Cenozoic active margin and shallow Cascades structure: COCORP results from western Oregon

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ABSTRACT

Two deep seismic reflection profiles totaling 98 km were recorded by the Consortium for Continental Reflection Profiling (COCORP) in western Oregon in 1984, providing some of the first onshore deep seismic reflection data from the forearc and arc of an active convergent margin. These data are of lower quality than most COCORP data, but when combined with other geological and geophysical data, they provide some useful insights into the subduction zone beneath Oregon and into extensional structures in the Cascades.

Line 1 crossed the Eocene sediments and underlying pillow basalts of the Coast Range and western Willamette Valley. Reflections from the Willamette Valley clearly define the lower and western boundary of flat-lying, rhythmically bedded Eocene sediments. These sediments appear to be underlain by as much as 8 km of seismically transparent Eocene pillow basalts. Layered reflections at depths of 8–16 km may indicate a remnant crust upon which the Eocene pillow basalts were erupted. East-dipping reflections at depths of 35–40 km may represent the décollement above the subducting Juan de Fuca plate.

Oregon line 2 crosses the Cenozoic volcanic arc terrains of the Western and High Cascades. Interpretation of COCORP seismic data suggests modification of existing models for normal faulting that postulate a symmetric graben or grabens in the Cascades of Oregon. In the High Cascades along COCORP Oregon line 2, reflections suggest a large, gently west-dipping half-graben with major offset only on the west side of the range. The seismic reflection data along with other geological and geophysical data suggest that the High Cascades were built on a series of blocks, rather than a single graben. The blocks were differentially faulted during the Pliocene. Seismic data from the Western Cascades indicate several normal faults with down-to-the-east offset, including a major, previously unidentified fault of probable Miocene age.

INTRODUCTION

During May–July of 1984, the Consortium for Continental Reflection Profiling (COCORP) carried out two deep seismic reflection surveys of a reconnaissance nature in western Oregon. On neither line was data quality particularly high. COCORP did not, therefore, complete a planned, long east-west transect in this area. Nevertheless, some useful data were obtained and are reported herein.

Oregon line 1 is a short, 35-km-long, deep seismic line in the Oregon Coast Range and western edge of the Willamette Valley (Figs. 1 and 2). Oregon line 2 is 63 km long with the west end near Blue River in the Western Cascades (Oligocene-Miocene) and extends eastward through the High Cascades (Pliocene–Recent) to just south of Sisters (Figs. 1 and 3). All exposed stratigraphic units in the Cascades are volcanic rocks of varying composition locally covered by Quaternary alluvium (Priest and others, 1983).

The northwestern United States is an active Cenozoic convergent margin, as evidenced by surface geology and ongoing volcanic activity (Fig. 1). Due to the lack of seismicity, however, little is known about the depth, geometry, and subduction process of the Juan de Fuca plate beneath the North American plate in Oregon.

On Oregon line 1, the data suggest that beneath the Oregon Coast Range, the top of the Juan de Fuca plate is 35 to 40 km deep and dips

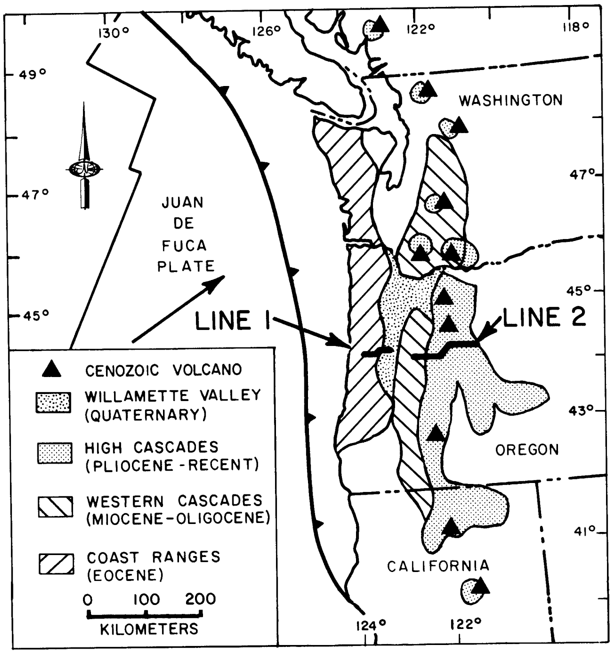

Figure 1. Regional physiographic-tectonic map of the northwestern American continent. The COCORP seismic reflection lines in Oregon are indicated by the heavy lines. Arrow on the Juan de Fuca plate shows direction of convergence. Figure modified from Priest (1986).

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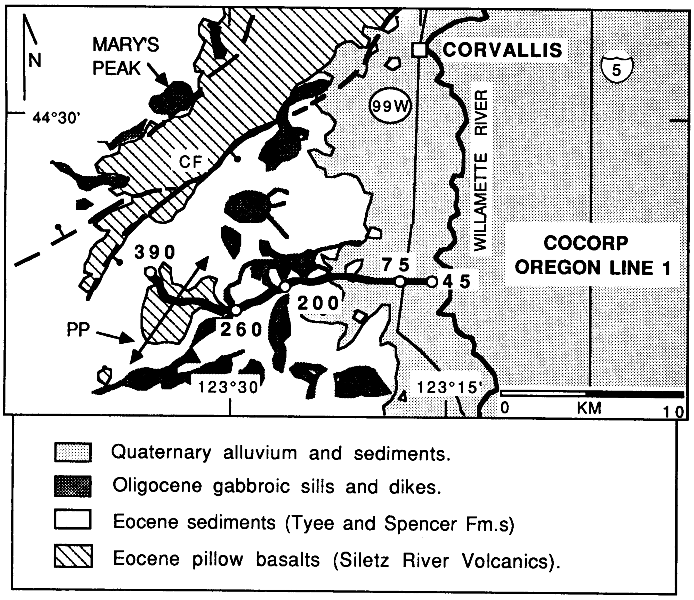


Figure 2. Geologic location map of Oregon line 1. Heavy line is the COCORP seismic reflection line; open circles are station locations. CF, Corvallis fault; PP, Prairie Peak anticline. Geology is from Vokes and others (1954) and Baldwin (1955).

approximately 15° to the east. Reflections from 8.0–16.0 km below the Coast Range suggest a layered, early Eocene crust of unknown lateral extent upon which Eocene pillow basalts were erupted. About 4.0 km of sedimentary sequences is imaged in the Willamette Valley.

The Cascade Range of Oregon forms the central part of the Cenozoic volcanic arc associated with the subduction of the Juan de Fuca plate (Atwater, 1970). Concurrent with subduction, the Cascades arc has experienced extension as evidenced by several prominent normal faults within the Cascades (see HCFZ and GRFZ, Fig. 3). Allen (1966) first recognized the linearity of faults with north-south trends on both sides of the High Cascades and proposed the presence of a High Cascades graben. Taylor (1981) further proposed that faulting within the High Cascades graben may have controlled late Pliocene to Recent volcanic activity. On the basis of regional gravity trends, Couch and others (1982a) suggested that the graben faulting extended into the Western Cascades.

Results from Oregon line 2 generally support the interpretations of Cenozoic normal faulting within the Cascades. Interpretation of reflections in the Western Cascades suggests several faults with down-to-the-east displacement, including a major, previously unknown normal fault of probable Miocene age. Along the line of the reflection survey, the data suggest the existence of a large half-graben in the High Cascades with major faulting on the west side of the range. The survey does not support, however, a model of symmetrical graben faulting flanking a long segment of the High Cascades (Allen, 1966; Taylor, 1981). Instead, the reflection data and other geologic data suggest that the High Cascades were built on a series of blocks differentially faulted during the Pliocene (?) within the area defined by Allen (1966) as the High Cascades graben.

