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A kinematic model for the formation of the Siletz-Crescent forearc terrane by capture of coherent fragments of the Farallon and Resurrection plates

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[1] The volcanic basement of the Oregon and Washington Coast ranges has been proposed to represent a pair of tracks of the Yellowstone hotspot formed at a mid-ocean ridge during the early Cenozoic. This interpretation has been questioned on many grounds, especially that the range of ages does not match the offshore spreading rates and that the presence of continental coarse clastic sediments is difficult to reconcile with fast convergence rates between the oceanic plates and North America. Updates to basement geochronology and plate motion history reveal that these objections are much less serious than when they were first raised. Forward plate kinematic modeling reveals that predicted basement ages can be consistent with the observed range of about 55–49 Ma, and that the entire basement terrane can form within about 300 km of continental sources for clastic sediments. This kinematic model indicates that there is no firm reason to reject the near-ridge hotspot hypothesis on the basis of plate motions. A novel element of the model is the Resurrection plate, previously proposed to exist between the Farallon and Kula plates. By including the defunct Resurrection plate in our reconstruction, we are able to model the Farallon hotspot track as docking against the Oregon subduction margin starting about 53 Ma, followed by docking of the Resurrection track to the north starting about 48 Ma. Accretion of the Farallon plate fragment and partial subduction of the Resurrection fragment complicates the three-dimensional structure of the modern Cascadia forearc. We interpret the so-called "E" layer beneath Vancouver Island to be part of the Resurrection fragment. Our new kinematic model of mobile terranes within the Paleogene North American plate boundary allows reinterpretation of the three-dimensional structure of the Cascadia forearc and its relationship to ongoing seismotectonic processes.

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1. Introduction

[2] Details of the formation of the mostly Eocene volcanic basement terrane in the Coast Ranges of Oregon and Washington, and southernmost Vancouver Island have remained enigmatic. Early plate tectonic models noting older ages for the basaltic rocks and foraminifera microfossils at the northern and southern ends of the terrane and younger ages at the center called for an interpretation of a captured fossil spreading ridge between the Kula and Farallon plates. *Duncan* [1982] suggested that many geologic aspects of the volcanic terrane are well explained if the terrane represents captured tracks of the Yellowstone hotspot on the Kula and Farallon plates. Later work showed two problems with these simple plate models [*Wells et al.*, 1984]. First, the difference between older and younger ages was larger than predicted by the relatively fast Kula-Farallon spreading rate. Second, the presence of proximal continental sediments interbedded with some of the older basalts conflicted with the large amounts of convergence predicted by the generally used kinematic models for motion of the Kula and Farallon plates [e.g., *Byrne*, 1979].

[3] Recent plate reconstructions [*Haeussler et al.*, 2003] suggested the formation of a new oceanic plate no later than the Paleocene, derived from the Kula plate, with plate motions more consistent with available geologic constraints. Although plate reconstructions may preclude Kula plate involvement during accretion of the Eocene terrane, Duncan's suggestion of an early record of Yellowstone hotspot activity remains viable. Whether the observed Miocene and younger volcanic progression of the Yellowstone hotspot represents a deep mantle plume, and if so, how long the plume has been active continues to be debated [e.g., *Fouch*, 2012]. If the terrane's geologic record provides good evidence that a mantle plume

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was responsible for near-ridge volcanism in the Eocene and if its track is consistent with the plume responsible for continental volcanism in the Miocene and younger, then interpretations that the plume first reached the shallow mantle at about 17 Ma [e.g., *Pierce and Morgan*, 2009] become less credible.

[4] Our goal in this paper is first to review updates to the geologic history of the Eocene volcanic terrane, especially its geochronology, and updates to the plate-motion record. Second, we use forward plate-kinematic modeling to repeat the test of whether geologic relations in the terrane are consistent with their formation as a near-ridge hotspot track. Because so much of the relevant oceanic plate record has been subducted, the details of this modeling are profoundly nonunique. Nonetheless, we find that straightforward plate models with a Paleogene Yellowstone hotspot near a spreading ridge axis are able to explain relations that previously appeared to be in conflict. Our model is one of many that might explain geologic relations in the northeastern Pacific in a plate tectonic framework. We do not evaluate possible alternatives, yet hope this result inspires further work in field relations, geochemistry, and geodynamics to investigate the possibility for a long-lived hotspot and further evidence of the formation and capture of oceanic microplates along convergent margins.

2. Geologic Setting

2.1. Siletz, Crescent, and Metchosin Volcanic Basement

[5] We primarily assess the volcanic basement terrane of the Coast Ranges of Oregon and Washington (Figure 1), namely thick accumulations of mostly lower Eocene submarine and subaerial oceanic basalt [e.g., *Snavely et al*., 1968] situated in the forearc region of the Cascadia subduction boundary. Stratigraphic nomenclature of the units that constitute this terrane has been evolving, with the term Siletz River Volcanics now generally applied to the oldest lavas (56–49 Ma, Table 1) throughout the Oregon Coast Range [*Wells et al*., 1994, 2000] and Crescent Formation often used for the oldest basalts (51–46 Ma) of western Washington [*Duncan*, 1982; *Moothart*, 1992; *Clark*, 1989; *Haeussler and Clark*, 2000; *Hirsch and Babcock*, 2009]. This more recent usage places the formerly separate Roseburg Formation of southwest Oregon [*Baldwin*, 1974; *Wells et al*., 2000] and the stratigraphically lowest basalts in the Tillamook area [*Wells et al*., 1994] as part of the Siletz River Volcanics. Many units that were either previously unnamed or grouped as the lower subset of the Grays River volcanics of *Duncan* [1982] in western Washington are now considered part of the Crescent Formation [*Wells and Coe*, 1985; *Moothart*, 1992; *Haeussler and Clark*, 2000]. The Metchosin Igneous Complex of *Massey* [1986] of southern Vancouver Island represents the northernmost outcrop of the Crescent volcanic suite. The Siletz River Volcanics, Crescent Formation, and Metchosin Igneous Complex are sometimes combined into the simpler informal term, Siletz terrane [e.g., *Snavely et al*., 1968; *Fleming and Tréhu*, 1999] although other terms including Siletzia [*Irving*, 1979], and Coast Range Volcanic Province [*Babcock et al*., 1992] have been used equivalently. We prefer the term Siletz-Crescent terrane when discussing the composite of all three units to emphasize that the Siletz River Volcanics and Crescent-Metchosin volcanic suite likely originated on different oceanic plates.

[6] Understanding of local structure and stratigraphy within the forearc basement units is often limited by poor exposure. One of the most complete stratigraphic sections is exposed along the Dosewallips River, eastern Olympic Peninsula (Figure 1), where *Babcock et al*. [1992] described an 8.4 km thick sequence of Crescent submarine basalt flows overlain by 7.8 km of subaerial basalt flows (see Table S1 in the supporting information) that together have been rotated to a near-vertical orientation. These substantial thicknesses are interpreted to be original because only minor faulting has been observed at this site; major faulting, which could have dramatically increased structural thickness, is not observed. On Vancouver Island, an incomplete sequence of Metchosin Igneous Complex rocks resembles the upper part of an ophiolite section with more than 3 km of gabbro and sheeted dikes overlain by submarine basalt flows, in turn overlain by subaerial basalt flows [*Massey*, 1986]. *Haeussler and Clark* [2000] observed a similar sequence near Bremerton, Washington. Farther south, exposed Crescent and Siletz River basalts are generally of submarine origin, in places overlain by subaerial basalts [*Snavely et al*., 1968]. For example, near Roseburg, Oregon, a sequence of Siletz River Volcanics is composed predominately of 4–6 km submarine flows capped by subaerial flows [*Wells et al*., 2000].

[7] Geophysical surveys provide more complete estimates of the thickness of the Siletz-Crescent terrane. The terrane is thickest in central western Oregon where it reaches 27 ± 5 km (Table S1) [*Tréhu et al*., 1994; *Graindorge et al*., 2003], much thicker than typical oceanic crust of 6–7 km. The terrane thins somewhat to the north where it is approximately 20 km thick beneath northwestern Oregon and southwestern Washington [*Fleming and Tréhu*, 1999; *Parsons et al*., 1999]. From northernmost Washington to southwestern Vancouver Island, the terrane is approximately 10 km thick [*Fleming and Tréhu*, 1999], only slightly thicker than normal oceanic crust. Offshore western Vancouver Island, the terrane is about 6 km thick [*Hyndman et al*., 1990].

[8] The western boundary of the Siletz-Crescent terrane, where exposed in the Olympic Peninsula, is a thrust contact with basalt of the Eocene Crescent Formation thrust over Eocene-Oligocene accretionary complex rocks of the Olympic Mountains basement terrane along the Hurricane Ridge fault [*Cady*, 1975; *Dragovich et al*., 2002; *Hirsch and Babcock*, 2009]. Magnetic anomalies [*Finn*, 1999; *Blakely et al*., 2005] permit confident tracing of this contact northward to offshore Vancouver Island and southward to offshore Oregon where it appears to be truncated by the high angle Fulmar fault [*Snavely et al*., 1980; *Fleming and Tréhu*, 1999].

[9] The eastern boundary of the Siletz-Crescent terrane, where exposed on Vancouver Island, is a thrust contact with Paleozoic and Mesozoic continental terranes thrust over Metchosin Igneous Complex rocks along the Leech River fault [*Massey*, 1986]. The eastern boundary is also a thrust contact where exposed near Roseburg, OR, with Mesozoic accretionary complex rocks thrust northwestward over Siletz River Volcanics along the Wildlife Safari fault [*Wells et al*., 2000]. Elsewhere, the eastern limit of the forearc terrane is buried under younger volcanic and sedimentary units.

