have concluded that the lower band of seismicity (Fig. 2) represents Benioff zone events within the Juan de Fuca slab.

In constructing our cross section (Fig. 2), we have assumed that the Benioff zone events are restricted to the upper part of the oceanic mantle because of the young age of the Juan de Fuca plate. The seismogenic part of the slab beneath western Washington and southern Vancouver Island has an age of about 6 to 12 Ma (Kulm et al., 1984). Intraplate seismicity in young lithosphere of this age is restricted to depths no greater than 14 to 20 km (Wiens and Stein, 1984, Fig. 7). The seismogenic part of the slab beneath Vancouver Island is at temperatures in excess of 450 °C (Lewis et al., 1988, Fig. 6), so it seems doubtful that the seismicity is from the crustal part of the slab. The mantle part of the slab, however, can deform seismically up to temperatures of 750 °C (Wiens and Stein, 1984). Our placement of the slab is also consistent with the depth to oceanic Moho determined by the refraction study of Taber and Lewis (1986) and teleseismic receiver function analysis of Owens et al. (1988) and Lapp (1987). A depth of 100 km is assumed for the top of the slab beneath the volcanic arc.

For comparative purposes, we have superimposed the position of the slab as determined for cross sections constructed across central Vancouver Island (A-A′) and southwestern Washington (C-C′) (Fig. 1). The position of the slab in these two sections is consistent with all available geophysical data (Benioff zone seismicity from Washington Regional Seismic Network catalog and unpublished catalog of G. C. Rogers [1985, written commun.]; refraction and gravity data from Kulm et al., 1984; Spence et al., 1985; reflection data from Clowes et al., 1987; Davis

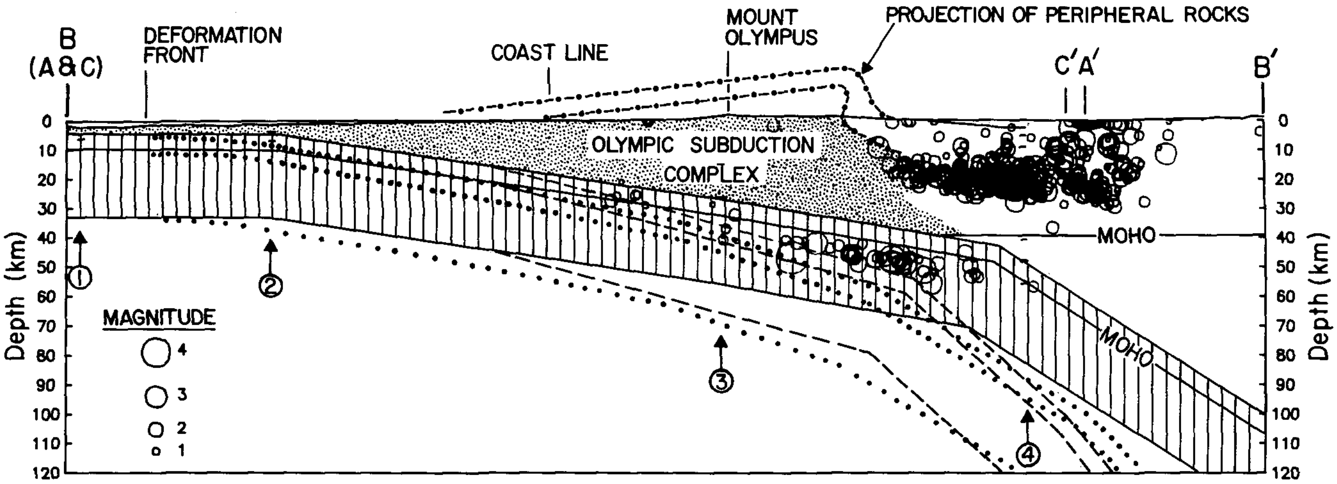

Figure 2. Structure cross section B-B′ across Olympic Mountains. Subducting Juan de Fuca slab is indicated with vertically ruled pattern. Open circles mark well-located earthquakes within 10 km of section line (events from Washington Regional Seismic Network catalog). Positions of slab in section A-A′ across southern Vancouver Island (dotted lines) and section C-C′ across southwestern Washington (dashed lines) are shown for comparison. These sections are superimposed using coastline as common point of alignment. Landward end of each section (A′, B′, C′) coincides with volcanic arc. Note that slab beneath Olympic Mountains is about 10 km shallower and arc is displaced about 65 km eastward relative to other sections. Position of slab in section B-B′ is based on well-located Benioff-zone earthquakes (shown on section) and velocity-depth data (numbered arrows) from following sources: 1, 2—projected from section D-D′ in Kulm et al. (1984); 3—section C-C′ in Kulm et al. (1984); 4—Langston and Blum (1977).

and Hyndman, 1989; teleseismic receiver function analysis from Lapp, 1987; Owens et al., 1988). The sections were superimposed using coastline intersections as a point of alignment. The deformation fronts in the sections also closely coincide in this construction.