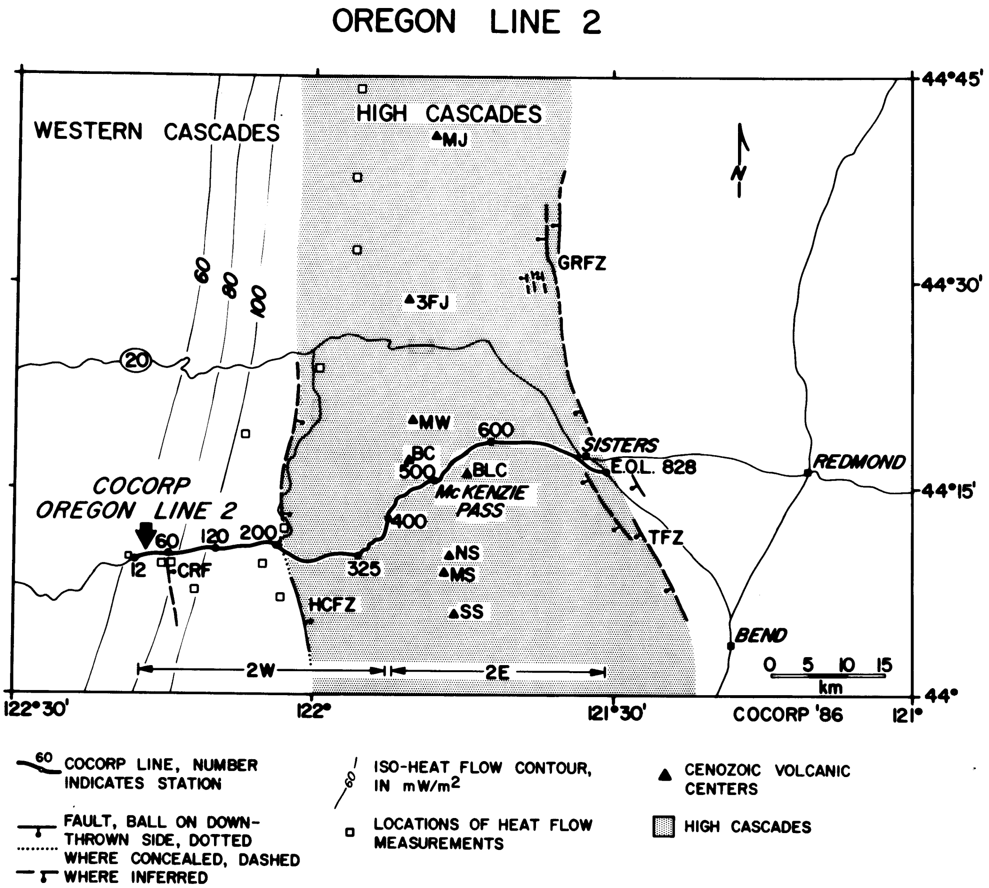


Figure 3. Location and tectonic map of Oregon line 2. Bars at bottom of figure indicate where line 2 was divided. MJ, Mount Jefferson; 3FJ, Three Fingered Jack; MW, Mount Washington; BC, Belknap Crater; BLC, Black Crater; NS, North Sister; MS, Middle Sister; SS, South Sister; CRF, Cougar Reservoir fault; HCFZ, Horse Creek fault zone; GRFZ, Green Ridge fault zone; TFZ, Tumalo fault zone. Heat-flow contours and measurement locations are from Black and others (1984).

Interpretable reflections in the Cascades are limited to the upper 3.0 s of the record. Background noise (for example, traffic) and poor signal penetration in the volcanic rocks of the Cascades apparently limit the depth to which interpretable reflections can be detected. Although the data have been extensively processed, there are few recognizable deeper reflections.

REGIONAL GEOLOGY AND TECTONIC SETTING

The northwestern United States is an area of active convergence with a convergence rate of about 4.0 cm/yr (Riddihough, 1984). Typically, such convergence rates result in areas of high seismicity and large earthquakes with well-defined Benioff zones (Heaton and Kanamori, 1984). A Benioff zone that dips between 10° and 15° to the northeast below the Puget Sound area of Washington has been identified on the basis of magnitude 4.0 or smaller earthquakes (Heaton and Kanamori, 1984). In western Oregon, however, a Benioff zone has not been defined, owing to insufficient seismicity. The lack of a well-defined Benioff zone may be due to aseismic subduction of the Juan de Fuca plate. Attempts have been made to estimate the depth of the Juan de Fuca plate in western Oregon on the basis of geophysical modeling, mostly with gravity, some refraction, and limited seismicity data. Depth estimates include 16, 18, 40–44, and 45 km by Berg and others (1966), R. W. Couch (1984, written commun.), Riddihough (1979), and Langston (1981), respectively.

The Coast Range

The Coast Range of Oregon is part of an uplifted forearc ridge between the Cascade volcanic arc and the subducting Juan de Fuca plate (Fig. 1). The Willamette Valley occupies a forearc setting between the Coast Range and the Cascades. The west end (stations 270–390) of Oregon line 1 crosses Eocene pillow basalts (Snavely and Wagner, 1963) that form the core of the Coast Range. The eastern two-thirds of the line (stations 45–260) crosses rhythmically bedded Eocene sandstones and siltstones.

In the Coast Range of central Oregon, exposed basement rocks are the lower and middle Eocene Siletz River Volcanics (SRV), which are mostly tholeiitic pillow basalts that grade locally to subaerial alkalic basalts (Snavely and Wagner, 1963; Snavely and others, 1968). The SRV have an estimated thickness of 3,000–7,000 m or greater, with the thickest sections at proposed eruptive centers in the Coast Ranges (Snavely and Wagner, 1963). Rhythmically bedded marine turbidite sandstones and siltstones of the Flournoy, Tyee, and Spencer Formations overlie the Siletz River Volcanics (Snavely and Wagner, 1963; Snavely and others, 1964; Baldwin, 1974).

Geochemical and geochronologic evidence suggests that the basalts of the Coast Ranges in Oregon and Washington were erupted as a series of seamounts adjacent to an oceanic spreading center (Duncan, 1982; Snavely and MacLeod, 1974). The ages of the eruptive centers in the Oregon Coast Range increase from 53 to 62 m.y. southward from the Oregon-Washington border (Duncan, 1982).

Alternatively, the basalts of the Coast Range may have been erupted during rifting of the continental margin. Paleomagnetic evidence suggests that northern Cordilleran terranes, for example, the Prince William and Chugach terranes of Alaska, were attached to the continent near Oregon approximately 62 m.y. ago and were subsequently rifted and transported northward (Wells and others, 1984). Wells and others (1984) envisioned an oblique rifting geometry perhaps analogous to the Gulf of California or the Andaman Sea north of Sumatra. The basalts of the Siletz River Volcanics were then erupted from a hot spot beneath the extensional basin forming behind the northward-moving terranes. Following eruption, a change in the direction of plate convergence caused the eruptive centers to be accreted to the new continental margin.

Interpretation of paleomagnetic data indicates that the eruptive centers were rotated 46° ± 13° clockwise following their eruption and formation and were accreted to the North American plate during the middle to late Eocene (Simpson and Cox, 1977; Magill and others, 1981; Duncan, 1982). In addition, a second phase of rotation, associated with Basin and Range extension, probably occurred from the Oligocene to the present with an additional 30° ± 10° of clockwise rotation (Magill and others, 1981). Following accretion, the subduction zone probably jumped from a position beneath or near the present position of the Cascades to its present position west of the Coast Range (Snavely and others, 1980).

Major uplift of the Coast Range probably began by late Oligocene–early Miocene (Snavely and others, 1980). Uplift was possibly associated with deep-seated ductile flow between the subducting slab and the Coast Range (Pavlis and Bruhn, 1983). Gentle folding accompanied the uplift. The Coast Range is extensively intruded by gabbroic sills and dikes of Oligocene age (Snavely and Wagner, 1961). There is some geodetic evidence of contemporary tilting from west to east of the Oregon Coast Range (Ando and Balazs, 1979).

The Cascades

In Oregon, the Cascades are divided into two physiographic provinces: the Western Cascades and the High Cascades. The Western Cascades province, lying between the Willamette Valley on the west and the High Cascades to the east, is a deeply dissected block of mostly Oligocene-Miocene volcanic rocks of varying composition and lithology. The Western Cascades were uplifted 4–5 m.y. ago and eroded to form much of the present topography (Priest and others, 1983). The High Cascades consist of a north-south belt, 20–30 km wide, of undeformed upper Pliocene–Recent lava flows and cones. The large stratovolcanoes that form the crest of the High Cascades were formed within the latest magnetic polarity epoch (690,000 yr). These composite cones rest upon a large platform of coalesced Pliocene-Pleistocene shield volcanoes (Taylor, 1981). Most of the large composite cones are of andesitic-silicic composition, whereas the platform is mainly basalt and basaltic andesite.

During the Miocene, the Western Cascades experienced faulting and minor folding (Peck and others, 1964; Priest and others, 1983). Faults have north and northwest trends with both normal and strike-slip motion. Most of the faults, including the Cougar Reservoir fault (CRF) (Fig. 3), had a down-to-the-east motion. Priest and Woller (1983) suggested that the CRF controlled major subsidence within the Western Cascades during the Miocene. These faults may have been active (or reactivated) as recently as the Pliocene (Hammond and others, 1980).

