Tectonics of the Neogene Cascadia forearc basin: Investigations of a deformed late Miocene unconformity

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ABSTRACT

The continental shelf and upper slope of the Oregon Cascadia margin are underlain by an elongate late Cenozoic forearc basin, correlative to the Eel River basin of northern California. Basin stratigraphy includes a regional late Miocene unconformity that may coincide with a worldwide hiatus ca. 7.5–6 Ma (NH6). The unconformity is angular and probably subaerially eroded on the inner and middle shelf, whereas the seaward correlative disconformity may have been produced by submarine erosion; alternatively, this horizon may be conformable. Tectonic uplift resulting in subaerial erosion may have been driven by a change in Pacific and Juan de Fuca plate motion. A structure contour map of the deformed unconformity and correlated seaward reflector from seismic reflection data clearly outlines deformation into major synclines and uplifted submarine banks. Regional margin-parallel variations in uplift rates of the shelf unconformity show agreement with coastal geodetic rates.

The shelf basin is bounded to the west by a north-south–trending outer arc high. Rapid uplift and possible eustatic sea-level fall resulted in the formation of the late Miocene unconformity. Basin subsidence and outer arc high uplift effectively trapped sediments within the basin, which resulted in a relatively starved abyssal floor and narrower Pliocene accretionary wedge, particularly during sea-level highstands. During the Pleistocene, the outer arc high was breached, possibly contributing to Astoria Canyon incision, the primary downslope conduit of Columbia River sediments. This event may have caused a change in sediment provenance on the abyssal plain ca. 1.3–1.4 Ma.

Keywords: accretionary wedges, Cascadia subduction zone, forearc basins, neotectonics, submarine fans, unconformity.

INTRODUCTION

The Cascadia continental shelf is underlain by a thick sedimentary sequence extending from the Eel River basin in the south to offshore Vancouver Island in the north. This forearc basin was probably continuous in the early to middle Eocene (Niem et al., *in* Christiansen and Yeats, 1992) but has subsequently been deformed and dissected into smaller basins, and eroded at its western margin. The basin stratigraphy contains several regional unconformities, suggesting a complex history of vertical tectonics, sedimentation, and eustatic sea-level change. An extensive middle to late Miocene unconformity was first reported by Kulm and Fowler (1974) and later by Snavely (1987), but the extent, origin, and deformation of this surface during the Pliocene and Quaternary have not been documented in detail. Cranswick and Piper (1992) produced a regional isopach map for the entire forearc basinal sequence on the Oregon margin. The late Miocene angular unconformity is easily identifiable in seismic reflection profiles across the Oregon continental shelf; however, it is less continuously traceable on much of the Washington shelf, and between central and southern Oregon. In these regions, the uncertainty of identifying and tracing the unconformity is due to poor stratigraphic control, nonangularity, faulting, or uplift and erosion. We therefore restricted our study, including the construction of a structure contour map of the unconformity, to the central Cascadia margin. The unconformity is commonly angular and therefore easily identifiable in this region, with stratigraphic control from exploration wells and seafloor samples. A younger regional unconformity, probable latest Pliocene or early Pleistocene age, is also recognized on the Oregon shelf, but it is less continuous than the late Miocene unconformity.

In this paper we discuss the age, origin, and deformation of the late Miocene unconformity and incorporate this information with the stratigraphy of the forearc basin, adjacent accretionary wedge, and abyssal plain to document the Neogene evolution of the Cascadia margin. Of particular interest are the tectonic and sedimentary interactions between the accretionary wedge and forearc basin and the influence of tectonics and eustatic sea-level fluctuations on sedimentation, erosion, and deformation. We also focus on variation of deformational patterns across and along the margin, and the extent of influence of basement lithology on basinal deformation.

METHODS

Seismic reflection profiles, shelf exploration well chronology, dated seafloor samples and cores, and Ocean Drilling Program (ODP) and Deep Sea Drilling Project (DSDP) drill site stratigraphy (Fig. 1) were used to map and interpret the late Miocene unconformity within the forearc basin. The unconformity and seaward correlative reflector were traced on seismic profiles throughout the central and northern Oregon margin to produce a structure contour map. A proprietary data set of multichannel migrated seismic reflection profiles used for the map was collected in two acquisition phases: (1) 1975, 46-channel, and (2) 1980, 96-channel, which form a network of north-south and east-west profiles

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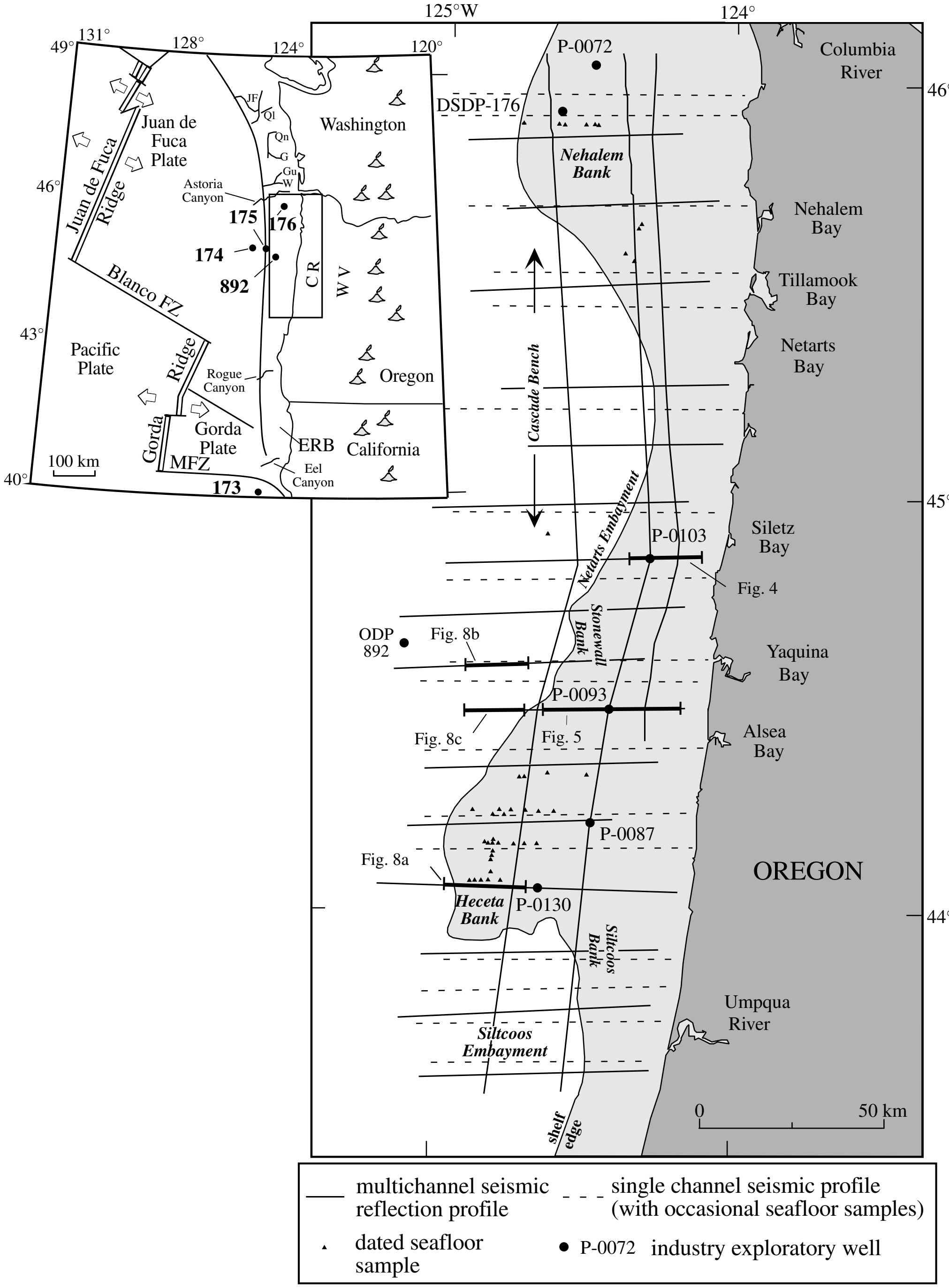

Figure 1. Tectonic setting of Cascadia subduction zone (inset) and study area indicating data sets used to identify late Miocene unconformity and determine stratigraphy of the forearc basin. Inset shows locations of relevant Ocean Drilling Program (ODP) and Deep Sea Drilling Project (DSDP) sites. WV—Willamette Valley, CR—Coast Range, ERB—Eel River basin, Washington submarine canyons: JF—Juan de Fuca; Ql—Quillayute; Qn—Quinault; G—Grays; Gu—Guide; W—Willapa. Main figure shows multichannel seismic (proprietary, solid) and single-channel data (dashed) used to trace the unconformity and locates dated samples at industry exploration wells, along single channel seismic lines (discontinuous samples along dashed lines) and at individual sample sites (triangles). Main structural features of the central Cascadia forearc basin are also labeled in italics. The Cascade Bench (Kulm and Fowler, 1974) extends between Heceta to Nehalem Bank along the upper slope of the Netarts Embayment. The shelf edge is defined here and in subsequent figures as a topographic break identified on seismic profiles. Bold lines represent seismic profiles illustrated in subsequent figures.

across the shelf and upper slope (Fig. 1). Additional seismic profiles (both single channel from Shell Oil Company (Fig. 1) and multichannel from U.S. Geological Survey [USGS] and industry) were used to confirm the position and depth of the unconformity in data gaps of the primary data set. The unconformity and a younger latest Pliocene or early Pleistocene unconformity and correlative reflectors were identified and traced out on each seismic profile. Sources of error in identifying the unconformity include tracing areas without angular truncation, regions where the surface is traced across uplifted regions and faults, and interpolation between seismic profiles. Where the unconformity could not be traced with confidence, data points were not included. The degree of uncertainty of age and position of the late Miocene unconformity were ranked based on these criteria to ascertain confidence and accuracy when mapping and interpreting this surface. The traced unconformities were digitized and converted to xyz values in UTM (Universal Transverse Mercator) coordinates and two-way traveltime. Traveltime was then converted to depth using published velocities of late Neogene units from refraction experiments (Shor et al., 1968) and from sonic well logs (Palmer and Lingley, 1989; Cranswick and Piper, 1992), loosely constrained by wide-angle seismic reflection and refraction data, which focused on deeper units (Tréhu et al., 1994; Parsons et al., 1998). From these data, the overlying stratigraphic section was divided into Quaternary and Pliocene units, separated by the younger (earliest Pleistocene) unconformity, and assigned velocities of 1.7 km/s and 2.1 km/s, respectively, for depth conversion. The ages of these two units are unconfirmed by biostratigraphic or radiometric age away from well or seafloor sample locations, and are therefore only approximate. Good agreement between calculated and measured depths at well locations supports this velocity model. Depth values represent depth to the unconformity below sea level rather than the seafloor.