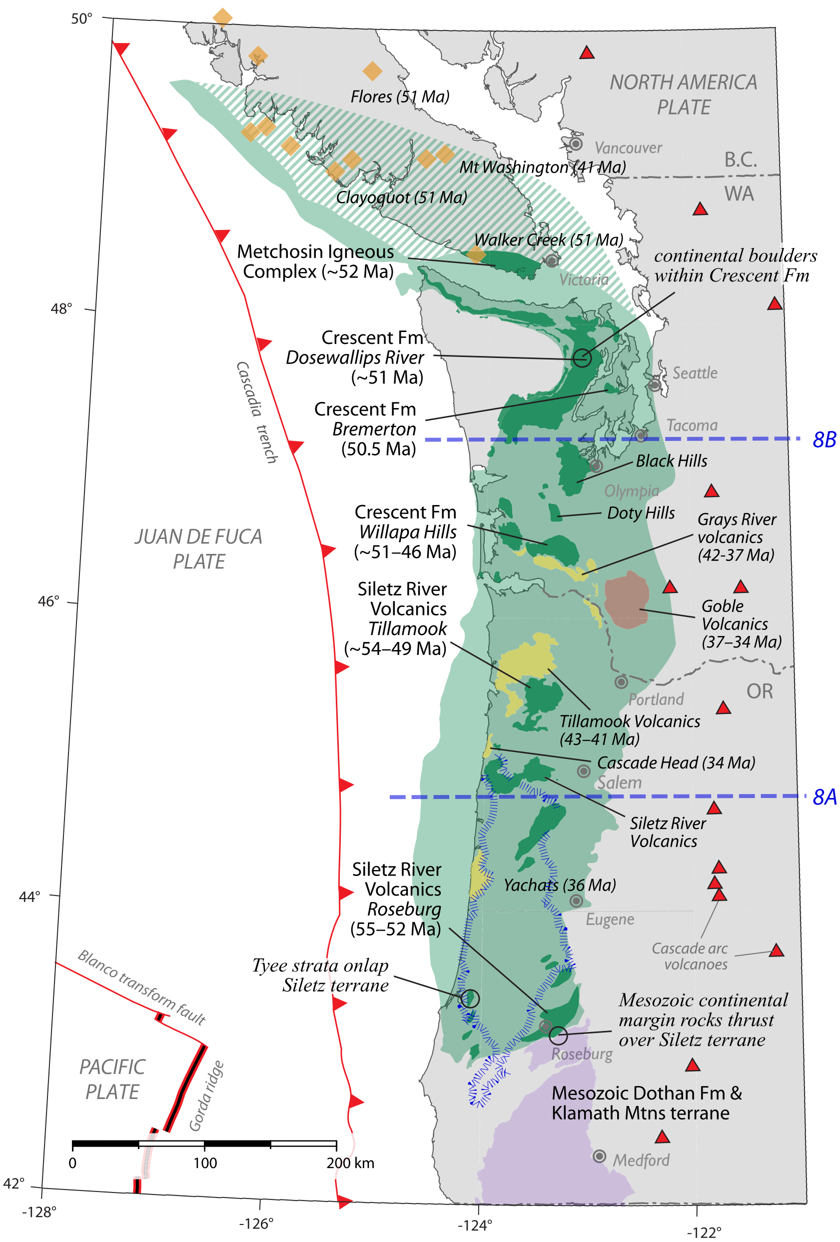


Figure 1. Map showing known distribution of Siletz and Crescent basaltic rocks (also known as the Coast Range Basalt Province) in outcrop and in the subsurface, along with radiometric age estimates for selected basaltic rocks. Note progression from older to younger ages—from both north and south edges toward center of composite basement terrane. Light green pattern denotes areal extent of Siletz and Crescent forearc terranes; light green diagonal lines denote subsurface extent of E layer; dark green pattern denotes areal extent of basaltic outcrops of the forearc terrane. Areal extent of Yachats, Cascade Head, Tillamook, and Grays River volcanic rocks denoted by yellow pattern; areal extent of Goble volcanic rocks denoted by red pattern; intrusive rocks denoted by orange diamonds. Light purple pattern denotes areal extent of the accreted Mesozoic Dothan Formation and Klamath Mountains terrane. Dashed blue line denotes outline of Tyee Formation. Hachured blue lines denote approximate location of profiles depicted in Figure 8. Plate boundaries from *Wilson* [2002]; Oregon volcanic outcrops from *Walker and MacLeod* [1991]; Washington volcanic outcrops from *Walsh et al.* [1987], *Phillips et al.* [1989], and *Schuster* [2005]; western edge of the Siletz-Crescent terrane on aeromagnetic data from *Blakely et al.* [2005]; southern end based on geologic data from *Walker and MacLeod* [1991]; eastern edge based on geologic data from *Snavely*, 1987; northern end of the terrane from this publication based on active source depiction of E layer from *Calvert et al.* [2003, 2006], *Graindorge et al.* [2003], *Nedimović et al.* [2003], and *Hyndman et al.* [1990]; outline of Tyee Formation from *Walker and MacLeod* [1991]; outline of Dothan Formation and Klamath Mountains terrane modified from *Walker and MacLeod* [1991], *Wells et al.* [2000], and *Irwin* [2003]. Transverse Mercator projection, WGS 84; standard parallel 128°W, centered at 46.8°N, 128°W, with standard parallel rotated 3.5° clockwise of vertical.

Table 1. Age Estimates of Selected Volcanic Members of the Siletz and Crescent Terranes, As Well As Volcanic and Intrusive Units of Postulated Slab-Window and Slab-Tear Events

(a) Siletz-Crescent Basement Terrane (*south to north*)		
Roseburg area (Siletz River Volcanics)	56–52 Ma (coccoliths)	*Wells et al*. [2000]
	64–59 Ma (K-Ar and Ar-Ar)	*Duncan* [1982]
Tillamook area (Siletz River Volcanics)	54–49 Ma (nannofossils)	*Wells et al*. [1994]
	57–50 Ma (K-Ar and Ar-Ar)	*Duncan* [1982]
Willapa Hills volcanics (Crescent Fm)	55–48 Ma (Ar-Ar)	*Moothart* [1992]
	50–45 Ma (benthic foraminifera)	*Moothart* [1992]; McDougall, *writ. comm*., 2011
	53–46 Ma (K-Ar)	*Duncan* [1982]
Black Hills volcanics (Crescent Fm)	56–51 Ma (K-Ar)	*Duncan* [1982]
Bremerton Igneous Complex (Crescent Fm)	50.5 Ma ([a]U-Pb)	*Haeussler and Clark* [2000]
	50–49 Ma (Ar-Ar)	*Clark* [1989]
	56–46 Ma (K-Ar and Ar-Ar)	*Duncan* [1982]
Dosewallips River area (Crescent Fm)	56–31 Ma (Ar-Ar)	*Hirsch and Babcock* [2009]
Northern Olympic Peninsula (Crescent Fm)	53–45 Ma (Ar-Ar)	*Hirsch and Babcock* [2009]
Metchosin Igneous Complex	52 Ma ([a]U-Pb)	R. Zartman, *unpubl. data in* Massey [1986]
	58–42 Ma (K-Ar and Ar-Ar)	*Duncan* [1982]
	58–39 Ma (K-Ar)	*Massey* [1986]
(b) Non-basement MORB- and OIB-Affinity Magmatism (*south to north*)		
Yachats Basalt	36–34 Ma (K-Ar and Ar-Ar)	*Parker et al*. [2010] *compilation*
Cascade Head Basalt	34–30 Ma (K-Ar and Ar-Ar)	*Parker et al*. [2010] *compilation*
Tillamook Volcanics	41–39 Ma (K-Ar)	*Wells et al*. [1994]
	42–39 Ma (nannofossils)	*Wells et al*. [1994]
	46–42 Ma (K-Ar)	*Magill et al*. [1981]
Grays River volcanics	42–37 Ma (Ar-Ar)	*Moothart* [1992]; *Kleibacker* [2001]; *McCutcheon* [2003]
Flores volcanics	51.2–50.5 Ma ([a]U-Pb)	*Irving and Brandon* [1990]
Clayoquot intrusions	51.2–48.8 Ma ([a]U-Pb)	*Madsen et al*. [2006]
Walker Creek intrusions	50.9–50.7 Ma ([b] U-Pb)	*Groome et al*. [2003]
Mt Washington (Catface) intrusions	41–35.3 Ma ([a]U-Pb and K-Ar)	*Isachsen* [1987]
(c) Non-Basement Arc-Affinity Volcanics		
Goble Volcanics	37–34 Ma (K-Ar)	*Phillips et al*. [1989] *compilation*
	45–32 Ma (K-Ar)	*Beck and Burr* [1979]

[a]U-Pb age estimates on zircon.
[b]U-Pb age estimates on monazite.
K-Ar and/or Ar-Ar age estimates on basalt include whole rock as well as plagioclase, hornblende, or biotite separates.

[10] Paleomagnetic investigations indicate that the Siletz-Crescent terrane has undergone a large tectonic rotation since its formation, especially in Oregon. *Simpson and Cox* [1977] determined a 68° clockwise rotation for the Siletz River Volcanics in the Siletz River area (Figure 2). *Wells et al*. [2000] reported a 79° rotation for Siletz River Volcanics in the Roseburg area. The overall history of rotation, including younger units has been summarized by *Wells and Heller* [1988].

[11] Geochemical analyses of Siletz, Crescent, and Metchosin lavas from the Roseburg area [*Wells et al*., 2000], the Siletz River area [*Snavely et al*., 1968], Willapa Hills [*Phillips et al*., 1989; *Moothart*, 1992], the eastern Olympic Peninsula [*Babcock et al*., 1992], and southern Vancouver Island [*Muller*, 1977, 1980] document its oceanic origin. The entire volcanic suite is dominated by lavas similar to mid-ocean ridge basalt (MORB) tholeiites, although alkalic basalts with ocean-island basalt (OIB) affinities represent a significant fraction of the Siletz River basalts in the Roseburg area [*Wells et al*., 2000] where a lower tholeiitic basalt unit is overlain by an upper, largely alkalic basalt unit. In the Dosewallips River section of the Crescent Formation, *Babcock et al*. [1992] documented an upward progression from normal (N-) to enriched (E-) MORB tholeiites in the submarine section, and then to E-MORB and OIB tholeiites in the subaerial section. They found the trace element Zr/Y ratio to be most diagnostic of source character, generally with values less than 3.0 in the lower section (indicative of normal MORB), progressing to above 3.0 in the more enriched upper section. The Willapa Hills lavas are tholeiitic and resemble the upper Dosewallips section with high Zr/Y ratios [*Phillips et al*., 1989], whereas the Metchosin lavas, also tholeiitic, resemble the lower Dosewallips section with lower Zr/Y ratios. In summary, the geochemistry of the stratigraphically lowest basalt flows is indicative of a mid-ocean ridge (hotspot) source, whereas the geochemistry of basalts up-section are indicative of an increasing contribution from a ocean island magma source.

[12] Recent analyses of multiple radiogenic isotopes from several basalt exposures amplify this interpretation. Siletz, Crescent, and Metchosin tholeiites are isotopically identical to early Columbia River Basalt lavas [*Pyle et al*., 2009], which are attributed to middle Miocene passage of the Yellowstone hotspot. *Pyle et al*. [2009] argued that the OIB component of the Siletz-Crescent basalts represents the plume head phase of the Yellowstone hotspot based on the large total volume of the Siletz-Crescent volcanic terrane, its relatively limited age range (discussed below), and radiogenic isotope systematics (e.g., $^{87}Sr/^{86}Sr$ 0.7031–0.7037),.

2.1.1. Geochronology of Volcanic Basement

[13] *McWilliams* [1978, 1980] first suggested an age progression within the Siletz-Crescent terrane, based on

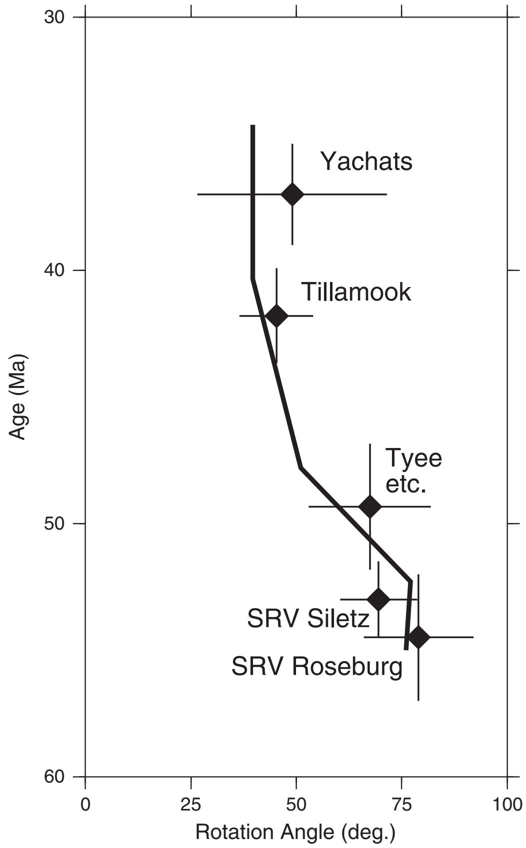


Figure 2. Paleomagnetic determinations of tectonic rotations for the older volcanic and sedimentary units of the Oregon Coast Range [*Simpson and Cox*, 1977; *Magill et al.*, 1981; *Wells et al.*, 2000], compared with our model (bold line) for transferring a Siletz microplate from the Farallon plate to the North American continental margin.

his interpretation of benthic foraminifera facies in sedimentary layers interbedded with basalt flows. He found the oldest ages (Penutian stage, ~early Eocene; Figure 3) in southwestern and central-western Oregon, and youngest ages, straddling the Ulatisian-Narizian boundary (~early to middle Eocene), near the Columbia River in northwestern Oregon and southwestern Washington. Benthic foraminiferal age assignments however have been notoriously difficult to correlate with global stratigraphic markers, and many boundaries have been suspected of being time-transgressive. *Poore* [1980], reviewing age assignments in California, found the Penutian-Ulatisian boundary to span several global age groups, rendering the age assignment for this boundary uncertain. Nonetheless, *Poore* [1980] found the Ulatisian-Narizian boundary in California to closely approximate the CP12-CP13 pelagic calcareous nannoplankton boundary, which in turn correlates to magnetic chron C21 in the earliest middle Eocene, ~48–47 Ma (Figure 3) [*Berggren et al.*, 1995; *Lourens et al.*, 2004]. The end of the Ulatisian in Oregon appears not to be significantly older than in California, as *Armentrout* [1981] noted co-occurrence of Ulatisian benthic foraminifera with planktonic foraminifera of the P10 zone and calcareous nannofossils of the CP12 zone in the Elkton Formation (formerly the Elkton Siltstone Member of the Tyee Formation) overlying the Siletz-Crescent terrane in western Oregon. Based on revised benthic foraminferal age assignments, the southern portion of the terrane is early Eocene and the central portion is earliest middle Eocene.