The High Cascades were built within a structural depression (termed the "High Cascades graben" by Allen, 1966, and Taylor, 1981) formed during early Pliocene subsidence about 5.5 m.y. ago along north- and northwest-trending faults (Hammond and others, 1980; Smith, 1985). Stratigraphic evidence on both sides of the High Cascades indicates that during Pliocene subsidence, a chain of shield volcanoes known as the "Plio-Cascades" subsided within the High Cascades graben (Taylor, 1981). The Plio-Cascades were subsequently buried by numerous flows that form the present platform in the High Cascades. In the region of the COCORP seismic reflection line, these bounding faults include the north-south Green Ridge fault zone (GRFZ) and the northwest-trending Tumalo fault zone (TFZ) and Horse Creek fault zone (HCFZ) (Fig. 3).

Geothermal studies by Blackwell (1982) and Brown and others (1980) have shown steep heat-flow gradients in the Western Cascades (Fig. 3). The locations of the heat-flow gradients are coincident with steep gravity gradients (Couch and others, 1982b). The gradients have a well-defined north-south trend that parallels the Western Cascade–High Cascade boundary in Oregon. Blackwell (1982) inferred the presence of one or more shallow (7–10 km) magma chambers beneath or adjacent to the axis of the High Cascades from modeling that incorporates both gravity and heat-flow data. The seismic penetration in this study, however, was not sufficiently deep to test Blackwell's (1982) model. The presence of numerous hot springs aligned along a 200-km length of the Western Cascade–High Cascade boundary may indicate the proximity of a magma body or other heat source to the surface (Blackwell, 1982).

DATA ACQUISITION AND PROCESSING

Oregon lines 1 and 2 are east-west seismic reflection transects (Fig. 1) recorded in the Coast Range and central Oregon Cascades, respectively, in the Spring and early Summer of 1984. Specific details of the acquisition and data processing sequence are outlined in Appendix 1.

Oregon line 1 was shot from east to west beginning in the Willamette Valley and is relatively straight until station 200. At station 270, the line crosses onto the Siletz River Volcanics (Fig. 2). The westernmost part of Oregon line 2 is nearly straight as it crosses the Western Cascades (stations 12–200, Fig. 3). Extending eastward from station 200, line 2 is more crooked as it crosses the rough, volcanic terrain of the High Cascades. The road ascending the crest of the High Cascades from the west was especially crooked, with a number of switchbacks due to the steep topography. Many of these switchbacks were eliminated by placing the cables and geophones cross-country in a straight line across the switchbacks.

In anticipation of high traffic noise, mantissa summing was employed during field recording for a portion of Oregon line 2 (stations 12–199). Mantissa summing is a data collection technique which has been shown to reduce the influence of strong noise sources, such as traffic (Sheriff, 1984; Klemperer and Brown, 1985).

Compared to many other COCORP surveys, the quality of data from Oregon line 2 was generally low, as illustrated by amplitude decay curves for near and far traces recorded in the Cascades (Fig. 4). The amplitude decay curves are compared with similar curves from a seismic data set that has good reflections (COCORP Washington line 2, Potter and others, 1986). In the curves from the Western and High Cascades (Figs. 4a and 4b, respectively), note the rapid decay of amplitude for trace 6 to an apparent background level at about 7.0 s, two-way traveltime (twtt). In the High Cascades, the absence of any amplitude on the far traces for stations 477–534, as indicated by the nearly straight line of trace 60, indicates almost a complete lack of source energy received at the far traces. In contrast, amplitude decay curves from the Washington data set never approach a background level for both the far and near traces (Fig. 4c). Seismic energy is still being recorded after 16 s (twtt).

On Oregon line 2, possible factors limiting the detectable energy returning from depth include (1) poor signal penetration due to poor source-to-ground coupling and/or attenuation of the seismic signal by the young basalts and (2) high levels of environmental noise such as traffic, wind, and so on. The reflections in the upper 2.0 s (twtt) are defined by the near traces, so that the low signal level of the far traces is unimportant to the shallowest data and hence to the interpretations presented in this paper.

Because of an abrupt bend in the seismic line, data collected at stations 367–416 were deleted and the line treated as two seismic sections during processing, 2W (stations 12–366) and 2E (stations 417–828) (Fig. 3). The west end of section 2E is offset 2 km to the north and 2 km east of the east end of section 2W. The changes in elevation (~800 m) and the offset between sections 2E and 2W required different reference levels for the two seismic sections. They are best aligned by matching the east end of 2W, at time 0 s (the top of the section), to 0.8 s (twtt) of 2E (this has been done in Fig. 5).

A variety of pre-stack processing techniques was used in attempts to enhance the data. These included FK filtering to attenuate low-frequency noise and surface waves (line 2W), deconvolution to remove short-period multiples, interactive refraction statics (line 2E), and velocity analysis. Because of the poor signal response in the far traces (especially on line 2E), near-source (500, 1,500, and 5,000 m) range stacks were used in an attempt to identify deeper events. The range stacks, however, did not reveal more or deeper events. A series of constant-velocity migrations of line drawings of the seismic sections were carried out to confirm aspects of the interpretations presented in this paper.

Post-stack coherency filtering proved very effective in enhancing previously identified weak reflections on Oregon line 1 (Fig. 6). A coherency filter recently developed at the COCORP (L. Zheng and L. D. Brown, unpub. data) improves linear and nonlinear events and reduces the artifacts associated with some coherency filters. Coherency filtering is used to attempt to remove incoherent noise, which frequently obscures weak reflections on crustal seismic sections, and to look for coherence of reflected signals across neighboring traces (L. Zheng and L. D. Brown, unpub. data).

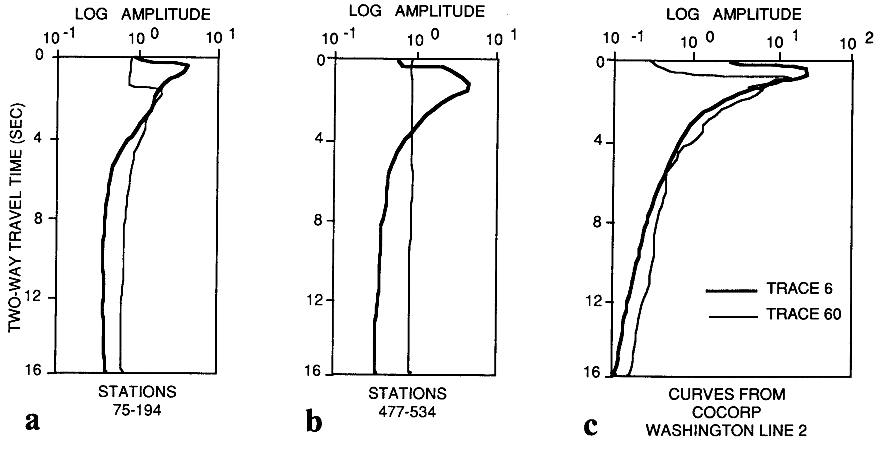


Figure 4. Amplitude decay curves from two areas of Oregon line 2 (4a and 4b), traces 6 and 60 (offsets of 0.9 and 6.0 km, respectively) summed and averaged. Similar curves for the same traces from COCORP Washington line 2 (Potter and others, 1986) are plotted at the same scale (4c). The data from Washington are very good. See text for discussion.

a. Stations 75–194 (in the Western Cascades), 100 files summed.

b. Stations 477–534 (over the crest of the High Cascades), 50 files summed.

c. 50 files summed from COCORP Washington line 2.

