

Geosphere

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Geosphere 2006;2;11-34
doi: 10.1130/GES00020.1

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Notes

Cenozoic to Recent plate configurations in the Pacific Basin: Ridge subduction and slab window magmatism in western North America

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ABSTRACT

Forearc magmatic rocks were emplaced in a semicontinuous belt from Alaska to Oregon from 62 to 11 Ma. U-Pb and ⁴⁰Ar-³⁹Ar dating indicates that the magmatism was concurrent in widely separated areas. Eight new conventional isotope dilution-thermal ionization mass spectrometry (ID-TIMS) U-Pb zircon ages from forearc intrusions on Vancouver Island (51.2 ± 0.4, 48.8 ± 0.5 Ma, 38.6 ± 0.1, 38.6 ± 0.2, 37.4 ± 0.2, 36.9 ± 0.2, 35.4 ± 0.2, and 35.3 ± 0.3 Ma), together with previous dates, indicate that southwestern British Columbia was a particularly active part of the forearc. The forearc magmatic belt has been largely attributed to ridge-trench intersection and slab window formation involving subduction of the Kula-Farallon ridge. Integration of the new and previous ages reveals shortcomings of the Kula-Farallon ridge explanation, and supports the hypothesis of two additional plates, the Resurrection plate (recently proposed) and the Eshamy plate (introduced herein) in the Pacific basin during Paleocene and Eocene time. We present a quantitative geometric plate-tectonic model that was constructed from 53 Ma to present to best account for the forearc magmatic record using ridge-trench intersection and slab window formation as the main causes of magmatism. The model is also in accord with Tertiary to present inboard magmatic and structural features.

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Keywords: tectonics, magmatism, geochronology, forearc, slab window, ridge subduction, western North America, Cordillera.

INTRODUCTION

Forearcs are typically amagmatic with low heat flow (Gill, 1981); however, subduction of a mid-ocean ridge imparts a thermal pulse into the forearc, which may result in near-trench magmatism (Marshak and Karig, 1977; DeLong et al., 1979; Sisson et al., 2003). As a ridge subducts, magmatic accretion ceases along the ridge axis and a gap called a slab window forms between the subducted parts of the two downgoing oceanic plates (Dickinson and Snyder, 1979; Thorkelson, 1996). Subducting ridges and the resulting slab windows have been linked to high heat flow, anomalous magmatism, and deformation in the overriding plate from forearc to backarc (Dickinson and Snyder, 1979; Hibbard and Karig, 1990; Barker et al., 1992; Sisson and Pavlis, 1993; Pavlis and Sisson, 1995; Kusky et al., 1997; Thorkelson, 1996; Lytwyn et al., 1997; Breitsprecher et al., 2003; Groome et al., 2003). Forearc areas affected by high heat flow and igneous activity are likely to have experienced ridge subduction and migration of either a ridge-trench-trench or ridge-trench-transform triple junction (DeLong et al., 1979; Dickinson and Snyder, 1979; Thorkelson, 1996; Lewis et al., 1997).

The Cenozoic subduction zone of western North America preserves forearc magmatic activity within a semicontinuous belt from Alaska southeastward to the coastal areas of British Columbia, Washington, and Oregon (Fig. 1). This chain of igneous rocks is spatially

and temporally complex and spans Paleocene to Miocene time. The most spatially and temporally coherent portion is the eastward-younging Sanak-Baranof Belt in southern to southeastern Alaska (Bradley et al., 1993; Haeussler et al., 1995; Bradley et al., 2003). The age progression has been attributed to the passage of an eastwardly migrating ridge-trench-trench triple junction related to the subduction of a mid-ocean spreading ridge in Paleocene to middle Eocene time (Hill et al., 1981; Bradley et al., 1993; Sisson and Pavlis, 1993). During this interval, forearc magmatism was also recorded farther south, on Vancouver Island and along the coasts of Washington and Oregon (Wells et al., 1984; Babcock et al., 1992; Groome et al., 2003). Subsequently, in late Eocene to Miocene time, forearc magmatism occurred more sporadically from Oregon to Alaska (Barnes and Barnes, 1992; Hamilton and Dostal, 2001; Kusky et al., 2003). Virtually all of these magmatic events have been regarded, by many workers, as products of ridge subduction (e.g., Babcock et al., 1992; Barnes and Barnes, 1992; Bradley et al., 1993; Sisson and Pavlis, 1993; Harris et al., 1996; Hamilton and Dostal, 2001; Groome et al., 2003; Kusky et al., 2003), although other tectonomagmatic environments, such as subduction-related volcanic arcs, rifted forearcs, leaky transforms, ocean islands, and mantle plumes, have also been proposed (Tysdal et al., 1977; Massey, 1986; Clowes et al., 1987; Massey and Armstrong, 1989; Babcock et al., 1992; Davis et al., 1995; Wells et al., 1984).

In this paper, we examine three suites of felsic forearc igneous rocks on Vancouver Island: the Mount Washington intrusions, Clayoquot intrusions, and Flores volcanics. Eight new U-Pb dates define two pulses of magmatism.

By integrating these findings with existing data from coastal areas of British Columbia, we can clarify the role of Vancouver Island in the history of near-trench magmatism from Alaska to Oregon. We also present a plate-tectonic model that supports and elaborates on the hypothesis of the Resurrection plate (Haeussler et al., 2003a) and includes the formation of another plate, the Eshamy plate, in the Pacific Basin. Plate boundaries have been selected to best fit the complicated pattern of forearc magmatism

in the northern Cordilleran subduction zone from 53 Ma to present.

SYNOPSIS OF CENOZOIC TO RECENT FOREARC MAGMATISM, ALASKA TO OREGON

Paleogene forearc magmatism forms a complex spatial and temporal pattern within the forearc areas of Alaska, British Columbia, Washington, and Oregon (Fig. 2; Table 1).

Magmatism occurred in two main pulses during the (1) Paleocene–early Eocene, and (2) late Eocene–Oligocene. The first pulse is represented in Alaska by intrusions of the Sanak-Baranof Belt, in British Columbia by the Walker Creek intrusions (50.9–50.7 Ma U-Pb), Metchosin Igneous Complex, Clayoquot intrusions (51.2–48.8 Ma U-Pb), and the Flores volcanic rocks (51.2–50.5 Ma U-Pb), all of which occur on Vancouver Island, and in Washington and Oregon by the Coast Range Basalt Province

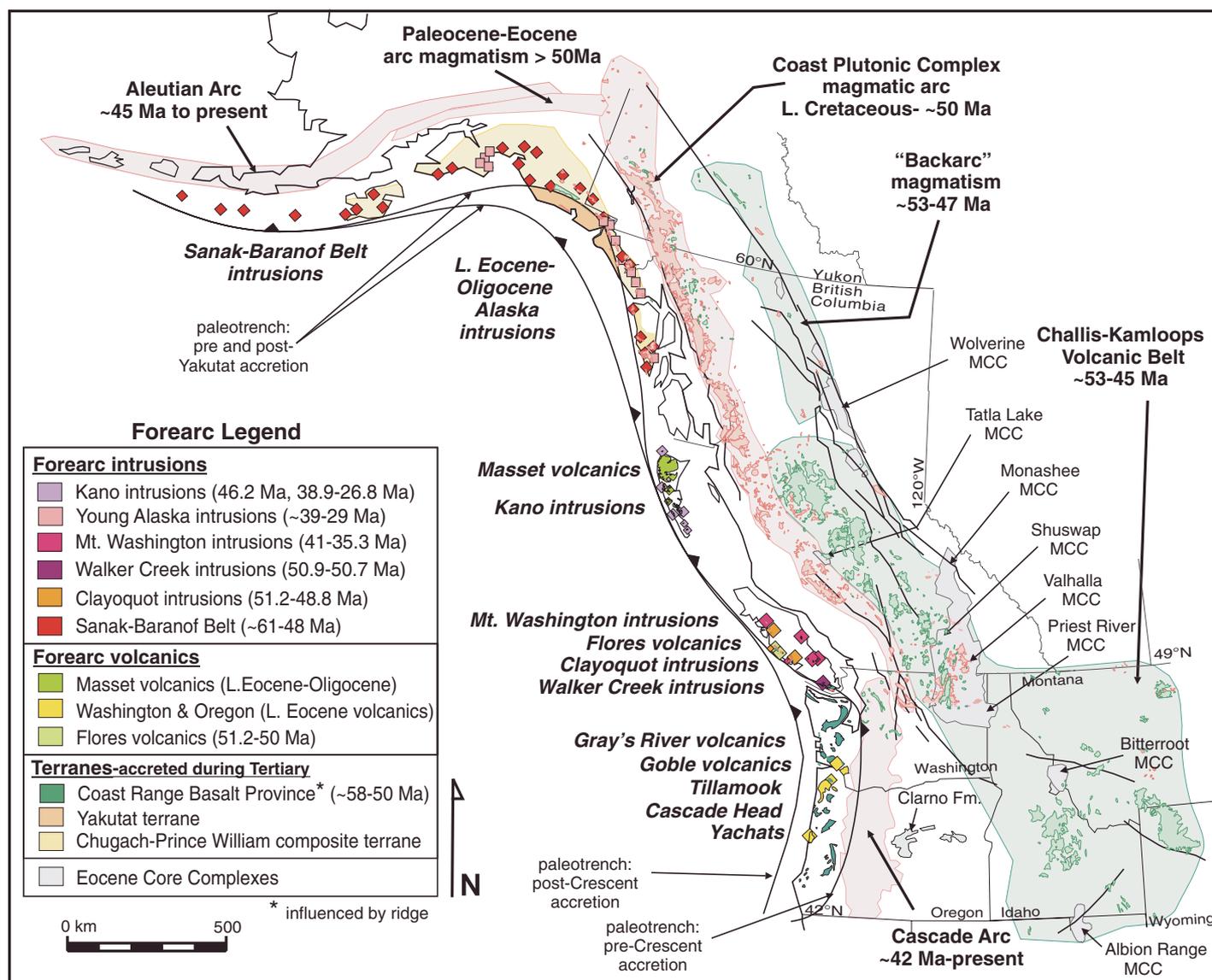


Figure 1. Paleocene to Oligocene forearc, arc, and backarc magmatism of the Pacific Northwest, USA. Also shown are the Coast Range Basalt Province, and the Chugach and Yakutat terranes, which accreted to the forearc area during Tertiary time. Arc magmatism is depicted in the pink fields and the Challis-Kamloops Belt and other Eocene extended-arc to backarc magmatism in the light-green fields. Within these fields, the intrusions are pink and volcanic rocks are green. Shown in light gray are selected metamorphic core complexes exhumed during Eocene time. MCC—metamorphic core complex. Where plutons and volcanic exposures are small, they are depicted as diamonds or squares. Major strike-slip faults are shown as thin black lines.

(ca. 58–50 Ma; Duncan, 1982). The second pulse of forearc magmatism is represented in Alaska by intrusions and volcanics in the Prince William Sound and St. Elias Mountain areas from ca. 42 to 30 Ma (Bradley et al., 1993; Kusky et al., 2003), and the Admiralty Island area from ca. 35 to 5 Ma (Ford et al., 1996). In British Columbia, the second pulse is represented by centers on both the Queen Charlotte Islands and Vancouver Island. On the Queen Charlotte Islands, the Kano plutonic suite ranges from 46 to 26 Ma U-Pb (Anderson and Reichenbach, 1991) and is considered coeval in part with the Masset volcanics, which may have persisted into Miocene time (46–11 Ma K-Ar whole rock; Hickson, 1991; Anderson and Reichenbach, 1991; Hamilton and Dostal, 1993). On Vancouver Island, the second pulse is represented by the Mount Washington intrusions, herein dated at 41–35.3 Ma by U-Pb zircon. To the south, the second pulse is represented by the Gray's River and Goble volcanics in Washington and the broadly synchronous Tillamook, Cascade Head, Cannery Hills, and Yachats basalt successions in

Oregon (Barnes and Barnes, 1992; Davis et al., 1995; ca. 44–34 Ma ^{40}Ar - ^{39}Ar).

Forearc Magmatism on Vancouver Island

Middle to Late Eocene

Paleogene forearc igneous rocks on Vancouver Island were the focus of this study (Fig. 3) and include from oldest to youngest: the accreted Metchosin Igneous Complex, the Walker Creek intrusions, Flores volcanic rocks, Clayoquot intrusions, and the Mount Washington intrusions.

The Metchosin Igneous Complex is correlative with the Coast Range Basalt Province (Massey, 1986) and has been interpreted as a partial ophiolite generated in an extensional setting (Massey, 1986). The Metchosin Igneous Complex was still forming at 52 ± 2 Ma based on a U-Pb zircon age from hornblende diorite (Yorath et al., 1999). The Walker Creek intrusions comprise a middle Eocene peraluminous tonalite-trondhjemite suite, which was emplaced within the Leech River Complex, a metamorphosed Cretaceous accretionary package juxtaposed against

the Wrangellian portion of Vancouver Island along the San Juan fault (Groome et al., 2003). The Walker Creek intrusions are interpreted to be products of forearc melts mixed with mid-ocean-ridge basalt (MORB) magmas during ridge-trench interaction (Groome et al., 2003). The Metchosin Igneous Complex accreted to the southwestern tip of Vancouver Island sometime during the middle Eocene, possibly as early as 50 Ma, causing deformation and rapid exhumation of the Leech River Complex from depths of ~10 km, as evidenced by rapid cooling of the Walker Creek intrusions (Groome et al., 2003).

North of the Leech River Complex, Eocene forearc igneous rocks on Vancouver Island include the Flores volcanic rocks and the Clayoquot and Mount Washington intrusions. The Flores volcanics are a suite of bimodal, subaerial, columnar to massive flows, debris-flow breccias, and ash-flow tuffs crosscut by undated felsic dikes (Irving and Brandon, 1990). The volcanic rocks are dominated by dacite but range from basaltic andesite to rhyolite in composition, with a SiO_2 gap between

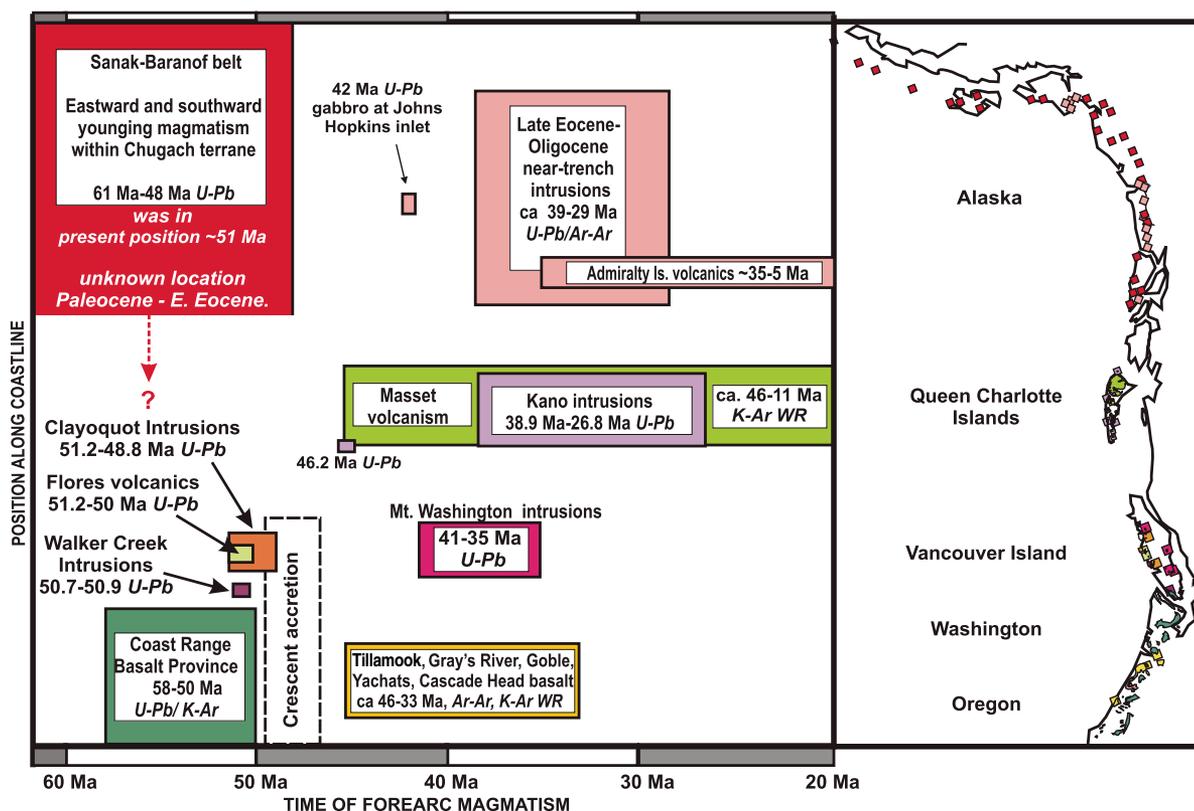


Figure 2. Tertiary forearc magmatism of coastal western North America from 61 Ma to 20 Ma. The vertical axis corresponds to geographic location and extent of magmatism along the coastline at right. The horizontal axis corresponds to time in Ma. Note synchronicity of forearc magmatism at different times in widely separated positions along the coast. Geochronology methods used to constrain timing of magmatism are shown in italics. Color of boxes matches specific locations of plutons depicted on the coastline. Alaska oroclinal bending occurred between 66 and 44 Ma (Hillhouse and Coe, 1994).