[14] *Duncan* [1982] documented the age progression within the Siletz-Crescent terrane by K-Ar and $^{40}Ar/^{39}Ar$ dating of several basalt exposures. Although oceanic basalts are far from ideal material for Ar dating and the internal consistency of his results was less than ideal, Duncan found a pattern broadly consistent with the results of *McWilliams* [1978, 1980], with the oldest ages of 64–59 Ma in southwestern Oregon and the youngest ages of 52–47 Ma in southwestern Washington (Table 1).

[15] Summarizing later work from south to north, *Wells et al.* [2000] reported coccolith microfossil ages for sedimentary interbeds within the Siletz River Volcanics in the Roseburg area (Figure 1) ranging from late Paleocene ("CP8?") in the core of the Roseburg anticlinorium to early Eocene (CP10) on the flanks (Figure 3). Furthermore, they found uniformly reversed magnetic polarity in 15 lava flows from three sites. Their integrated interpretation of fossil ages and magnetic polarity clearly identified magnetic chrons C24r and C23r (~56–52 Ma), although these ages are younger than the radiometric ages reported by *Duncan* [1982]. In the Tillamook area of northwestern Oregon (Figure 1), *Wells et al.* [1994] reported sedimentary interbeds in the Siletz River Volcanics containing calcareous nannoplankton from zones CP9b and CP11 (~54–49 Ma). This age range is consistent with three of five radiometric ages reported by *Duncan* [1982], with the other two Duncan ages slightly older.

[16] *Moothart* [1992] reported benthic foraminifera facies with ages of Ulatisian to early Narizian (Figure 3) for sedimentary interbeds in the Crescent Formation. He also reported $^{40}Ar/^{39}Ar$ ages of igneous rocks of the Crescent Formation in the Willapa Hills area of southwestern Washington (Figure 1). Two Crescent pillow basalts gave ages ~55–52 Ma, though slightly low $^{36}Ar/^{40}Ar$ intercepts (near 0.0030) suggest minor Ar retention, therefore slightly old ages. One sill, intruded into the overlying McIntosh Formation, gave a plateau age of 48.7 ± 0.5 Ma with a consistent inverse-isochron age of 48.6 ± 4.5 Ma and an atmospheric $^{36}Ar/^{40}Ar$ intercept of 0.0033, consistent with an unbiased value.

[17] *McDougall* [2007] has recommended an overhaul of the western U.S. benthic foraminiferal zonation, based on revised taxonomy and preferential use of species correlated to global markers. Her revised boundaries for the Eocene in California are generally younger than previous workers [e.g., *Poore*, 1980]. McDougall reviewed the species reported by *Moothart* [1992] and considers the Crescent Formation sites to have an early Ulatisian age under her revised zonation (K. McDougall, *written comm.*, 2011), rather than the Ulatisian to early Narizian ages reported by Moothart (Figure 3). She correlates these sites to planktonic foraminiferal zones P9-P10 (~50–45 Ma; magnetic chrons C22-C21), or in the best constrained sample, late P9 (~49 Ma). These numeric ages are reasonably consistent with the earlier biostratigraphic work of *McWilliams* [1978, 1980] and radiometric work of *Duncan* [1982].

[18] *Haeussler and Clark* [2000] reported a $^{238}U/^{206}Pb$ zircon age of 50.5 ± 0.6 for leucogabbro of the Bremerton Igneous Complex of *Babcock et al.* [1992] in northern Washington (Figure 1). In the same area, *Clark* [1989]

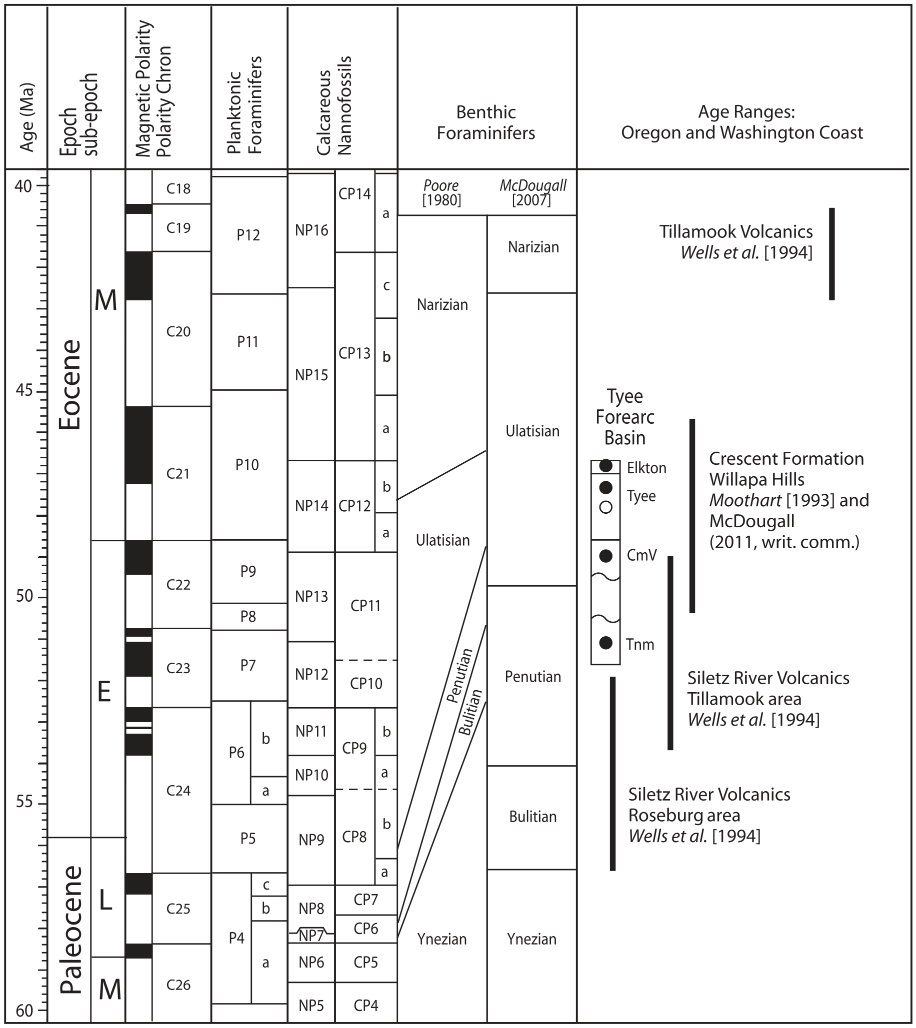


Figure 3. Age correlation chart modified from *Gradstein et al.* [2004] showing age range of key volcanic units, based on combined interpretation of radiometric, paleontologic, and magnetic polarity data discussed in the text. Age ranges do not show the full span of K-Ar or Ar-Ar data in most cases, but do show ranges consistent with paleontologic and magnetic data; therefore, they may be longer than the true range. Age constraints for the Tyee Basin stratigraphic section sampled by *Simpson and Cox* [1977] based on compilation of polarity data (circles) and paleontologic data reported by *Wells et al.* [1994]. (CmV = Camas Valley Fm; Tnm = Tenmile Fm)

obtained, through Duncan, ^{40}Ar/^{39}Ar total-fusion ages of 50.4 ± 0.6 Ma and 49.2 ± 0.8 Ma, and an isochron age of 50.3 ± 1.5 Ma from a single sample of leucogabbro (Table 1). These ages are within the scatter of previous ages reported by *Duncan* [1982] for northern Washington. Paleomagnetic samples from the area are exclusively reversed [*Beck and Engebretson*, 1982]. The zircon age and reversed polarity suggest intrusion during magnetic chron C22r [*Gradstein et al.*, 2004].

[19] *Babcock et al.* [2006] and *Hirsch and Babcock* [2009] listed new ^{40}Ar/^{39}Ar ages by Duncan from Crescent lavas on the Olympic Peninsula, Washington (Table 1). In the Dosewallips River section, eastern Olympic Peninsula (Figure 1), ages considered reliable included 51.0 ± 4.7 Ma and 50.5 ± 1.6 Ma from the upper subaerial section. The lower submarine section yielded an age of 56.0 ± 1.0 Ma (showing "evidence of substantial argon retention") and a highly discordant age of 31.2 ± 0.2 from the base of the submarine section. Paleomagnetic results from subaerial flows at Mount Walker, about 15 km north along strike from the Dosewallips River site, are mostly normal with a single reversed sample from the middle of the section [*Babcock et al.*, 1992]. Ages near 51 Ma with magnetic polarity that is mostly, but not exclusively, normal, are weakly suggestive of polarity chron C23n (Figure 3). Two sites in Crescent pillow basalt on the northern Olympic Peninsula yielded ages of 52.9 ± 4.6 Ma and 45.4 ± 0.6 Ma [*Hirsch and Babcock*, 2009].

[20] *Massey* [1986] reviewed radiometric dating of the Metchosin Igneous Complex of southern Vancouver Island. Ten K-Ar ages are scattered over the range of 55–39 Ma (Table 1). A U-Pb zircon age of 52 ± 2 Ma is attributed to unpublished data of R. Zartman (reported in *Massey* [1986]). We prefer the zircon age because of the vastly greater robustness of U-Pb dating relative to K-Ar dating.

[21] In summary, the age-progression interpretations of *McWilliams* [1980] and *Duncan* [1982] indicating older ages

at the south and north ends of the Siletz-Crescent terrane and younger ages in the center are broadly supported by more recent dating efforts. The youngest ages remain in southwestern Washington, and possibly northwestern Oregon, and the oldest ages are in southwestern Oregon. A discrepancy between *Duncan*'s [1982] Ar ages (64–59 Ma) and *Wells et al.*'s [2000] stratigraphic ages (~56–52 Ma) near Roseburg is the only case where more recent ages are not within the scatter of early ages. Because the Roseburg sample areas are not overlapping, it is possible that both age ranges are correct. For our interpretation of the accumulation of a thick basaltic section near Roseburg (Table 1 and Table S1), we favor the younger biostratigraphic ages, because if both are correct, the younger ages represent the time of thickening of the basaltic section during seamount formation. Also, if one of the age ranges is wrong, we find it much more plausible that the Ar ages are too old, due to a mechanism such as Ar retention during submarine eruption, rather than the biostratigraphic ages are too young. In a magnetic context for comparison to plate tectonic models, the oldest ages, near Roseburg, are C24r, the youngest ages, near the Columbia River, are C21 or C22n, and northwestern Washington and Vancouver Island zircon ages are intermediate, C22r to C23 (Figure 3). Some previous studies have interpreted scatter in Ar ages as indicating long duration of volcanism [e.g., *Hirsch and Babcock*, 2009]. We prefer to interpret the scatter as an indication of the limitations of Ar dating on oceanic basalt. The integrated interpretation above shows a limited overall age range of about 55–49 Ma, with few individual sites having a clearly demonstrated age range longer than 1 Myr.

2.1.2. Sedimentary Units Interbedded With Volcanic Basement

[22] In addition to age constraints, sedimentary interbeds within the Siletz-Crescent terrane provide important clues regarding its paleogeographic setting. *Cady* [1975] and *Babcock et al.* [1992] described continentally derived turbidites as occurring below and interbedded throughout the submarine section of the Crescent Formation on the Olympic Peninsula. A conglomerate is interbedded with pillow basalt within the Crescent Formation ~10 km above the base of the section on the east side of the Olympic Mountains [*Cady*, 1975]. This conglomerate includes boulders of hornblende-biotite quartz diorite that are up to 3 m in diameter, implying proximity to the continent during the later phase of volcanism. *Cady* [1975] attributed these boulders to plutons in the Cascade range to the east based on an approximate radiometric age of 65 Ma from a boulder and a similarity between petrographic features (F. W. Cater, 1966, *oral comm.*, in Cady [1975]).