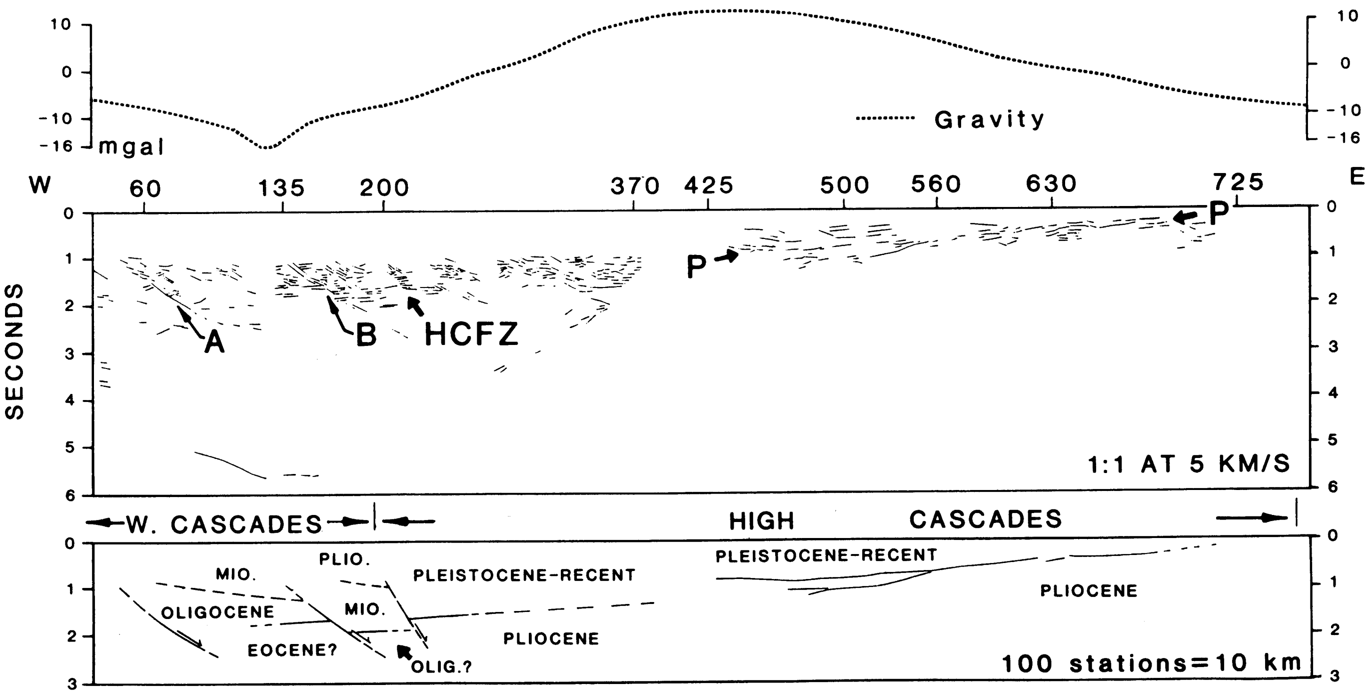


Figure 5. Line drawing of seismic reflection data on Oregon line 2 with a residual gravity profile (Couch and others, 1982b). At the bottom is an interpretive structural section. Age assignments are inferred from geologic mapping by Brown and others (1980), Flaherty (1981), Taylor (1981), and Peterson and others (1976). A may be an extension of the Cougar Reservoir fault, whereas B is a Miocene(?) fault. The Horse Creek fault zone (HCFZ) appears to be well imaged. Horizon P probably represents the boundary between the lower Pliocene and the upper Pliocene–Pleistocene. The westward dip of P indicates a half-graben within the High Cascades. A time shift of +0.8 s has been applied to reflections for line 2W (stations 12–370) owing to different reference levels, required by large elevation changes in the Cascades, used in processing. The time shift allows for correct structural interpretations.

SEISMIC DATA AND DISCUSSION

Oregon Coast Range and Willamette Valley

Horizontal to subhorizontal reflections from the Willamette Valley sedimentary rocks define a zone between 0.0 and 1.8 s (twtt) below stations 70–225. The zone thins to the west above a set of east-dipping reflections that have an unmigrated dip of approximately 15° (see D, Figs. 6 and 7). At 1.0–2.0 s (twtt), zone D is underlain by a series of weak, west-dipping reflections beneath station 220. Farther west, between stations 280 and 320, at 0.5–2.5 s (twtt), another set of weak, west-dipping reflections is imaged beneath surface exposures of the Siletz River Volcanics.

The zone from 2.5 to 5.0 s (twtt) is characterized by horizontal to gently west-dipping reflections (see OC, Figs. 6 and 7). The strongest of these reflections defines the base of the zone at about 5.0 s (twtt). These laterally persistent horizontal reflections contrast with the variable dips of the overlying reflections from 0.0 to 2.5 s (twtt).

The deepest reflections recorded are a set of east-dipping reflections from 10.0 to 11.0 s (twtt) and are most prominent beneath stations 250–350 (Figs. 6 and 7). These reflections dip about 15° to the east after a constant-velocity (6 km/s) migration and have a depth of approximately 35–40 km. The proximity of the Corvallis fault (CF, Fig. 2) to the western half of the seismic line suggested that the east-dipping reflections at 10.0–11.0 s may actually be reflections from the fault rather than from structures in the vertical plane. Calculations of two-way traveltimes using rock velocities of 3.5–5.0 km/s, which were derived from the seismic data, indicate that any reflections from the Corvallis fault would begin arriving at 2.8–4.0 s (twtt) for station 200, with two-way traveltimes decreasing for stations farther to the west and closer to the fault (Fig. 2).

Reflections from the Willamette Valley define a lower and western boundary for the flat-lying sediments of the Tyee Formation, which are mapped along the line east of station 260 by Vokes and others (1954). The east-dipping reflections of D define the western limit of horizontal reflections and presumably represent the westernmost extent of the Tyee Formation. Locally, a 1-km-wide, east-southeast–dipping gabbroic dike is mapped as being nearly coincident with the SRV-Tyee contact, with a small section of Tyee sediments between the dike and SRV (Figs. 2 and 7b) (Vokes and others, 1954). A preferred interpretation is that D represents a sediment-dike contact. Alternatively, D may represent the contact between Siletz River Volcanics and overlying Tyee sediments. A third possibility is that D could be from an east-dipping normal fault bounding the western Willamette Valley. Geologic mapping by Vokes and others (1954) and Gandera (1977) in the western Willamette Valley, however, has not identified any displacement of the late Eocene stratigraphy. If D does represent a fault, offset must have occurred prior to deposition of the Eocene sediments. The base of shallow subhorizontal reflections at 2.0 s at the east end of Oregon line 1, using a velocity of 4.0 km/s, suggests a maximum thickness of about 4.0 km for the Eocene Willamette Valley sediments at this locality.

The Siletz River Volcanics are characterized by few continuous reflections (Figs. 6 and 7). Eocene basalts on Vancouver Island (the Met-