The unconformity surface was initially contoured by hand within a CAD (computer aided design) system to prevent the introduction of artifacts between the discrete lines of data points. The resulting hand-contoured xyz data were converted to a TIN (arc/info triangulated irregular network). The TIN allows the irregularly distributed data to be densified, producing a more regular data set with reduced potential for artifacts. The xyz TIN file was finally regridded (continuous curvature surface, grid cell = 300 m) in GMT (generic mapping tools software; Wessel and Smith, 1991), and converted to a shaded relief image. This file was also imported into and georeferenced in a GIS (geographic information system) to compare with other data sets such as swath bathymetry, sidescan sonar, gravity and magnetic anomalies, a neotectonic map, and previous structural interpretations.

CASCADIA FOREARC BASIN STRATIGRAPHY

The late Miocene unconformity is within a sequence of forearc sediments of middle Eocene to Pleistocene age on the Oregon continental shelf and late Miocene to Pleistocene age on the Washington shelf (Fig. 2; Snavely, 1987; Palmer and Lingley, 1989; Christiansen and Yeats, 1992, Snavely and Wells, 1996). The stratigraphy of the Eel River basin, northern California, is also illustrated in Figure 2 (Ingle, 1987; Clarke, 1992; McCrory, 1995). Basinal sediments overlie Eocene oceanic basalt of the Siletzia terrane on the central Oregon margin and Eocene to middle Miocene mélange and broken formation of the ancient accretionary complex on the Washington and northern Oregon margin (Fig. 2), and presumably west of the seaward Siletzia margin on the central Oregon margin. Middle Miocene basalts on the coast and inner shelf have an identical geochemical signature to the most far-traveled Columbia River Basalt Group flood basalts, and are thought to represent equivalent coastal and submarine invasive flows into water-saturated sediments, as indicated by peperite dikes and sills (Beeson et al., 1979). Coastal basalts are palagonitized breccias and pillow lavas, with subaerial flows. The Miocene shoreline was close to its present-day position. An early Pleistocene (possibly latest Pliocene) unconformity is identified in both stratigraphy and in seismic reflection records (e.g., Kulm and Fowler, 1974; Palmer and Lingley, 1989).

The seaward extent of basinal sediments is marked by an outer arc high on the current outer shelf and upper slope (Fig. 3). The unconformity reaches the seafloor on the innermost shelf, east of which Miocene and older sediments crop out on the seafloor. The central Cascadia forearc basin is currently almost entirely filled (see bathymetric base map of Fig. 3). Holocene hemipelagic sedimentation has been minimal compared to the Pleistocene glacial period. During the Pleistocene, submarine canyons transported the majority of sediments directly to the abyssal plain and submarine fans, such as the Astoria fan (Nelson, 1976).

RESULTS

Age of the Unconformity

When initially described, the age of this regional Cascadia unconformity was tentatively placed at the middle-late Miocene boundary (Kulm and Fowler, 1974). Benthic foraminifera on either side of the unconformity from oil-exploratory wells on the continental shelf (S.D. Drewry et al., 1993, Minerals Management Service [MMS] unpublished work) provide an approximate age ranging between middle Miocene and earliest Pliocene. Benthic foraminiferal stages are those of Kleinpell (1938) and Mallory (1959), originally defined in Californian stratigraphy. These stages are time transgressive and document water depth more accurately than age, based on comparisons with more reliable coccolith stratigraphy (Bukry and Snavely, 1988). Therefore, these stages are indicative of age only in a general way. In the following section, we attempt to constrain further the age of this unconformity while considering the limitations of the biostratigraphy.

The youngest sediments underlying the unconformity are Mohnian (P-0130, P-0103 exploration wells, Oregon shelf) and Delmontian (P-0072, P-0075, Washington-Oregon border, P-0155, P-0141, P-0150, Washington shelf) from benthic foraminifera (S.D. Drewry et al., 1993, MMS unpublished work). Delmontian fauna indicate late Miocene age, but may be equivalent to part of the middle and late Miocene Mohnian Stage and therefore somewhat older (Barron, 1986). Fauna immediately overlying the unconformity are the Repettian to Wheelerian Stage of Natland (1952) (Pliocene to Pleistocene) with one example of Miocene Mohnian fauna identified (we suggest that this is reworked sediment).

Results from the microfossil chronologies of hiatuses in worldwide drill holes suggest a correspondence in the Miocene-Pliocene section, with global tectonic and climatic origins (Barron and Keller, 1983; Keller and Barron, 1987). Two prominent late Miocene unconformities occur at 10.5–9.2 Ma and 7–6 Ma, corresponding to Neogene hiatuses NH4 and NH6, respectively. Both hiatuses are present at ODP Site 892 on the Oregon upper continental slope (Fig. 1), dated as 11.4–9.0 Ma and 7.45–6.2 Ma (Fourtanier and Caulet, 1995). The regional nature of the angular unconformity on the Oregon (and Washington) shelf suggests that it may be related to one of these two global hiatuses. The lack of late Miocene fauna above the unconformity and the probable presence of late Miocene sediments beneath the unconformity supports a correlation with NH6 (7–6 Ma). This hypothesis is supported by the presence of two unconformities within the middle to late Miocene of offshore Washington (Fig. 2; Palmer and Lingley, 1989; McNeill et al., 1997), the younger (between the Montesano and Quinault Formations) being the likely correlative of the Oregon regional unconformity, based on interpretation and comparison of seismic reflection records and well stratigraphy. The older angular unconformity in Washington separates Eocene to middle Miocene mélange and broken formation and the overlying Montesano Formation. The late Miocene unconformity and its correlative seaward disconformity or conformity are likely to be time-transgressive in a seaward direction across the margin.

Regional and Global Correlations of the Late Miocene Cascadia Unconformity

Post-middle Miocene Willamette Valley (and Tualatin basin) sediments are predominantly homogeneous fine-grained fluvial and lacustrine (Yeats et al., 1996), and are therefore unlikely to reveal any significant hiatus or unconformity without accurate age control (Wilson, 1998, and 1998, personal commun.). However, a late Miocene unconformity is recognized locally in the Oregon Coast Range (Armentrout, 1980; Baldwin, 1981) and in the western Cascades (Hammond, 1979).

Evidence for the two late Miocene hiatuses also exists on the Californian continental margin. Both the NH4 and NH6 hiatuses are identified in DSDP drill site 173 (Fig. 1). The stratigraphy of the onshore Eel River basin (Fig. 1) reveals an unconformity and hiatus between the lower to middle Miocene Bear River beds and the overlying late Miocene Pullen Formation, which is the

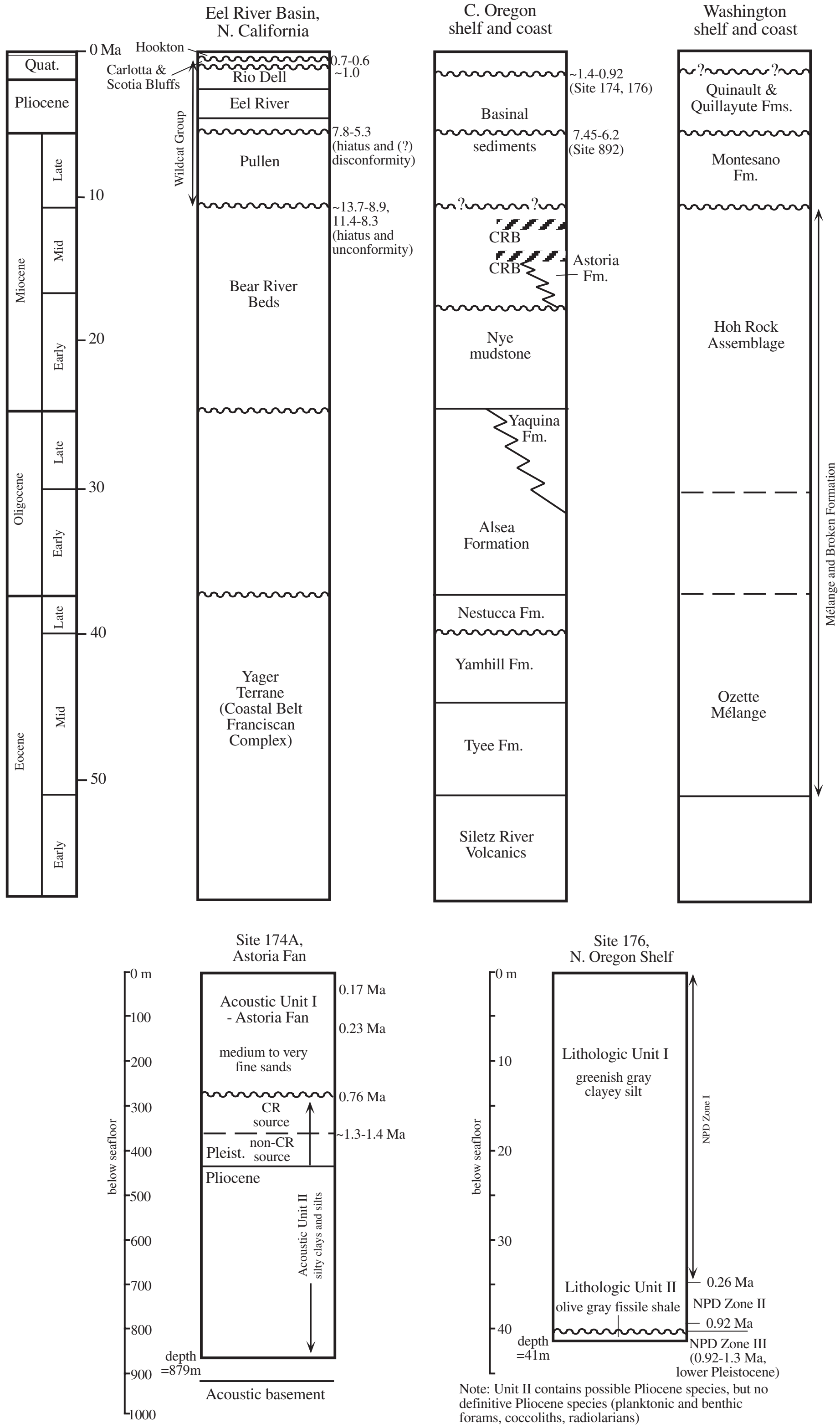

Figure 2. Generalized stratigraphy of the Eel River basin, northern California (Ingle, 1987; Clarke, 1992; McCrory, 1995), central Oregon coast and shelf (Snavely, 1987; Christiansen and Yeats, 1992; Snavely and Wells, 1996), and the Washington coast and shelf (Palmer and Lingley, 1989; McNeill et al., 1997). Also shown are stratigraphic columns of Deep Sea Drilling Project (DSDP) Sites 174 on the edge of the Astoria fan and 176 on the northern Oregon outer shelf of Nehalem Bank (locations in Fig. 1; Kulm et al., 1973; Schrader, 1973). Note that the stratigraphy is illustrated as depth columns. Ages for Site 174 are based on planktonic foraminifera (Ingle, 1973; Goldfinger et al., 1996a) and interpolation between known ages. Ages for Site 176 are based on diatoms (North Pacific diatom zones) with reference to other microfossils, including benthic and planktonic foraminifera, calcareous nannoplankton, and radiolarians reported in Kulm et al. (1973). CR—Columbia River, CRB—Columbia River Basalt, Pleist.—Pleistocene, NPD—North Pacific diatom. Ages of significant unconformities and hiatuses for each region are indicated.

basal unit of the Wildcat Group (Fig. 2; Barron, 1986; Clarke, 1992; McCrory, 1989, 1995). This is coeval with an unconformity throughout California. The Wildcat Group includes a hiatus and possible disconformity (McCrory, 1995) at 7.8–5.3 Ma (NH6), although Clarke (1992) found the Wildcat Group to be largely conformable in this area. However, Clarke (1992) mentioned an angular unconformity between the Miocene and Pliocene formations just north of Cape Mendocino in the southernmost part of this basin. This younger hiatus also corresponds to the top of the Monterey Formation of California and the Mohnian-Delmontian boundary (Keller and Barron, 1987), although the type Delmontian stage may be equivalent to the Mohnian (Barron, 1986). The NH6 event is associated with a period of major climatic and oceanographic change (including eustatic sea-level fall, Barron, 1986), coincident with an increase in terrigenous material, increased glaciation and general cooling, and intensified bottom current circulation and dissolution throughout the Pacific basin (Ingle, 1978; Keller and Barron, 1987). It is also coincident with widespread angular unconformity or disconformity throughout California, suggesting a coeval tectonic event (Barron, 1986).