TABLE 1: SUMMARY OF FOREARC IGNEOUS ROCK SUITES FROM ALASKA TO OREGON INCLUDING AGE RANGES, ROCK TYPES, AND SUMMARIES OF PREVIOUSLY PUBLISHED INTERPRETATIONS

Locality	Name of forearc magmatic suite	Notes	Summary of previously published interpretations	Rock type	Age (Ma)	Dating method(s)	Age references
Alaska							
Sanak Island to Baranof Island	Sanak-Baranof Belt	Eastward younging well-defined age progression of forearc intrusions from 63 Ma on Sanak Island to 48 Ma on Baranof Island	Geochemical-isotopic and structural studies consistent with triple junction migration and slab window formation (Hill et al., 1981; Barker et al., 1992; Bradley and Kusky, 1992; Harris et al., 1996; Pavis and Sisson, 1995; Lyrwijn et al., 1997, 2000)	Tonalite, granodiorite, granite, trondhjemite, gabbroic plutons	63–48	Mostly U-Pb and Ar-Ar	Haeussler et al., 1995; Bradley et al., 1993, 2000
Prince William Sound, Glacier Bay, John's Hopkins Inlet	Miner's Bay pluton, Nellie Juan pluton, Terentiev (gabbro)	These plutons are part of small magmatic suite, not yet studied in detail	Have been attributed to ridge-trench intersection enabled by a jump-back of the Resurrection-Kula ridge during plate reorganization event at ca 40 Ma (Kusky et al., 2003)	Quartz-feldspar porphyry, biotite-hornblende granite, gabbro, granite	29.2–42	K/Ar biotite, Ar-Ar U-Pb	Nelson et al., 1999; L. Snee, 1992, personal commun., in Kusky et al., 2003; Sisson et al., 2003
Tkope-Portland Peninsula, southeast Alaska	Tkope-Portland Peninsula magmatic belt	These two suites are synchronous and range from 35–5 Ma	Based on chemical arguments, Ford and Brew (1996) attributed Admiralty Island suite to asthenospheric upwelling in a slab window environment	Felsic to intermediate Basalt to andesite	35–5	K-Ar	Pfaffner et al., 1994
Admiralty Island, Alaska	Admiralty Island volcanics						
British Columbia							
Queen Charlotte Islands (Haida Gwaii)	Kano intrusions	Kano intrusions form well-dated northward younging plutonic suite, coeval with Masset volcanism and Tertiary dyke swarms (Southern and Jessop, 1991; Anderson and Reichenbach, 1991; Hamilton and Dostal, 1993)	Hamilton and Dostal (2001) used chemical and tectonic arguments to ascribe the Masset volcanics to heterogeneous asthenospheric melting in a slab window setting	Monzodiorite, granodiorite, diorite	46.2–26.8, younging from south to north	U-Pb zircon	Anderson and Reichenbach, 1991
	Masset volcanics and related dyke swarms			Mafic to felsic bimodal suite, dominated by basalt	46–11	K-Ar whole rock	Hyndman and Hamilton, 1993
Vancouver Island	Metchosin Igneous Complex	Accreted to southern tip of Vancouver Island as part of the Coast Range Basalt Province, possibly as early as 50 Ma (Groome et al., 2003)	Partial ophiolite developed in an extensional setting (Massey, 1986)	Basalt, gabbro, hornblende diorite	52 ± 2	U-Pb zircon	Yorath et al., 1999
	Walker Creek intrusions	Intruded into the Leech River Complex, southern Vancouver Island	The Walker Creek intrusions are interpreted to be products of forearc melts mixed with MORB magmas during ridge-trench interaction (Groome et al., 2003)	Tonalite, trondhjemite, and granodiorite	50.9 ± 0.6 to 50.7 ± 1.9	U-Pb monazite	Groome et al., 2003
	Flores volcanics	Erupted in a forearc position, coeval with the Clayoquot intrusions	Leaky transform volcanism (Massey and Armstrong, 1989)	Dominantly dacite, also basaltic andesite to rhyolite	51.2 ± 0.4 to 50.5 ± 0.5	U-Pb zircon	Irving and Brandon, 1990
	Clayoquot intrusions	Intrudes heterogeneous Wrangellia terrane which comprises most of Vancouver Island. Emplaced in forearc position	Forearc position considered anomalous	Tonalite, trondhjemite, and granodiorite with subordinate granite and quartz monzodiorite	48.8 ± 0.5 to 51.2 ± 0.5	U-Pb zircon	This publication
	Mt. Washinton intrusions				35.3 ± 0.3 to 41 ± 1	U-Pb zircon	This publication and Isachsen, 1987
Northwest US							
Western Washington, western Oregon	Coast Range Basalt Province	Formed offshore, not as part of the forearc. Large, thick, accreted oceanic plateau. Includes the Melchoshin Igneous Complex, Crescent terrane, Siletz terrane	Several hypotheses exist for the origin of the Coast Range Basalt Province. Most call for a combination of hotspot activity with influence from a mid-ocean ridge (eg., Wells et al., 1984; Babcock et al., 1992)	Dominantly basalt with interbedded nearshore continental sediments	62–49	K-Ar	Wells et al., 1984
Oregon	Tillamook basalt	Erupted onto continental margin, subsequent to accretion of the Coast Range Basalt Province	Davis et al. (1995) attributed these suites to extensional volcanism and rifting at ca. 42 Ma	Basalt	45–41.5	Ar-Ar, K-Ar	Magill et al., 1981; Haeussler et al., 2003
Southwest Washington	Goble and Gray's River volcanics	Erupted onto continental margin, subsequent to accretion of the Coast Range Basalt Province		Basalt	~40	K-Ar	Beck and Burr, 1979; Wells, 1981
Oregon	Yachats basalt, Cascade Head-Cannery Hills basalt	Erupted onto continental margin, subsequent to accretion of the Coast Range Basalt Province	Davis et al. (1995) attributed these suites to a later event, similar to the event which led to the Tillamook, Gray's River, and Goble basalts. Barnes and Barnes (1992) attributed the peculiar alkalic geochemistry of the Yachats and Cascade Hills basalts to melting in a ridge subduction-slab window environment	Basalt	~37–34	K-Ar, Ar-Ar	Snively and MacLeod, 1974; McElwee and Duncan, 1982; Barnes and Barnes, 1992; Davis et al., 1995

Note: Dated rocks from these suites were used to constrain the tectonic model. Note that most suites have been attributed to slab window effects by previous workers. MORB—mid-oceanic-ridge basalt

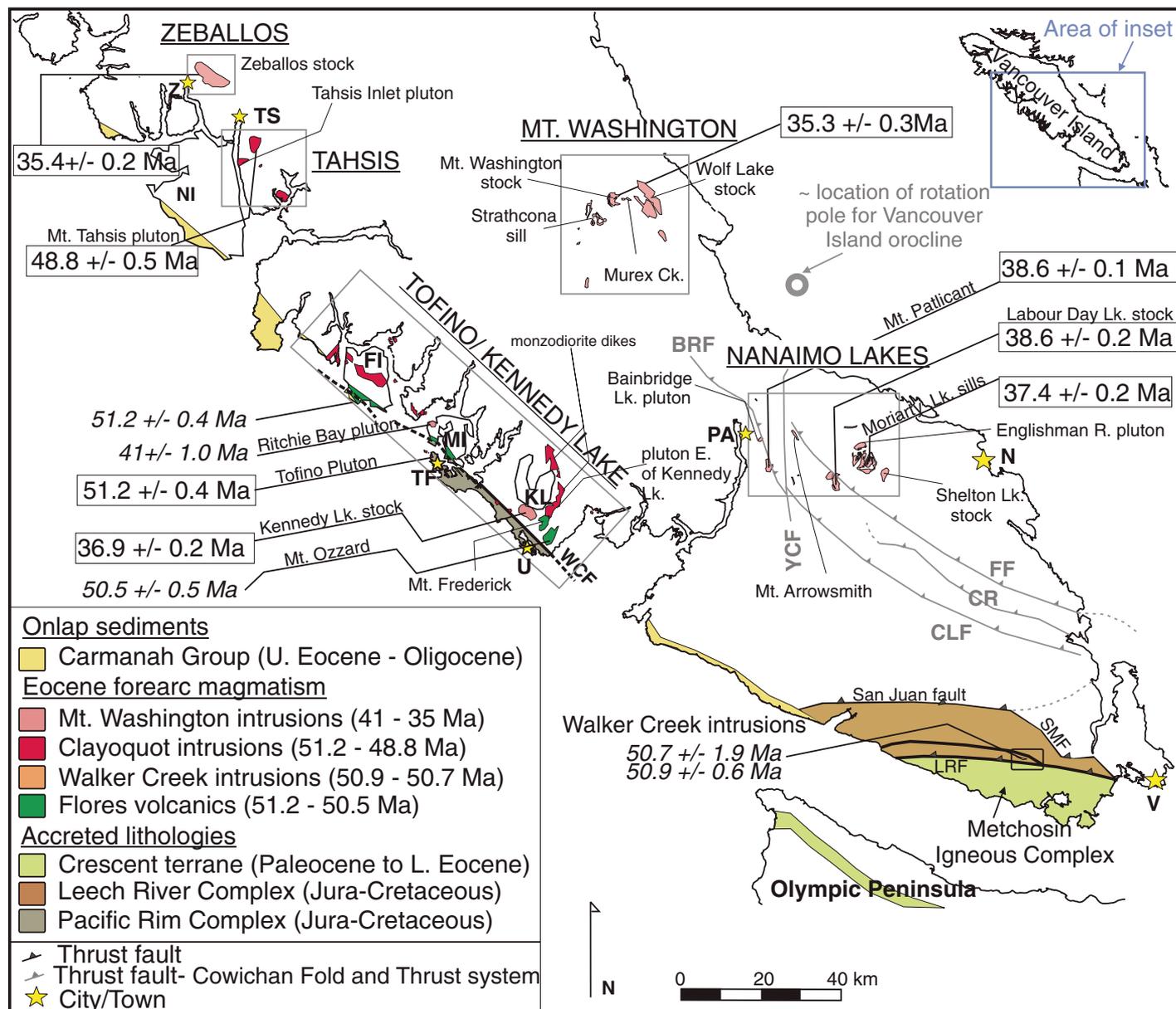


Figure 3. Location map of Eocene forearc magmatism, structures, and terranes that were accreted to Vancouver Island during Tertiary time. Eocene intrusions of the Mount Washington and Clayoquot suites and the Flores volcanics are shown in areal groupings that demonstrate similar intrusive styles, petrography, and geochemistry. U-Pb ages and pluton names are displayed for select intrusions visited in this study. New U-Pb dates obtained for this study are shown in boxes. Dates shown in italics are previously published ages. Structures: BRF—Beaufort range fault, FF—Fulford fault, YCF—Yellow Creek fault, CLF—Cowichan Lake fault, CR—Chemainus River fault, SMF—Survey Mountain fault, LRF—Leech River fault, WCF—West Coast fault. Also shown is the approximate location of the pole of oroclinal bending on southern Vancouver Island possibly related to crescent accretion. Structures shown in gray are related to the Cowichan fold-and-thrust system. Cities/towns: Z—Zeballos, TS—Tahsis, TF—Tofino, U—Ucluelet, PA—Port Alberni, V—Victoria, N—Nanaimo. Other abbreviations: KL—Kennedy Lake, MI—Meares Island, FI—Flores Island, NI—Nootka Island.

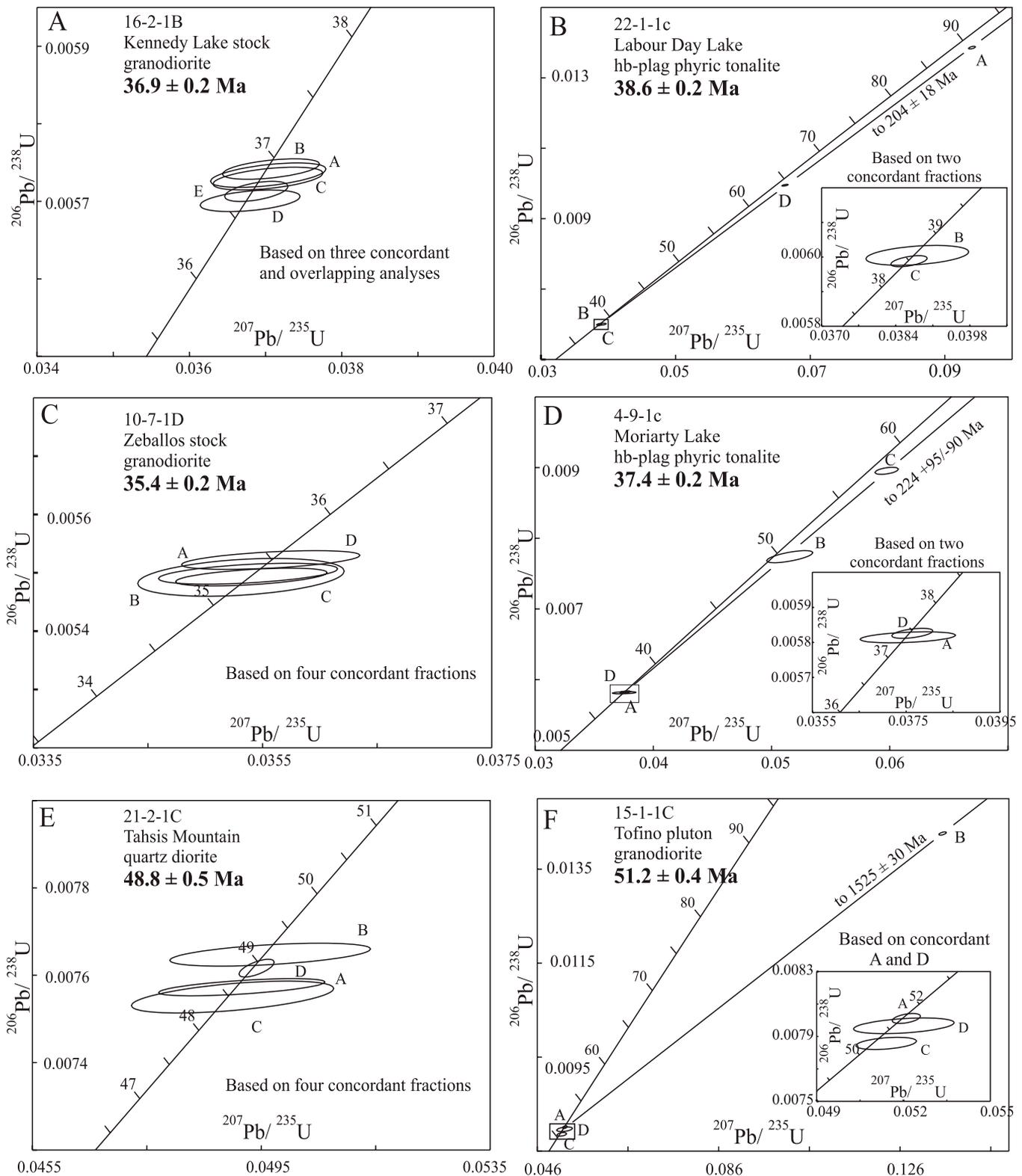


Figure 4. Concordia plots of zircon fractions used for U-Pb age determinations from the Pacific Centre for Isotopic and Geochemical Research (PCIGR). Insets show close-ups of ellipses and fractions used to make the age determination. Age of inheritance is provided on the divergent line pointing toward the upper intercept. **A:** Kennedy Lake stock. **B:** Labour Day Lake intrusion. **C:** Zeballos stock. **D:** Moriarty Lake sill complex. **E:** Tahsis Mountain. **F:** Tofino pluton. **B** and **D** show evidence of Jurassic inheritance, possibly from the island intrusions. **Sample F** intrudes the Pacific Rim Complex and demonstrates Precambrian inheritance. **A–D** belong to the Mount Washington Intrusive Suite. **E** and **F** are Clayoquot intrusions.