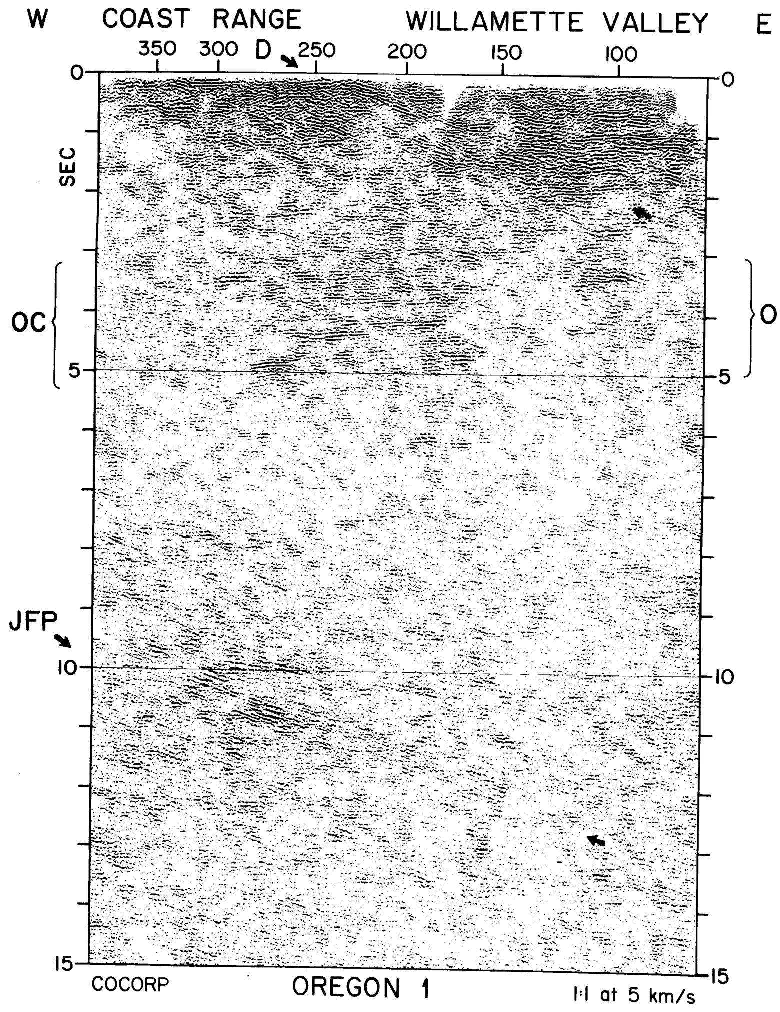
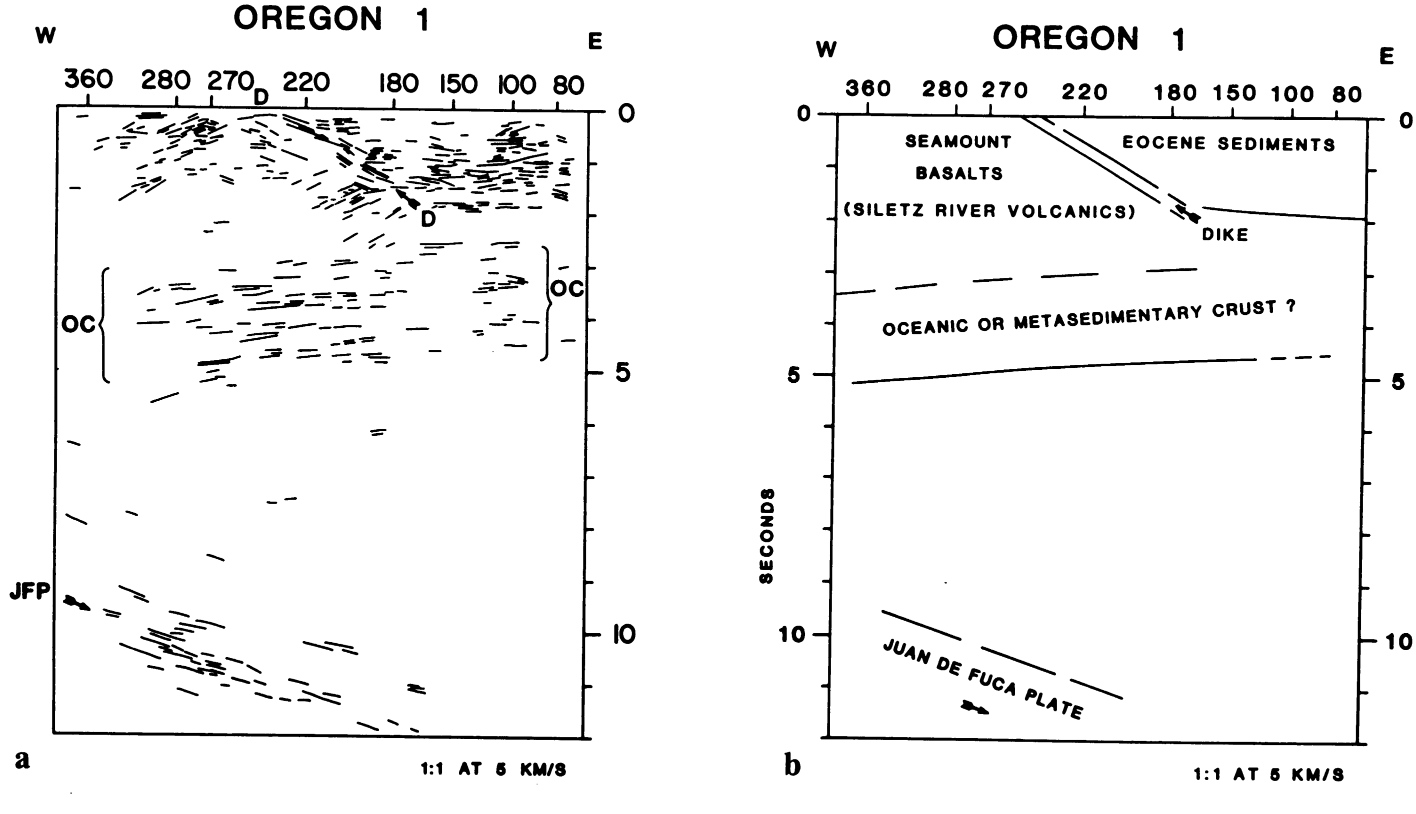


Figure 6. Coherency-filtered seismic section of Oregon line 1. Note the east-dipping reflection D that may define the base of Eocene sediments overlying the Eocene pillow basalts. Reflections D project updip to the surface position of a large east-dipping dike nearly coincident with the sediment-basalt contact. The weak, gently west-dipping reflections OC, from 2.5–5.0 s, may be from Eocene crust underneath the pillow basalts of the Oregon Coast Range. The deepest reflections may be from the décollement zone above the subducting Juan de Fuca plate (JFP). See text for further discussion.

Figure 7. a. Line drawing of seismic reflection data on Oregon line 1. For description of abbreviations, see Figure 6.

b. Interpretive cross section of Oregon line 1. See text for discussions of other possible interpretations.



chosin Formation), which are equivalent to the Siletz River Volcanics, have a similar seismic character on recent LITHOPROBE reflection surveys (Green and others, 1986). Discontinuous, weak west-dipping reflections below stations 280–360 may be representative of minor deformation and folding in the SRV, which is evidenced locally by the northeast-southwest–trending Prairie Peak anticline (see PP, Fig. 2). The Siletz River Volcanics are underlain by a 2.5-s zone of horizontal to gently west-dipping reflections from 2.5 to 5.0 s (see OC, Figs. 6 and 7) at depths of 8.0–16.0 km. If this change in seismic character corresponds to the base of the Siletz River Volcanics, it suggests a maximum thickness locally of about 8 km for these volcanics. The seismic character of these volcanics also suggests that they thin eastward below the increasingly thicker Willamette Valley sediments and above the horizontal reflections OC. This interpretation is supported by a change in the Bouguer gravity anomaly from +50 mgals in the Coast Range to –20 mgals at the east end of line 1 that suggests a significant decrease in the crustal mass along the Coast Range–Willamette Valley boundary.

An umambiguous interpretation of reflections OC is not possible without well-constrained seismic velocity control. They may represent underplated sediments or oceanic crust from a Cenozoic subduction zone (Green and others, 1986; Yorath and others, 1985). Alternatively, reflections OC may come from the crust upon which the Siletz River Volcanics were erupted, either oceanic crust of the Farallon plate (Duncan, 1982) or a modified crust formed in an extensional basin during continental rifting (Wells and others, 1984; Nicolas, 1985).

Vancouver Island, 460 km to the north, is in a tectonic setting similar to that of western Oregon and Washington, one of ongoing subduction beneath Tertiary accreted terranes. Seismic reflection data recorded by LITHOPROBE on Vancouver Island show an area of high reflectivity at traveltimes similar to those of reflections OC beneath seismically transparent Eocene basalts equivalent to the Siletz River Volcanics of the Oregon Coast Range (Green and others, 1986; Clowes and others, 1987; Yorath and others, 1985). The zone of high reflectivity has been interpreted as an underplated slab of oceanic lithosphere (Green and others, 1986). Green and others (1986) suggested that "the strongly laminated character (of the reflection zone) could result from layered sediments, intercalated volcanics and sediments, layered igneous basement rocks, or tectonic structures induced by underthrusting," with a favored interpretation of tectonically underplated mafic oceanic crust.

Clowes and others (1987) interpreted the area of high reflectivity to represent underplated, imbricated slices of subduction-complex sediments that may include slices of mafic oceanic crust. This interpretation is based on an updip projection of the reflections to an exposure on the Olympic Peninsula in Washington of the Olympic core rocks, an uplifted subduction complex of marine sediments that has been thrust under Eocene basalts (the Crescent Volcanics) (Clowes and others, 1987; Tabor and Cady, 1978). In the Oregon Coast Range, the base of the Eocene basalts (the Siletz River Volcanics) is not exposed, and it is not known if similarly underthrusted sediments are currently below the basalts.

If an underplated slab of older oceanic crust is present beneath the Oregon Coast Range, reflections from it might be expected to dip in a direction more or less parallel to that of the subducting plate, as has been interpreted on the LITHOPROBE data (Green and others, 1986). Reflections OC (Fig. 6), however, have a gentle westerly dip, which is opposite to that of the east-dipping subducting plate. If an underplated oceanic slab is represented by reflections OC, then it has apparently been rotated from east dipping to west dipping since emplacement. The absence of significant westward tilting or rotation of the overlying Eocene basalts and sediments does not support the presence of a rotated, underplated oceanic slab beneath the Oregon Coast Range. The gentle west dip may be an Eocene feature indicating flexure of the crust due to loading by the voluminous Siletz River Volcanics.