[23] In the Roseburg area, *Wells et al.* [2000] also reported conglomerates with up to boulder-sized clasts derived from a Klamath Mountains source interbedded with the Siletz River basalts. Distinctive clasts include cherts, greenstones, and metagraywackes. Thick conglomerate interbeds are found both in the oldest basalts and ~5 km up-section in the youngest basalts, implying proximity to the continent throughout the eruptive sequence.

2.2. Umpqua and Tyee Basin Sedimentary Rocks

[24] In southwestern Oregon, the lower to middle Eocene Umpqua and Tyee Basin strata overlying Siletz River Volcanics have long been studied for the record they provide on regional tectonics. The following summary, including new age data and revisions to stratigraphic nomenclature, is simplified from *Wells et al.* [2000].

[25] *Simpson and Cox* [1977] interpreted a vertical-axis clockwise rotation of 64° for these Eocene sedimentary rocks (Figure 2) nearly as large as (and within the uncertainty of) what they found for older Siletz River Volcanics. The oldest sedimentary unit sampled by Simpson and Cox is the Tenmile Formation (formerly a member of the Lookingglass Formation). This turbidite unit is locally tightly folded. Its biostratigraphic age straddles the CP10-CP11 calcareous nannofossil boundary [*Wells et al.*, 2000], which combined with the unit's normal magnetic polarity, indicates polarity chron C23n and an age of 51–52 Ma (Figure 3). Marine mudstones of the Camas Valley Formation (formerly a member of the Flournoy Formation), overlying the Tenmile Formation in angular unconformity, yielded CP11 nanofossils [*Wells et al.*, 2000] and normal polarity, either C23n or C22n. The overlying Tyee Formation grades upward from turbidites to fluvial-deltaic deposits. Tyee strata contain few fossils, but yielded normal magnetic polarities overlying reversed polarities. The widespread Tyee Formation is only gently deformed. Marine mudstones of the Elkton Formation, overlying the Tyee Formation, yielded CP12b coccoliths [*Wells et al.*, 2000] and normal magnetic polarity, indicating a rather narrow age range in lower C21n, ~47 Ma (Figure 3).

[26] *Heller and Ryberg* [1983] interpreted compositional and structural changes observed in this overlying sedimentary sequence to record the process from accretion of the Siletz terrane to establishment of a stable forearc at the beginning of the middle Eocene. In particular, the gently deformed Tyee Formation overlies both the tightly folded Siletz River Volcanics of the Roseburg area and the adjacent Mesozoic Dothan Formation and Klamath Mountains terrane (Figure 1), and is not offset by the thrust contact between them [e.g., *Heller and Ryberg*, 1983; *Wells and Heller*, 1988; *Wells et al.*, 2000], thus providing a key geologic constraint on the timing and location of initial contact between Siletz terrane and the Eocene continental margin. *Heller and Ryberg* [1983] and *Wells and Heller* [1988] interpreted this field relation as indicating that accretion of the entire "seamount terrane" was completed prior to Tyee turbidite deposition. We point out that Tyee deposition may well overlap with eruption of the youngest Siletz-Crescent terrane lavas in the Willapa Hills area (Figure 3) to the north.

2.3. Younger Igneous Units

2.3.1. Tillamook Volcanics and Grays River Volcanics

[27] Another geologic constraint on where and when the oceanic fragments docked against the continental margin comes from volcanic rocks that have intruded into and erupted through the Siletz-Crescent terrane near Tillamook, Oregon (Figure 1). As summarized by *Wells et al.* [1994], the Tillamook Volcanics comprise a submarine to subaerial sequence of tholeiitic to alkalic flows of mostly high-TiO_2 basaltic composition. Radiometric ages are scattered, with preferred ages roughly 42–44 Ma. Interbedded marine strata contain calcareous nannofossils assigned to zone CP14a. Paleomagentic results of *Magill et al.* [1981] showed exclusively reversed directions rotated about 46° clockwise relative to North America (Figure 2).

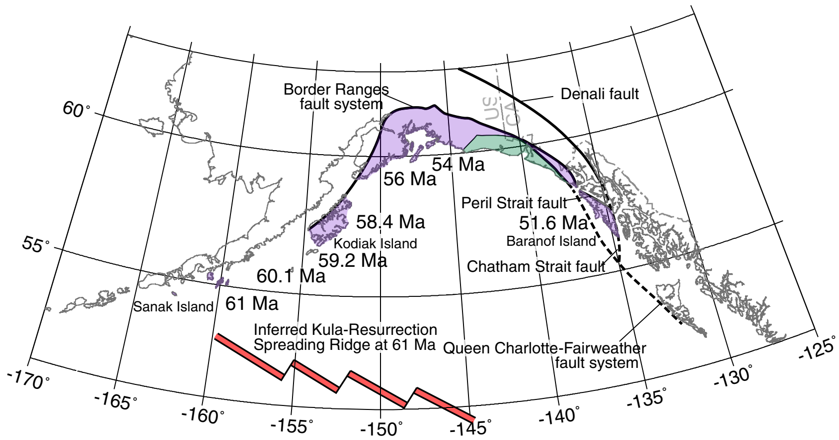


Figure 4. Map depicting the mobile terranes of southern Alaska mentioned in the text, after *Haeussler et al*. [2003], *Bradley et al*. [2003], and *Cowan* [2003]. The Upper Cretaceous Chugach-Prince William terrane denoted by the light purple pattern; the lower Eocene Yakutat terrane denoted by the light green pattern. Radiometric ages are from MORB-affinity forearc intrusions in the Sanak-Baranof magmatic belt [*Bradley et al*., 2003; *Farris and Paterson*, 2009], and are attributed to eastern migration of the Kula-Resurrection spreading ridge in Paleogene time.

Combining the nannofossil zone and the reversed polarity suggests an age of ~41 Ma, during polarity chron C19r (Figure 3), several million years younger than the underlying Siletz River Volcanics.

[28] The Tillamook Volcanics have been correlated with the 42–37 Ma Grays River volcanics of southwest Washington (Figure 1), described in detail by *Chan et al*. [2012]. Chan et al. interpreted the tholeiitic to mildly alkalic (OIB) geochemistry and isotopic values of the Grays River volcanics as reflecting a magma source in a hotspot-influenced slab-window setting.

2.3.2. Yachats and Cascade Head Basalts

[29] Two basaltic sequences of late Eocene age are exposed along the central Oregon coast (Figure 1), the ~36 Ma tholeiitic Yachats Basalt and the ~34 Ma alkalic Cascade Head Basalt [*Parker et al*., 2010, and references therein]. Both sequences are associated with nearby alkalic syenite and camptonite intrusions. The isotopic signature of these units is that of enriched MORB (e.g., $^{87}Sr/^{86}Sr$ ranges 0.7033–0.7036).

2.3.3. Goble Volcanics and Other Arc Volcanics

[30] East of the Grays River volcanics in southwestern Washington are the Goble Volcanics (Figure 1), a suite of predominantly basalitc andesite lavas of probable late Eocene age. Chemistry, age data, and relations to nearby geologic units are summarized by *Phillips et al*. [1989]. Preferred K-Ar ages range from 37 to 34 Ma. Composition is transitional between tholeiitic and calc-alkaline, so the suite is easily distinguished from slightly older Grays River flows (formerly treated as part of Goble) by higher SiO_2 and lower TiO_2 values. Phillips et al. interpreted the Goble Volcanics to be part of the early Cascade volcanic arc in western Washington and Oregon.

2.4. Yakutat Terrane, Southeastern Alaska

[31] *Davis and Plafker* [1986] argued that the basaltic rocks dredged from the southern margin of the Yakutat terrane in offshore southeastern Alaska (Figure 4) correlate with the Siletz-Crescent terrane. They based this correlation on the early to middle Eocene age of the basalts inferred from associated fossils and scattered K-Ar dates, on strong geochemical similarities especially to Metchosin basalts on Vancouver Island and lower Crescent basalts from the Olympic Peninsula, and on associated clastic strata indicating deposition near a continental margin. Subsequent seismic refraction surveys indicated that mafic igneous crust for the Yakutat terrane ranges from 17 km thickness in the northwest to 30 km in the southeast [*Christeson et al*., 2010; *Worthington et al*., 2012]. Recent GPS-derived velocities [*Elliott et al*., 2010] confirm earlier inferences from regional geophysics [e.g., *Plafker et al*., 1978] that the Yakutat terrane currently has only minor motion relative to the Pacific plate and ~50 mm/yr motion relative to the North American plate.

2.5. Yellowstone-Snake River Plain Volcanism

[32] Our kinematic model includes a track for a Paleogene Yellowstone hotspot. The fundamental cause of volcanism in the Yellowstone-Snake River Plain system, however, has been the subject of long and vigorous debate, recently summarized by *Pierce and Morgan* [2009] and *Fouch* [2012]. One group favored a deep-seated mantle plume source for volcanism and uplift as proposed by *Morgan* [1971], while others emphasized inconsistencies between observed geology and the predictions of a simple plume model. Some of these workers favored convection in the upper mantle triggered by complexities in the fate of the subducted Juan de Fuca plate [e.g., *James et al*., 2011]. Evidence in favor of the classic deep plume model includes the age progression of rhyolitic volcanic centers in the Snake River Plain consistent with the rate predicted by the reference frame defined by the age progression of Hawaii and other volcanic centers [*Armstrong et al*., 1975], the high $^{3}He/^{4}He$ ratio of Yellowstone-Snake River Plain lavas indicative of an undepleted source [*Graham et al*., 2009], and the large positive geoid anomaly centered on Yellowstone [*Smith and Braile*, 1994]. Evidence against a Hawaii-style plume includes lava compositions indicative of magma

generation primarily in continental mantle [*Carlson and Hart*, 1987; *Leeman et al*., 2009]. Seismic tomography models have been cited as evidence both for and against a low-velocity column representing a high-temperature plume, although most interpretations of recent data favored such a column plunging steeply to the northwest from Yellowstone, Wyoming to a depth of at least 400–500 km [e.g., *James et al*., 2011].

[33] Even among workers who support the interpretation of a classic mantle plume, there is disagreement as to how long the plume has been present in the shallow mantle. Many favored a plume head first reaching the shallow mantle at the time of widespread flood-basalt volcanism starting at ~17 Ma [e.g., *Pierce and Morgan*, 2009], while others interpreted the lack of back-arc volcanism prior to 17 Ma as simply a result of the Juan de Fuca slab trapping the plume below the slab thereby preventing earlier shallow melting [*Duncan*, 1982; *Geist and Richards*, 1993]. *Parker et al*. [2010] interpreted strongly alkalic flows (e.g., Cascade Head Basalt) and intrusions of late Eocene age in the Oregon Coast Range as evidence for a newly formed mantle plume. Following *Duncan* [1982], several workers favored an interpretation that the Yellowstone hotspot played a major role in forming the thick basalt sequences of early Eocene age in the Oregon and Washington Coast Ranges in a setting near a mid-ocean ridge [e.g., *Murphy et al*., 2003], discussed in detail below. We test the plausibility of this interpretation in our kinematic model by explicitly defining a Yellowstone track based on plate circuit parameters.

3. Tectonic Models

3.1. Previous Tectonic Interpretations

[34] *Duncan* [1982], elaborating on a previous interpretation of *Simpson and Cox* [1977], proposed that the age progression in the Siletz and Crescent terranes results from capture of tracks of the Yellowstone hotpsot that formed on the Farallon and Kula plates, respectively. *Wells et al*. [1984] argued that the age progression cannot result from capture of hotspot tracks embedded in coherent plate fragments, because plate motions of both the Farallon and Kula plates, with respect to the hotspot frame, were much faster than the motion suggested by the gradient of Duncan's ages. Additionally, the presence of continentally derived conglomerate interbedded with the basalts seemed difficult to reconcile with high convergence rates of the Kula and Farallon plates relative to the North American plate. *Wells et al*. [1984] proposed one alternative model in which volcanism associated with the Yellowstone hotspot occurred on transform faults during plate-motion reorganization in the North Pacific between 61 and 48 Ma. They also proposed a second alternative in which the hotspot triggered rifting of the continental margin, allowing accumulation of thick basaltic sequences in rift basins. *Babcock et al*. [1992] favored this latter interpretation, citing terrigenous sediments interbedded with Crescent basalts throughout the Dosewallips River section, lack of lower crust and upper mantle ophiolite sections, rarity of upper crust ophiolite sections, and conformable relations between Crescent basalts and overlying terrigenous marine sediments as evidence against interpretations based on accreted oceanic crust.