54 and 62 wt%. The mafic varieties are dominantly metaluminous, and the more felsic rocks are either metaluminous or peraluminous. The Mount Washington and Clayoquot suites are hypabyssal, dominantly hornblende-plagioclase phyrlic intrusions composed mainly of metaluminous tonalite, trondhjemite, and granodiorite, with subordinate granite and quartz monzodiorite. SiO₂ levels in the two suites range from 55 to 74 wt%, with an average of 65 wt%.

The Mount Washington and Clayoquot intrusive suites were formerly collectively known as the Catface Intrusions (Muller and Carson, 1969), but were divided into eastern and western belts based on geographic separation (Carson, 1973). The two belts were further refined and given names according to a temporal separation defined by K-Ar ages (Massey, 1995). In that definition, the Mount Washington intrusions were regarded as having been generated during a ca. 35–47 Ma event affecting the eastern periphery and northwestern portions of Vancouver Island. The Clayoquot intrusions were grouped as an older suite (ca. 45–67 Ma) on the west coast. The close proximity and crosscutting field relations of some Clayoquot intrusions with the Flores volcanics suggest that the two suites were coeval (Massey, 1995).

METHODS

U-Pb Methodology

Six plutons were dated at the Pacific Centre for Isotopic and Geochemical Research (PCIGR) in the Department of Earth and Ocean Sciences, University of British Columbia, employing conventional isotope dilution–thermal ionization mass spectrometry (ID-TIMS). Analytical techniques were provided in Friedman et al. (2001). Interpreted ages, zircon descriptions, isotopic systematics, and U-Pb analytical data are listed in Tables 2 and 3. Data are plotted on standard concordia diagrams, with error ellipses and discordia intercepts shown at the 2σ level of uncertainty (Fig. 4). Results for all samples processed at PCIGR include multiple concordant and overlapping results, allowing for all interpreted magmatic ages to be based upon ²⁰⁶Pb/²³⁸U data. Results for three of the samples indicate the presence of inherited Pb components. Upper intercepts of discordia lines fit through these data sets give estimates of the average ages of these inherited components in the analyzed grains. Two additional samples were dated at the Royal Ontario Museum using the conventional ID-TIMS U-Pb methods outlined by Krogh (1973, 1982) (Tables 2 and 4). These samples also give concordant and overlapping results, allowing igneous crystallization

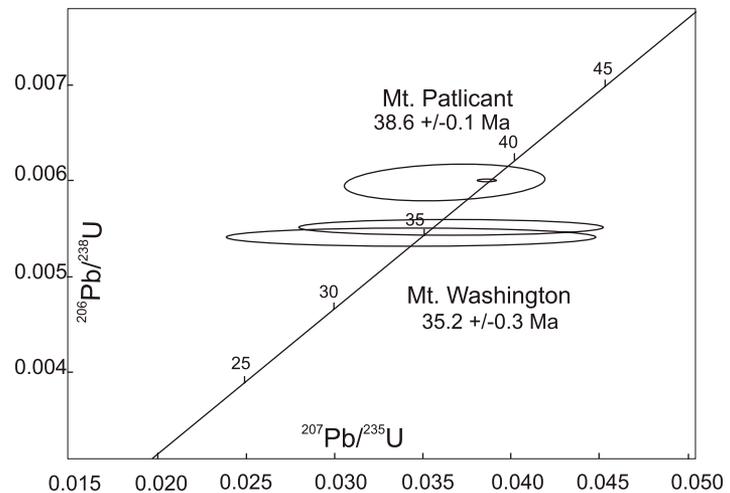


Figure 5. Concordia diagram for U-Pb zircon results from the Royal Ontario Museum. Mount Patlicant is located in the Nanaimo Lakes area and has an interpreted U-Pb age of 38.6 ± 0.1 Ma. Mount Washington stock is located in the Mount Washington area and is dated at 35.2 ± 0.3 Ma. Both are concordant.

ages to be determined primarily from high-precision ²⁰⁶Pb/²³⁸U data (Fig. 5).

New U-Pb Geochronology

Eight new U-Pb ages on zircon were obtained for plutons of the Clayoquot and Mount Washington intrusive suites (Tables 2, 3, and 4). These suites were selected for dating in order to define the two Tertiary magmatic pulses on Vancouver Island and to create a tightly constrained geochronological framework for Vancouver Island forearc magmatism. Tertiary ages had previously been assigned to these two suites based on stratigraphic arguments and K-Ar methods, which provided less-reliable geochronological constraints.

Two Clayoquot intrusions and six Mount Washington intrusions were dated (Fig. 3). Individual intrusions were selected for dating based on geography, the existence of previous age dates, and economic importance. Spatial distribution was important in order to obtain a representative suite of ages for the entire island. Ages were obtained from plutons that were previously dated by K-Ar methods under the hypothesis that cooling histories could be determined. It was discovered, however, that published K-Ar ages were unreliable, as most new U-Pb ages were younger than the K-Ar ages. The study included two stocks of economic importance, the Zeballos stock, and the Mount Washington stock, both of which host past-producing mines.

The new U-Pb zircon ages from this study supplant the older K-Ar dates and reveal that Massey's (1995) temporal division, which

divided the intrusive suites, requires minor revision. Plutons on Vancouver Island ranging in age from 51.2 to 48.8 Ma are herein regarded as the Clayoquot intrusions and are broadly coeval with the Flores volcanic rocks (51.2–50.5 Ma; Irving and Brandon, 1990). The Clayoquot intrusions are found only in the western areas of central Vancouver Island and on small islands off the west coast (e.g., Flores Island; Fig. 3). The Mount Washington intrusive suite ranges in age from 41.0 to 35.3 Ma and includes all Tertiary plutons within the eastern clusters, plus some intrusions on western Vancouver Island and small islands off the west coast (e.g., Meares Island; Fig. 3).

Knowledge of the timing of Tertiary forearc magmatism on Vancouver Island is critical, not only in the context of local tectonics and the geologic history of Vancouver Island and western Canada, but also the tectonic history of the northern Cordilleran subduction zone and the Pacific Basin. Vancouver Island is situated in a central position along the forearc of the subduction zone, directly south of the very well-constrained age trends in Alaska and Queen Charlotte Islands, and immediately north of less well-dated forearc magmatism in Washington and Oregon. When the age data from each region are viewed as a group, they reveal that synchronous forearc magmatism occurred in widely separated geographic areas of the subduction zone throughout the Tertiary (Fig. 2), more commonly than previously recognized. This has serious implications for the number of mid-ocean ridges intersecting the continental margin during the Tertiary, and hence the number of oceanic plates in the Pacific Basin. The

TABLE 2. ZIRCON DESCRIPTIONS, SYSTEMATICS, AND U-PB AGE INTERPRETATIONS FROM PCIGR

Sample #	Location notes	Location (NAD 83)		Split ¹ Size: b-axis (mm)	Description of selected zircon grains	Systematics and/or interpretation	Age and intrusive suite
		Eastings	Northing				
15-1-1c	Beach outcrop in Tonquin park, Tofino. Intrudes Pacific Rim complex. Tofino/Kennedy lakes area	10287440	5447955	N2, 149	Four fractions analyzed; orangish-pink, clear; most selected grains are elongate, (l/w ~4–8), euhedral, and contain tubes parallel to the c-axis.	Age based on mean ²⁰⁶ Pb/ ²³⁸ U date of two concordant/overlapping fracts; one slightly younger interpretation as Pb loss; error based on total overlap of these on concordia. Fraction B contains inheritance; ~1526 upper intercept for regression through all data; MSWD = 3.15.	51.2 ± 0.4 Clayoquot
21-2-1C	Pluton on NE side of Tahsis Mtn., in the Tahsis area, just south of Zeballos	9672953	5527575	N2, 149	Four fractions analyzed; pale pink, clear; l/w ~3–6; all selected grains have c-axis parallel tubes.	Age based on mean ²⁰⁶ Pb/ ²³⁸ U date and total overlap of four concordant analyses.	48.8 ± 0.5 Clayoquot
22-1-1C	Large stock located just west of Labour Day Lake, in the Nanaimo Lakes area	10393002	5442464	N5, 134	Four fractions analyzed; pale pink, clear; l/w ~3–6; square to slightly rounded x-sections; about 50% of selected grains contain c-axis parallel tubes.	Age based on mean ²⁰⁶ Pb/ ²³⁸ U for 2 concordant/overlapping analyses. Two with inheritance; regression through all data gives 204 ± 18 Ma upper intercept. MSWD = 0.79.	38.6 ± 0.2 Mt. Washington
4-9-1C	Sill complex located just northeast of Moriarty Lake in the Nanaimo Lakes area	10400092	5443048	N2, 149	Four fractions analyzed; pale tan, clear; l/w 3–6, most with c-axis tubes.	Age based on mean ²⁰⁶ Pb/ ²³⁸ U for 2 concordant/overlapping analyses. Two with inheritance; regression through all data gives 224 + 95/-90 Ma upper intercept. MSWD = 0.10.	37.4 ± 0.2 Mt. Washington
16-2-1C	Pluton located just south of Kennedy Lake in Tofino–Kennedy Lake area	10312005	5432461	N2, 149	Five fractions analyzed; amber to tannish, clear, l/w 4–8; Selected grains are elongate prisms and laths.	All concordant; although two are slightly younger, with possible very minor Pb loss. Age based on mean ²⁰⁶ Pb/ ²³⁸ U date of three concordant and overlapping analyses.	36.9 ± 0.2 Mt. Washington
10-7-1D	Sample taken from Zeballos stock, NW Vancouver Island	9657889	5544765	N1, 134	Four fractions analyzed; vivid pink, clear; moderate to elongate prisms (l/w 3–6) and rare stubby multifaceted grains.	All concordant; age based on mean ²⁰⁶ Pb/ ²³⁸ U and total overlap on concordia.	35.4 ± 0.2 Mt. Washington
24-5-1D*	Sample taken from Mt. Washington, NE Vancouver Island	10334683	5515188		Two fractions analyzed; elongate laths to short prisms	Age based on mean ²⁰⁶ Pb/ ²³⁸ U and total overlap on concordia.	35.2 ± 0.2 Mt. Washington
8-3-1G*	Sample taken from Mt. Patilicant, Port Alberni area	10375032	5445870		Two fractions analyzed; elongate prisms to stubby to round anhedral grains	All concordant; age based on mean ²⁰⁶ Pb/ ²³⁸ U and total overlap on concordia.	36.6 ± 0.6 Mt. Washington

¹N2: Zircons nonmagnetic on Franz magnetic separator at field strength of 1.8A and side slope of 2°; N5, at 5° side slope. Front slope for all: 20°.

*Samples analyzed at the Royal Ontario Museum.

MSWD—mean square of weighted deviates; PCIGR—Pacific Centre for Isotopic and Geochemical Research.

new age determinations obtained in this study were therefore not treated on a local scale, but instead were integrated with previous published data for forearc magmatism of the northern Cordilleran subduction zone and used to anchor a Pacific Basin-wide tectonic interpretation.

TECTONIC MODEL

Necessity of a New Tectonic Model

Forearc magmatism is prevalent in the Cenozoic history of Alaska, British Columbia, Washington, and Oregon (Fig. 2), and most of these forearc igneous provinces have been attributed to ridge-trench interactions, or the effects of a nearby ridge. Most of the Paleogene tectonic reconstructions for the Pacific Basin involve the Farallon, Kula, and Pacific plates (e.g., Engebretson et al., 1985; Stock and Molnar, 1988; Lonsdale, 1988; Norton, 1995) and typically attribute forearc magmatism to subduction of the Kula-Farallon or Kula-Pacific ridges (e.g., Byrne, 1979; Thorkelson and Taylor, 1989; Babcock et al., 1990; Sisson and Pavlis, 1993). The pattern of forearc magmatism, however, strongly suggests that these ridges could be responsible for only part of the forearc igneous activity and that at least one additional mid-ocean ridge was subducting beneath the forearc at the same time (Haeussler et al., 1995, 2003a), a realization that serves as a cornerstone in our model.

The Resurrection Plate

The “additional” plate in the Pacific Basin, called the Resurrection plate, subducted beneath North America in Paleogene time (Miller et al., 2002; Haeussler et al., 2003a; Bradley et al., 2003). The magnetic record of the Resurrection plate has been completely subducted, and its existence in the Pacific Basin was first rationalized by the occurrence of synchronous middle Eocene ridge-related magmatism in Alaska, British Columbia, and Washington (Haeussler et al., 2003a).

According to Haeussler et al. (2003a), the Resurrection plate was situated between the Kula and Farallon plates, bordered to the north by the Kula-Resurrection ridge, and to the south by the Farallon-Resurrection ridge. The plate was constructed such that the Kula-Resurrection ridge intersected the continental margin near Alaska and generated the eastward-younging magmatism of the Sanak-Baranof Belt; concurrently, the Farallon-Resurrection ridge intersected the trench near Vancouver Island and Washington and produced the Coast Range Basalt Province (Haeussler et al., 2003a) and possibly the Walker Creek intrusions.