If the Siletz River Volcanics were erupted upon an oceanic crust during the early Eocene (Duncan, 1982), then the laterally continuous reflections from 2.5 to 5.0 s may represent layering in the lower part of this oceanic crust. Layering within the oceanic crust, although not commonly interpreted on seismic reflection profiles, has been suggested on recent profiles from the Atlantic Ocean by the North Atlantic (NAT) Study Group (1985). More typically, the oceanic crust has often been seismically identified as a 2-s (twtt) zone characterized by numerous diffractions from the top of the crust and discontinuous horizontal reflections from the Moho (NAT Study Group, 1985; Nasu and others, 1982). The strong diffractions come from a very irregular topography of pillow basalts, in most cases overlain by water and poorly compacted sediments of much lower density. Perhaps burial of the early Eocene oceanic crust by the basalts of the Siletz River Volcanics, of similar density, would eliminate the point sources of the numerous diffractions. The absence of strong diffractions may enhance the possibility of seeing reflections from within the oceanic crust, with the strongest reflections coming from the Moho transition zone (Collins and others, 1986).

A third possibility is supportive of the continental rifting model of Wells and others (1984) briefly described earlier in this paper. The layered reflections OC may represent metasedimentary rocks, forming a crustal type described by Nicolas (1985), which is formed in areas of continental rifting with high rates of sediment influx.

As continental rifting occurs, the new rift basin is filled with large volumes of sediments. These sediments are intruded by basalt sills and plugs that do not reach the surface to form a "normal" oceanic crust. The sediments are subsequently metamorphosed by the heat flux from the underlying mantle, resulting in a well-layered metasedimentary crust (Nicolas, 1985). The end result is an anomalous serpentinized upper mantle in direct contact with the sediments.

Wells and others (1984) suggested, on the basis of paleolatitude indicators and plate-motion modeling, that the Eocene seamount basalts were erupted from a hot spot overridden by an extensional (rift) basin formed during continental rifting. The presence of coarse, continentally derived sediments interlayered in the Eocene seamount basalts (Snavely, 1984) indicates that sediments were indeed being deposited during formation of the seamounts and may have contributed to the formation of a metasedimentary crust during the early Eocene opening of the extensional basin. Further thickening of the metasedimentary crust may have been accomplished by the underplating of a deep crustal sill complex, as has been suggested for the Hawaiian-Emperor seamount chain (Watts and others, 1985). A deep crustal sill complex may also contribute to the layered appearance of reflections OC.

Velocity data suggest that the zone OC may be some type of oceanic crust. Teleseismic and unreversed refraction data (see CR and COR, Fig. 8) in the Coast Range suggest an average velocity of about 6.6 km/s for zone OC, which is comparable to velocities of 6.2–7.3 km/s for young oceanic crust and slower than the 5.65 km/s for a metasedimentary crust (Nicolas, 1985).

The 6.6 km/s velocity, however, does not exclude the two models proposed in this paper. Underplated oceanic crust, in the absence of extensive subduction-complex sediments, would by its nature have similar velocities. The 5.65 km/s may or may not be a representative velocity for a metasedimentary crust. Nicolas (1985) reported the metasedimentary crustal velocity for only one of the nine basins from which he describes his model. The model describes the metasedimentary crust in terms of a presently active rift zone, whereas zone OC is no longer a zone of active rifting and has undergone burial since its formation. Deeper burial would

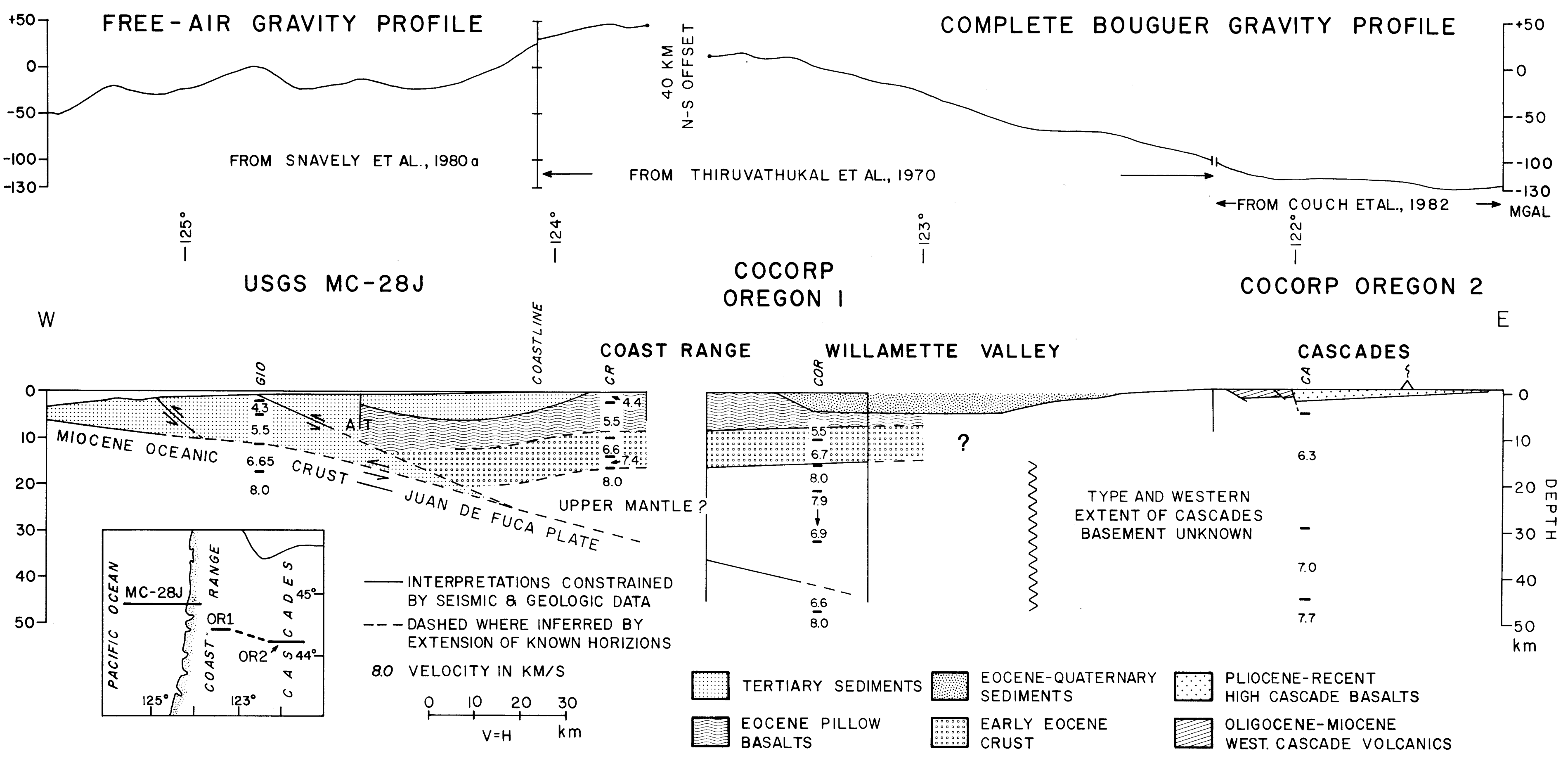


Figure 8. Regional transect of western Oregon, continental margin–volcanic arc. COCORP reflection profiles are combined with the U.S. Geological Survey continental margin transect MC-28J (Snavely and others, 1980a). Dashed lines represent an extrapolation between seismic reflection data sets. Inferred horizons were further constrained by velocity profiles by Shor and others (1968), G10; Berg and others (1966), CR; Langston (1977), COR; and Leaver and others (1984), CA. Profiles G10 and CA were reversed refraction profiles, whereas CR was unreversed. COR was derived from teleseismic P-wave conversions. The offshore gravity data are free-air, and onshore data are Bouguer.

result in higher pressures, a higher degree of metamorphism, and a corresponding increase in velocity. The refraction survey by Berg and others (1966) (see CR, Fig. 8) reports that zone OC is underlain by a thin layer with a velocity of 7.4 km/s (slower than mantle but faster than oceanic crust), which in Nicolas' model would correspond to the anomalous serpentinized upper mantle.

More and better-constrained velocity data could help resolve some of the uncertainties of the proposed models for the zone OC. Future velocity work would need to reverse the refraction profiles and concentrate on defining velocities in the lower crust and upper mantle.

If either of the latter two models is correct, then the strong reflections at 5.0 s (twtt) (Fig. 6) may represent the present and pre-accretion Eocene Moho beneath the Siletz River Volcanics. Five seconds (twtt) corresponds to a depth of about 16 km and agrees well with the results of Berg and others' (1966) refraction profile that suggested that the present base of the crust beneath the Coast Ranges is 16 km deep.

The deepest reflections from Oregon line 1 occur at depths of 35–40 km (10.0–11.0 s, Fig. 6), dip eastward, and may be from the zone of décollement above the Juan de Fuca plate (JFP). The east-dipping set of subparallel reflections may indicate that the décollement is not a sharp boundary but rather a zone of underplated, imbricated sediments and/or mafic oceanic crust as has been interpreted on LITHOPROBE data from Vancouver Island (Green and others, 1986). This interpretation is consistent with interpretations of other geophysical observations with respect to the depth and attitude of the Juan de Fuca plate by Langston (1981) and Riddihough (1979), but not with work done by R. W. Couch (1984, written commun.).