On the Alaskan margin, two unconformities are present, middle-late Miocene and latest Miocene (Fisher et al., 1987), that may be the equivalents of NH4 and NH6. The earlier unconformity is thought to be associated with tectonic erosion resulting from subduction of the Kula-Pacific ridge at 10 Ma (Fisher et al., 1987).

Mechanism of Erosion: Subaerial vs. Submarine

The unconformable surface is angular on much of the inner and middle shelf (Figs. 3 and 4). However, the seaward correlative reflector, predominantly on the outer shelf and upper slope, is nonangular with respect to older strata and may be disconformable and/or conformable (Figs. 3 and 5). Evidence of eastward-onlapping beds above the unconformity indicates that the surface sloped gently westward about 1°, with local dips to 5°. Original relief of as much as 20 m on the unconformity was due to incomplete planation of Miocene folds. The erosional surface can therefore be treated as originally subhorizontal and largely planar. Two seismic reflection profiles on the inner shelf indicate that the middle Miocene Columbia River Basalts (16.5–12 Ma) were also truncated during this erosional event (Figs. 4 and 5).

The angularity of the landward unconformity initially suggests that it was formed by subaerial rather than submarine erosion. However, sediments immediately above and below the unconformity were deposited at water depths consistent with the outer shelf (outer neritic) to lower slope (lower bathyal), 100–3000 m water depth (Kennett, 1982), as indicated by benthic foraminifera. The majority of sediments are fine to medium grained (clay, silt, and silty sandstone) with little indication of deposition in a shallow-marine or high-energy environment. Sample transects across the late Miocene unconformity (along Shell Oil Company single-channel profiles indicated in Fig. 1) indicate possible evidence of shallow-marine sedimentation (carbonaceous debris, volcanic pebbles, and cobbles), but these sediments may have been transported. Sediments at the unconformable base of the Wildcat Group (ca. 10 Ma) and at the 7.8–5.3 Ma hiatus within the Wildcat Group of the onshore Eel River section (Fig. 2), in southern Cascadia, are also lower bathyal with no indication of in situ shallow-marine sedimentation (McCrory, 1995).

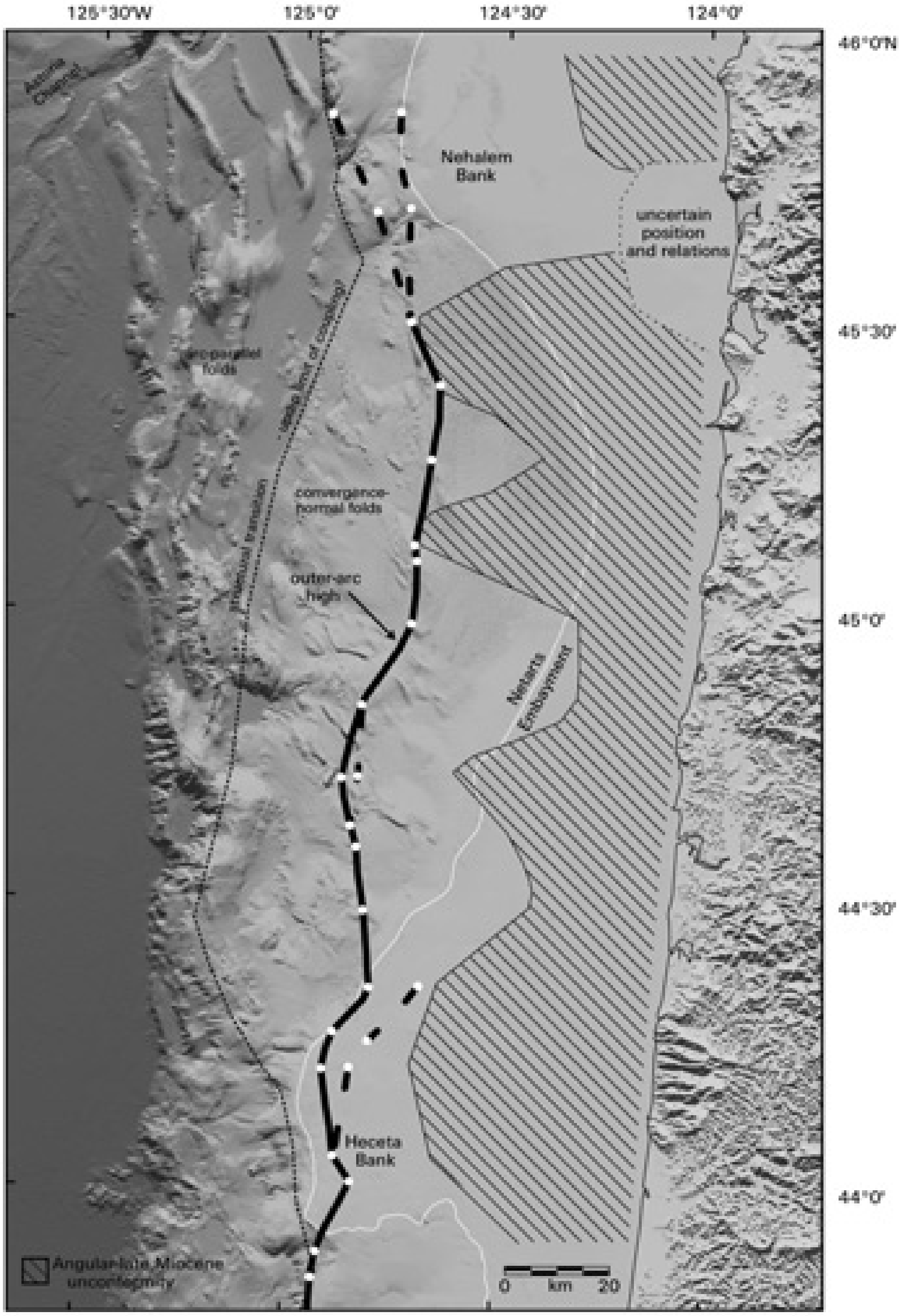

Figure 3. Extent of the late Miocene unconformity and correlative seaward conformity or disconformity overlying shaded relief bathymetry base map of the Oregon margin. Diagonal lines indicate where the unconformity is clearly an angular unconformity. Elsewhere, the correlative reflector is nonangular or partially angular but can still be traced as a continuous reflector to the outer arc high. White line represents present-day shelf break. Thick black line indicates the position of the outer arc high marking the seaward extent of the forearc basin, white dots represent positions on seismic profiles, dashed lines indicate alternative outer arc high positions. Thin dashed line represents a topographic break and change in fold orientation which may indicate the seaward extent of interplate coupling (Goldfinger et al., 1996b).

Subaerial Erosion

If erosion was subaerial, the Cascadia margin underwent a rapid change in relative sea level, a result of eustatic sea-level change and/or vertical tectonic motion. Assuming that eustatic sea level could have fallen a maximum of ~150 m, tectonic uplift is still required to bring deep-marine preerosional sediments to sea level, with the exception of those indicating outer shelf or uppermost slope conditions. Sediments immediately

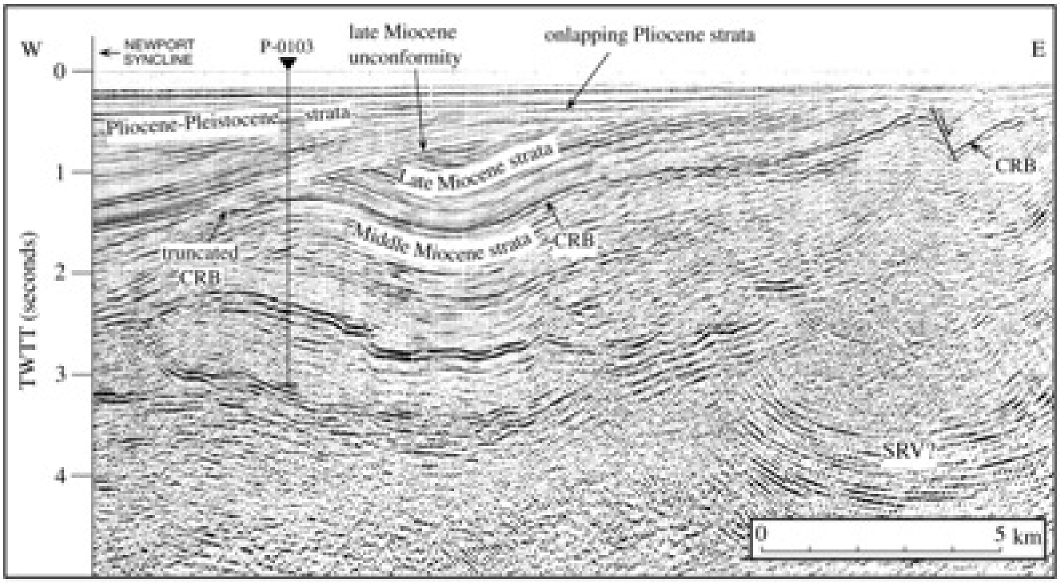

Figure 4. East-west multichannel seismic reflection profile west of Siletz Bay (located in Fig. 1) showing the angular nature of the late Miocene unconformity on the inner to middle shelf. This profile also shows the probable truncation of the Columbia River Basalt flow (CRB) just seaward of well P-0103, suggesting subaerial erosion. Pliocene sediments onlap the unconformity at very low angles, suggesting a gently westward dipping surface at the time of erosion. SRV—Siletz River Volcanics. After Yeats et al. (1998).

overlying the unconformity indicate water depths as shallow as outer neritic to upper bathyal (~100–500 m) (Kulm and Fowler, 1974; S.D. Drewry et al., 1993, MMS unpublished work). If sea level is allowed to rise a potential maximum of 150 m, no posterosional tectonic subsidence is required for the minimum water depths of these bathymetric zones. However, two industry wells (P-0150 and P-0141 on the Washington shelf) encountered middle-lower bathyal water depths immediately above the unconformity, which yields a significantly higher posterosional subsidence rate if these sediments were in situ. Using paleo-water depths and ages of sediments at oil-exploratory well sites above and below the unconformity and an age of 7.5–6.2 Ma for erosion (ODP drill site 892, Fourtanier and Caulet, 1995), uplift and subsidence rates can be approximated. Calculations yield reasonable values of tectonic uplift and subsidence rates, of the order of tens to hundreds of meters per million years, comparable to uplift rates determined for the Oregon outer shelf by Kulm and Fowler (1974) from seafloor samples. As the amount of section eroded is unknown, some error is introduced in estimations of the amount and onset of uplift.