[35] There are several reasons to revisit the simple model of forming the Siletz and Crescent terranes as a pair of hotspot tracks that accreted as relatively coherent units, previously rejected by *Wells et al*. [1984]. First, additional dating discussed above casts doubt on the reliability (or at least its relevance) of the oldest ages presented by *Duncan* [1982]. A revised age range of 55–49 Ma for the terranes is consistent with much faster spreading rates than the original range of 64–47 Ma.

[36] Second, the model of a Kula-Farallon ridge interaction with the continental margin needs to be revised, considering evidence that Paleogene plate reorganization in the North Pacific may have included formation of new plates and microplates broken from the Kula and Farallon plates. *Haeussler et al*. [2003], citing evidence that a spreading ridge swept eastward across southern Alaska about 60–50 Ma (Animations S1 and S2), proposed the existence of a Resurrection plate between the Kula and Farallon plates, with the Kula-Resurrection spreading ridge reaching North America in Alaska and the Farallon-Resurrection spreading ridge reaching North America in the vicinity of Oregon and Washington, at least 60–50 Ma. If the Resurrection plate existed, the spreading rate on the Farallon-Resurrection boundary near Oregon may have been significantly slower than the rate expected for the Kula-Farallon boundary and more compatible with the rates required by the geologic record.

[37] Third, spreading rate information for the Northeast Pacific has been updated for the Kula-Pacific plate pair by *Lonsdale* [1988a], and for the Farallon-Pacific plate pair and the descendant Juan de Fuca (Vancouver)-Pacific plate pair by *Caress et al*. [1988]. Importantly, Juan de Fuca-Pacific spreading rates after 52 Ma are much slower than Farallon-Pacific spreading rates prior to 53 Ma. In addition to the unavailability of this work to *Wells et al*. [1984], most (perhaps all) more recent studies have not properly incorporated these plate revisions.

[38] The *Haeussler et al*. [2003] and *Bradley et al*. [2003] model for the existence of a Paleogene Resurrection plate is not the only plate configuration that might explain geologic relationships in southern Alaska and the Pacific Northwest region of North America. Owing to the fragmentary nature of terranes along the continental margin and evidence that many terranes are far-traveled, many aspects of the plate tectonic evolution of the Paleogene northeast Pacific remain unresolved. In particular, plate reconstructions by *Breitsprecher et al*. [2003] and *Madsen et al*. [2006] provide alternative models for multiple oceanic plates in the northeast Pacific. If the Chugach terrane of southern Alaska was along the Pacific Northwest in the Paleogene, there may not be a need for a Resurrection plate [*Cowan*, 2003]. Our kinematic model incorporates limited offset of Chugach terrane and more interior terranes after 60 Ma based on a thorough review by *Farris and Paterson* [2009] favoring limited young offset.

[39] The Resurrection plate model results in two ridge systems (Kula-Resurrection and Resurrection-Farallon) that effectively take up the spreading rate previously manifested by one faster spreading ridge (Kula-Farallon). This revision to previous plate tectonic models is integral to our reconstruction of the Siletz and Crescent terrane as representing an accreted oceanic ridge segment.

3.2. Hotspot Influence

[40] Perhaps Iceland astride the Mid-Atlantic ridge (separating the North American and Eurasian plates) is a

Table 2. Revised Plate Motion Rates Used in Kinematic Model

Polarity Chron Range	Age (Ma GTS2004)				
(a) Farallon/Juan de Fuca motion relative to North America					
		Velocity at 42°N, 122°W		Velocity at 35°N, 119°W	
		Speed (mm/yr)	Az. (deg)	Speed (mm/yr)	Az. (deg)
C27y-C24y	61.6–52.6	115	37	118	40
C23o-C21o	51.9–47.2	64	45	71	50
C23o-C21o	47.2–39.5	74	38	87	45
(b) Kula plate motion relative to North America					
		Velocity at 55°N, 160°W		Velocity at 55°N, 135°W	
		Speed (mm/yr)	Az. (deg)	Speed (mm/yr)	Az. (deg)
C30o-C27y	67.7–61.6	45	340	139	0
C27y-C24r	61.6–54.3	95	326	112	350
C24r-C24y	54.3–52.6	404	298	526	333
C23o-C21o	51.9–47.2	90	335	93	356

close modern analog to the postulated Juan de Fuca-Resurrection ridge segment that accreted to western North America during Eocene time. Iceland is an emergent segment of the Mid-Atlantic ridge owing in part to its proximity to the Iceland hotspot, which provides for increased production of mid-ocean ridge basalts. Icelandic basalts contain a mixture of MORB and OIB (i.e., hotspot) affinities, as do basalts within the Siletz-Crescent terrane. The emergent Icelandic landmass is approximately 550 km long by 300 km wide, including several smaller islands, and reaches to more than 2000 m above sea level. In comparison, the Siletz-Crescent terrane is about 700 km long by 150 km wide. The portion of the Siletz-Crescent terrane that was emergent during active volcanism, however, is much smaller and appears to have been restricted to isolated volcanic centers (Table S1).

[41] The Galápagos hotspot tracks, emanating from the Cocos-Nazca spreading ridge system in the Central East Pacific, provide another modern example of hotspot-spreading ridge interaction. The Cocos track on the Cocos plate is characterized by a 1000 km long by 300 km wide aseismic ridge that stands 2000 m above the seafloor [*Werner et al.*, 2003]. Seamounts composed mainly of alkali basalt lavas rise more than 1000 m above the ridge. The Carnegie track on the Nazca plate, including the emergent Galápagos Islands and the Carnegie Ridge, is also a 1000 km long by 300 km wide aseismic ridge, covered with seamounts that rise several hundred meters above the ridge. Volcanic rocks from these aseismic ridges are mainly composed of mid-ocean ridge basalts (tholeiites) whereas the overlying seamounts are mainly oceanic island basalt (alkali basalts) [*Werner et al.*, 2003], similar to the basaltic sequence of the Siletz and Crescent terranes. The Galápagos hotspot tracks are much thicker than normal oceanic crust, thus generating buoyant crust that resists subduction, and in the case of the Cocos track indents the Central American continental margin [*LaFemina et al.*, 2009]. We envision a similar scenario for the Eocene accretion of the Yellowstone hotspot tracks.

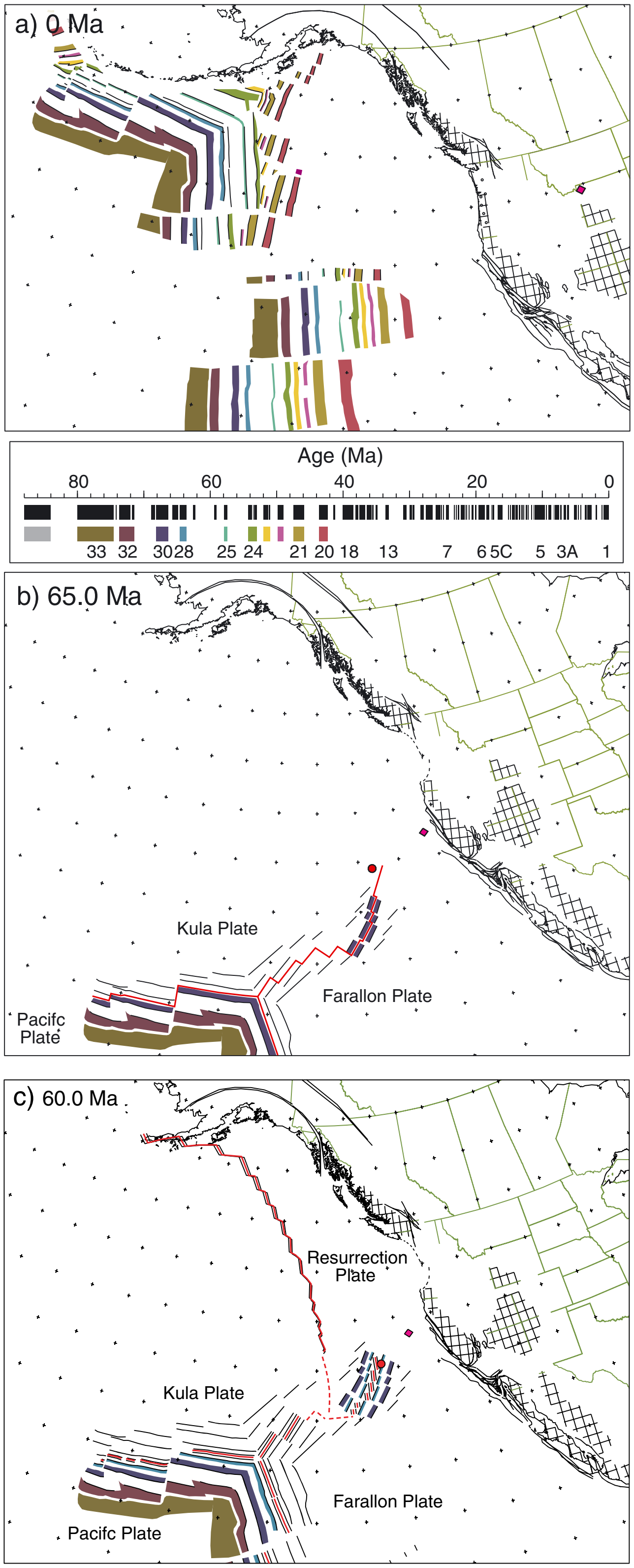
3.3. Revised Kinematic Framework

[42] Our goal for this contribution is to test the viability of a hotspot-track origin for the Siletz and Crescent terranes by constructing a quantitative kinematic plate model that incorporates the new geologic, paleomagnetic, and plate tectonic constraints discussed above. The framework of the model is Pacific-North America relative motion calculated from the plate circuit through Africa and Antarctica, updated slightly from the work of numerous prior authors including *Stock and Molnar* [1988]. North America-Africa motion is from *Klitgord and Schouten* [1986], Africa-East Antarctica motion is from *Nankivell* [1997], West Antarctica-East Antarctica motion is new from *Wilson and Luyendyk* [2009], and Pacific-West Antarctica motion is from *Cande et al.* [1995]. Model parameters are presented in Table S2.

[43] Within the Pacific basin, Farallon-Pacific motion for C25 (56 Ma) and older is slightly updated from *Rosa and Molnar* [1988] to maintain assumed constant half-spreading rate over shorter time intervals than they considered. Motion younger than C25 follows *Caress et al.* [1988], with the Juan de Fuca plate separating from the Farallon plate north of the Mendocino fracture zone early in C23r (52.3 Ma), but is refined to be consistent with unpublished, more recently collected data obtained through the National Geophysical Data Center (http://www.ngdc.noaa.gov/). Following *Menard* [1978], Caress et al. referred to the new plate as the Vancouver plate, but for simplicity and adherence to precedence conventions we call it the Juan de Fuca plate. Kula-Pacific motion is updated from previous work by using spherical geometry to quantify the spreading rate variations documented by *Lonsdale* [1988a]. This revision is a moderate change from the work of *Rosa and Molnar* [1988], who averaged spreading rates over time intervals where motion turns out not to have been constant. Revised convergence velocities for Farallon/Juan de Fuca and Kula plates relative to North America are presented in Table 2.