Crustal cross sections constructed by Langston (1981) and Couch are both very near COCORP Oregon line 1, but the profile of Riddihough (1979) is about 2° to the north. Langston (1981) concluded from the observation of teleseismic P-wave conversions that the subducting JFP would have a Moho at about 44 km depth near Corvallis, Oregon. R. W. Couch (1984, written commun.) and Riddihough (1979) both have used modeling of gravity and refraction data to estimate the depth and configuration of the JFP beneath the Coast Ranges, with differing results. Couch modeled the Moho of the JFP at a very shallow 18 km beneath the Coast Range. Modeling by Riddihough (1979) shows a depth of about 40 km for the Moho, which is more consistent with the seismic reflection results of 35–40 km for the presumed reflections from the subducting JFP.

An alternate suggestion is that the deep reflections may be from the Moho of the descending JFP. This implies that the décollement above the JFP is not seismically distinct, as indicated by the few reflections between 5.0 and 10.0 s (see Figs. 6 and 7). This interpretation, however, is not supported by Langston's (1977) crustal velocity model for the Corvallis area (see COR, Fig. 8).

Detailed seismic refraction data coincident with the COCORP reflection profile could help to discriminate amongst these several possibilities by providing deep seismic velocity control.

Oregon Cascades

Reflections on the western half of Oregon line 2 (stations 12–370) are characterized by numerous discontinuous events mostly confined to the upper 3.0 s (Fig. 5). (Note that all times referenced to west of station 370 have been time shifted +800 ms, as explained earlier.) Many of the short discontinuous reflections are horizontal to gently west dipping and appear truncated by two zones of relatively continuous east-dipping events (see A and B, Fig. 5). East of event B, beneath stations 200–220, another set of weak, gently west-dipping reflections terminates near the surface exposure of the steeply east-dipping Horse Creek fault zone (HCFZ). These terminations are inferred to mark the subsurface expression of the Horse Creek fault zone (see HCFZ, Fig. 5). One weak, east-dipping reflection is imaged between 5.0 and 6.0 s under stations 75–140. Farther east, below stations 320–375, reflections between 1.0 and 1.8 s are horizontal to gently east dipping.

Reflections beneath the High Cascades are mostly confined to the upper 1.5 s. The most prominent reflection, event P, has a gently westward dip and is traceable eastward from approximately station 440 at about 1.0 s (twtt) to a near-surface position at the east end of the line (Fig. 5). Other reflections include a set of discontinuous, subhorizontal events beneath event P in the vicinity of stations 620–670 and a series of horizontal and dipping events from 0.2–0.6 s (twtt) under the axis of the High Cascades (stations 475–525).

Features A and B are interpreted to be reflections from east-dipping normal fault planes. These faults are interpreted on the basis of geology and seismic character as outlined below. The fault plane of the steeply east-dipping HCFZ is not imaged; however, the termination of several horizontal reflections is suggestive of the location of this important fault zone.

Reflection A projects to the surface at about station 30 and may be related to the Cougar Reservoir fault (CRF), thought to have controlled major subsidence during the Miocene (Priest and Woller, 1983). The CRF has been mapped 5 km south of station 60 and has been inferred to extend northward with a north-south strike (Priest and others, 1983); however, there appears to be no expression of the CRF on the reflection data at station 60. This may be due to (1) the steep (70°) dip of the fault as defined by Priest and Woller (1983), (2) a lack of continuous, reflective markers to define offset or truncation, or (3) the fact that the strike of the CRF curves northwest and intersects the COCORP line at station 30 (that is, A is the CRF).

If the CRF does intersect the COCORP line at station 60, it must be truncated by or sole into fault A because A does not appear to be offset near station 60. The apparent migrated dip of A is 45° to the east, although it is not possible to determine its true dip and strike without three-dimensional control. If fault A is a northwest extension of the CRF, it would strike N30°W and have a true dip (after migration) of about 50° to the northeast. The amount of offset along fault A is difficult to determine without the presence of reflective horizons or geologic data that can be correlated to known horizons. The deepest age assignments for the Western Cascades shown in Figure 5 are speculative, as indicated by the question marks. The shallower age assignments were given by integration of the seismic data with geologic mapping in the area by Brown and others (1980) and Flaherty (1981).

A steep residual gravity gradient with a north-south trend led Couch and others (1982a) to propose a major north-south fault, with down-to-the-east motion, at a position generally coincident with the CRF. In the immediate vicinity of the COCORP line, however, the steep residual gravity gradient is absent (Couch and others, 1982b). There is an anomalous residual gravity low between faults A and B (Fig. 5). The east side of the residual gravity low gives way to a steep residual gravity gradient at a position coincident with the position of fault B (approximately station 130). This low could suggest the presence of low-density Eocene sedimentary and volcanic rocks in the footwall, below a relatively thin cover of Oligocene and Miocene volcanic rocks, which are juxtaposed against Oligocene-Miocene basalts on the east side of the fault (Flaherty, 1981). This would imply that higher-density volcanics have replaced lower-density sediments, which have been displaced downward on the east side of the fault.

Inferred fault B projects to the surface at about station 135 and is defined by east-dipping events that truncate numerous horizontal and gently west-dipping reflections (Fig. 5). Fault B has an apparent migrated dip of about 40°. The location of fault B is coincident with the easternmost extent of Oligocene rocks and the axis of a north-south syncline as mapped by Brown and others (1980). West of station 130 and to the north of the seismic line, the top of the Oligocene volcanic rocks are mapped over a large area at an elevation of about 900 m, where they are overlain by Miocene volcanic flows. East of station 130 and west of the HCFZ, Oligocene units are absent and Miocene units are mapped to elevations as low as 610 m. The basal west-dipping reflections (stations 155–200) at 1.9 s (twtt) may represent faulted Oligocene rocks overlain by flows of Miocene basalt and basaltic andesite that ponded against and/or draped over an east-facing scarp (Fig. 5). Subsequent flows in late Miocene and Pliocene time could have then completed burial of this fault scarp.

The proposed relationship between fault B and the residual gravity low previously discussed suggests a significant amount of normal offset for the fault. The lowermost reflections east of fault B (Fig. 5) are at about 2.2 s (about 4 km), whereas immediately west of fault B, they are at about 1.6 s (about 2.2 km), suggesting an offset of about 2 km.

The HCFZ is mapped as a steeply east-dipping fault zone that strikes north-south (Flaherty, 1981; Brown and others, 1981; Jan, 1967). The HCFZ correlates directly with the steep truncation of weak, gently west-dipping, layered reflections beneath stations 200–220 (Figs. 5 and 8). The actual amount of offset along the HCFZ is not obvious from the seismic data; however, a minimum offset of 615 m has been estimated by Flaherty (1981). Most of the layered reflections that are truncated east of the HCFZ probably represent upper Pliocene–Pleistocene flows from the Plio-Cascades that were emplaced against the east-facing Pliocene fault scarp. This interpretation is based on field evidence (Flaherty, 1981; Taylor, 1981) that shows lower Pliocene volcanic rocks to be structurally much higher on the west side of the HCFZ relative to upper Pliocene–Pleistocene volcanic rocks on the east side. The base of the layered upper Pliocene–Pleistocene(?) reflections is 1.6 s or about 2 km depth (Figs. 5 and 8). If these basal reflections correlate with the base of the lower Pliocene units mapped at the surface on the west side of the HCFZ, an offset of at least 2 km is indicated.

Reflections farther east (stations 220–360) of the HCFZ appear to have a lower limit of about 1.6 s (Fig. 5) and may also represent the base of upper Pliocene–Pleistocene volcanic flows. They continue eastward to the end of line 2W (below station 370). Several weak reflections that dip westward beneath stations 250–360 on an unmigrated section appear to project updip to a position coincident with the easternmost extent of shallow reflections on line 2W. This would, perhaps, suggest a structural relationship such as a fault. After migration, however, these weak reflections are moved east of station 370, implying that a structural relationship is not a valid interpretation. The depth and nature of these lowermost reflections suggest that they can be traced eastward and can be correlative with reflection P beneath the High Cascades (Fig. 5).