The absence of a regressive sequence underlying the late Miocene unconformity can be explained by erosion during the hiatus. Several hypothesized mechanisms could explain the absence of a shallow-marine stratigraphic section overlying the unconformity: (1) the deposited section was sufficiently thin (low sedimentation rates) that it was not encountered during relatively infrequent well sampling; (2) an underlying shallow-marine section was eroded by submarine erosion as subsidence occurred; (3) subsidence rates or rates of sea-level rise were sufficiently rapid to prevent deposition of a thick shallow-marine section; (4) the locations of wells were such that they encountered thinner sedimentary sections. Mechanism 1 is possible, although there is little evidence of any transgressive sequence overlying the unconformity. Mechanism 2 is possible if deposition was minimal. Evidence of rapid sea-level rise following the last glacial maximum suggests that mechanism 3 is not uncommon (Fairbanks, 1989; Bard et al., 1996) and is likely if combined with the effects of mechanisms 1 or 2. Most industry wells are located on the flanks or crests of anticlines, where accumulation rates would have been lower than within synclinal basins. However, lack of significant topography on the erosional surface makes this hypothesis unlikely regionally. We favor a combination of mechanisms 1, 2, and 3 to account for absence of shallow-marine deposits, where a mechanism of subaerial erosion is proposed.

Submarine Erosion

Submarine erosion and periods of reduced sedimentation rates or nondeposition have been proposed as causes of angular unconformities where evidence of a shallow-marine environment is absent (e.g., Yeats, 1965; van Andel and Calvert, 1971; Teng and Gorsline, 1991). To determine whether bottom currents on the Cascadia margin would be capable of eroding this stratigraphic section, we estimate current velocities required to erode sediments of certain age and lithology, and compare these with present-day near seafloor velocities. Sediments underlying the unconformity range in age from early to middle-late Miocene. Consolidation and cementation of the older sediments would likely increase the velocity required for erosion, but the importance of these factors at the time of erosion is relatively unknown for the pre-Pliocene Cascadia sediments. Lithologies immediately above and below the unconformity are predominantly silt, with silty sandstone and clay and little sand. Minimum velocities (measured 1 m above seafloor, Miller et al., 1977) required to erode unconsolidated sediments, without consideration of cohesion of clays and silts, are sand 25–80 cm/s; silt 15–25 cm/s; and clay ~10–20 cm/s. Estimated values for cohesive clays and silts would be ~25–100 cm/s or possibly greater (P. Komar, 1998, personal commun.).

We examine the California Current, the predominant erosional shelf-slope current of this region (Hickey, 1989), as a present-day analogue of currents responsible for submarine erosion during the late Miocene. Measurements on the Oregon and Washington margins (J. Huyer and R. Smith, 1998, personal commun.; Hickey, 1989, respectively) indicate midslope to outer shelf near seafloor velocities regularly reaching ~10–20 cm/s, with maxima of 20–50 cm/s, but locally ranging to 70 cm/s on the mid-shelf (Smith and Hopkins, 1972). These recent current velocities suggest that submarine abrasion of the strata (of consolidation, based on age, and sediment type determined above) by the California Current on the outer shelf and upper slope is theoretically possible but less likely if silts and clays were cohesive and consolidated, and in the absence of abrasive coarse-grained sediments. It also seems unlikely that submarine currents would have sufficient strength to erode the Columbia River Basalt sills, which are clearly truncated in Figures 4 and 5.

Positions and velocities of submarine currents may vary significantly with fluctuations in sea

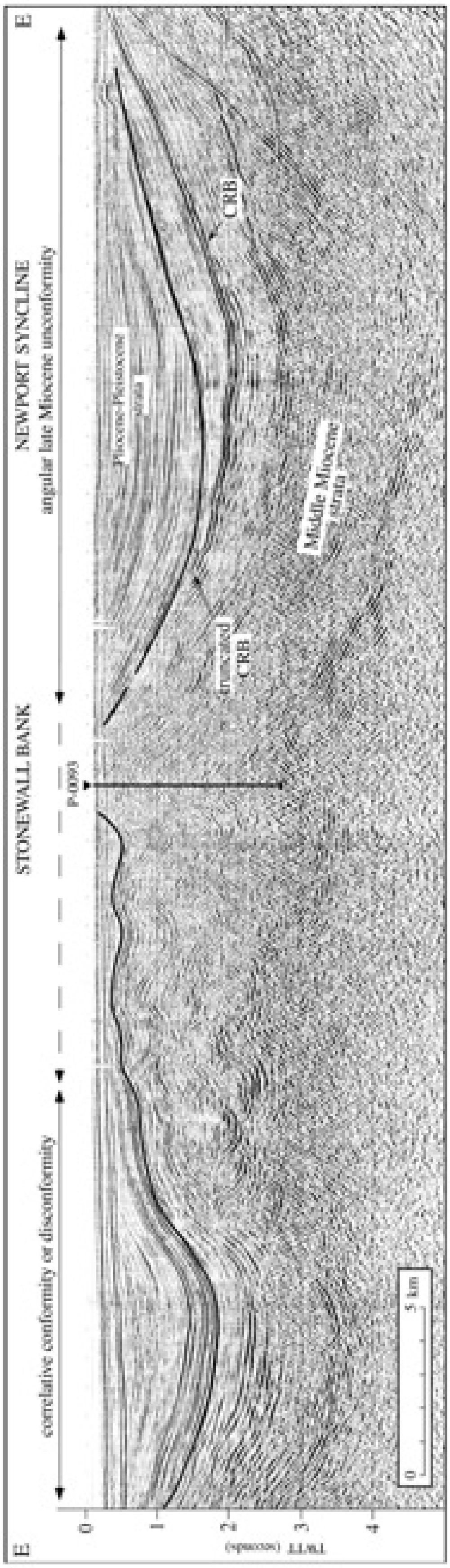

Figure 5. East-west multichannel seismic reflection profile across the continental shelf at Stonewall Bank west of Alsea Bay (Fig. 1). The late Miocene unconformity (bold black line) is angular landward of Stonewall Bank anticline, slightly angular just west of the bank where small wavelength folds are truncated, and conformable or disconformable seaward of these folds. This profile typifies the character of the unconformity from east to west. Note that the unconformity truncates the late Miocene Columbia River Basalt flow (CRB) just east of well P-0093. Pliocene sediments onlap landward at a very low angle at the eastern end of the profile and are parallel to the unconformity elsewhere, indicating little relief on the unconformity at the time of erosion. After Yeats et al. (1998).

level. Currents would shift farther seaward as eustatic sea level fell, and deep-sea erosion may have intensified somewhat during glacial periods, as suggested on the Chilean margin (Thornburg and Kulm, 1987). Considering a feasible increase in velocities, it still seems unlikely that sediments of the lithology, cohesion, and age indicated here could be eroded.

We hypothesize that this unconformity was eroded subaerially where angular truncation and basalt erosion occur. Seaward of this region (Fig. 3), erosion may have been submarine, or the reflector may represent a conformable contact. Despite the conformable appearance of this reflector in the seaward part of the basin, we hypothesize that a hiatus is still present, as at Site 892 on the continental slope.

Driving Mechanism of Erosion: Tectonic Uplift and Eustatic Sea-Level Fall

Significant hiatuses or unconformities are the result of regional or global tectonic effects or eustatic sea-level change. If the Cascadia late Miocene unconformity was, in part, the result of subaerial erosion, as suggested here, tectonic uplift is necessary, but sea level may be a contributing factor. Worldwide late Miocene eustatic lowstands (Haq et al., 1987) occurred at 10.5 (major), 8 (minor), 6.5, and 5.5 Ma, two of which coincide with global hiatuses at 10–9 Ma and 7–6 Ma. Ingle (1978) also linked the Pacific basin 7–6 Ma hiatus to a major global climatic event.

Potential causes of rapid vertical motions at an active margin include subduction erosion (ultimately leading to subsidence), underplating of subducted sediments (leading to uplift), changes in sedimentation rate, or a change in the direction, dip, or convergence rate of the subducting plate. Subduction erosion might result from the subduction of a basement feature such as a seamount chain or ridge (e.g., von Huene and Lallemand, 1990), but would not produce a regionally synchronous unconformity as observed here. Marine magnetic anomaly maps (Geophysics of North America, CD-ROM, 1987) show that there is also little evidence of regional basement highs underlying the margin or its conjugate oceanic basement west of the Juan de Fuca Ridge in appropriate positions for late Miocene subsidence. Underplating may result from the decollement stepping down landward due to changes in the physical properties of the subducted sediments.

The older hiatus (ca. 10.5 Ma) is closely linked with Pacific plate motion changes, observed in a change in orientation of the Hawaiian island-seamount chain, and rotation of the convergence vector between the Juan de Fuca and North American plates to more normal, leading to increased thrusting and uplift along this margin (McCrory, 1995). Wilson et al. (1984) identified a 10° clockwise rotation of the Juan de Fuca plate motion (relative to North America) at 8.5 Ma, immediately preceding NH6, a possible result of separation of the Gorda plate from the Juan de Fuca plate. A large clockwise rotation of Juan de Fuca relative motion also occurred slightly later, at 5.89 Ma (chron 3A, Wilson, 1993) and may be connected to the separation of the Explorer plate from the Juan de Fuca plate. During this period, changes in Pacific plate motion are also recorded, which may have driven rotation of Juan de Fuca plate motion. Motion of the Pacific plate relative to North America in the northeast Pacific became more northerly (a clockwise rotation) ca. 8 Ma or chron 4 (Atwater and Stock, 1998). Barron (1986) suggested evidence of a tectonic event linked to Pacific plate motion change at 6 Ma (Jackson et al., 1975).