These sections show that the slab beneath the Olympic Peninsula has a shallower dip relative to areas to the north and south. This result was first documented by Crosson and Owens (1987) and Weaver and Baker (1988), using Benioff zone seismicity. They concluded that the Juan de Fuca plate was bowed up in a large arch beneath the Olympic Mountains. The relief associated with this arch is about 10 km, based on section B-B′, which parallels the crest of the arch. We argue that the formation of this arch produced much of the local structural relief currently observed in the Olympic Mountains (Fig. 3). Our projection of the Peripheral rocks over the Olympic subduction complex (Fig. 2) is based on this interpretation.

DISCUSSION

The arch in the Juan de Fuca slab has been attributed to the unusual curvature of the Cascadia margin (Crosson and Owens, 1987; Weaver and Baker, 1988). The curvature associated with most subduction zones is concave toward the arc, which results in the slab forming a downward-facing arch as it is subducted (Le-Pichon et al., 1973, p. 227). A commonly invoked analogue is the shell of a ping-pong ball, which produces a similar curvature when dimpled. The Cascadia margin, however, has the opposite sense of curvature (concave side facing seaward); as a result, the slab is forced to arch upward as it is subducted. The slab deforms as it moves through this arched region, which probably accounts for the high density of Benioff zone events beneath the eastern Olympic Mountains and Puget Sound (Chiao and Creager, 1989). Another indication of the arch is the greater distance of the modern volcanic arc from the trench in the vicinity of the arch (Figs. 1 and 2).

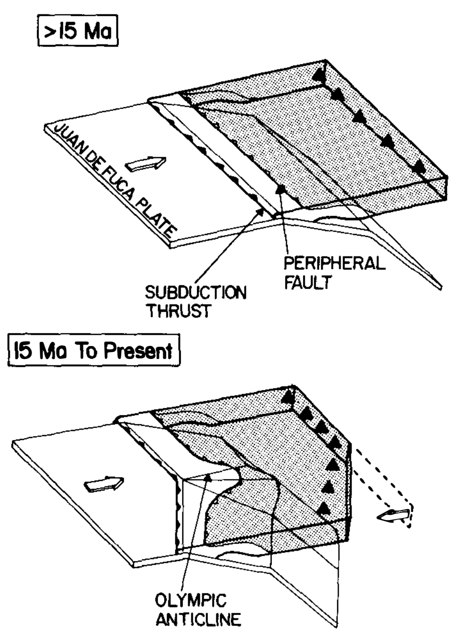

Figure 3. Schematic illustration of development of slab arch beneath Olympic Mountains uplift.

We postulate that the slab arch formed when extension in the Basin and Range province caused the southern Cordilleran margin to swing westward, thus producing the present curvature of the Cascadia margin. Restorations based on paleomagnetic data (Frei, 1986; Wells and Heller, 1988) show a relatively straight, northwest-trending margin prior to the middle Miocene. The major phase of Basin and Range extension is recognized as being middle Miocene to recent in age (e.g., Wernicke et al., 1987). Thus, the margin was changing shape at the same time that the Olympic uplift was forming. Paleomagnetic evidence also indicates that the hinge point associated with the formation of the curved margin was located in the Olympic Mountains area. Paleomagnetic data show that Oligocene volcanic rocks of the Cascade arc from northern California to Washington are substantially rotated (e.g., Wells and Heller, 1988), whereas lower Eocene volcanic rocks on southern Vancouver Island show no rotation (Irving and Brandon, 1990).

CONCLUSIONS

Fore-arc uplift at convergent margins is usually attributed to subduction underplating and within-wedge shortening. These factors have no doubt been important during the evolution of the Olympic subduction complex. As shown in Figure 2, the lawsonite-bearing rocks are currently 30 km above the top of the Juan de Fuca slab. Continued underplating and within-wedge shortening have moved these rocks upward from their original site of accretion. However, we argue that the unusually large amount of structural relief present in the Olympic Mountains is a local effect caused by formation of the slab arch. We infer that the slab arch was formed by middle Miocene to Holocene Basin and Range extension, which is consistent with stratigraphic and fission-track evidence of erosional denudation of the Peripheral rocks and the Olympic subduction complex during late Miocene time. Our example illustrates the important role that slab geometry can have on the tectonic development of a convergent margin, and shows that changes in the configuration of an active continental margin can produce seemingly unrelated and widely separated effects.