[44] Rotation parameters for the hypothesized Resurrection plate are new to this study (Table S2). Details are of course both speculative and nonunique. Desirable properties of its motion are a northward velocity of the southeastern part of the plate adjacent to Washington relative to North America, maximizing the available time for continentally derived sediments to interbed with Crescent lavas, and significant clockwise rotation of the plate, so the northern part of the plate has fast eastward motion relative to the Kula plate, allowing the relatively fast sweep of the plate boundary across southern Alaska inferred by *Haeussler et al.* [2003].

[45] To fit the high rotation rate recorded for the Siletz River Volcanics, Tyee Formation, and Tillamook Volcanics (Figure 2), we assume that a distinct Siletz microplate separates from the northern end of the Farallon plate, to be eventually captured in the Oregon Coast Range. For model simplicity, we assume this separation occurs at the same



time as the Juan de Fuca plate separates from the Farallon plate, although either earlier or slightly later is also possible. To satisfy constraints including rotation history and geologic relations near Roseburg, the Siletz microplate must have a rotation pole location in southwest Oregon relative to North America.

[46] Our North America reconstruction builds from that of *Wilson et al*. [2005], who modeled plate configurations back to 30 Ma and north to ~49°N. Additional restorations, which we treat as subsequent to the 65-40.5 Ma scope of the current model, include ~35–30 Ma extension in eastern Nevada and adjacent areas (e.g., review by *Dickinson* [2002]) and ~100 km of late Cenozoic right-oblique transpressional motion on the Denali-Fairweather fault system of southern Alaska and adjacent western Canada. Active motion within North America during the scope of the model includes extension ~55–48 Ma in southeastern British Columbia and northeast Washington [*Dickinson*, 2002], and ~100 km of right-lateral strike-slip motion on the Fraser-Straight Creek fault of Washington and British Columbia [*Tabor et al*., 1984], with fault activity bounded between 47 and 34 Ma [*Coleman and Parrish*, 1991]. We shift the position of the Klamath Mountains ~30 km inland to avoid having the Siletz microplate pass under them on its way to the Oregon coast.

[47] Because several authors have proposed that the Yellowstone hotspot is responsible for the thickness and some of the geochemical characteristics of the Siletz and Crescent terranes [e.g., *Duncan*, 1982; *Pyle et al*., 2009], we include a hotspot reference frame in our model. Our time frame is almost certainly too old for an approximation of mutually fixed hotspots to be valid [e.g., *Tarduno et al*., 2003]. We offer a simple depiction of the uncertainty in the hotspot frame by presenting both the position of the Yellowstone hotspot as described by the motion of Atlantic

Figure 5. (a) Location map showing elements of the kinematic model in their present positions. Extant isochrons denoted by colored polygons. Grid pattern denotes continental regions modeled as rigid blocks; green lines denote state and provincial boundaries. (b) A schematic map showing the initial Pacific-Kula-Farallon triple junction configuration of kinematic model at 65 Ma. Colored isochron shading shows observed isochrons on Pacific plate or inferred isochrons adjacent to proposed ridge in region interacting with Oregon-Washington-British Columbia coast. Magenta diamond denotes location of Yellowstone hotspot in Pacific plate reference frame [*Lonsdale*, 1988b]; red circle denotes alternative location of hotspot in Atlantic reference frame [*Müller et al*., 1993]. Thin red lines denote location of active spreading ridges. (c) A schematic map showing the Pacific-Kula-Farallon triple junction configuration of kinematic model at 60 Ma including initial Resurrection-Kula boundary. (d) A schematic map showing the Resurrection-Kula boundary connecting triple junctions with Farallon plate and North American plate in southern Alaska at 50 Ma and location of Siletz microplate. (e) A schematic map showing a close-up of the plate configuration at 47.5 Ma after the Crescent microplate broke from Resurrection plate. Heavy black lines denote break along western edge of Crescent and Siletz microplates; small black circles denote a schematic representations of basaltic outcrops.

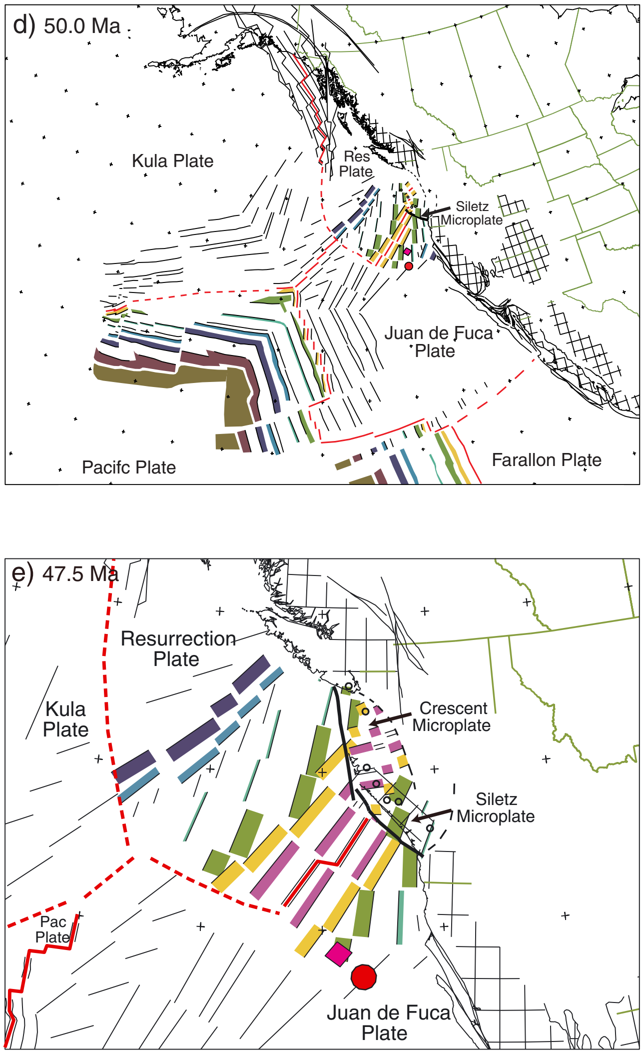


Figure 5. (continued)

hotspots relative to the North American plate from *Müller et al*. [1993], and alternatively the position as described by the motion of Pacific hotspots relative to the Pacific plate [*Lonsdale*, 1988b].

3.4. Model Evolution

[48] For oceanic plates, we generate magnetic isochrons as old as C32y (71 Ma) by assumed symmetric spreading. The shape of the Kula-Farallon boundary (Figure 5b) is chosen such that the ridge axis will intersect North America near Oregon and Washington after 55 Ma. We model the Resurrection plate as separating from the Kula plate (Figure 5c) at C27y (60.9 Ma), although different early histories are also possible. A series of left-stepping transform faults on the Kula-Resurrection ridge facilitates the eastward sweep of the plate boundary across southern Alaska (Animation S1). Northward motion of the Resurrection plate is set to be either similar to or slower than northward motion of the Kula plate.

[49] As outlined above, the Farallon plate breaks up during C23r (52.3 Ma; Figure 5d). Our Siletz microplate is only moving slowly relative to North America in southwestern Oregon, and must be converging with the Juan de Fuca plate on its southwest side. Because of the rapid rotation rate we specify to fit Siletz paleomagnetic data, convergence must continue for several million years between the Siletz microplate and North America at the latitude of northern Oregon, though the longitudinal position of the boundary is poorly known.

[50] Our Crescent microplate separates from the Resurrection plate somewhat later, at ~48 Ma (Figure 5e), eventually to be captured in the Washington Coast Range and constitutes the southward-younging part of the composite forearc terrane. Rotation of the Crescent microplate is significantly less than the Siletz microplate (~22–40° clockwise; *Magill et al*. [1981]; *Wells and Helle*r [1988]). Final docking of the Crescent microplate relative to adjacent parts of greater North America may have been achieved at Vancouver Island by ~40 Ma. The slower motion of the Resurrection plate, as compared with the Kula plate, and the captured Siletz and Crescent microplate fragments resolve previous issues with the rate of basalt age progression and the amount of plate convergence discussed above.

[51] Subducted oceanic fragments have been stranded beneath the North American continental margin elsewhere. The Monterey microplate (derived from the Farallon plate), for example, rotated, stalled, and tore beneath central California in the Miocene [*Wilson et al*., 2005]. The Guadalupe and Magdalena microplates (also derived from the Farallon plate) rotated, stalled, and tore beneath Baja California in the Miocene [*McCrory et al*., 2009] as well. These microplate fragments can be mapped from the sea floor eastward beneath the continental margin. In these particular cases, the stranded fragments marked the end of subduction—with the final microplate fragments too buoyant to subduct completely—followed by plate reorganization as a transform boundary. Whereas off Vancouver Island, in our scenario, subduction was reestablished after accretion of the Crescent microplate, initially between the Resurrection and North American plates. With the final demise of the Resurrection plate, a second reorganization occurred (likely before 35 Ma), this time establishing (ongoing) subduction of the Juan de Fuca plate beneath the entire Pacific Northwest continental margin from northern California to Vancouver Island.

[52] Death of the Juan de Fuca-Resurrection spreading segment after docking did not preclude continued subduction of the down-dip Juan de Fuca and Resurrection slabs owing to negative buoyancy of oceanic lithosphere older than ~10 Myr; however, new oceanic lithosphere would no longer have been created [e.g., *Thorkelson*, 1996]. If we assume a continuation of Juan de Fuca-Resurrection spreading parameters for the descending slabs, such motion predicts slow opening of a narrow slab window between Tillamook, Oregon and Willapa Hills, Washington starting about 44 Ma. This prediction is consistent with the timing and location of the Tillamook Volcanics and the Grays River volcanics (Figure 1), and compatible with the isotopic composition of the Grays River volcanics, which indicate magma with mantle plume and slab-window signatures [*Chan et al*., 2012]. Such slab windows, derived from subducting ridge segments, typically fill with upwelling asthenosphere and are often associated with volcanism [e.g., *Thorkelson*, 1996; *McCrory and Wilson*, 2009]. Modern analogs include the subduction of the Cocos-Nazca ridge beneath Costa Rica [e.g., *Johnston and Thorkelson*, 1997], the East Pacific Rise beneath Mexico

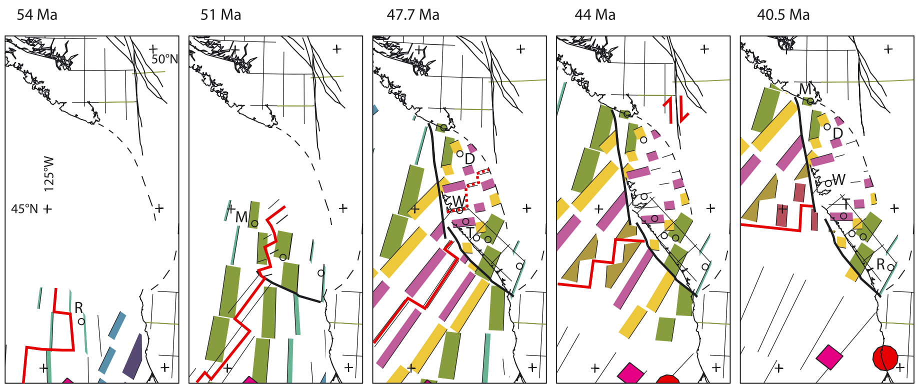


Figure 6. Close-up view of model evolution for the Oregon-Washington Coast Range area. Small black circles show Siletz-Crescent outcrop landmarks tracked from their formation on oceanic crust to their capture positions in North America (M, Metchosin; D, Dosewallips River; W, Willapa HIlls; T, Tillamook; R, Roseburg). None of the sites formed far from the continent, consistent with the presence of clastic sediments interbedded with basalts. Thin red lines denote location of active spreading ridges; red arrows denote location of active strike-slip faulting. (See Figure 5 for description of other symbols.)

[e.g., *Bourgois and Michaud*, 2002], and the Nazca-Antarctic ridge beneath southernmost South America [*Breitspecher and Thorkelson*, 2009]. In fact, magmatism inboard from the Cascade arc in the Pacific Northwest region has been attributed to an earlier slab window sequence *Breitsprecher et al.* [2003], which initiated in the Cretaceous, and the Sanaf-Baranof magmatic belt is attributed to a slab window sequence [*Farris and Paterson*, 2009], associated with the eastward sweep of the Kula-Resurrection ridge along the Alaskan trench in the Paleogene.