The NH6 event has also been loosely linked to other more recent major tectonic events in the Pacific basin ca. 5 Ma, including the opening of the Gulf of California and the inception of the modern San Andreas system (Normark et al., 1987; Barron, 1986), although this may be a separate younger event. Riddihough (1984) suggested that a more significant clockwise rotation in Juan de Fuca plate motion occurred between 4 and 3 Ma rather than ca. 5 Ma, which coincides with the change in Pacific plate motion at 3 Ma noted by Wessel and Kroenke (1997), a change probably driven by the collision of the Ontong Java plateau. The initial significant phase of this last event may have begun in the latest Miocene (Kroenke et al., 1986). These younger events may be unrelated to late Miocene uplift on the Cascadia margin, particularly in light of the absence of resolvable Pliocene plate motion changes in recent Pacific–North America plate reconstructions (Atwater and Stock, 1998).

These apparent correlations suggest that changes in Pacific and Juan de Fuca plate motion around 8–6 Ma may have caused rapid uplift and erosion on the Cascadia margin, with contemporaneous eustatic sea-level fall.

Deformation of the Late Miocene Unconformity

Major Neogene deformational patterns of the central and northern Oregon margin can be determined from the structure contour map of the late Miocene unconformity (Fig. 6) and relative uplift rates compared with those onshore.

Preserved Depositional Centers

The most prominent syncline preserving basinal sediments within the central Oregon shelf basin is the Newport syncline (20 km west of Newport) and its extension to the northwest, here named the Netarts syncline, which underlies the present-day continental slope (Fig. 6). Both synclines are coincident with a prominent gravity low (Fleming and Tréhu, 1999). The late Miocene unconformity reaches its greatest depth of ~2500 m below sea level within the Newport syncline. A shallower syncline to the southwest also has a northwesterly trend (Fig. 6). These two elongate synclines are separated by the rapidly uplifted Stonewall Bank, which is underlain by an active northwest-trending blind reverse fault (Yeats et al., 1998). Middle Miocene and older rocks are exposed within the core of Stonewall Bank anticline. Prior to the predominantly late Pliocene (2–3 Ma) and younger uplift of Stonewall Bank (Yeats et al., 1998), strata within the Newport syncline and the syncline west of the bank were probably connected. Shallower synclines are also evident at the northern and southern ends of the gridded data set, underlying the currently uplifted Nehalem Bank (part of the Astoria basin) and between Heceta and Siltcoos Bank (underlying the prominent geomorphic embayment, which we refer to as Siltcoos Embayment, Fig. 1).

We hypothesize that the effect of sedimentary loading on the surface would only accentuate topography induced by tectonic deformation. If loading were a significant factor we would expect greatest effects close to the Columbia River mouth (present-day and ancestral), whereas the deepest syncline (Newport) is located well to the south (~150 km).

Late Neogene Structural Trends

The general trend of structures deforming the late Miocene unconformity is between northwest and north-northwest on the middle to outer shelf and upper slope (Fig. 6). This trend is in agreement with latest Quaternary structural trends in the same region (Goldfinger et al., 1992, 1997; McCaffrey and Goldfinger, 1995). These structures reflect some degree of plate coupling beneath the shelf and upper slope. West of the outer arc high, a seaward transition from convergence-normal (northwest-southeast, domain B) to arc-parallel (north-south, domain A) fold trends (Fig. 7, thin dashed line in Fig. 4) coincides with a topographic break. We interpret this to be a structural boundary or backstop, representing a change in principal stress direction between domains B and A (Fig. 7). Goldfinger et al. (1992, 1996b) suggested that this backstop may also be related to the updip limit of the seismogenic plate boundary.

Recent investigations of active structures deforming the innermost shelf and coastal region (domain C, Fig. 7) suggest that east-west fold axes may be dominant (McNeill et al., 1998). This is in agreement with a north-south regional maximum horizontal compressive stress for the onshore northwestern United States (e.g., Zoback, 1992). Examples of this change can be seen in deformation of the inner shelf unconformity, such as the Nehalem Bank fault (McNeill et al., 1998) and the probable southeastern extension of the Stonewall Bank anticline (Fig. 6), with a change of trend from northwest-southeast to ~east-west as they approach the inner shelf and coastline. Yeats et al. (1998) suggested that the Stonewall anticline (and blind reverse fault) plunges to the southeast. However, the abrupt southern termination of the Newport syncline suggests a structural origin, possibly related to the landward extension of the Stonewall Bank structure.

Tectonic Influence of the Siletzia Terrane

The seaward margin of the Siletzia terrane was modeled by Tréhu et al. (1994) and Fleming and Tréhu (1999) using magnetic anomaly and seismic reflection data, and positioned from well stratigraphy (Snavely et al., 1980; Snavely and Wells, 1996). The exact position of this boundary is only locally constrained and is probably known within ±5 km regionally (A. Tréhu, 1999, personal commun.). Its approximated position from these data is indicated by the purple line in Figure 6. Considering the abrupt westward change in basement lithology from the basalt of the Siletzia terrane to marine strata (Snavely et al., 1980; Snavely and Wells, 1996) underlying the continental shelf and upper slope of offshore Oregon, there is minimal regional structural change in deformation across this boundary, as interpreted from deformation of the late Miocene unconformity (Fig. 6). The Siletzia margin does not control the position of the outer arc high, with a possible exception at Heceta Bank (Fig. 6). The transition from seaward convergence-driven east-west compression to landward north-south regional compression does not coincide with the Siletzia margin, as suggested by Fleming (1996), but occurs fairly consistently on the inner to middle shelf between central Oregon and Washington (Fig. 7; McNeill et al., 1998). With the possible exceptions of Heceta Bank and Stonewall Bank, deformation above the Siletzia backstop does not appear to affect or control uplift of the submarine banks. It appears that the Siletzia margin does not represent as significant a structural backstop as has previously been hypothesized. However, the wavelength of compressional folding changes over the seaward margin of Siletzia, with shorter wavelength folds over the margin relative to fold wavelength on Heceta Bank to the east and the accretionary wedge to the west (Tréhu et al., 1994, 1995; Yeats et al., 1998; C. Hutto, 1998, personal commun.).

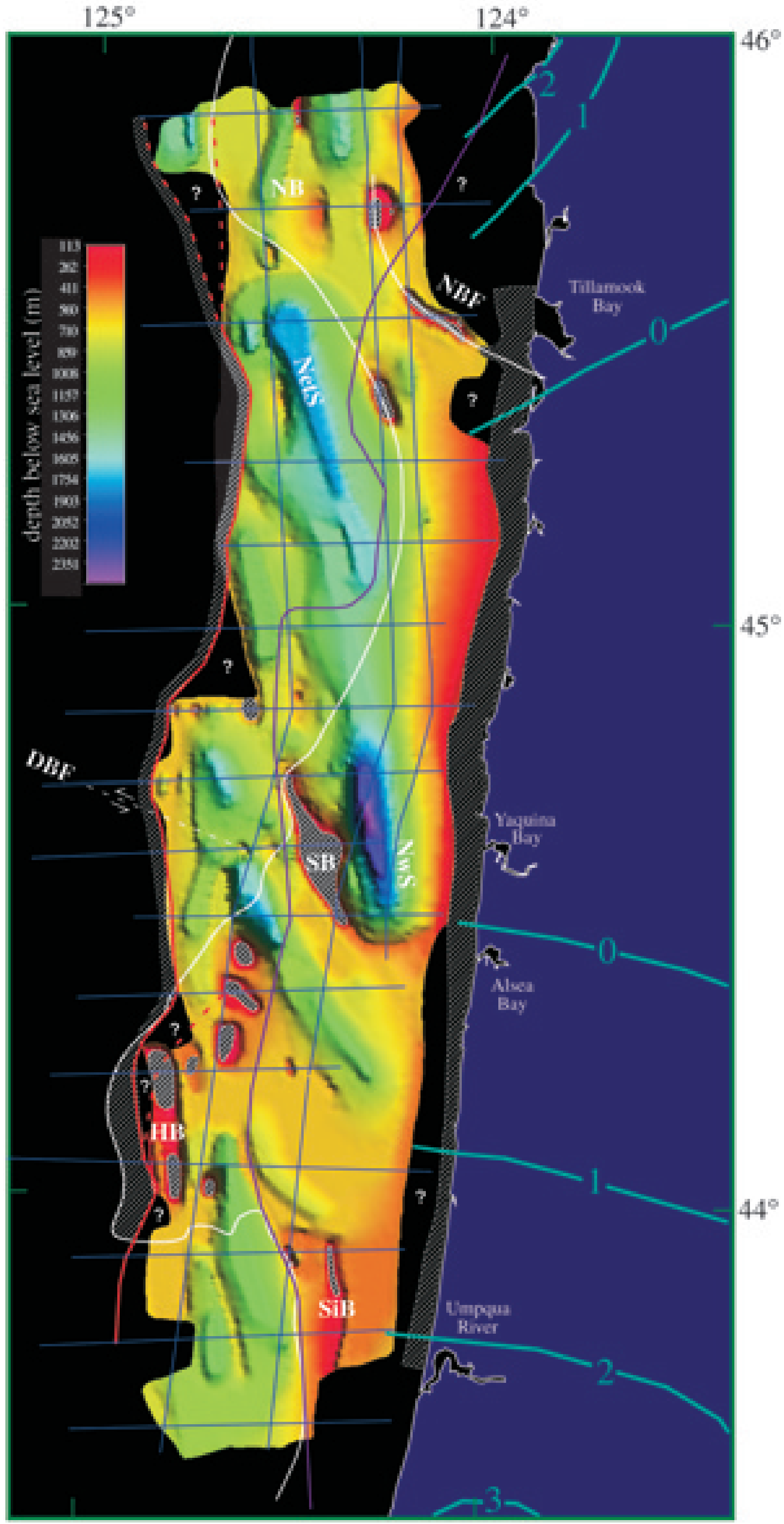

Figure 6. Shaded relief structure contour map of the late Miocene unconformity on the central Cascadia shelf and upper slope. Depth in meters below sea level. Major structural and topographic features are labeled: HB—Heceta Bank; SiB—Siltcoos Bank; SB—Stonewall Bank; DBF—Daisy Bank fault; NBF—Nehalem Bank fault; NB—Nehalem Bank; NwS—Newport syncline; NetS—Netarts syncline. Black areas surrounding the structure contour map represent regions where the position of the unconformity are uncertain, except the region west of the outer arc high, where the unconformity was either not present or has been eroded. The positions of seismic profiles used to produce the structure contour map are shown. Seaward margin of the Siletzia terrane (purple line), outer arc high (red line), and geodetic uplift contours from Mitchell et al. (1994, light blue lines) are also included. White line represents the shelf break.

Onshore and Offshore Margin-Parallel Deformation Rates

Of particular interest in Cascadia are possible connections between short- and long-term uplift rates and potential implications for the extent and location of interplate coupling on the subduction interface. Geodetic uplift rates, determined from repeated highway releveling in the past 70 yr (Mitchell et al., 1994), suggest that coupling may be variable along strike in central Oregon. While these data show that most coastal locations in Cascadia are rising, and tilting landward, the central Oregon Coast Range from about 44.5N to 45.5N appears to be doing neither (Fig. 6). The rates determined from geodetic work are high and suggest long-wavelength deformation, such that most investigators attribute them to the elastic response of the upper plate to interplate coupling.