REFERENCES CITED

Bigelow, P.K., 1987, The petrology, stratigraphy and basin history of the Montesano Formation, southwestern Washington and southern Olympic Peninsula [M.S. thesis]: Bellingham, Western Washington University, 263 p.

Brandon, M.T., 1989, Deformational styles in a sequence of olistostromal melanges, Pacific Rim Complex, western Vancouver Island, Canada: Geological Society of America Bulletin, v. 101, p. 1520–1542.

Brandon, M.T., Cowan, D.S., and Vance, J.A., 1988a, The Late Cretaceous San Juan thrust system, San Juan Islands, Washington: Geological Society of America Special Paper 221, 81 p.

Brandon, M.T., Miller, D.S., and Vance, J.A., 1988b, Fission track dates for initiation and uplift of the Cenozoic subduction complex of the Olympic Mountains, NW Washington [abs.]: Geological Society of America Abstracts with Programs, v. 20, p. 145.

Chiao, L-Y., and Creager, K.C., 1989, Kinematic deformation of the subducted Cascadia Slab [abs.]: Eos (Transactions, American Geophysical Union), v. 70, p. 1333.

Clowes, R.M., Brandon, M.T., Green, A.G., Yorath, C.J., Sutherland Brown, A., Kanasewich, E.R., and Spencer, C., 1987, LITHOPROBE—Southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections: Canadian Journal of Earth Sciences, v. 24, p. 31–51.

Crosson, R.S., and Owens, T.J., 1987, Slab geometry of the Cascadia subduction zone beneath Washington from earthquake hypocenters and teleseismic converted waves: Geophysical Research Letters, v. 14, p. 824–827.

Davis, E.E., and Hyndman, R.D., 1989, The structure and tectonic history of the northern Cascadia subduction zone: Recent accretion and deformation: Geological Society of America Bulletin, v. 101, p. 1465–1480.

Deer, W.A., Howie, R.A., and Zussman, J., 1986, Disilicates and ring silicates: Rock-forming minerals, Volume 1B: New York, Longman Scientific and Technical, 629 p.

Frei, L.S., 1986, Additional paleomagnetic results from the Sierra Nevada: Further constraints on Basin and Range extension and northward displacement in the western United States: Geological Society of America Bulletin, v. 97, p. 840–849.

Frost, B.R., 1980, Observations on the boundary between zeolite facies and prehnite-pumpellyite facies: Contributions to Mineralogy and Petrology, v. 73, p. 365–373.

Glassley, W.E., Whetten, J.T., Cowan, D.S., and Vance, J.A., 1976, Significance of coexisting lawsonite, prehnite, and aragonite in the San Juan Islands, Washington: Geology, v. 4, p. 301–302.

Gleadow, A.J.W., and Brooks, C.K., 1979, Fission track dating, thermal histories, and tectonics of igneous intrusions in East Greenland: Contributions to Mineralogy and Petrology, v. 71, p. 45–60.

Hawkins, J.W., Jr., 1967, Prehnite-pumpellyite facies metamorphism of a graywacke-shale series, Mount Olympus, Washington: American Journal of Science, v. 265, p. 798–818.

Heller, P.L., Tabor, R.W., and Suczek, C.A., 1987, Paleogeographic evolution of the United States Pacific Northwest during Paleogene time: Canadian Journal of Earth Sciences, v. 24, p. 1652–1667.

Irving, E., and Brandon, M.T., 1990, Paleomagnetism of the Flores volcanics: Vancouver Island in place by Eocene time: Canadian Journal of Earth Sciences, v. 27, p. 811–817.

Kulm, L.D., and 12 others, editors, 1984, Western North American continental margin and adjacent ocean floor off Oregon and Washington: Atlas 1 (Ocean Margin Drilling Program, Regional Atlas Series): Woods Hole, Massachusetts, Marine Science International, 32 sheets.

Langston, C.A., and Blum, D.E., 1977, The April 29, 1965 Puget Sound earthquake and the crustal and upper mantle structure of western Washington: Seismological Society of America Bulletin, v. 67, p. 693–711.

Lapp, D.B., 1987, The subduction geometry beneath western Washington from deconvolved teleseismic P-waveforms [M.S. thesis]: Seattle, University of Washington, 70 p.

LePichon, X., Francheteau, J., and Bonnin, J., 1973, Plate tectonics: Amsterdam, Elsevier, 300 p.

Lewis, T.J., Bentowski, W.H., Davis, E.E., and Hyndman, R.D., 1988, Subduction of the Juan de Fuca plate: Thermal consequences: Journal of Geophysical Research, v. 93, p. 15,207–15,225.