PACIFIC BASIN PLATE MOTIONS AND NORTHERN CORDILLERAN SLAB WINDOWS: 53–0 MA

TABLE 3. U-PB ANALYTICAL DATA FOR SAMPLED TERTIARY MAGMATIC ROCKS FROM VANCOUVER ISLAND

Fraction ¹	Wt (mg)	U ² (ppm)	Pb ³ (ppm)	²⁰⁶ Pb ⁴ / ²⁰⁴ Pb	Pb ⁵ (pg)	²⁰⁸ Pb ⁶ (%)	Isotopic ratios (1σ, %) ⁷			Apparent ages (2σ, Ma) ⁷		
							²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb
JM02-15-1-1C; Clayoquot Suite granodiorite; UTM: Zone 10, E 0287433, N 5447736												
A, 5	0.059	244	2.0	2389	3	10.5	0.00801 (0.19)	0.0521 (0.47)	0.04716 (0.42)	51.4 (0.2)	51.6 (0.5)	58 (20)
B, 16	0.045	180	2.6	1815	4	10.7	0.01426 (0.12)	0.1354 (0.30)	0.06886 (0.25)	91.3 (0.2)	128.9 (0.7)	895 (10)
C, 6	0.045	300	2.5	963	7	14.2	0.00786 (0.24)	0.0514 (1.0)	0.04745 (0.96)	50.5 (0.2)	50.9 (1.0)	72 (45/46)
D, 16	0.027	243	2.0	720	4	10.4	0.00797 (0.30)	0.0520 (1.7)	0.04736 (1.6)	51.1 (0.3)	51.5 (1.7)	67 (73/77)
JM02-21-1C; Mt. Washington Suite, Zeballos area intrusion; quartz diorite; UTM: Zone 09, E 0673066, N 5527162												
A, 3	0.082	187	1.4	653	11	9.5	0.00757 (0.13)	0.0492 (1.5)	0.04708 (1.4)	48.6 (0.1)	48.7 (1.4)	53 (66/68)
B, 6	0.090	276	2.1	445	27	11	0.00765 (0.17)	0.0497 (1.8)	0.04709 (1.7)	49.1 (0.2)	49.2 (1.7)	54 (79/83)
C, 7	0.040	341	2.7	231	30	13	0.00755 (0.24)	0.0490 (1.8)	0.04707 (1.7)	48.5 (0.20)	48.6 (1.7)	53 (79-/83)
D, 15	0.075	421	3.3	1004	16	13.4	0.00762 (0.13)	0.0494 (0.31)	0.04706 (0.24)	48.9 (0.1)	49.0 (0.3)	52 (12)
JM02-22-1-1C; Mt. Washington Suite, Labour Day Lake intrusion; hbl-plag porphyry; UTM: Zone 10 E 0393096, N 5442262												
A, 7	0.021	697	9.4	4317	3	7.1	0.01386 (0.13)	0.0940 (0.27)	0.04921 (0.22)	88.7 (0.2)	91.3 (0.5)	158 (10)
B, 7	0.046	155	0.9	815	3	9.2	0.00600 (0.24)	0.0388 (1.2)	0.04687 (1.2)	91.3 (0.5)	38.7 (0.9)	43 (56/58)
C, 14	0.050	315	2.2	1327	4	22.2	0.00599 (0.13)	0.0387 (0.44)	0.04681 (0.39)	38.5 (0.1)	38.5 (0.3)	40 (19)
D, 25	0.045	393	4.1	1070	11	14.8	0.00995 (0.12)	0.0663 (0.34)	0.04830 (0.28)	63.8 (0.2)	65.1 (0.4)	114 (13)
JM02-4-9-1C; Mt. Washington Suite, intrusion east of Moriarity Lake; hbl-plag porphyry; UTM: Zone 10 E 0398963, N 5444131												
A, 5	0.052	132	0.8	755	3	9.1	0.00581 (0.13)	0.0375 (1.4)	0.04682 (1.3)	37.4 (0.1)	37.4 (1.0)	40 (62/64)
B, 7	0.045	269	2.1	121	59	9.4	0.00774 (0.55)	0.0515 (1.9)	0.04830 (1.6)	49.7 (0.6)	51.0 (1.9)	114 (73/76)
C, 10	0.040	172	1.6	872	5	11.2	0.00895 (0.28)	0.0597 (0.83)	0.04838 (0.74)	57.5 (1.0)	58.9 (1.0)	118 (35)
D, 22	0.045	251	1.4	1269	3	9.9	0.00583 (0.12)	0.0376 (0.58)	0.04685 (0.53)	37.4 (0.1)	37.5 (0.4)	42 (25/26)
JM02-16-2-1B; Mt. Washington suite, intrusion south of Kennedy Lake, granodiorite; UTM: Zone 10, E 0312094, N 5432040												
A, 5	0.086	181	1.1	477	12	14.8	0.00573 (0.14)	0.0370 (1.0)	0.04685 (0.96)	36.9 (0.1)	36.9 (0.7)	41 (45/47)
B, 7	0.077	237	1.4	696	10	13.4	0.00574 (0.12)	0.0371 (0.86)	0.04683 (0.81)	36.9 (0.1)	37.0 (0.6)	40 (38/39)
C, 9	0.122	209	1.3	606	16	13.9	0.00573 (0.13)	0.0370 (0.97)	0.04688 (0.92)	36.8 (0.1)	36.9 (0.7)	43 (43/45)
D, 9	0.067	163	1.0	1130	3	15.9	0.00570 (0.12)	0.0368 (0.89)	0.04682 (0.85)	36.6 (0.1)	36.7 (0.6)	40 (40/41)
E, 10	0.070	168	1.0	917	5	15.2	0.00571 (0.11)	0.0369 (0.56)	0.04682 (0.52)	36.7 (0.1)	36.8 (0.4)	40 (25)
JM02-10-7-1D; Mt Washington Suite, Zeballos stock, granodiorite; UTM: Zone 9, E 0658002 N 5544352												
A, 13	0.136	106	0.6	569	9	13.2	0.00549 (0.14)	0.0354 (0.93)	0.04675 (0.87)	35.3 (0.1)	35.3 (0.7)	36 (41/42)
B, 30	0.180	99	0.6	845	7	14.2	0.00550 (0.16)	0.0354 (1.1)	0.04666 (1.0)	35.3 (0.1)	35.3 (0.8)	32 (49/51)
C, 16	0.125	110	0.6	1121	4	14.7	0.00549 (0.30)	0.0353 (1.3)	0.04663 (1.2)	35.3 (0.2)	35.2 (0.9)	30 (57/59)
D, 25	0.122	121	0.7	489	11	14.8	0.00552 (0.15)	0.0356 (1.1)	0.04672 (1.0)	35.5 (0.1)	35.5 (0.8)	35 (49/50)

¹All zircon grains selected for analysis were strongly abraded prior to dissolution. Capital letter designation is fraction identifier, followed by the number of grains analyzed in that fraction. See Table 2 for descriptions of analyzed zircons and interpretations; see Table 1 for interpreted ages.

²U blank correction of 1 pg ± 20%; U-fractionation corrections were measured for each run with a double ²³³U-²³⁵U spike (about 0.004/amu).

³Radiogenic Pb.

⁴Measured ratio corrected for spike and Pb fractionation of 0.0037/amu ± 20% (Daly collector), which was determined by repeated analysis of NBS Pb 981 standard throughout the course of this study.

⁵Total common Pb in analysis based on blank isotopic composition.

⁶Radiogenic Pb.

⁷Corrected for blank Pb (2–8 pg, throughout the course of this study), U (1 pg) and common Pb concentrations based on Stacey-Kramers model Pb (Stacey and Kramers, 1973) at the interpreted age of the rock or the ²⁰⁷Pb/²⁰⁶Pb age of the rock.

The Resurrection plate was first modeled to be completely subducted by 50 Ma to explain the shutoff of arc magmatism in British Columbia at ca. 50 Ma (Haeussler et al., 2003a). However, during late Eocene to Oligocene time, forearc magmatism still occurred in southwest Alaska, on Vancouver Island, the Queen Charlotte Islands, and in Washington and Oregon. These forearc magmatic centers have generally been regarded as the result of ridge-processes or ridge-trench interaction and slab window formation in individual study areas (Barnes and Barnes, 1992; Hamilton and Dostal, 2001; Kusky et al., 2003; Groome et al., 2003; Madsen et al., 2003) but have never been unified in a regional tectonic

model in context with early Eocene to Oligocene forearc magmatism. The persistence of concomitant forearc magmatism in widely separated areas until Oligocene time suggests that a successful ridge-subduction model for forearc magmatism would require at least two mid-ocean ridges and four oceanic plates remaining in the Pacific Basin through the Oligocene.

Purpose and Methodology

Modeling of Pacific Basin plate tectonics from 53 Ma to the present (Figs. 6–14) was undertaken to test the possibility that all of the forearc magmatic centers of the Northern

Cordilleran subduction zone could have been caused by ridge subduction. The model was constructed under several assumptions: (1) spreading at all mid-ocean ridges was symmetrical, except where asymmetry was preserved in the magnetic record; (2) the capture of the Kula plate by the Pacific plate occurred at chron 18r, or ca. 40 Ma (Lonsdale, 1988; recalibrated to the time scale of Cande and Kent, 1995); (3) hotspots were essentially fixed relative to each other between 40 Ma and the present; and (4) ridges and transform faults intersect at 90° angles. Using these constraints, the configurations of the plate boundaries were adjusted to fit sites of ridge subduction with forearc mag-

TABLE 4: ID-TIMS U-PB ANALYTICAL DATA FOR TWO SAMPLES DATED AT THE ROYAL ONTARIO MUSEUM

No. Anal.	Description	Wt. (mg)	U (ppm)	Th/U	Pbrad (pg)	Pbcom (pg)	Corrected for mass discrimination, common Pb (assumed to be all blank), and spike										Error Correl.		
							$\frac{206\text{Pb}}{204\text{Pb}}$	$\frac{207\text{Pb}}{204\text{Pb}}$	$\frac{206\text{Pb}}{238\text{U}}$	2 sig	$\frac{207\text{Pb}}{235\text{U}}$	2 sig	$\frac{206\text{Pb}}{238\text{U}}$	2 sig	$\frac{207\text{Pb}}{235\text{U}}$	2 sig		$\frac{207\text{Pb}}{206\text{Pb}}$	2 sig
JM02-24-5-1D Mount Washington																			
1	BP01 4 Ab zr, lpr	0.002	300	0.41	3.3	4.6	63.27	17.689	0.00542	0.00009	0.0344	0.0105	0.0460	0.0140	34.87	0.58	—	—	0.02820
2	BP02 4 Ab zr, spr	0.02	80	0.37	8.8	10.2	73.40	18.272	0.00552	0.00008	0.0366	0.0087	0.0481	0.0114	35.49	0.49	—	—	0.02360
JM02-8-3-1G Patricant Mountain																			
1	BP04 1 Ab zr, spr	0.005	517	0.47	15.9	1.0	1048.9	63.92	0.00600	0.00001	0.0387	0.0006	0.0467	0.0007	38.57	0.09	—	—	0.21471
2	BP06 4 Ab zr, spr	0.02	937	0.45	11.5	6.6	127.9	20.54	0.00599	0.00019	0.0370	0.0052	0.0448	0.0062	38.50	1.23	—	—	0.24970

Note: zr—zircon grain; Ab—abraded; lpr—long prismatic; spr—short prismatic; Disc.—percent discordance for the given $^{207}\text{Pb}/^{206}\text{Pb}$ age. Pbrad is total radiogenic Pb. Pbcom is total measured common Pb assuming the isotopic composition of laboratory blank: 206/204—18.221; 207/204—15.612; 208/204—39.360 (errors of 2%). Blank U calculated on 0.1 pg. Th/U calculated from measured radiogenic $^{206}\text{Pb}/^{208}\text{Pb}$ ratio and $^{207}\text{Pb}/^{206}\text{Pb}$ age assuming concordance. Daily ion counting mass discrimination correction factor is 0.1% per amu. VG554 instrument thermal mass discrimination correction factor is 0.1% per amu using silica gel on rhenium filaments. Uranium decay constants are from Jaffey et al. (1971).

matism. In the early model iterations, the plate configurations and Resurrection plate vectors were largely speculative and based solely on the forearc magmatic record, as most evidence of mid-ocean ridges north of the Great Magnetic Bight has been subducted (Atwater, 1970). The model becomes increasingly constrained into Oligocene time, since the configuration of the Pacific-Farallon ridge is preserved in the sea-floor magnetic record. Between ca. 20 Ma and present, the Pacific-Farallon plate boundary configuration, and the location of the ridge-trench triple junction was reconstructed from the magnetic record after Wilson (1988).

The shapes of the resultant slab windows are idealized, assuming no thermal erosion of the plate edges, no spherical shell strain, and no microplate development or intraplate deformation other than that illustrated in Figures 6–14. Our model indicates that slab windows grew and migrated beneath North America throughout the Cenozoic and played a central role in Cordilleran evolution. Several tectonic models were created but rejected prior to establishing the best-fit model, presented herein. Rejected models incorporating only the Kula, Farallon, and Pacific oceanic plates interacting with North America were clearly unsuccessful in recreating the forearc magmatic record.

The best-fit model introduced in the following involves the Pacific, Farallon, Kula, and Resurrection plates. Our model shows the Resurrection plate separating into two microplates in the middle Eocene, in the same way that the Farallon plate divided into smaller plates (the Juan de Fuca and Rivera plates) in the Neogene. The southern microplate retains the name Resurrection plate, while the northern microplate is herein named the Eshamy plate, after Eshamy Bay in Alaska.

The plate model spans the time interval of 53 Ma to the present and was constructed in 1 m.y. increments on a Mercator projection using vectors calculated from published stage poles, recalibrated to the time scale of Cande and Kent (1995). The model was generated in two parts based on availability and suitability of published plate circuit rotation poles and hot-spot rotation poles. The first part of the model spans 53–39 Ma (Movie 1) and was constructed with respect to North America. Vectors for the Farallon and Pacific plates were calculated from the plate circuit rotation poles of Norton (1995). Kula vectors were calculated from poles of Lonsdale (1988), taking into account the proportion of asymmetrical spreading preserved in the magnetic record. Stein et al. (1977) documented similar asymmetry in spreading at the Pacific-Antarctic ridge, indicating that the proposed amount of Kula-Farallon asymmetry

is reasonable. Resurrection plate vectors were constrained by the vectors of the bounding Kula and Farallon plates (i.e., with azimuths of motion intermediate between those of the Kula and Farallon plates) and the requirement that the Farallon-Resurrection and Kula-Resurrection ridges meet at a realistic $\sim 120^\circ$ triple junction with the Kula-Farallon ridge. Kula rotation poles of Norton (1995) were not used, because they are not available for times after 53 Ma and could not fulfill the needs of our model, which is based on a Kula plate “death” (fusion to the Pacific plate) at chron 18r, or ca. 40 Ma (Engelbretson et al., 1985; Lonsdale, 1988).

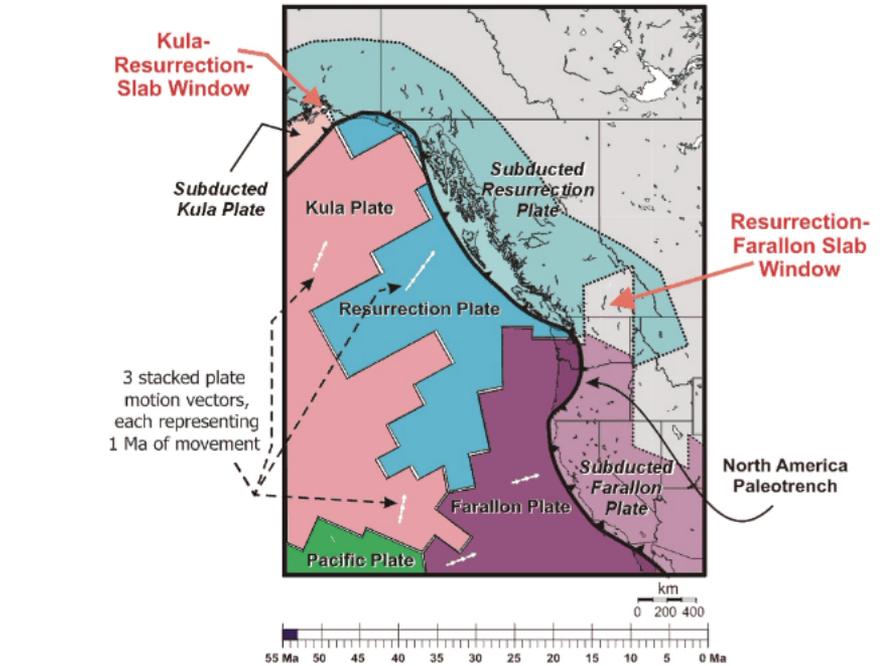
The second part of the model spans the time interval 39 Ma to the present (Movie 2) and was calculated with respect to hotspots based on the stage poles of Engelbretson et al. (1985). Vectors calculated from the stage poles of Engelbretson et al. (1985) demonstrate that the Pacific-Farallon ridge migrated approximately northward and remained in an approximately stable position west of the trench during the interval 39 Ma to present, which is consistent with the detailed reconstructions of Wilson (1988) and the present position of the Juan de Fuca ridge. Plate circuit rotation poles (Norton, 1995) were available for this interval, but were not used because the vectors calculated from these poles resulted in an unacceptably large migration of the model Pacific-Farallon ridge toward the North America trench, causing the ridge to completely subduct during the Oligocene. The discrepancy between calculated vectors and ridge behavior from late Eocene to present time may be related to the complicated history of ridge propagation experienced by the Juan de Fuca (Pacific-Farallon) ridge beginning at ca. 33 Ma (Wilson, 1988). Fixed-hotspot tectonic reconstructions, particularly those which relied on the Hawaiian-Emperor seamount chain (e.g., those of Engelbretson et al., 1985), have been called into question as more recent work indicates that the Hawaiian hotspot changed its azimuth of motion relative to other hotspots and the North American plate at ca. 43–40 Ma, when the bend in the Hawaii-Emperor seamount chain formed (Norton, 1995; Tarduno and Cottrell, 1997). For that reason, we restricted our use of the hotspot-framework poles to time frames subsequent to the 40 Ma apparent swerve in the Hawaiian plume, with the understanding that the hotspot reference frame serves as a reasonable basis for evaluating late Eocene through Recent plate motions.