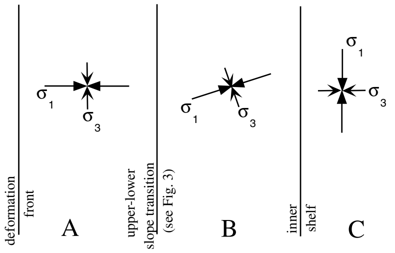

Figure 7. The change in principal stress orientations across the Cascadia margin as determined from fold trends measured from seismic reflection, bathymetry, and sidescan sonar data. Domain A (between the deformation front and upper-lower slope transition) is characterized by east-west compression or margin-parallel fold axes; domain B (between the mid slope and inner shelf) is characterized by northeasterly compression; and domain C (landward of the mid shelf) is characterized by north-south compression.

The overall long-term pattern of deformation of the central Oregon upper slope and shelf can be described as two major uplifts to the north and south separated by a broad structural downwarp. This pattern is evident in both modern bathymetry (Fig. 3) and deformation of the late Miocene unconformity in the form of the Newport and Netarts synclines (Fig. 6). A similar pattern is suggested by short-term geodetic uplift rates (Mitchell et al., 1994; Fig. 6) where 0–1 mm/yr uplift rates are observed on the central Oregon coast. There appears to be a correlation between long-term and short-term deformation rates on the Oregon margin, suggesting that the geodetic signature may not be entirely elastic and that not all strain is recovered during the earthquake cycle. Alternatively, this correlation may be purely coincidental.

Onshore, variations in long-term uplift over several time scales are recorded by Pleistocene marine terrace elevations (West, 1986; West and McCrumb, 1988; Kelsey et al., 1994), incision rates of Coast Range streams (Personius, 1995), and Coast Range topography (Kelsey et al., 1994; Personius, 1995). Correlations have also been observed by Kelsey et al. (1994) between Coast Range topography, marine terrace elevations, and geodetic uplift over the entire margin length, but the correlation of these data sets with offshore deformation patterns determined here is currently unclear. The implications for local versus regional strain contribution to these deformation patterns and degree of plate coupling may be revealed by future results from ongoing global positioning system (GPS) measurements (e.g., Goldfinger et al., 1998).

DISCUSSION

Evolution of the Neogene Cascadia Forearc

The geometry and stratigraphy of the Cascadia forearc are due to the interplay between sea-level change, sedimentation, and tectonics. We attempt to determine how the shelf forearc basin has developed through time by assessing these factors in a regional and global context.

Preserved sediments of the central Cascadia Neogene-Quaternary forearc basin are bounded to the east by the Coast Range and to the west by a discrete outer arc high, which is parallel to the deformation front on the shelf and upper slope (Fig. 3). Locally the outer arc high is a broad and complex anticlinal structure. The position of the landward edge of the outer arc high, i.e., the seaward limit of basinal sedimentation, or the most prominent structural and/or topographic high, is shown in Figure 3. At Heceta Bank, two prominent highs have been identified, although the seaward of these two structures is preferred as the main outer arc high due to the presence of basinal sediments between them (Fig. 8A). This current outer arc high is a structural high (Fig. 8, A–C) the formation of which caused basinal sediments to pond behind it. Growth strata suggest that uplift of this outer arc high began in the early-mid Pliocene on much of the central-northern Oregon margin (Fig. 8, B and C), but possibly slightly later on Heceta Bank to the south (Fig. 8A). Stratigraphic ages in the extreme seaward portion of the basin are uncertain; therefore, the exact timing of uplift cannot be determined. However, we are fairly certain that this outer arc high postdates the late Miocene unconformity. Prior to uplift on this structure, the forearc basin extended farther west (Fig. 9, A–C), as indicated by pre-Pliocene sediments thickening slightly toward the outer arc high (Fig. 8), and was bounded by an older outer arc high. This pre-Pliocene outer arc high is not recognized in any seismic lines. The western edge of the older basin was truncated, and eroded sediments were probably accreted or subducted. Truncation and erosion or slumping at the edge of the shelf basin is supported by the incorporation of Miocene (and possibly Eocene) bathyal sediments into the second accretionary ridge seaward of the basin (now at 700 m water depth, ODP Site 892; Fourtanier and Caulet, 1995; Fourtanier, 1995; Caulet, 1995; Zellers, 1995). We therefore hypothesize that the most recent outer arc high (Fig. 3) represents a truncational structural boundary against which the mid-Pliocene to Pleistocene accretionary wedge was built (Fig. 9D). This truncation boundary may be represented by a region of complex, probably faulted, stratigraphy just west of the outer arc high (e.g., Fig. 8C).

The accretionary wedge is too narrow to incorporate late Miocene to recent accreted sediments based on recent sedimentation and accretion rates (Westbrook, 1994), suggesting that either sedimentation and accretion rates have increased dramatically during the Quaternary or that a significant erosional event, possibly a result of subduction erosion, removed much of the late Miocene and early Pliocene wedge. The same erosional event may have been responsible for erosion of the seaward edge of the forearc basin. If the shelf forearc basin originally extended farther seaward, the accretionary wedge was either west of its current position or narrower. The former is contrary to the general assumption that the margin and wedge have gradually built westward with time, whereas the latter may be a reflection of relative sediment budgets of the basin and the accretionary wedge, with alternating growth of the wedge and filling of the forearc basin, controlled by growth of the outer arc high and fluctuating sea level. Basin filling and growth of the outer arc high in conjunction with sea-level change and sediment input therefore control the rate of growth of the accretionary wedge.

To the east, the Willamette Valley basin was formerly partially connected to the offshore basin prior to the major late Miocene uplift of the Oregon Coast Range. Following uplift of the Coast Range (underway at the time of Columbia River Basalt flows to the coast ca. 16.5–15 Ma with major uplift beginning after 15 Ma; Yeats et al., 1996), the two basins were separated (Fig. 9, B and C). The formation of such a double forearc basin–outer arc high is unusual on continental margins, but is also present on the northeast Alaskan margin, where seaward (Sanak, Shumagin, Tugidak, Albatross, Stevenson) and landward (Cook Inlet and Shelikof Strait) basins are separated by the uplifted Kodiak Island and Kenai Peninsula (von Huene et al., 1987).

Submarine Bank Morphology and Origin

The positions of uplifted submarine banks and intervening embayments control the morphology of the current Oregon shelf break, which postdates the Pliocene formation of the shelf basin outer arc high. Miocene rocks below the unconformity are exposed at these banks, which include Nehalem, Stonewall, Heceta, and Siltcoos Banks (Fig. 6), and Coquille Bank to the south.

The most plausible mechanisms for localized submarine bank uplift appear to be (1) uplift above a subducted or accreted basement feature or (2) uplift above an active reverse fault or fault zone. The latter is the preferred origin of the active Stonewall Bank (Yeats et al., 1998) and probably drives the uplift of Siltcoos Bank, west of the Umpqua River (Fig. 1), which has a trend

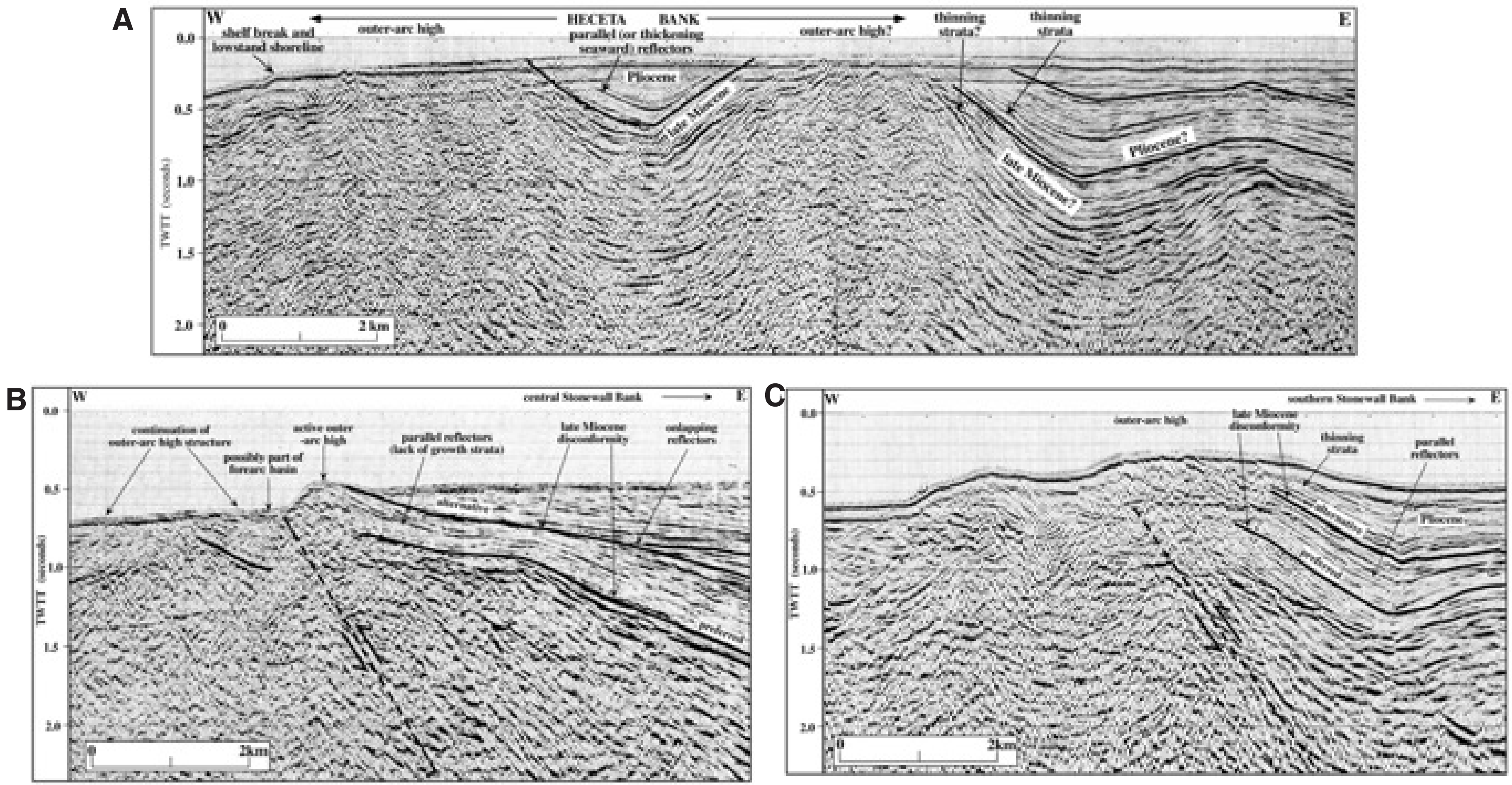

Figure 8. (A) East-west multichannel seismic reflection profile across Heceta Bank showing the outer arc high of the Neogene forearc basin (located in Fig. 1). On the continental shelf, the outer arc high has been truncated during Pleistocene lowstands, and lacks topographic relief in contrast to the upper slope (see B and C). The outer arc high is composed of two parallel anticlines (see Fig. 3), the outermost of which we interpret as the outer arc high due to a small basin between them. On this profile, there is little evidence of growth strata in early (-mid?) Pliocene sediments in the smaller basin, suggesting uplift of the seaward major outer arc high some time in the mid Pliocene. Thinning of strata of the main basin suggests that major growth of the eastern outer arc high followed formation of the late Miocene unconformity with minor growth in the late Miocene (pre-Pliocene strata show some thinning and erosional truncation). The exact age of sediments is unknown. From this profile and others across Heceta Bank, we interpret that major growth of the main outer arc high in this region began in the middle Pliocene.