An eastward projection of reflection P to the surface places it at a position equivalent to the Pliocene-Pleistocene boundary (Peterson and others, 1976). An interpretation of reflection P suggests that it represents the boundary between the upper layers of Pliocene units associated with the Plio-Cascades of Taylor (1981) and the base of the Pleistocene High Cascades platform. Stratigraphically, this boundary represents a change from the ash-flow tuffs of the lower Pliocene to the basalt flows of the upper Pliocene–Pleistocene (Taylor, 1981; Peterson and others, 1976). Immediately east of the east end of line 2 (east of Sisters, Oregon), ash-flow tuffs form the top of the lower Pliocene Deschutes Formation, whereas to the west, Pleistocene basalts dominate. An exact structural boundary has not been identified, owing to the recent flows which cover the area (and the Tumalo fault zone) (Peterson and others, 1976; Smith, 1985).

The shallow depth of reflection P near the Tumalo fault zone suggests that the eastern High Cascades boundary, near Sisters, has experienced very little offset, much less than the offset of as much as 3,000 m documented 20 km to the north along the Green Ridge fault zone (Taylor, 1981; Conrey, 1985). Near the seismic line, the Tumalo fault zone (see TFZ, Fig. 3) has no escarpment and is not imaged on the COCORP survey (although a steep fault might not be seen, owing to limited CDP coverage at the east end of 2E). This interpretation suggests that faulting along the length of the High Cascades is not continuous in magnitude or strike. Recent studies of volcanic stratigraphy north of the Green Ridge fault resulted in a similar conclusion (Yogodzinski, 1985).

Shallow reflections from 0.2–0.6 s in the area of McKenzie Pass (stations 495–500) may be related to repeated overlapping of volcanic flows from volcanoes near the pass such as Belknap Crater, Black Crater, and North Sister.

The Cascades have a prominent geothermal gradient (Blackwell, 1982) that might be explained in part by the regional structure. The regional heat flow increases from west to east along the line (Fig. 3). The increase in heat flow may be facilitated by the presence of faults that juxtapose rocks of different lithology and higher heat capacity. The faults themselves may act as a conduit for heat convection, as evidenced by the presence of hot springs along the HCFZ. The pronounced heat-flow gradient in the Western Cascades occurs across the series of the faults identified by the seismic reflection data and may imply that faulting is a controlling factor of the heat flow (Figs. 3 and 5).

The amount and type of faulting along the length of the High Cascades is well documented in only a few places (Smith, 1985). The presence of faults in some places has led to the assumption that faulting is continuous along the length of the boundaries of the range (Allen, 1966; Taylor, 1981). North-south vent alignments have been cited as evidence of major faulting within the range and especially along the east side (Taylor, 1981). The reflection data, however, seem to indicate that at the latitude of the COCORP line, the High Cascades have experienced subsidence in a half-graben along the west side of the range during the Pliocene. Although some faulting appears to have occurred within the range, most of the offset was along the HCFZ, on the western boundary of the High Cascades, with minor offset along the eastern-bounding TFZ.

North of the east end of the COCORP line, evidence of significant down-to-the-west offset along the Green Ridge fault zone (GRFZ) has been used to suggest the presence of a major bounding fault along the length of the east side of the High Cascades (Taylor, 1981; Conrey, 1985; Smith, 1985). Conrey (1985) has identified several faults with down-to-the-west motion associated with the GRFZ and has shown that the faulting has had scissor-like motion, with the faults having an increasing amount of offset to the north along the GRFZ. The apparent hinge point for the scissors is just south of the GRFZ and north of the COCORP line. Conrey's (1985) results show minor offset just north of the east end of line 2E and are consistent with the conclusion that little offset has occurred in the immediate vicinity of the COCORP line.

The seismic results, combined with field data, suggest that there is no single, long, continuous master fault adjacent to the High Cascades and no evidence for a simple High Cascades graben. Rather, along a north-south trend, a Pliocene platform was broken into a series of normal-faulted blocks upon which the High Cascades were built. Each of these blocks has experienced varying degrees of faulting, with some subsiding as half-grabens (as shown on the reflection data). Subsidence of these blocks need not be thought of as having strictly down-to-the-east or down-to-the-west motion. The scissor motion of the GRFZ produced northerly dips of stratigraphic units within the High Cascades (Conrey, 1985).

CONCLUSIONS

The seismic reflection data, when viewed in a regional context, provide constraints to understanding and proposing models for the regional structure of the convergent boundary along the northwestern United States. A regional transect depicting the possible crustal structure of western Oregon combines the results of this reflection survey, a U.S. Geological Survey (USGS) continental-margin transect (Snavely and others, 1980a), and other geophysical and geological data and is shown in Figure 8.

In Figure 8, the Juan de Fuca plate is interpreted to dip 15° to the east beneath the Oregon Coast Range and is constrained by the 15° dip of the deep reflections at 35–40 km depth on Oregon line 1, which is projected updip to the top of the Juan de Fuca plate as identified on USGS MC-28J (Snavely and others, 1980a). These deep reflections probably come from a décollement zone above the subducting Juan de Fuca plate.

The reflection data support the model of a two-layered, 16-km-thick crust (Berg and others, 1966; Langston, 1977) beneath the Oregon Coast Range, composed of 8 km of Eocene pillow basalts (Siletz River Volcanics) underlain by approximately 8 km of what may be the original crust upon which these basalts were erupted. This crust may be oceanic crust or a unique metasedimentary crust formed during early Eocene rifting of the North American continental margin. This two-layered crust is inferred to extend westward beneath the continental shelf (Fig. 8); however, its eastern extent beneath the Willamette Valley is not known. Velocity profiles CR (Berg and others, 1966) and COR (Langston, 1977) provide an additional constraint for extending the 16-km-thick crust westward from Oregon line 1.

A Bouguer gravity anomaly high of +50 mgals over the Coast Range (Fig. 8) may be due to a mass excess in both the crust and a shallow upper mantle (Thiruvathukal and others, 1970). The relatively small, although broad, −20 mgals free-air gravity anomaly over the marginal basin off the coast (Fig. 8) is most probably due to the presence of the Eocene crust and pillow basalts beneath the basin. The Bouguer gravity anomaly decreases eastward over the Willamette Valley and the Cascades and is probably reflective of an increasingly thicker crust to the east (Leaver and others, 1984).

Identifiable reflections were limited to the shallow crust on Oregon line 2. The shallow results of Oregon line 2 are shown in Figure 8 and support the general concept of normal faulting within the Cascades. Seismic reflection data from Oregon line 2W exhibit a series of down-to-the-east normal faults with a combined offset of as much as 4 km or more. In the High Cascades, the reflection data have suggested the presence of a large, shallow half-graben along Oregon line 2 with most of the offset along the western boundary. Furthermore, the data suggest that in general, the High Cascades are built on a series of blocks that were faulted independently during the early Pliocene.

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APPENDIX 1. ACQUISITION AND PROCESSING INFORMATION

Field parameters
- 5 vibrator source array, 100 m long
- Vibrate every station, 8 sweeps per vibration point (VP)
- 8–32 Hz upsweep, 32 s sweep, 48 s total recording time
- 16 s record
- 96 channel recording system, nominal offsets 400–9,900 m
- 100 m group interval, 100 m 24 geophone linear array

Processing sequence: Oregon line 1
- Demultiplex
- Vibroseis correlate
- Trace amplitude balance—1,000 ms window
- "Crooked line" geometry
- Deconvolution—33 ms lag
- Mute refracted arrivals
- Trace edit
- Datum statics—to 80 m at 4,000 m/s
- Common midpoint sort (gather)
- Velocity analysis
- Normal moveout (NMO)
- Post-NMO (stretch) mutes
- Stack—nominal 48-fold, 50 m trace spacing
- Coherency filter

Processing sequence: Oregon line 2W
- Demultiplex
- Vibroseis correlate
- Trace amplitude balance—1,500 ms window
- "Crooked line" geometry
- FK filter
- Deconvolution—spatially varying 32–56 ms lag, two zones
- Mute refracted arrivals
- Trace edit
- Datum statics—to 338 m at 3,800 m/s
- Common midpoint sort (gather)
- Velocity analysis
- Normal moveout (NMO)
- Post-NMO (stretch) mutes
- Stack—nominal 48-fold, 50 m trace spacing

Processing sequence: Oregon line 2E
- Demultiplex
- Vibroseis correlate
- Trace amplitude balance—1,000 ms window
- "Crooked line" geometry
- FK filter
- Deconvolution—spatially varying 55–70 ms lag, two zones
- Filter 8–32 Hz bandpass
- Trace edit
- Interactive refraction statics
- Common midpoint sort (gather)
- Velocity analysis
- Normal moveout (NMO)
- Post-NMO (stretch) mutes
- Stack—nominal 48-fold, 50 m trace spacing

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