Complications to Modeling: Northward Terrane Transport

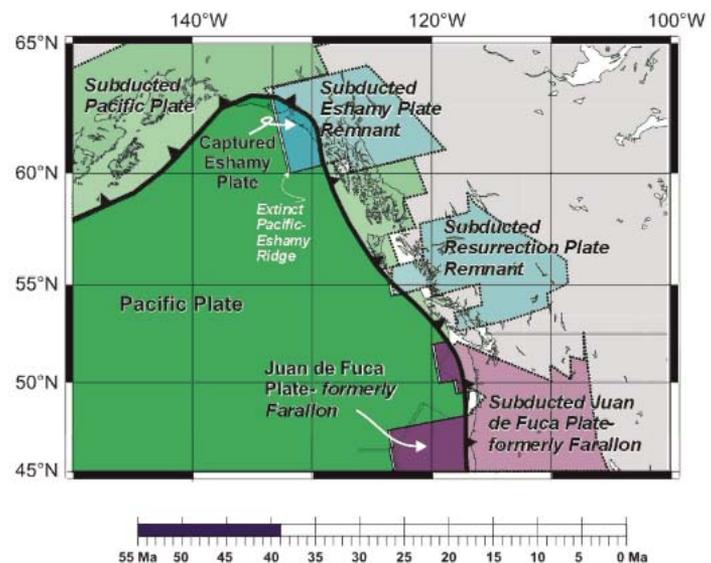
Complications to tectonic modeling in the Pacific Basin exist where the geographical

location of magmatism is drawn into question, such as in the cases of the Chugach–Prince William terrane and the Yakutat terrane. The Chugach–Prince William terrane (Fig. 1), which contains the 61–48 Ma intrusions of the Sanak–Baranof Belt, was likely affected by northward transport during the Cretaceous and Tertiary and was possibly also involved in oroclinal bending between 66 and 44 Ma (Coe et al., 1989; Bradley et al., 2003) or oroclinal orogeny before 45 Ma (Johnston, 2001). The Yakutat terrane contains near-trench intrusions of comparable age. Both terranes were at uncertain locations during the Eocene, moved independently of each other, and docked against mainland Alaska at different times within the Tertiary (Davis and Plafker, 1986; Roeske et al., 2003).

For purposes of our model, it is important to elucidate the amount of coastwise transport experienced by the Chugach–Prince William terrane during the interval between 53 Ma and the present. From Cretaceous until early Eocene time, the Chugach–Prince William terrane likely migrated northward on an oceanic plate or as a forearc sliver from an original position located several hundred to thousands of kilometers south of its current location (Irving et al., 1996; Cowan et al., 1997; Cowan, 2003; Bradley et al., 2003; Pavlis and Sisson, 2003; Roeske et al., 2003). Paleomagnetism of the Resurrection Peninsula ophiolite within the Chugach–Prince William terrane suggests northward migration of as much as $13 \pm 9^\circ$ since 57 Ma (Bol et al., 1992), although it is uncertain how much of the movement took place between formation of the ophiolite crust at 57 Ma and its incorporation into the Chugach–Prince William terrane by 53 Ma (Kusky and Young, 1999; Bradley et al., 2000, 2003). Movement along the Border Ranges fault, the inboard fault along which the Chugach–Prince William terrane migrated, may have been largely complete by 51 Ma (Johnson and Karl, 1985; Little and Naeser 1989; Roeske et al., 2003). Estimates of movement along the Border Ranges fault indicate that between Cretaceous and middle Eocene time (85–51 Ma), the fault accommodated between 700 and >1000 km of displacement (Roeske et al., 2003). Roeske et al. (2003) concluded that this dextral movement ceased by ca. 51 Ma, based on ^{40}Ar – ^{39}Ar ages of sericite formed during strike-slip faulting. A further constraint on the motion of the Chugach–Prince William terrane is provided by an undeformed near-trench pluton, which cross-cuts and pins a segment of the Border Ranges fault at 51 Ma (Johnson and Karl, 1985). Little and Naeser (1989) provided sedimentary evidence that movement of the Border Ranges fault has been on the order of tens of kilometers since early Eocene time. Hillhouse and Coe (1994)



Movie 1: Tectonic model for the Pacific Basin and northwestern North America from 53 Ma to 39 Ma. If you are viewing the PDF, or if you are reading this offline, please visit www.gsapubs.org or <http://dx.doi.org/10.1130/GES00020.s1> to view the animation.



Movie 2: Tectonic model for the Pacific Basin and northwestern North America from 39 Ma to present day. If you are viewing the PDF, or if you are reading this offline, please visit www.gsapubs.org or <http://dx.doi.org/10.1130/GES00020.s2> to view the animation.

used paleomagnetic data on Tertiary lithologies in the inboard Wrangellia terrane to conclude that subsequent to oroclinal formation (between 66 and 44 Ma; Coe et al., 1989), the trench-parallel translation of outboard terranes was nearly

complete. Miller et al. (2002) examined the inboard Denali and Iditarod–Nixon Fork faults and determined that there has been 38 km of dextral slip on the western limb of the Denali fault since 40 Ma and only 134 km since 85 Ma.

Displacement estimates for the eastern limb of the Denali fault, however, are larger, ~450 km since Cenozoic time (Eisbacher, 1976; Nokleberg et al., 1985). The Iditarod–Nixon Fork fault has undergone 88–94 km of dextral displacement since 65 Ma. Taken together, geological and some paleomagnetic evidence suggests that the Chugach–Prince William terrane was near its present position on a regional scale during the interval of the proposed tectonic model (53 Ma to present). Current geodetic data show movement of the Cascadia forearc with respect to North America, suggesting that forearc fixity may have varied during the intervals in question (Wells et al., 1998).

The Yakutat terrane is a composite oceanic-continental package that moved northward during the Tertiary. Estimates for northward movement of the Yakutat terrane during the Tertiary range from 5° (Plafker, 1983) to 30° of latitude (Bruns, 1983). The Yakutat terrane began accreting in the Gulf of Alaska during Miocene time (Bruns, 1983; Plafker, 1983; Plafker et al., 1994). Fission-track data suggest that accretion possibly began between 19 and 14 Ma in the northwest and southeast, respectively (O’Sullivan and Currie, 1996; Sisson et al., 2003). Continued convergence between the Yakutat terrane and North America is evident from seismological and uplift studies (Mazzotti and Hyndman, 2002) and global positioning system (GPS) data (Fletcher and Freymueller, 1999).

Regional Tectonic Synthesis

The following chronology highlights the main aspects of plate kinematics in the Pacific Basin and the effects of ridge-trench intersection on the North American plate, from early Eocene to Recent. The early history of plate tectonics in the Pacific Basin is not as well constrained, and the reader is directed to previous works by Atwater (1970), Engebretson et al. (1985), Stock and Molnar (1988), Lonsdale (1988), Babcock et al. (1992), Wells et al. (1984), Haeussler et al. (2003a), Groome et al. (2003), and Breitsprecher et al. (2003).

Because the mid-ocean ridges of the Pacific Basin were subducting for millions of years prior to 53 Ma (Thorkelson and Taylor, 1989; Haeussler et al., 2003a), it was necessary to construct a ridge-transform and slab window configuration that would be “inherited” by the tectonic model. For the Farallon-Resurrection slab ridge, Cretaceous to Paleocene ridge subduction would have created windows of uncertain geometry beneath British Columbia and the northwestern United States. In southern British Columbia and the northern United States, an

approximate position for the slab window was geochemically defined by Breitsprecher et al. (2003). For the first frame of our model, we modified the middle Eocene slab window shape and position from Breitsprecher et al. (2003) to be consistent with the plate motion vectors calculated for this study. This slab window was positioned underneath the Challis-Kamloops volcanic belt to account for anomalous arc, within-plate, and alkaline magmatism (Ewing, 1981; Thorkelson and Taylor, 1989; Breitsprecher et al., 2003). The slab window related to the Kula-Resurrection ridge would have been created underneath British Columbia, Alaska, and Yukon Territory, although the location of the Kula-Resurrection slab window is uncertain, because the Chugach–Prince William terrane, which contains the Sanak-Baranof Belt intrusions, moved northward during Cretaceous to early Eocene time. For this model, schematic windows were drawn underneath Alaska, Yukon, and British Columbia according to the vectors calculated for 53 Ma.

53–50 Ma

From 53 to 50 Ma (Fig. 6), the Resurrection plate was located off the coast of British Columbia and was bordered to the north and west by the Kula plate and to the south and east by the Farallon plate (Haeussler et al., 2003a). This configuration satisfies the general requirement for two widely separated ridge-trench intersections. Along the continental margin, two different sets of ridge-trench intersections existed, one near Alaska and one west of Vancouver Island. In Alaska, the segmented Kula-Resurrection ridge migrated south and east along the Chugach–Prince William terrane accretionary complex, generating the well-documented eastward-younging age progression in the Sanak-Baranof Belt (Bradley et al., 1993). A series of Kula-Resurrection ridge-trench triple junctions was formed as the segmented ridge-transform geometry intersected the curved Alaskan trench. The multiple intersections created fraternal slab windows and short-lived microplates. The subducted portions of the Kula and Resurrection plates were modeled to be dipping at 26°, and the subducted portions of the Farallon plate dipping at 11°. The Kula-Resurrection ridge intersections in Alaska provided a heat source for the high-temperature, low-pressure metamorphism of the Chugach metamorphic complex (e.g., Sisson and Pavlis, 1993; Pavlis and Sisson, 2003) and are consistent with both the forearc magmatic record and the observed lull in Alaskan arc magmatism from 55 to 40 Ma (Wallace and Engebretson, 1984). Ridge-trench interaction may also help to explain the older than 53 Ma

obduction of the 57 Ma Resurrection Peninsula ophiolite into the Chugach terrane accretionary complex (Bradley et al., 2003).

The ridge-trench intersection located near Vancouver Island involved the Farallon-Resurrection ridge. The Farallon-Resurrection ridge-trench intersection and associated slab window is coincident with the Flores volcanics (51.2–50 Ma), the Clayoquot intrusions (51.2 Ma), and the Walker Creek suite (50.7–50.9 Ma). The Farallon-Resurrection ridge-trench intersection can also be tied to the high-temperature, low-pressure metamorphism of the Leech River Complex at 51 Ma (Groome et al., 2003).

The Farallon-Resurrection slab window emanated from the ridge-trench triple junction near Vancouver Island and extended below southern British Columbia, Washington, Montana, Idaho, and Wyoming. The presence of this slab window below these areas is roughly coincident with the alkalic, within-plate, and adakitic magmatism of the middle Eocene Challis-Kamloops Belt (cf. Thorkelson and Taylor, 1989; Hole et al., 1991; Johnston and Thorkelson, 1997; Breitsprecher et al., 2003). Challis-Kamloops volcanism began at ca. 53 Ma and extended from southern Idaho and Wyoming to northern British Columbia (Fig. 1). Igneous rocks in the southernmost area of the belt are highly alkaline, within-plate basalts, e.g., Montana Alkaline Province and Absaroka volcanics. The middle of the belt in southern British Columbia and the northern United States represents a zone of geochemical transition. The transition zone displays several geochemical types, including adakitic, alkalic, within-plate, and arc rocks (Ewing, 1981; Breitsprecher et al., 2003). Volcanic complexes within the transition zone include the Kamloops, Princeton, and Penticton Groups and the Colville Igneous Complex. The northern part of the belt exhibits extended-arc to backarc geochemistry (Ootsa Lake and Endako Groups and Clisbako Lake volcanic rocks; Ewing, 1981; Breitsprecher et al., 2003). The spatial and geochemical relationships between transition zone magmas and those of within-plate affinity suggest a transition from a slab edge to a slab window environment (Thorkelson and Taylor, 1989; Breitsprecher et al., 2003).

Challis-Kamloops magmatism occurred during widespread pull-apart basin formation and core complex exhumation in British Columbia and northern Washington (Fig. 1). Core complex exhumation included the Tatla Lake Complex from 53 to 46 Ma (Friedman and Armstrong, 1988), the Monashee Complex from ca. 63 to 45 (Parrish et al., 1988), the Shuswap Complex from ca. 60 to 50 Ma (Tempelman-Kluit and Parkinson, 1986; Johnson and Brown, 1996; Vanderhaeghe and Teyssier, 2003), and the

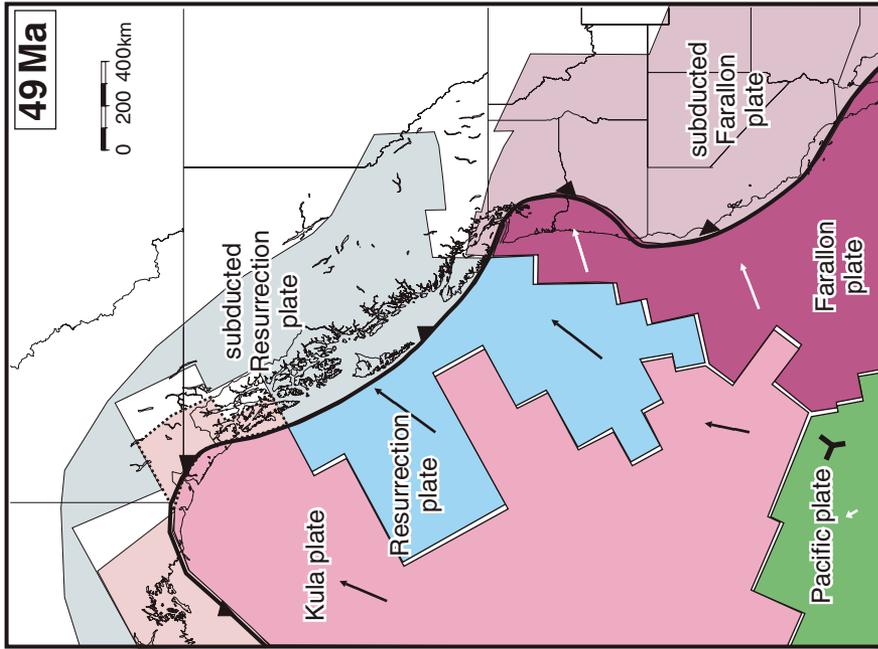


Figure 7. Time slice from the tectonic model showing the best-fit plate configurations at 49 Ma, after 4 m.y. of forward plate movement. The vectors on the oceanic plates represent 3 m.y. of plate motion. The black Y-shaped feature represents the abandoned Kula-Farallon-Pacific triple junction, which is now preserved as the Great Magnetic Bight near Alaska.