(B and C) East-west multichannel seismic reflection profiles across the upper slope of the central Cascadia forearc (locations in Fig. 1). Both profiles cross the structural outer arc high on the upper slope west of Stonewall Bank where its relief has not been truncated during late Pleistocene lowstands. B crosses the outer arc high west of central Stonewall Bank and C crosses at the southern end of the bank. In both profiles, the position of the late Miocene disconformity (correlative to the landward angular unconformity) is uncertain due to poor age control and nonangularity. Preferred and alternative positions are given. Preferred positions are both overlain by parallel strata, suggesting that growth of this outer arc high did not begin until the mid-late Pliocene (indicated by seaward thinning in the upper part of the basin). Alternative positions are overlain by thinning strata suggesting that, alternatively, growth of the outer arc high began soon after the late Miocene erosional event, in the early Pliocene. In both B and C, the outer arc high is not a simple anticline but a broader structure composed of more than one anticline: this is common in other regions of the forearc basin. Possible positions of reverse faults are also shown, underlying the high, and associated with uplift of the underlying mélange (discontinuous reflectors).

similar to surrounding active structures. Stonewall Bank overlies the western edge of the basaltic Siletzia terrane, which may be related to its uplift mechanism. Fleming (1996) suggested that the Stonewall Bank blind reverse fault may be a reactivation of the late Eocene Fulmar fault of Snavely (1987), which truncates the seaward margin of the Siletzia terrane. However, Stonewall Bank anticline trends obliquely (northwest) to the north-south–trending edge of Siletzia (Fig. 6); therefore, the two faults are probably unrelated.

Heceta and Nehalem Banks are much larger features than Stonewall or Siltcoos Banks, with probable differing origins. The Siletzia margin and an accreted basement ridge (Fleming and Tréhu, 1999) underlie the eastern edge of Heceta Bank, and thus may contribute to the uplift of this bank (although these features are landward of the highest bank uplift rates from Kulm and Fowler, 1974). Heceta Bank is the only location where the uplifted basin outer arc high is located on the current continental shelf; to the north and south, the high is located on the present upper slope or close to the shelf break. A Siletzia backstop may contribute to pronounced uplift associated with the outer arc high at Heceta Bank. On the northern Oregon shelf, the outer arc high maintains an approximately N-S trend across the upper slope and shelf break, whereas the Siletzia boundary trends northeasterly and landward (Fig. 6; Fleming and Tréhu, in press).

Nehalem Bank is of comparable size to Heceta Bank. It is located farther seaward of Siletzia (Fig. 6), and uplift is less pronounced than it is at Heceta Bank (Kulm and Fowler, 1974). The shelf basin outer arc high is located just seaward of Nehalem Bank; the bank is underlain by as much as 1.5 km of post-Miocene basinal sediments. Miocene sediments are only exposed locally at the seafloor. The uplift mechanism of this bank remains poorly understood.

Indications of Paleo-Shelf Edge Positions

We hypothesize that the early Pleistocene shelf edge following near-complete filling of the forearc basin was located above the north-south–trending outer arc high. Folded Miocene and Pliocene units that were uplifted at the outer arc high within the Netarts Embayment (Fig. 8, B and C) are truncated at the seafloor, suggesting subaerial erosion. At the beginning of the Pleistocene, when the basin filled, and during significant sea-level lowstands, the entire basin surface was probably abraded. Subsequent to subsidence within geomorphic embayments (Netarts and Siltcoos) and formation of the current shelf edge, the outer arc high was abraded on the continental shelf during more recent Pleistocene lowstands (Fig. 8A), but not on the slope within shelf embayments, where

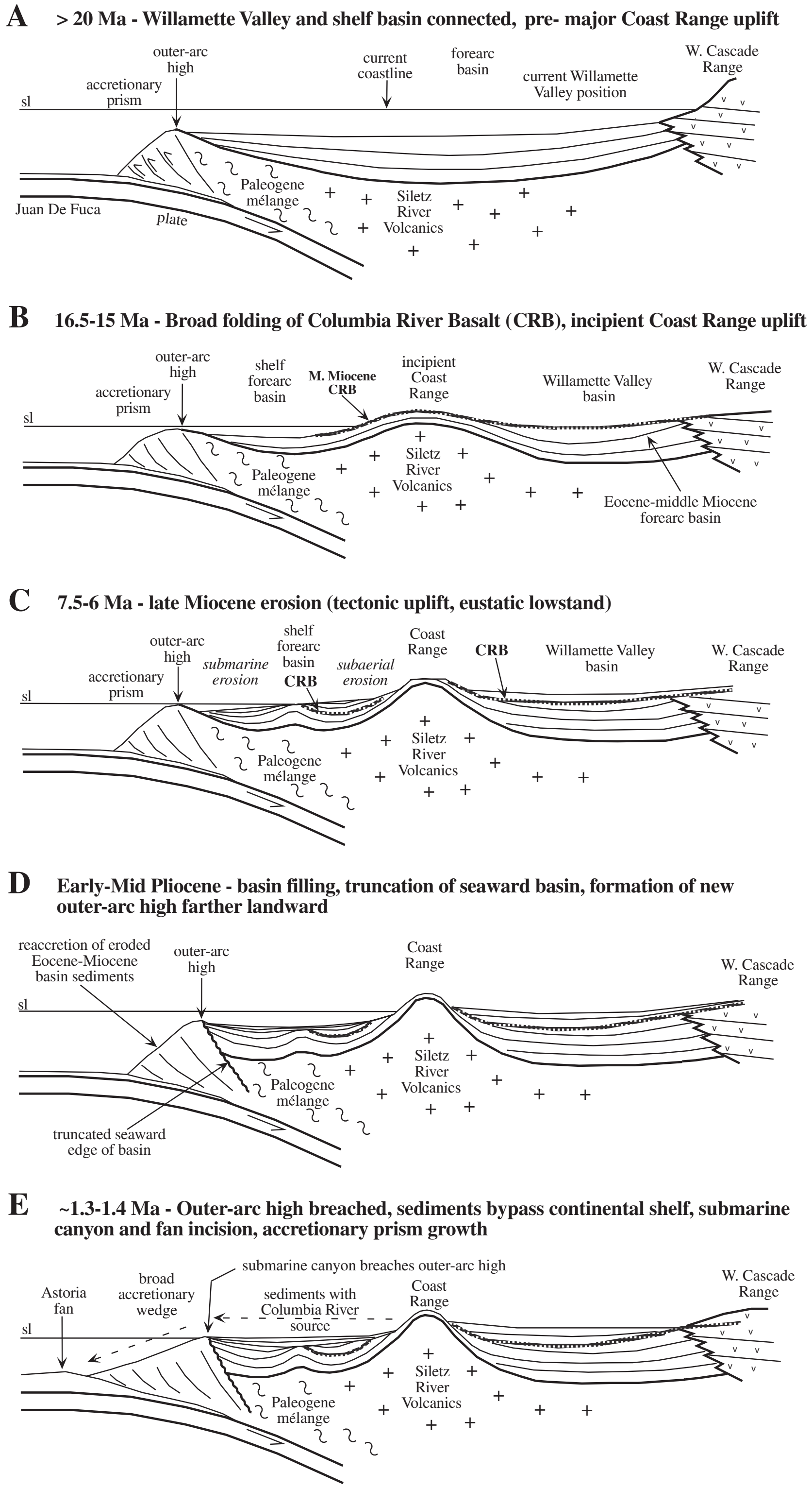

Figure 9. Neogene tectonic history of the central Cascadia forearc depicted by time slice cross sections of the central Oregon forearc from the Cascades to the deformation front. Stratigraphy and topographic features are generalized for the central Oregon forearc. (A) The Willamette Valley and shelf basins connected prior to major uplift of the Coast Ranges (with local highlands existing). (B) Columbia River Basalt emplacement and incipient Coast Range uplift. (C) Erosion of the late Miocene unconformity. (D) Pliocene basin filling and truncation of the seaward edge of the offshore basin. (E) Shelf basin filling and submarine fan formation in the Pleistocene.

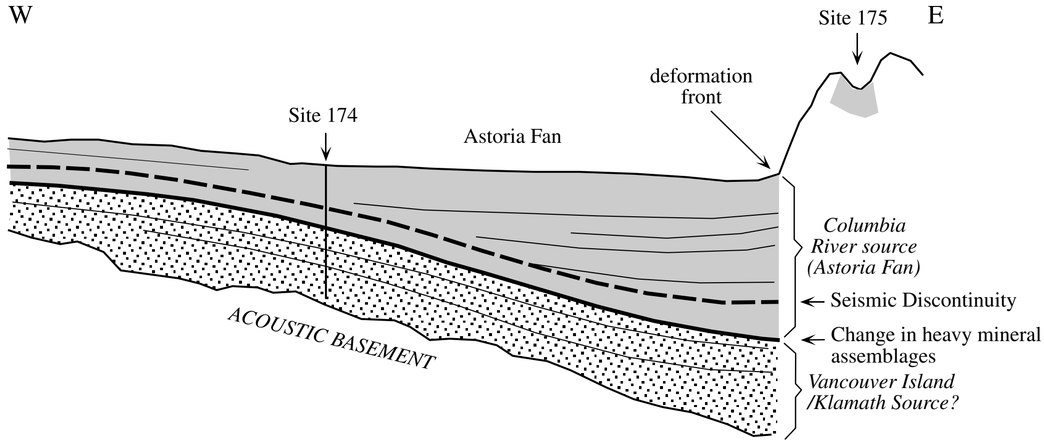

Figure 10. East-west seismic reflection profile across the Astoria fan and lower accretionary wedge at Deep Sea Drilling Project (DSDP) Site 174 (after von Huene and Kulm, 1973). A seismic discontinuity represents the base of the Astoria fan sands (0.76 Ma); however, this is underlain by a change in heavy mineral assemblages. This marks the transition from a younger Columbia River source to an older more distal source (possibly Vancouver Island to the north or Klamath Mountains to the south). We hypothesize that the onset of Columbia River source sedimentation coincides with initial breaching of the forearc basin outer arc high and inception of the Astoria submarine canyon. This transition is dated as about 1.3-1.4 Ma.

it still has topographic expression (Fig. 8, B and C). This topographic high is the outer edge of the Cascade Bench of Kulm and Fowler (1974) (Figs. 1 and 4). Topographic relief of the outer arc high (Fig. 8, B and C) must have therefore formed subsequent to early Pleistocene abrasion. A similar submarine bench exists north and south of Coquille Bank and on the southern Oregon upper slope (Klamath Plateau of Kulm and Fowler, 1974). We hypothesize that the seaward edge of these benches represents the southern continuation of the outer arc high, which may extend as far south as the Eel River basin (Fig. 1). The present shelf break was subsequently formed as a result of continued submarine bank uplift and subsidence within intervening synclinal embayments.