4. Discussion

[53] Our primary goal has been evaluating whether the objections raised by *Wells et al.* [1984] to *Duncan*'s [1982] model for a near-ridge hotspot origin for the Siletz-Crescent terrane are still valid. As outlined above, the primary objections were that the range of Duncan's ages was much greater than predicted by the Kula-Farallon spreading rates and that the presence of coarse continental sediments interbedded with lavas was difficult to reconcile with the very rapid convergence rates between either the Kula or Farallon plate and North America. The strength of these objections, especially the first, has been reduced by revisions to the geochronology, indicating an age range for the terrane of about 5–8 Myr, with only the Roseburg area of southwest Oregon demonstrably older than 53 Ma, when plate motions were fastest.

[54] In detail, our model has most of the Siletz-Crescent terrane form at or near the spreading boundary between the Siletz microplate, which we identify following a suggestion of *Simpson and Cox* [1977], and the Resurrection plate of *Haeussler et al.* [2003]. We consider the introduction of both of these plates to be plausible geologically. The Siletz microplate is reasonable as a buoyant fragment of thick oceanic crust, rotating due to substantial coupling to North America at its southeast end and partial coupling with the Juan de Fuca plate at its northwest end. Arguments in favor of the existence of the Resurrection plate have been advanced by *Haeussler et al.* [2003] and *Bradley et al.* [2003]. For purposes of fitting the Siletz-Crescent ages, any oceanic plate moving northward at moderate rate more slowly than the Kula plate can serve as the conjugate to the Siletz microplate. We find no difficulty in using the Resurrection plate to satisfy both the progression of 60–50 Ma intrusive ages across southern Alaska that originally motivated the identification of the plate, and the ages of Crescent and Metchosin basalts in Washington and British Columbia.

[55] With the revised chronology and additional plates introduced, coarse conglomerates interbedded with lavas near Roseburg and in eastern Olympic Peninsula present much less of a conflict with the near-ridge hotspot model than when considered by *Wells et al.* [1984]. At Roseburg, separation of the Siletz microplate allows minimum convergence between the oldest lavas and North America. With the Resurrection plate moving northward relative to North America, it is possible for the Olympic Peninsula conglomerates to accumulate close to the continent and move northward for 2–3 Myr before accreting to the continent.

4.1. Fitting Observations of Age Progression and Convergence to Continent

[56] If the Oregon and Washington Coast Ranges basalts formed at a single segment on the Farallon-Resurrection spreading ridge near the Yellowstone hotspot, a full spreading rate of 60–80 km/Myr for this postulated segment is easily reconciled with evidence for the existence of the Resurrection plate outlined by *Haeussler et al.* [2003]. Furthermore, a ridge segment is easily reconstructed near Yellowstone hotspot at about 57 Ma (Figure 6). Migration of the ridge axis away from the hotspot could be at a speed of about 50 km/Myr, leading to several million years of decreasing but significant influence by the hotspot on volume and geochemistry of the axial basalts. A reorganization of the subduction boundary along the western edge of North America at about 48 Ma could coherently transfer the thick oceanic crust formed at this spreading segment to greater North America at about the location of the current Oregon and Washington Coast Ranges. A prolonged subduction reorganization during establishment of Juan de Fuca subduction beneath the Siletz microplate could explain rotation of the older elements of the Oregon Coast Range (i.e., Siletz River Volcanics) within the forearc as documented by *Simpson and Cox* [1977] and *Wells and Heller* [1988].

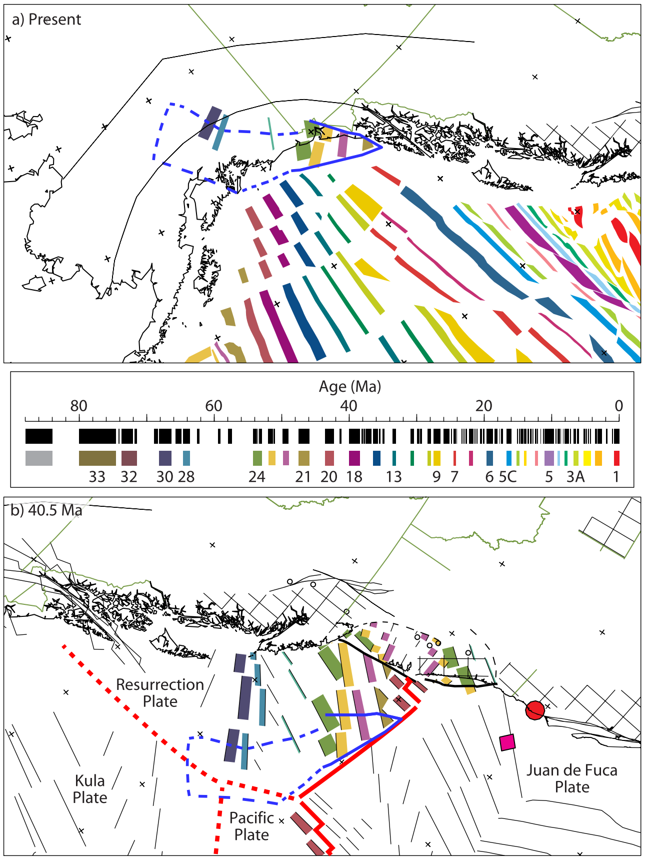


Figure 7. Schematic maps depicting the origin of the Yakutat terrane mafic basement as a captured remnant of the Resurrection plate. Present geometry (a) includes outline of Yakutat terrane (bold blue line) and its subducted continuation as inferred by *Eberhart-Phillips et al*. [2006]. Yakutat isochrons are from 40.5 Ma model (b), assuming both Kula and Resurrection plates were captured by the Pacific plate at 40.5 Ma (dying boundaries dashed red lines) and have moved rigidly with Pacific plate since then. This simple model provides an explanation for the similarity of age, thickness, and geochemistry between Yakutat dredged basalts and Crescent Formation lavas [*Davis and Plafker*, 1986]. (See Figure 5 for description of other symbols.).

[57] We recognize three Eocene to Oligocene magmatic provinces in the Pacific Northwest: (1) accreted oceanic basement and affiliated seamounts (Siletz-Crescent); (2) younger, MORB-affinity lavas built on basement (Tillamook and Grays River); (3) post accretion Cascade arc volcanics (Goble and other, more inland suites). The closest modern analog for the Siletz-Crescent terrane may be the Cocos and Carnegie ridges associated with the Galápagos hotspot, with the frequently alkalic and OIB-affinity lavas of the Galápagos Islands [e.g., *White et al*., 1993] corresponding to the Siletz River Volcanics of Oregon and the more tholeiitic, MORB-affinity Cocos Ridge [e.g., *Verma and Schilling*, 1982] corresponding to the Crescent and Metchosin Formations. Our forward model shows the hotspot tending to remain under the Farallon/Juan de Fuca plate (Animations S1 and S2), much as the Galápagos hotspot has recently stayed beneath the Nazca plate [e.g., *Wilson and Hey*, 1995].

[58] Intrusive and volcanic rocks north of the Metchosin Igneous Complex such as the 51–49 Ma Clayoquot intrusions of *Madsen et al*. [2006] and Flores volcanics of *Irving and Brandon* [1990] (Figure 1 and Table 1), and metamorphism of the Leech River accretionary complex ca. 51 Ma [*Groome et al*., 2003; *Madsen et al*., 2006] have been attributed to heating during ridge subduction [*Breitsprecher et al*., 2003]. Our model suggests a minor refinement of Breitsprecher et al.'s interpretation, in that these intrusive and volcanic rocks formed in a slab window at the continuation of the ridge between the Resurrection plate and Siletz microplate.

[59] Volcanic rocks younger than ~45 Ma require a separate mechanism than a full mid-ocean ridge segment. We model the Juan de Fuca and Resurrection plates as continuing to separate under the margin near the present-day Columbia River subsequent to docking of the Siletz microplate. *Chan et al*. [2012] presented detailed evidence that the Grays River volcanics (and by analogy, Tillamook Volcanics) formed at a hotspot-influenced slab window. Our model indicates no difficulty in having the Juan de Fuca-Resurrection spreading ridge reach the continent and form a slab window in the area of the modern Columbia River for the relevant time around 40 Ma. As the Juan de Fuca plate moves northeastward, the slab window would move far under the continent and subduction would be reestablished in the area no later than about 35 Ma, consistent with occurrence of the arc-affinity Goble Volcanics [*Phillips et al*., 1989].

[60] Although not part of the original motivation for our model, the similarity between Yakutat basalts and Crescent basalts noted by *Davis and Plafker* [1986] is potentially explained by their having previously been adjacent parts of the Resurrection plate (Figure 7). The separation distance is large enough that, if previously adjacent, they must have been moving apart for most of the time since their formation. Under the simple model that the Resurrection plate was captured by the Pacific plate at the same time as the Kula plate at ~40 Ma [*Lonsdale*, 1988a], the distance between Yakutat and Crescent may have been only a few hundred kilometers at 50 Ma, with both formed near the Yellowstone hotspot.

4.2. Model Nonuniqueness

[61] The proposed plate reconstruction is inherently nonunique because the entire Juan de Fuca-Resurrection ridge system appears to have been subducted—except for the Siletz-Cresecent and perhaps the Yakutat and Resurrection Peninsula fragments. Nonetheless, we have constructed a quantitative kinematic model based on the most recent geologic and plate tectonic constraints. This model is intended as a test of whether a previously discarded tectonic model might in fact be plausible. We are able to fit essentially all available constraints, and point out below where further revisions may be appropriate. Certainly, even if the Resurrection plate existed, details of its motion will remain obscure for the foreseeable future. We emphasize, however, that a nonunique model is capable of refuting the argument that plate motion data are inconsistent with forming the Siletz-Crescent terrane at a near-ridge hotspot.

4.3. Subduction Zone Jump and Slab Remnants

[62] *Hyndman et al*. [1990] interpreted the Tofino Basin, an elongate trough filled with Eocene sedimentary strata beneath the present-day continental slope off Vancouver Island, to be a relict subduction trench. This interpretation implies that the subduction boundary was abandoned when the trench jumped

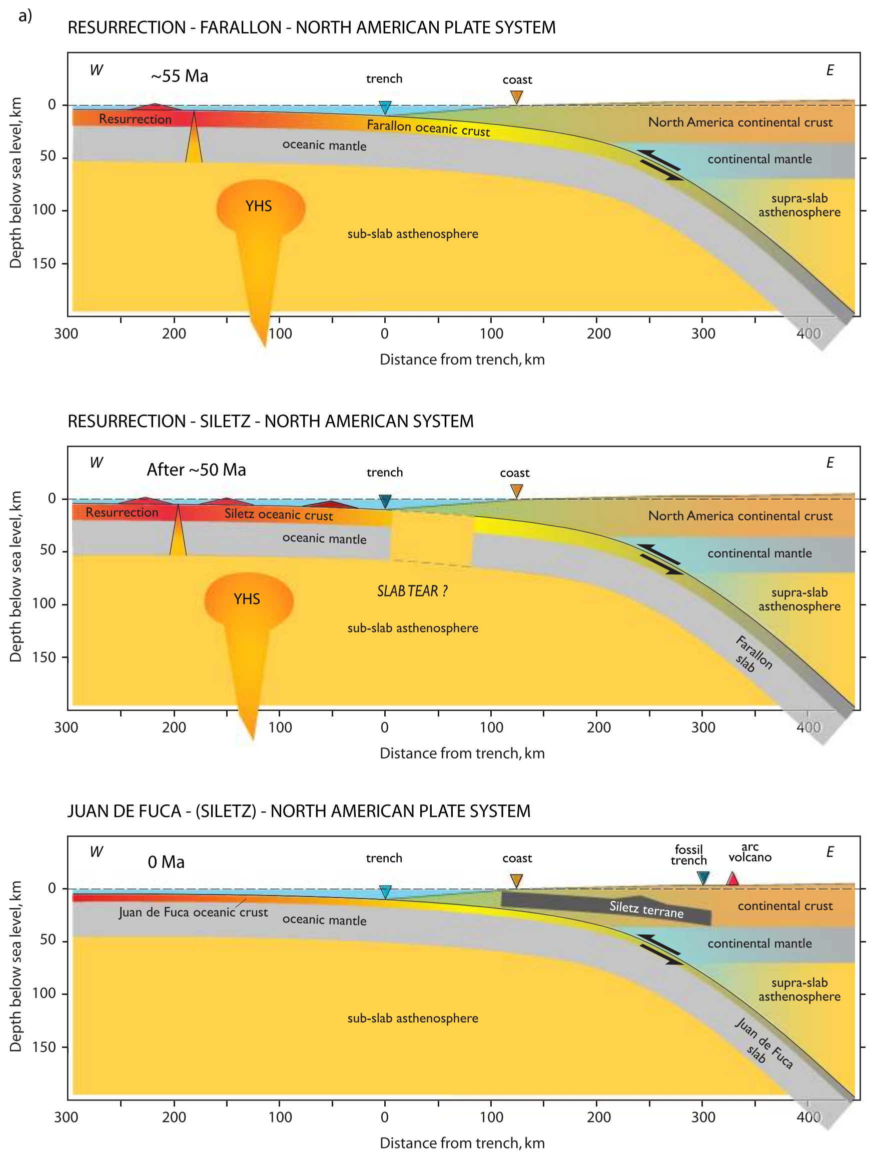


Figure 8. (a) Schematic cross-section showing break-off and accretion of Siletz microplate (derived from Farallon/Juan de Fuca plate. (b) Schematic cross-section showing break-off and accretion of Crescent microplate (derived from Kula/Resurrection plate). See Figure 1 for approximate locations of these profiles.