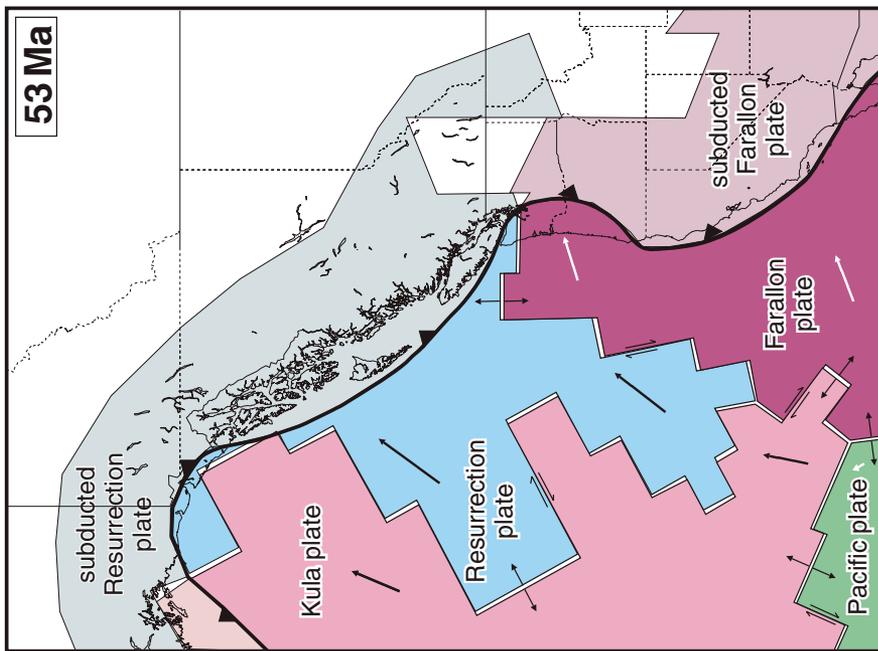


Figure 6. The first frame of the tectonic model at 53 Ma, which represents a best-fit plate configuration for the Pacific Basin developed to account for the forearc magmatic record from Alaska to Oregon. The oceanic plates include the Kula plate, Resurrection plate, Pacific plate, and Farallon plate. The oceanic portions of these plates are labeled to the west of the trench (represented by the toothed line) and the subducted components are to the east and north of the trench. Subducted oceanic crust was geometrically modeled to move at the same vectors as the attached plates in the ocean basin. Arrows on the plates represent vectors for 3 m.y. of plate motion. The vectors were calculated from published stage poles of Norton (1995) and Lonsdale (1988), and Resurrection plate vectors were estimated for this study as discussed in the text. Subducted slabs are diagrammatically shown to terminate at the 300 km isobath on this and all subsequent figures. White areas to the east and north of the trench represent mantle. Mantle-filled gaps between subducted plates are slab windows. The 53 Ma model frame displays the generalized slab windows beneath Alaska and British Columbia and the Pacific Northwest that were inherited by this model.

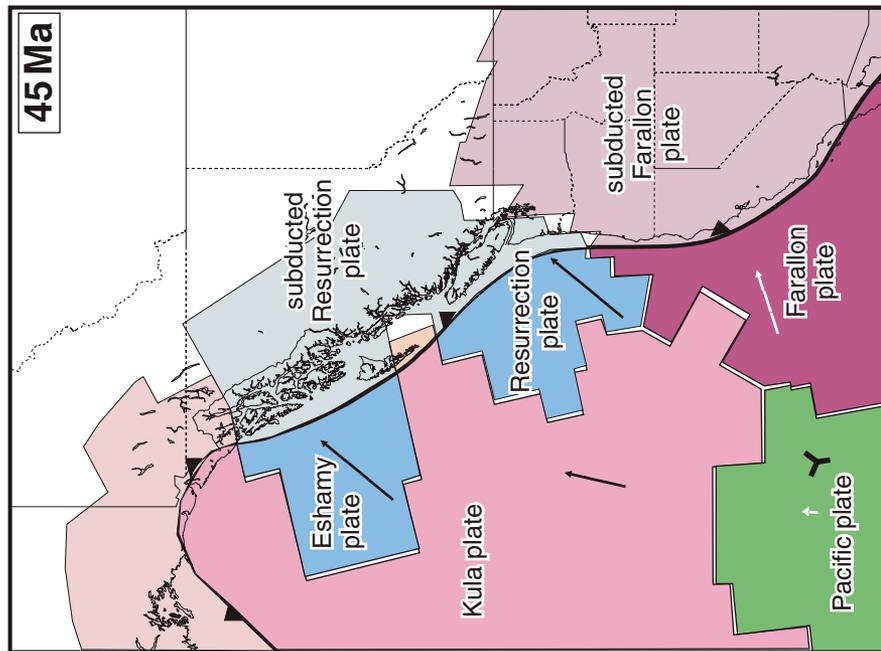


Figure 8. Time slice from the tectonic model showing the best-fit plate configurations at 45 Ma. A plate reorganization event at 47 Ma slightly altered the ridge-transform geometries. Vectors on the oceanic plates represent 3 m.y. of plate movement. A large promontory on the Kula plate intersected the trench at 47 Ma and divided the Resurrection plate. This created the Eshamy plate to the north of the promontory, and a Resurrection plate remnant to the south. The paleotrench near Washington and Oregon made an onboard jump at ca. 48 Ma due to accretion of the Coast Range Basalt Province (shown as instantaneous for the purposes of the model).

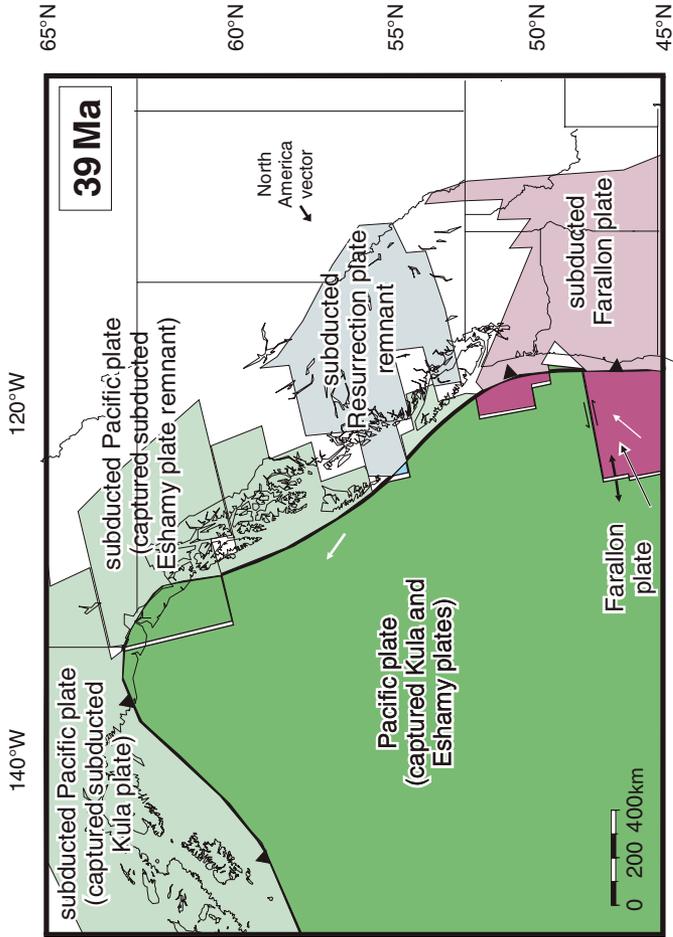


Figure 9. Time slice of the tectonic model at 39 Ma. This is the second phase of the tectonic model and was performed with respect to hotspots using the stage poles of Engebretson et al. (1985). At 40 Ma, the Kula and Eshamy plates became fused to the Pacific plate. The Pacific plate began to move in approximately transform motion with respect to the Queen Charlotte transform and initiated extension in the Queen Charlotte Basin (discussed in text). Note the small North America vector. The subducted Resurrection remnant is assumed to move at the same vector as before it was fully subducted. The newly extinct ridge in the Gulf of Alaska is still able to impart a thermal pulse into the forearc of Alaska. Vectors represent 3 m.y. of plate motion.

Valhalla Complex from 58–56 until 52 Ma (Parrish et al., 1988). The Okanagan metamorphic core complex in southern British Columbia and northern Washington was exhumed between 52 and 45 Ma (Harms and Price, 1992; Parrish et al., 1988). In Idaho, the Priest River Complex began exhumation at ca. 50 Ma (Doughty and Price, 1999). The Eocene regional extension, assumed high heat flow and core complex exhumation observed in these inboard areas may be manifestations of the shifting Farallon-Resurrection slab window. This interpretation is consistent with the model reconstructions and location of the slab window as suggested by Breitsprecher et al. (2003) and Dostal et al. (2003).

50–45 Ma

From 50 to 45 Ma (Fig. 7), forearc magmatism occurred in Alaska, the Queen Charlotte Islands, Vancouver Island, and Oregon. In Alaska, the Kula-Resurrection ridge continued to migrate south and east and caused emplacement of the ca. 50–48 Ma intrusions on Baranof Island, which complete the Sanak-Baranof Belt age progression. Farther south, at 50–49 Ma, the Farallon-Resurrection ridge-trench triple junction was situated in the vicinity of Vancouver Island, coincident with emplacement of a 48.8 Ma Clayoquot intrusion in the Tahsis area. From 49 to 45 Ma, the Farallon-Resurrection triple junction and slab window migrated southward to underlie parts of Washington and Oregon. Tillamook volcanism began in Oregon at ca. 45 Ma.

At 47 Ma, the Resurrection plate started to divide into two smaller plates, although their subducted parts may have remained connected. This splitting of the Resurrection plate occurred when a large ridge-transform promontory intersected the trench near the southern Queen Charlotte Islands, giving rise to a northern Eshamy plate and a southern Resurrection plate located near Vancouver. For modeling purposes, the Eshamy plate is assumed to have moved according to the same vector as the Resurrection plate during this interval, as they were joined at depth. As a large promontory of the Farallon-Resurrection plate boundary intersected the trench, a small slab window was generated, leading to formation of the 46.2 Ma Carpenter Bay pluton in the southern Queen Charlotte Islands.

Inboard, the Farallon-Resurrection slabs extended below southern British Columbia, Washington, and Oregon at different times. The presence of the slab window below these areas can be linked to the continuation of alkaline arc to within-plate magmatism in the Challis-Kamloops Belt, and regional extension and core complex formation. Challis-Kamloops

volcanism occurred in both the northern and southern regions of the belt, but was beginning to wane, and by 45 Ma was essentially complete. The Wolverine core complex underwent exhumation from 50 to 42 Ma, and the Shuswap Complex underwent a second exhumation event ca. 45 Ma (Vanderhaeghe et al., 2002). The Okanagan core complex in southern British Columbia and northern Washington was exhumed by 45 Ma (Harms and Price, 1992; Parrish et al., 1988). In Idaho, the Priest River Complex was becoming exhumed between 50 and 45 Ma (Doughty and Price, 1999), and the Tatla Lake core complex finished exhumation at 46 Ma (Friedman and Armstrong, 1988). Normal arc magmatism, represented by Tertiary plutons in the Coast Plutonic Complex, became markedly reduced at ca. 50 Ma.

The forearc tectonic history between 50 and 45 Ma is complicated by the accretion of the Coast Range Basalt Province. Underplating of this large basalt province may have caused an outboard jump of the paleotrench near Oregon, Washington, and Vancouver Island. On southern Vancouver Island, subduction-accretion of the Metchosin Igneous Complex portion of the basalt province is linked to the generation of the Cowichan fold-and-thrust system of England and Calon (1991) and formation of an Eocene orocline (Johnston and Acton, 2003). Accretion of the Metchosin complex may have begun as early as 50 Ma, as suggested by rapid cooling and exhumation of the Leech River Complex (through 400 °C by 45 Ma; Groome et al., 2003). Timing of accretion is less well constrained in Washington and Oregon, but likely occurred at approximately the same time.

45–40 Ma

Between 45 and 40 Ma (Fig. 8), forearc magmatism was concurrent in Alaska, on Vancouver Island, and in Oregon. The Kula-Eshamy ridge intersected the trench near the St. Elias Mountains and provided the heat source for the creation of a 42 Ma gabbro body near John's Hopkins Inlet. Farther south, the Farallon-Resurrection ridge was migrating northward from its position near Oregon where the Tillamook volcanics were being erupted (ca. 45–42 Ma) to a position near Vancouver Island. By 41 Ma, the Farallon-Resurrection ridge was located offshore of Meares Island. Subduction of the Farallon-Resurrection ridge led to the Mount Washington intrusive event, marked by the emplacement of the 41 Ma Ritchie Bay pluton. An important feature of this time interval is the first intersection of the Kula-Farallon ridge with the North American trench. The Kula-Farallon ridge had been located offshore in the Pacific

Basin at all earlier times and first intersected the continental margin in Oregon at 40 Ma. The slab windows generated in Washington and Oregon in this interval by both the Farallon-Resurrection and Kula-Farallon ridges were small and concentrated near the trench. The onset of magmatism in the Cascade volcanic arc at ca. 42 Ma was apparently unaffected by these windows. Initiation of Cascade volcanism may have been coincidental with steepening of the subducted Farallon plate subsequent to tectonic underplating of the Coast Range Basalt Province.

Inboard, the slab window was positioned under parts of central and southern British Columbia, and at 40 Ma was located under most of Washington. Challis-Kamloops Belt magmatism ceased in British Columbia at ca. 45 Ma, as did most inboard extension and core complex exhumation. This progressive magmatic shutoff may have been related to thermal re-equilibration of the crust with upwelling mantle in the slab window and a less tensional stress regime. Alternatively, the shutoff may be an artifact of sparse dating. In the northwestern United States, numerous dates of ca. 35 Ma have been obtained for upper flows in the Challis-Kamloops Belt, but in southern British Columbia, there is a paucity of reliable data. A few late Eocene to early Oligocene ages have been obtained in British Columbia. Lamprophyre dikes in the Nass River and Whitesail Lake areas have ages as young as ca. 33–35 Ma. Breitsprecher (2002) obtained a ca. 35 Ma Ar-Ar whole-rock age for a lava flow in the Kamloops Group.

40–35 Ma

Tectonic reconstructions indicate that ca. 40 Ma (Chron 18r) was a time of major plate reorganization that resulted in the fusion of the Kula plate to the Pacific plate (Engebretson et al., 1985; Lonsdale, 1988). This event, known as the “death” of the Kula plate (Engebretson et al., 1985; Lonsdale, 1988), resulted in the Kula plate moving according to the Pacific plate Euler pole (Engebretson et al., 1985; Lonsdale, 1988), i.e., approximately parallel to the coastline of southeastern Alaska. According to our tectonic model, most of the Resurrection plate had been subducted by this time, however, the Eshamy plate and a small microplate near Queen Charlotte Islands remained (Fig. 9). These microplates are assumed to have fused to the Pacific plate during Kula death and hence began to move as part of the Pacific plate at 40 Ma.

Despite the death of the Kula and Eshamy plates at 40 Ma, forearc magmatism continued in Alaska, on Queen Charlotte Islands, Vancouver Island, and in Washington and Oregon. Magmatism in Alaska was scattered between

the Prince William Sound area and the St. Elias Mountains and Baranof Island areas. In our tectonic model, the young extinct ridge associated with the Eshamy plate was located ~500 km from the main locus of late Eocene to Oligocene

forearc magmatism in Prince William Sound. This recently fossilized ridge would experience thermal erosion of the thin, young slab along the subducting ridge crest, which could still generate forearc melting (cf. Severinghaus and Atwa-

ter, 1990; Thorkelson and Breitsprecher, 2005). If the extinct Eshamy-Kula ridge was instead located in Prince William Sound, melting associated with this northward-migrating feature would be a plausible mechanism for generating the localized pulse of magmatism in that area.

Another group of late Eocene to Oligocene to Miocene forearc plutons is preserved in the St. Elias Mountains and on Baranof Island in Alaska, which spatially and temporally coincide with the Admiralty Island volcanics. These plutons and volcanics are poorly dated, but magmatism may have persisted into Miocene time. The tectonic model shows that a small extinct slab window generated at the formerly active Eshamy-Kula boundary may have migrated under this region between 40 and 30 Ma. This model slab window was 100 km wide and was positioned between two crustal-scale breaks, which were formerly active as Eshamy-Kula oceanic transform faults.