Nitsch, K.H., 1972, Das P-T-X_{CO_2} stabilitatsfeld von lawsonit: Contributions to Mineralogy and Petrology, v. 34, p. 116–134.

Owens, T.J., Crosson, R.S., and Hendrickson, M.A., 1988, Constraints on the subduction geometry beneath western Washington from broadband teleseismic waveform modelling: Seismological Society of America Bulletin, v. 78, p. 1319–1334.

Shouldice, D.H., 1971, Geology of the western Canadian continental shelf: Canadian Society of Petroleum Geologists Bulletin, v. 19, p. 405–424.

Snavely, P.D., Jr., Niem, A.R., MacLeod, N.S., Pearl, J.E., and Rau, W.W., 1980, Makah Formation—A deep marginal basin sequence of late Eocene and Oligocene age in the northwestern Olympic Peninsula, Washington: U.S. Geological Survey Professional Paper 1162-B, 28 p.

Spence, G.D., Clowes, R.M., and Ellis, R.M., 1985, Seismic structure across the active subduction zone of western Canada: Journal of Geophysical Research, v. 90, p. 6754–6772.

Stewart, R.J., 1974, Zeolite facies metamorphism of sandstone in the western Olympic Peninsula, Washington: Geological Society of America Bulletin, v. 85, p. 1139–1142.

Taber, J.J., and Lewis, B.T.R., 1986, Crustal structure of the Washington continental margin from refraction data: Seismological Society of America Bulletin, v. 76, p. 1011–1024.

Taber, J.J., and Smith, S.W., 1985, Seismicity and focal mechanisms associated with the subduction of the Juan de Fuca plate beneath the Olympic Peninsula, Washington: Seismological Society of America Bulletin, v. 75, p. 237–249.

Tabor, R.W., and Cady, W.M., 1978a, The structure of the Olympic Mountains, Washington—Analysis of a subduction zone: U.S. Geological Survey Professional Paper 1033, 38 p.

——1978b, Geologic map of the Olympic Peninsula: U.S. Geological Survey Map I-994, scale 1:125,000.

Thompson, A.B., 1971, P_{CO_2} in low-grade metamorphism; zeolite, carbonate, clay mineral, prehnite relations in the system CaO-Al_2O_3-SiO_2-CO_2-H_2O: Contributions to Mineralogy and Petrology, v. 33, p. 145–161.

Vance, J.A., 1968, Metamorphic aragonite in the prehnite-pumpellyite facies, northwest Washington: American Journal of Science, v. 266, p. 299–315.

Walsh, T.J., Korosec, M.A., Phillips, W.M., Logan, R.L., and Schasse, H.W., 1987, Geologic map of Washington-southwest quadrant: Washington Division of Geology and Earth Resources Geologic Map GM-34, 28 p., two sheets, scale 1:250,000.

Weaver, C.S., and Baker, G.E., 1988, Geometry of the Juan de Fuca plate beneath Washington and northern Oregon from seismicity: Seismological Society of America Bulletin, v. 78, p. 264–275.

Wells, R.E., and Heller, P.L., 1988, The relative contribution of accretion, shear, and extension to Cenzoic tectonic rotation in the Pacific Northwest: Geological Society of America Bulletin, v. 100, p. 325–338.

Wells, R.E., Engebretson, D.C., Snavely, P.D., Jr., and Coe, R.S., 1984, Cenozoic plate motions and the volcano-tectonic evolution of western Oregon and Washington: Tectonics, v. 3, p. 275–294.

Wernicke, B.P., Christiansen, R.L., England, P.C., and Sonder, L.J., 1987, Tectonomagnetic evolution of Cenozoic extension in the North American Cordillera, *in* Coward, M.P., Dewey, J.F., and Hancock, P.L., eds., Continental extensional tectonics: Geological Society of London Special Publication 28, p. 203–221.

Wiens, D.A., and Stein, S., 1984, Intraplate seismicity and stresses in young oceanic lithosphere: Journal of Geophysical Research, v. 89, p. 11,442–11,464.

ACKNOWLEDGMENTS

Supported by National Science Foundation Grant EAR-8707442, ARCO, and Yale University. We thank R. W. Tabor for his thin-section collection, and R. S. Ludwin, R. S. Crosson (Washington State), and G. C. Rogers (Vancouver Island) for earthquake data. Unpublished fission-track data discussed herein are the result of a collaborative project with M. K. Roden, D. S. Miller, and J. A. Vance. We appreciate the comments and challenges from R. W. Tabor, R. S. Ludwin, S. Box, and T. Foster.

Manuscript received December 11, 1989
Revised manuscript received July 25, 1990
Manuscript accepted August 17, 1990