about 50 km westward [*Hyndman et al*., 1990] to its current location. This postulated trench jump occurred about 48 Ma, during the same time period in which we model the Crescent microplate separating from the Resurrection plate and beginning to accrete to the North American plate (~48 to 40 Ma). Petroleum boreholes drilled into Tofino Basin strata bottom in volcaniclastic or volcanic rocks attributed to the Crescent terrane [*Shouldice*, 1971; *Narayan et al*., 2005]. Paleobathymetry data from the boreholes indicate the Crescent rocks were uplifted as much as 5 km from marine bathyal depths in late Eocene time [*Shouldice*, 1971; *Clowes et al*., 1987; *Hyndman*, 1995; *Narayan et al*., 2005] after the trench jump, followed by about 5 km of subsidence [*Hyndman*, 1995] down to their current bathyal depth. As discussed above, we interpret these Crescent rocks to represent a detached fragment of the former Resurrection plate.

[63] The Crescent terrane is about 40 km wide beneath Vancouver Island [*Hyndman et al*., 1990], including the so-called E layer (which has a seismic velocity structure that mimics the dipping structure of the actively subducting Juan de Fuca plate in the subsurface). We postulate that when this fragment detached from the Resurrection plate and stalled beneath the margin (Figure 8b), the microplate separated from the Resurrection plate both along its up-dip and down-dip edges, with the down-dip, less buoyant Resurrection slab continuing to subduct. We do not model a ridge segment in the vicinity during the associated thermal event, but rather suggest that the edge of the Siletz-Resurrection slab window had reached Vancouver Island by this time. The thermal event was followed by rapid cooling in the middle Eocene, which *Groome et al*. [2003] attributed to onshore uplift and unroofing during continued underthrusting of the Crescent terrane beneath the Leech River Complex of *Fairchild and Cowan* [1982] along the Leech River fault. Cooling may also reflect sealing off of the postulated slab gap after the Resurrection plate reestablished a subduction geometry beneath the Crescent fragment.

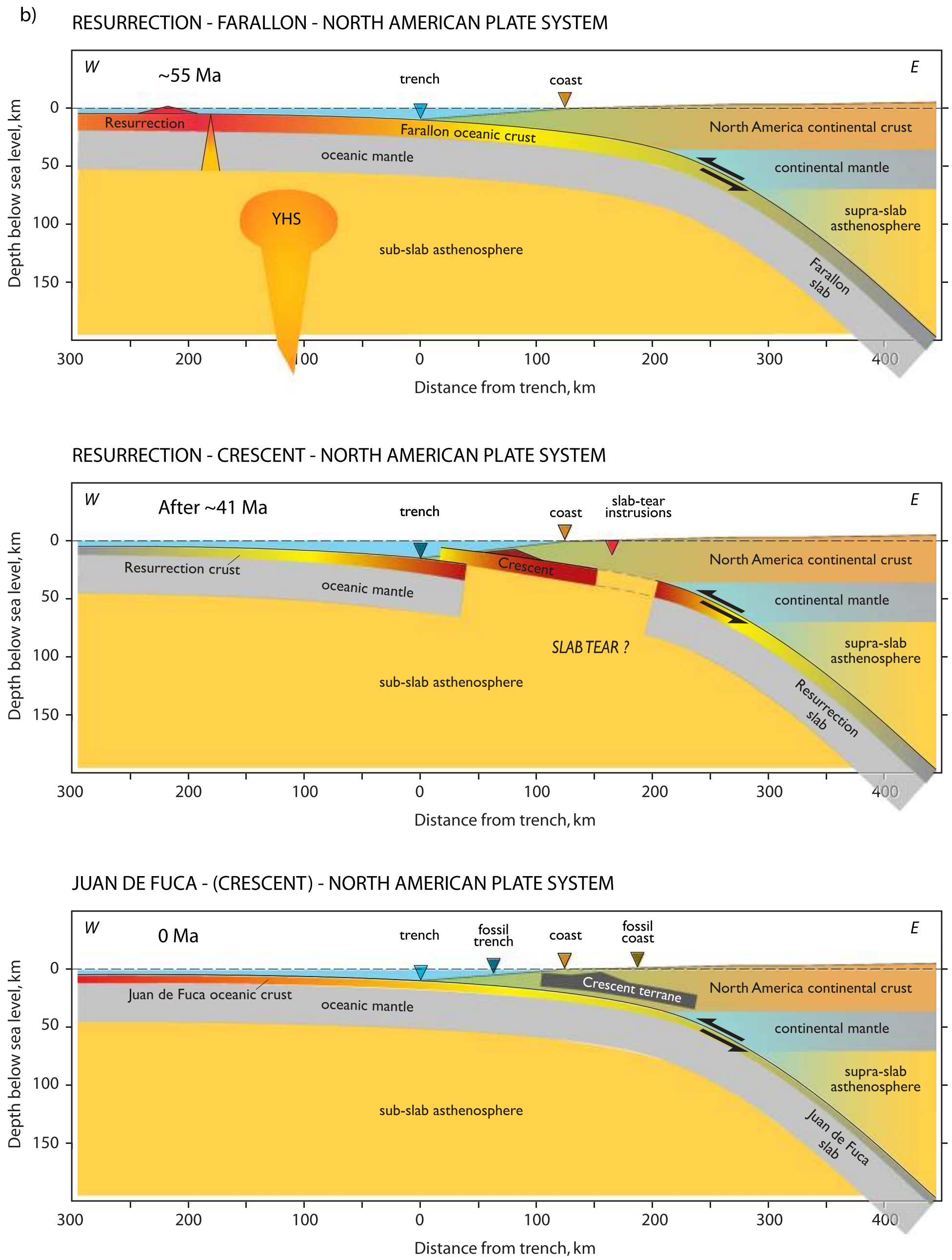


Figure 8. (continued)

[64] Another magmatic event occurred ~41 Ma, (i.e., Mt. Washington intrusions of *Isachsen* [1987]), perhaps during the final demise of the Resurrection plate. A second plate reorganization would have occurred when the subducting oceanic plate shifted from the Resurrection to the Juan de Fuca plate, this time establishing (ongoing) subduction of the Juan de Fuca plate beneath the Vancouver Island-Washington continental margin. This reorganization occurred more recently than our 65-40.5 Ma reconstruction, but likely before 35 Ma when Cascada arc volcanism initiated.

4.4. Hotspot Speculations

[65] A distinct change in terrane character occurs in the vicinity of Willapa Bay, Washington, corresponding to our location of the boundary between the Crescent (Resurrection-derived) and the Siletz (Farallon-derived) microplates. Basalt exposures on Vancouver Island and around the lower part of the Olympic Peninsula are modified by low-grade metamorphism [*Duncan*, 1982] to zeolite and pumpellite-prenhite facies [*Parsons et al.*, 1999], implying shallow burial (<10 km) and subsequent unroofing. We interpret evidence of shallow burial to indicate partial Crescent microplate subduction beneath the continental margin (Figure 8b); however, metamorphism may also have resulted from thermal events associated with subsequent intrusive episodes discussed above. Regardless, after the Crescent microplate broke from the Resurrection plate and its subduction stalled, a new Resurrection trench formed seaward of the Crescent fragment, preserving the former subduction geometry, consistent with observed subsurface relations beneath southwestern Vancouver Island and southwestern Washington.

[66] In contrast, the Siletz microplate lacks a metamorphic signature of shallow burial, and instead appears to have docked against the continental margin (Figure 8a), which then subsided, shifting from a subaerial to a marine bathyal environment [e.g., *Wells et al*., 2000], before being uplifted and partially unroofed by erosional processes. The former trench associated with the Farallon plate may be represented by the Wildlife Safari fault in southwestern Oregon [*Wells et al*., 2000], but otherwise is presumed to be buried beneath the eastern edge of the Willamette Valley [*Madsen et al*., 2006] or under the modern arc [*Wells et al*., 1998], in other words, arc-ward of the accreted microplate fragment.

[67] Our kinematic reconstruction predicts that the older portions of both accreted microplates (i.e., northernmost Crescent and southernmost Siletz) should be thicker than the younger portions owing to their closer proximity to the Yellowstone hotspot during formation. Although available thickness data are not well resolved (Table S1) and may not represent original thicknesses, the Resurrection-derived fragment generally seems to be only somewhat thicker than typical oceanic crust, 6–10 km, with up to another 5 km of seamount or ocean island basalts superimposed on top in discrete areas. In contrast, the Siletz-derived fragment is more than three times as thick as typical oceanic crust (generally >20 km) based on subsurface estimates. These differences in crustal thickness may have contributed to differences in inferred accretion processes, with stalled subduction in the north and lateral docking in the south.

5. Summary

[68] Our kinematic model depicting motion of the Kula, Farallon, and Juan de Fuca plates relative to the Pacific plate is based on magnetic isochrons on the Pacific plate, slightly updated from previous work. Pacific-North America motion is derived from the global circuit, again slightly updated from previous work. The position of the Paleogene Yellowstone hotspot is assumed to be fixed to the Pacific hotspot frame of *Lonsdale* [1988b] with adjustments for revised dating.

[69] By explicitly including the defunct Resurrection plate in our Paleogene reconstruction of the western North America plate boundary, we are able to account for the origin of the enigmatic Siletz and Crescent basement terranes. These terranes are remnants of a Juan de Fuca-Resurrection ridge segment that accreted to the North America subduction margin from about 53 to 40 Ma. The ridge segment accreted sequentially, first an intact Siletz block in the south, followed by an intact Crescent block to the north.

[70] The successful tectonic evolution of the Siletz-Crescent terrane in our model supports the existence and development of the Resurrection plate. In addition, our model supports the viability of a long-lived Yellowstone hotspot, originating prior to the Miocene Columbia flood basalts. We attribute subsequent volcanism between Tillamook, Oregon and Willapa Hills, Washington, starting ~44 Ma, to the formation of a slab window that resulted from continued divergence of the Resurrection and Juan de Fuca plates beneath the continental margin.

[71] We envision future revisions to the kinematic model as new geologic, radiometric, and paleomagnetic constraints become available. For example, rotation of the 44–41 Ma Tillamook Volcanics [*Magill et al*., 1981] is not included in the present model. This rotation may represent in situ structurally controlled rotation during an extensional forearc phase. We encourage testing of kinematic model predictions such as the basic implication that the northern and southern portions of the basaltic forearc terrane are derived from different oceanic plates.

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