The capture of the Kula plate by the Pacific plate initiated highly oblique subduction to transform-style movement along the Queen Charlotte fault, a marked contrast to the previous convergence by the Resurrection and Kula plates. The postulated change in plate motion to more oblique subduction may be linked to the observed initiation of extension in the Queen Charlotte Basin (Hyndman and Hamilton, 1993). Basin formation was coeval with emplacement of several Kano plutons, formation of dike swarms, and eruption of voluminous Masset lava and tephra (Hyndman and Hamilton, 1993). This magmatism was also coeval with slab window formation, which may have aided extension in the area and was responsible for the enriched (E)-MORB geochemistry of the Masset basalts (Hamilton and Dostal, 2001). Kano intrusions in this interval range from 39 to 35 Ma in a northward-younging age progression. The northward-younging trend suggests diachronous emplacement associated with a migrating slab window. A large slab window grew underneath British Columbia between 40 and 35 Ma as the subducted Resurrection plate foundered into the mantle and the Pacific-Farallon slab window began to open. The gradual opening of the Pacific-Farallon slab window beneath Vancouver Island resulted in the broad northward-younging magmatic trend displayed by the Mount Washington intrusions. This trend began on southern Vancouver Island with emplacement of the 38.6 Ma Nanaimo Lakes area intrusions and culminated with the ca. 35 Ma Zeballos and Mount Washington area intrusions on northern Vancouver Island.

In Washington and Oregon, voluminous forearc volcanism with mafic, tholeiitic to alkalic compositions occurred in a synextensional setting

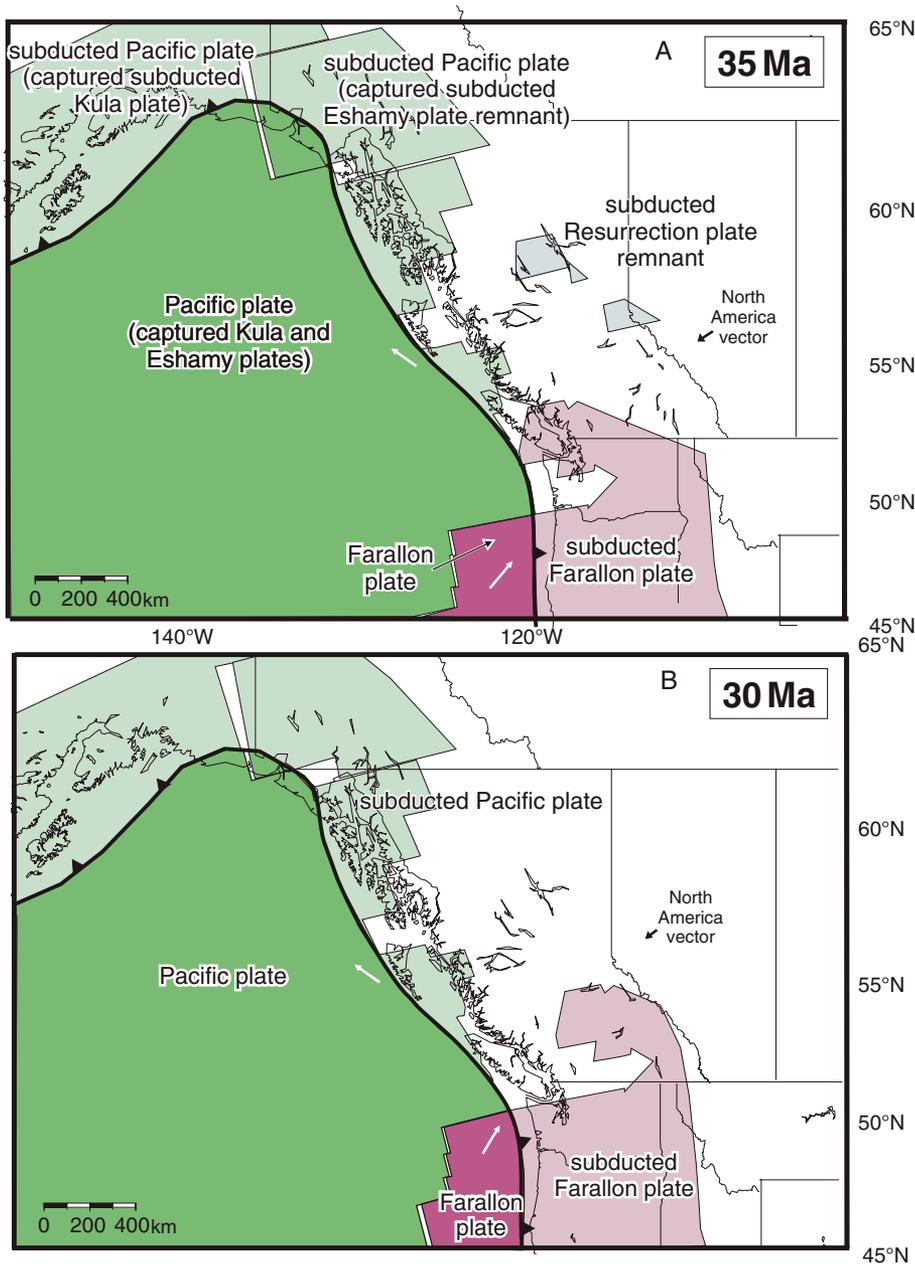


Figure 10. Time slices of the tectonic model at (A) 35 Ma and (B) 30 Ma. In frame A, slab windows are opening beneath Washington, Vancouver Island, and Queen Charlotte Islands, and asthenosphere underlies most of British Columbia. The subducted portion of the Resurrection plate has foundered into the mantle since 39 Ma. The extinct ridge in the Gulf of Alaska remains in a stationary position during subduction. In Frame B, the slab window, which was underlying Washington, has migrated north to underlie southern British Columbia. The Resurrection plate remnants have equilibrated with the mantle, and a vast slab window is present beneath British Columbia. Late Oligocene to Recent magmatism is beginning inboard of the trench. Arrows shown represent 3 m.y. of plate motion.

throughout this interval (Barnes and Barnes, 1992; Davis et al., 1995). This intraplate-character forearc magmatism is represented by the Cascade Head, Cannery Hills, Yachats, Goble, and Gray's River basalts. A small, trench-localized Pacific-Farallon slab window was present below Washington and Oregon throughout this interval and can be linked to the forearc magmatism through anomalous geochemical signatures. This forearc magmatism was synchronous with Cascades arc-related magmatism and was dominantly submarine but was subaerial near the top of the succession.

35–25 Ma

Between 35 and 25 Ma, the last vestiges of the captured Eshamy plate were subducted, and, aside from the Admiralty Island area, forearc magmatism ceased in Alaska (Fig. 10). Farther south, the Pacific-Farallon slab window migrated northward away from Oregon, where forearc magmatism was also shutting off. Northward slab window migration was facilitated by subduction of a long Pacific-Farallon transform segment. In the forearc area, the Pacific-Farallon slab window lay beneath Vancouver Island and the Queen Charlotte Islands.

Forearc magmatism was widespread and voluminous on the Queen Charlotte Islands, as represented by the Kano intrusions, Masset volcanics, and Tertiary dike swarms. The Kano intrusions were emplaced between 35 and 26.8 Ma in a northward-younging suite (Anderson and Reichenbach, 1991). Masset volcanism and related dike swarms were active throughout this interval. All were emplaced in an extensional to transtensional setting, evidenced by north-south-trending dike swarms and plutons, and graben-filling volcanics. Extension may have been the combined result of transform motion and slab window migration underneath the area. The E-MORB character of several Masset basalts provides geochemical evidence that subslab mantle came into contact with North American lithosphere via the slab window (Hamilton and Dostal, 2001). Northward younging of the Kano intrusions is consistent with northward slab window migration.

On Vancouver Island, forearc magmatism was complete by 35 Ma (Madsen et al., 2003); however, the tectonic model suggests that a slab window was present underneath this area throughout most of the 35–25 Ma interval. Mount Washington suite intrusions and Flores volcanics are commonly crosscut by late-stage dikes of unknown age, which are possibly related to the persistent presence of this heat source. These dikes were emplaced after solidification of the plutons they crosscut. During this interval, the slab window no longer extended

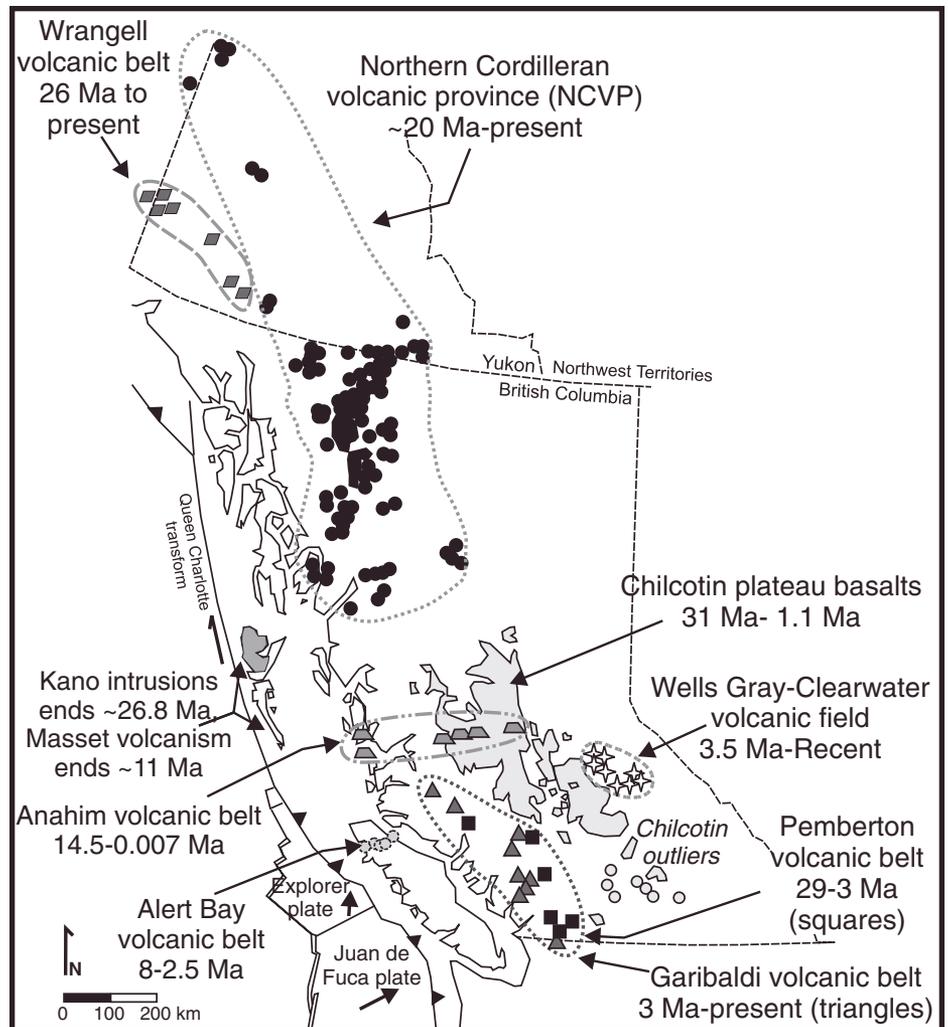


Figure 11. Late Oligocene to present forearc, arc, and within-plate magmatism in British Columbia, Yukon, and southeast Alaska. Volcanic fields are outlined. Individual volcanic centers of the northern Cordilleran volcanic province, Anahim belt, Wells Gray-Clearwater, and Wrangell fields are depicted by shapes within outlined fields. The Chilcotin Plateau is represented by the gray field in interior British Columbia. Arc magmatism is represented by squares and triangles of the Pemberton and Garibaldi belts, respectively. Oligocene to present forearc magmatism is found on Queen Charlotte Islands (Kano intrusions and Masset volcanics) and northern Vancouver Island (Alert Bay volcanic belt). The volcanic fields are described in the text. This figure is provided for comparison with the late Oligocene to present slab window reconstructions in Figures 10 and 12–14.

beneath Washington and Oregon. This configuration is consistent with cessation of forearc magmatism in that region at ca. 33 Ma (Barnes and Barnes, 1992) and the continuation of arc magmatism of the Cascades chain. By 25 Ma, the Pacific-Farallon ridge had reached a stable position in Queen Charlotte Sound.

As forearc magmatism waned, arc-related and anomalous transitional to alkalic and within-plate basaltic magmatism began to flourish (Fig. 11). Subduction-related magmatism of the Pemberton volcanic belt began at ca. 29 Ma

(K-Ar whole rock; Souther and Yorath, 1992). East and north of the Pemberton volcanic belt, the Chilcotin Plateau basalts commenced their long eruptive history. The Chilcotin basalts are composed of a series of transitional to alkalic basalt flows that began erupting at ca. 31 Ma (K-Ar whole rock; Mathews, 1989) and now cover over 50,000 km² of south-central British Columbia. Some Chilcotin group basalts are geochemically and isotopically similar to Pacific seamounts and appear to have been generated from partial melting of an oceanic

mantle-type source beneath British Columbia (Bevier, 1983). Other flows within the Chilcotin Plateau are geochemically more akin to transitional and backarc volcanism (Bevier, 1983). Chilcotin basalt geochemistry suggests that the alkalic and within-plate basalts were generated from melting of a mantle source with little crustal influence. The transitional basalts are backarc type and have a small crustal influence that may have been provided by the nearby subducted Juan de Fuca slab (Bevier, 1983).

The onset of subduction-generated arc volcanism in the southern Pemberton belt at ca. 29 Ma implies that the Juan de Fuca slab was present below southern British Columbia during this time interval. This also coincides with a subducted Juan de Fuca slab edge beneath the Chilcotin Plateau. Initiation of Pemberton belt arc magmatism occurred in a northward-younging succession beginning with the Chilliwack batholith (straddling Washington–British Columbia border) at 29 Ma. Farther north within the belt, the Coquihalla volcanics erupted from ca. 21 to 22 Ma, followed by 16–17 Ma volcanism near Pemberton. The Salal Creek stock in the northern area of the belt has been dated at ca. 8 Ma, and the northernmost Pemberton volcanic

center in the Franklin Glacier area is dated at 6.8 Ma (Baadsgaard et al., 1961; Richards and White, 1970; Richards and McTaggart, 1976; Wanless et al., 1978). The northward younging of Pemberton arc volcanic activity suggests that a slab window edge was migrating northward underneath British Columbia between at least the onset of arc volcanism at 29 Ma and eruption of the northernmost volcano at 6.8 Ma. The relationship between the Chilcotin Plateau and Pemberton volcanic belt indicates that central British Columbia was proximal to a slab edge or above a slab window at different times within this interval. This magmatic relationship provides a rough estimate of where the slab window edge may have been in British Columbia at ca. 30 Ma. This location is consistent with the model reconstructions (Stacey, 1974; Thorkelson and Taylor, 1989), which suggest that the slab window lay beneath most of British Columbia and Yukon, with a southern boundary situated underneath the Chilcotin Plateau.

25–15 Ma

The name Juan de Fuca plate applies to a remnant of the former Farallon plate after ca. 28 Ma.

The name change was necessary as the Farallon plate became segmented when the East Pacific Rise intersected the continental margin offshore of California. The northern segment became known as the Juan de Fuca plate, and the southern remnant, the Rivera plate.

Between 25 Ma and 15 Ma, the Pacific–Juan de Fuca (formerly Pacific–Farallon) ridge–trench intersection remained in a semiconstant position in Queen Charlotte Sound, while the Pacific–Juan de Fuca ridge underwent a series of ridge-propagation events (Fig. 12; Wilson, 1988). In the forearc, the Queen Charlotte Islands were situated above the slab window but in close proximity to, or partially overlying the edge of, the subducted Pacific plate. On the Queen Charlotte Islands, synextensional Masset volcanics and related dikes were apparently still being emplaced, according to the youngest igneous date of 11 Ma (K–Ar whole rock).

Farther inboard, the Pacific–Juan de Fuca slab window was located under the northern half of British Columbia and the Yukon Territory, and arc volcanism and ubiquitous alkalic mafic volcanism commenced. In Alaska, arc volcanism of the Wrangell volcanic belt began at ca. 26 Ma, followed by the transitional and alkalic volcanism of the Wrangell volcanic field at ca. 18 Ma. Alkalic and transitional Wrangell lavas have geochemical signatures consistent with melting and mixing between E-MORB, normal (N)-MORB, and slab-derived components and have been described as leaky transform magmatism (Skulski et al., 1991). However, a preferred depiction is that this volcanic succession was generated above the northeastern edge of the Pacific plate, i.e., the northwestern margin of the large Pacific–Juan de Fuca slab window, which was situated underneath most of northern British Columbia and Yukon at that time (Thorkelson and Taylor, 1989).