Timing of Pleistocene Shelf Basin Filling and Submarine fan Progradation

The Astoria Canyon, the primary conduit for Columbia River sediments, is the only major submarine canyon on the central Cascadia slope (Nelson et al., 1970; Carlson and Nelson, 1987), in contrast to the Washington slope, which is dissected by several prominent canyons (Barnard, 1978; McNeill et al., 1997). On the southern Oregon slope, the Rogue Canyon (Fig. 1) acts as a conduit for Rogue River sediments from the Klamath Mountains. The base of the Pleistocene Astoria submarine fan occurs at 264 m below the seafloor at Site 174A (Figs. 1, 2, and 10), and the age at that depth is 0.76 ± 0.5 Ma based on interpolation between dextral coiling events (Goldfinger et al., 1996a; Ingle, 1973). However, the base of the Astoria fan is time transgressive, such that oldest fan sediments have now been accreted.

At Site 174A, a change in heavy mineral assemblages occurs at 370 m below the seafloor (Figs. 2 and 10); that event was initially dated as ca. 2 Ma, representing the beginning of Columbia River sediment provenance (Scheidegger et al., 1973). Prior to this time, heavy minerals suggest a source from the north (British Columbia) or south (Klamath Mountains). The same change in sediment provenance is observed at Leg 146 sites (e.g., Site 892) on the Cascadia slope (Chamov and Murdmaa, 1995), and late Pleistocene sediments at Site 175 (Fig. 1) also have a Columbia River source (Fig. 10; Scheidegger et al., 1973; von Huene and Kulm, 1973). Scheidegger et al. (1973) suggested that the mineralogy change at Site 174 may have resulted from translation of the Juan de Fuca plate to the northeast toward the Columbia River source. However, translation in this short period of time (~1–2 m.y.) would only result in 50–100 km of northeasterly migration (Chamov and Murdmaa, 1995; Westbrook, 1994). Columbia River suspended sediments were transported far southward during the late Pleistocene, as indicated by the clay mineralogy of sediments in the Escanaba Trough, more than 500 km to the south (McManus et al., 1970; Fowler and Kulm, 1970). Vallier et al. (1973) also identified Pleistocene turbidite sands with Columbia River source in the Escanaba Trough, DSDP Site 35, with a transition from Columbia River to Klamath source observed downcore. In the same region, at ODP Site 1037 (Fouquet et al., 1998), Brunner et al. (1999) also identified latest Pleistocene Columbia River turbidites and correlated turbidite ages with jokuhlaups of glacial lake Missoula, Washington. Late Pleistocene to present-day heavy minerals of sand fractions from the Cascadia channel and southeastern Cascadia basin (Duncan and Kulm, 1970) also indicate Columbia River provenance. This suggests that a more southerly position of Site 174 during the early Pleistocene cannot account for a significant change in mineralogy. Diagenesis of older sediments was also proposed as a possible cause of the mineralogy change (Chamov and Murdmaa, 1995); however, Scheidegger et al. (1973) found little evidence of chemical alteration of the heavy mineral assemblages at Sites 174 and 175.

As an alternative hypothesis, we suggest that this change in mineralogy, ca.. 1.3–1.4 Ma (see following for details of age determination) coincides with the near-complete filling of the shelf basin in Oregon and breaching of the outer arc high. The resulting increased sedimentation rates may have resulted in a transition from a trench-confined fan system to a much broader fan and progradation out to Site 174, as modeled by Schweller and Kulm (1978) and described in the central and southern Chilean trench by Thornburg and Kulm (1987). This contrasts with margins where trench sedimentation rates are low and sediments are confined to the trench with dominant axial transport, e.g., the Middle America Trench (Underwood and Bachman, 1982; Underwood and Moore, 1995), or starved trenches such as north Chile (Thornburg and Kulm, 1987). These events may also coincide with the initial development of the current Astoria Canyon, as suggested by von Huene and Kulm (1973). This change may represent the onset of distal turbidite deposition or levee-overbank deposits at Site 174, followed by westward progradation of the Astoria fan to Site 174 by 0.76 Ma. The earlier, more landward submarine fan deposits have since been accreted and incorporated into the two thrust ridges closest to the deformation front (Kulm and Fowler, 1974). An alternative explanation for a transition in sediment provenance is sediment trapping behind an accretionary ridge (within a slope basin) or behind an oceanic basement feature (Underwood and Bachman, 1982). However, seismic reflection data reveal no suitable basement candidates.

At Site 176, on the seaward edge of the shelf basin (Fig. 1), Pleistocene to possible late Pliocene sediments indicate a Columbia River source (Scheidegger et al., 1973). The presence of Pleistocene sediments within the uppermost shelf basin at Site 176 (Fig. 2) indicates that sedimentation continued within the shelf basin following initiation of the early Pleistocene Astoria fan system. Prior to 1.4 Ma, sedimentation rates on the abyssal plain and slope should have been somewhat reduced, although they were also strongly controlled by fluctuating sea level.

Unfortunately the timing of the change in mineralogy and sediment source was not redated along with the Astoria fan base (Goldfinger et al., 1996a). However, its age can be more accurately determined using Site 174A stratigraphy and assumptions concerning sedimentation rates. An assumed constant sedimentation rate for the period between the Pliocene-Pleistocene boundary and initial Columbia River source sediments yields an age of ca. 1.3–1.4 Ma for the mineralogy change (Fig. 2). This is a maximum age for this transition, because sedimentation rates at Site 174 may have increased through the Pleistocene. We believe that the delay between the onset of Pleistocene glaciation (ca. 2 Ma) and the change in mineralogy provenance at Site 174 supports a cause of this change in addition to increased sedimentation rates and direct sedimentation at the head of the canyon related to a eustatic lowstand. However, Nelson (1976) suggested that the older fan deposits may be located northwest of Site 174, considering the leftward migration of channels during fan history, which may be an alternative explanation for this delay.

The early Pleistocene unconformity, identified on the Washington, Oregon, and northern California shelf (Fig. 2) may be coincident with this mineralogy change and hypothesized breaching of the outer arc high. This unconformity is probably at least 0.92 Ma at Site 176 on the northern Oregon shelf (Fig. 2, Kulm et al., 1973), earliest Pleistocene on the Washington shelf (Palmer and Lingley, 1989), and ca. 1 Ma marking the top of the Rio Dell Formation in Eel River basin stratigraphy off northern California. These tentative correlations suggest regional control such as regional tectonic uplift and/or eustatic sea-level fall. A slightly younger angular unconformity (0.7–0.6 Ma), clearly identifiable in stratigraphy onshore (McCrory, 1995) and in seismic records offshore (Clarke, 1992), marks the end of marine sedimentation within the onshore part of the Eel River basin (the upper boundary of the Carlotta and Scotia Bluffs Formations, McCrory, 1995), coincident with a significant tectonic event throughout the Pacific basin (Ingle, 1986).

Prior to the Pleistocene, we suggest that fluctuations in basin sedimentation and slope basin–abyssal plain sedimentation were controlled by eustatic change and ponding behind the forearc basin outer arc high. Absence of preserved pre-Pleistocene fan stratigraphy prevents the determination of details of this forearc history.

CONCLUSIONS

The late Cenozoic central Cascadia forearc basin provides evidence for a complex history of tectonics and sedimentation. Prior to initial uplift of the Oregon Coast Range between 16.5 and 15 Ma, when eruptions of the Columbia River Basalt Group flowed to the coast through broad valleys (Beeson et al., 1989), a wide forearc basin extended from the Cascade foothills to the offshore outer arc high and accretionary wedge, with local highlands in the Oregon Coast Range. More extensive uplift of the Oregon Coast Range after Columbia River Basalt eruption subsequently separated the Willamette Valley from the shelf basin (Fig. 9, A–C).

Large volumes of sediments accumulated within the offshore forearc basin during sea-level highstands and ponded behind a former, now-eroded, outer arc high. Tectonic uplift and possible eustatic sea-level fall resulted in formation of a regional unconformity ca. 7.5–6 Ma (Fig. 9C), recognized as a worldwide hiatus (NH6 of Keller and Barron, 1987). The unconformity is angular close to shore, but is disconformable or possibly conformable in the seaward part of the basin. The same hiatus is observed in the Eel River basin of the southern Cascadia margin, but without a widespread angular unconformity. This hiatus may be associated with tectonic uplift resulting from clockwise rotations of Pacific and Juan de Fuca (relative to North America) plate motions around 8–6 Ma.

Truncation and erosion of the seaward forearc basin and former outer arc high, probably in the early to mid Pliocene, reduced the width of the basin. Basinal sediments were reaccreted and incorporated into the accretionary wedge. The more landward present-day outer arc high, which bounds the seaward edge of the forearc basin, was uplifted at this time (Fig. 9D). The truncation event may have been a result of subduction erosion due to basement anomalies on the Juan de Fuca oceanic basement or oversteepening and failure of the margin. A comparable example of massive slope failure in the Pleistocene is described by Goldfinger et al. (2000) on the southern Oregon margin.

During Pliocene basin subsidence, the majority of terrigenous sediments from the Columbia River and other minor fluvial Coast Range sources ponded behind the outer arc high and failed to reach the abyssal plain during eustatic highstands. The accretionary wedge at this time was probably narrower with lower accretion rates. Eustatic lowstands led to canyon incision, fan formation, and slope basin–abyssal plain sedimentation. The initially planar late Miocene unconformity was deformed by localized subsidence (e.g., Newport syncline) and submarine bank uplift (e.g., Stonewall and Heceta Banks) throughout the Pliocene and Quaternary.

At about 1.3–1.4 Ma (extrapolated from Site 174A Astoria fan stratigraphy), the outer arc high was breached following near-complete basin filling, and sediments bypassed the shelf and accumulated in slope basins and on the abyssal plain. This may have coincided with or accentuated incision of the Astoria submarine canyon and progradation of the Astoria fan system, with resulting rapid growth of the Pleistocene accretionary wedge off Oregon (Fig. 9E). This event may also be represented by an early Pleistocene unconformity throughout much of the Cascadia forearc basin.

Comparisons between deformation of the unconformity and onshore geodetic uplift rates indicate similarities. The validity of these correlations and their driving mechanisms may be resolved further by future GPS results.

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