In southern British Columbia, the subduction-driven arc magmatism of the Pemberton volcanic belt continued. East and north of the Pemberton arc, new voluminous mafic and alkalic magmatism of nonsubduction character was activated. This new inboard activity included the northern Cordilleran volcanic province (Edwards and Russell, 1999) and the Cheslatta Lake suite portion of the Chilcotin basalts (Anderson et al., 2001). The Cheslatta Lake suite comprises a mafic igneous suite of alkaline to transitional basalt and began erupting in mid-Miocene time, ca. 21 Ma. Cheslatta magmatism is dominantly alkalic, and some flows demonstrate near primitive-mantle melt compositions and/or ultramafic mantle-derived xenoliths (Anderson et al., 2001). These magmas have been interpreted as melts of an ocean-island-type mantle source.

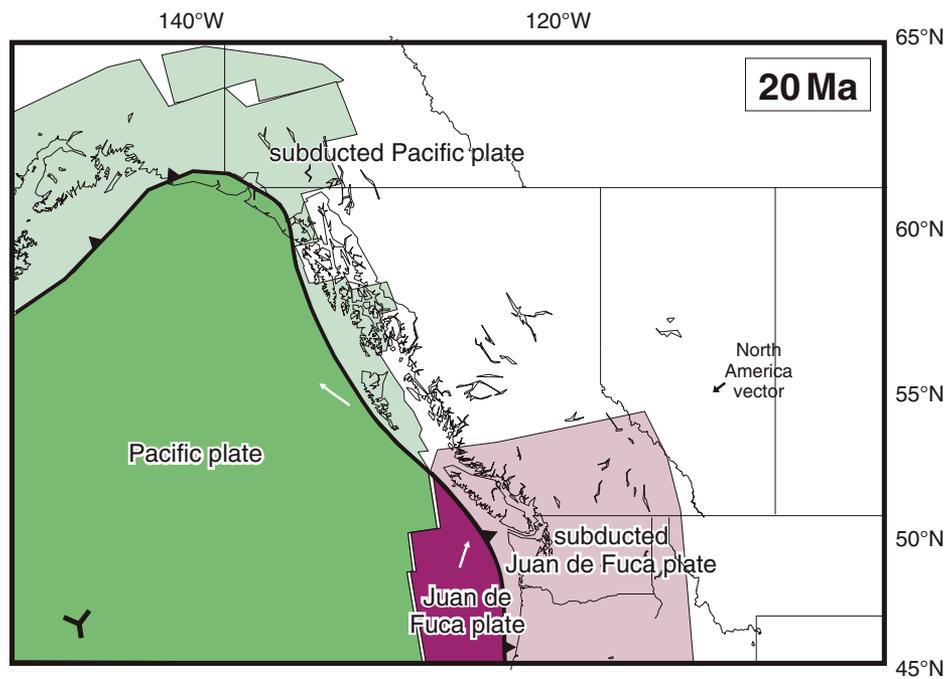


Figure 12. Time slice of the tectonic model at 20 Ma. After 28 Ma, the Farallon plate is referred to as the Juan de Fuca plate. The black Y-shaped feature represents the Great Magnetic Bight now preserved offshore of Alaska. During this interval, a large slab window was present under northern British Columbia and parts of Yukon Territory, and was responsible for the anomalous inboard magmatism shown in Figure 11. Vectors represent 3 m.y. of plate motion

The northern Cordilleran volcanic province (Fig. 11) comprises a large volcanic belt that extends across the Yukon Territory and northern British Columbia between the Tintina and Denali faults and includes volcanic centers previously assigned to the Stikine volcanic belt (Edwards and Russell, 1999). Volcanism associated with the northern Cordilleran volcanic province began at ca. 20 Ma in northern British Columbia and Yukon, but was minor in this time interval. The volcanic rocks are dominantly alkali olivine basalt, hawaiite with lesser nephelinite, basanite, peralkaline phonolite, trachyte, and comendite. The most Mg-rich and alkalic basalts of the volcanic province have trace-element characteristics consistent with an ocean-island basalt source and have been compared to young basalts of the U.S. Basin and Range Province (Edwards and Russell, 1999, 2000). The northern Cordilleran volcanic province has been described as the product of incipient continental rift magmatism above a slab window, with eruptive events governed by periods of compression and tension of the continental margin related to subtle changes of the motions of the Pacific plate relative to North America (Edwards and Russell, 1999, 2000).

15–5 Ma

From 15 to 5 Ma, the Pacific–Juan de Fuca ridge continued to propagate, and its triple junction with North America migrated slowly northward in the area between the southern tip of Queen Charlotte Islands and northern Vancouver Island (Fig. 13). In the forearc area, the eastern portion of the Queen Charlotte Islands was underlain by the Pacific–Juan de Fuca slab window, but the western areas were possibly underlain by a small protrusion of subducted Pacific plate. Queen Charlotte Basin extension was completed in this interval, and the Queen Charlotte fault became more transpressive starting at ca. 6 Ma (Rohr et al., 2000).

During this interval, Vancouver Island was dominantly underlain by the subducted slab of the Juan de Fuca plate. However, at ca. 9–8 Ma, the southern border of the window migrated south during a ridge adjustment. The migration of the slab window below Vancouver Island was concurrent with the onset of volcanism of the Alert Bay volcanic belt at ca. 8 Ma on northern Vancouver Island. The Alert Bay belt forms a trench-normal linear trend of alkalic volcanics and minor intrusions that persisted until ca. 2.5 Ma. Armstrong et al. (1985) attributed this volcanic belt to plate-edge-related magmatism generated near the subducted edge of the Juan de Fuca plate.

In inboard areas, the Pacific–Juan de Fuca slab window was situated beneath northern British Columbia and Yukon Territory. A new magmatic

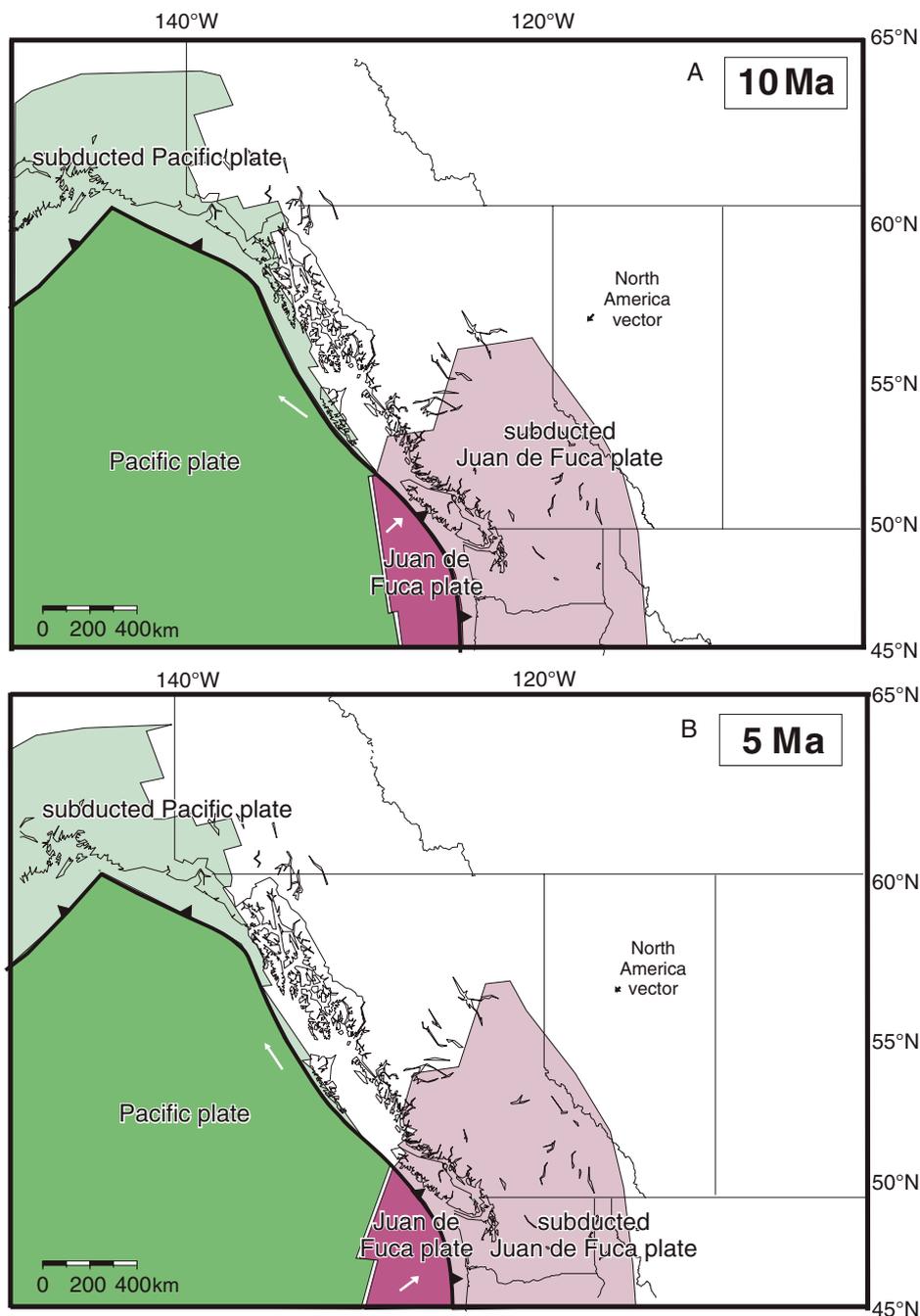


Figure 13. Time slices from the tectonic model at (A) 10 Ma and (B) 5 Ma. A large, stable slab window present underneath northern British Columbia has persisted until present. Queen Charlotte Islands are possibly underlain by a small protrusion of Pacific plate, but the plate boundary may instead be transpressional. Regardless, Queen Charlotte Islands are dominantly underlain by asthenosphere and not subducted Pacific slab. Vectors displayed on plates represent 3 m.y. of plate motion.

chain, the Anahim volcanic belt, was established while mafic mantle-derived magmatism of the Chilcotin Group and the northern Cordilleran volcanic province persisted. The Anahim volcanic belt is a west-east-trending chain of alkaline

basalts and associated peralkaline differentiates that began erupting at ca. 14.5 Ma in the west and progressed until ca. 7 ka in the east (K-Ar; ages summarized in Bevier, 1989). Anahim belt basalts are mantle-derived and have little to no crustal

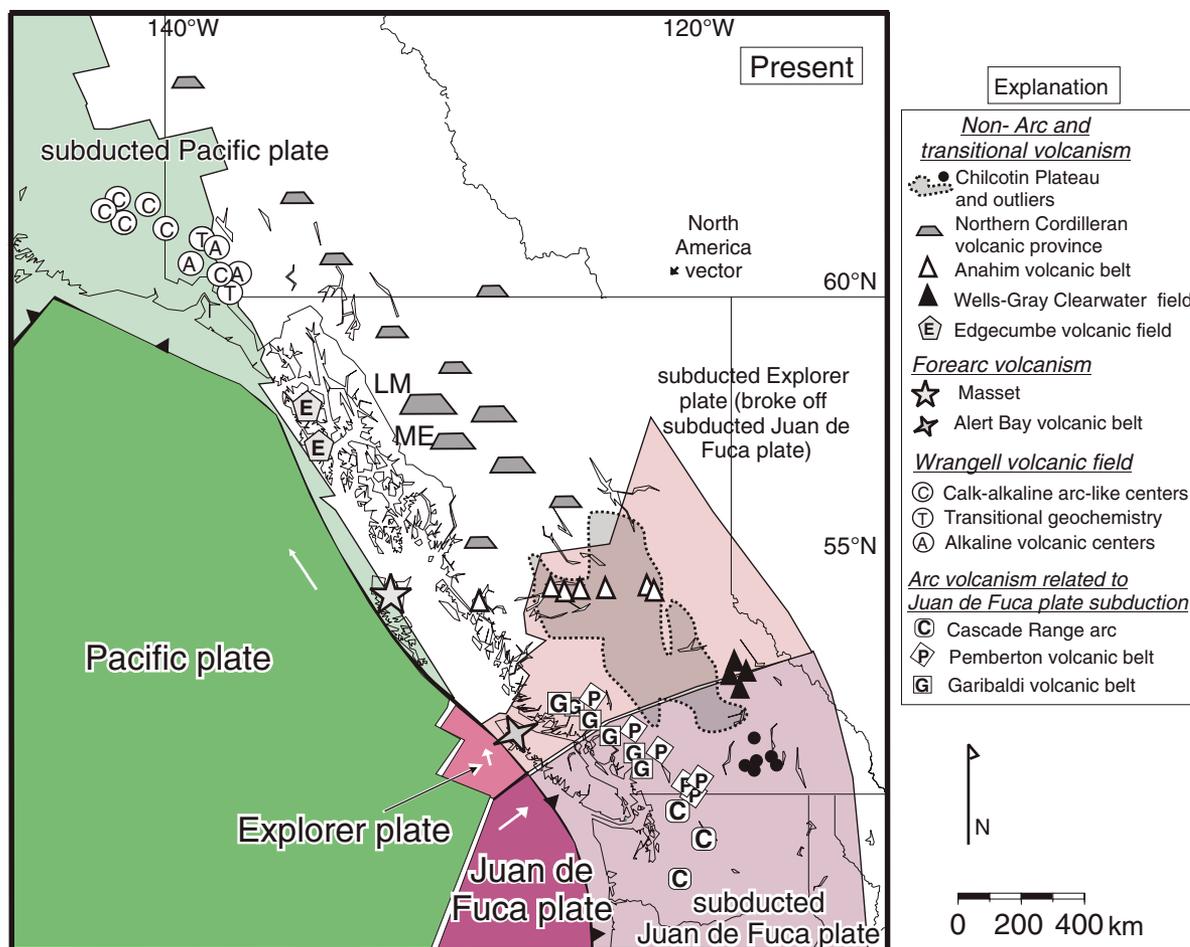


Figure 14. The final frame of the tectonic model, which depicts the present-day tectonic setting of Alaska, Yukon Territory, British Columbia, Washington, and Oregon. At 4 Ma, the Explorer plate broke free from the Juan de Fuca plate, possibly along the small circle defined by the Juan de Fuca Euler pole, as depicted at top. Superimposed on this diagram are forearc and inboard magmatic features. These magmatic features are late Oligocene to Recent in age. LM—Level Mountain, ME—Mount Edziza

signature. Sr and Pb isotopes suggest that the Anahim volcanic belt was derived from a mantle source similar to that which generated Pacific Ocean seamounts (Bevier, 1989). The age progression and linear trend of magmatism is a possible result of hotspot activity (Bevier et al., 1979; Bevier, 1989); however, the Anahim belt has also been described as an edge effect of the subducted Juan de Fuca plate in the mantle (Stacey, 1974). The slab window model indicates that subducted plate-edge effects are a plausible explanation for the generation of the Anahim belt.

Between 15 and 5 Ma, the Chilcotin Plateau basalts and the northern Cordilleran volcanic province experienced pulses of activity. Increased activity in the Chilcotin Plateau occurred at 16–14 Ma and 9–6 Ma (Mathews, 1989). The central part of the northern Cordilleran volcanic province exhibited a magmatic pulse from 8 to 4 Ma (Edwards and Russell, 1999).

5–0 Ma

From 5 Ma to present, ridge propagation and fracture development complicated the tectonic history of the Juan de Fuca ridge. At ca. 4 Ma, the Juan de Fuca plate fractured along the Nootka fault zone during an interval of ridge propagation, reorientation, asymmetric spreading, and transform elongation (Riddihough, 1977, 1984; Wilson, 1988; Botros and Johnson, 1988). The northern section of the Juan de Fuca plate became the Explorer plate (Fig. 14). After this separation, the two plates moved independently, with the Juan de Fuca plate continuing to subduct with a similar vector as before and the Explorer plate progressively slowing as it jammed against the North American plate and later becoming partially coupled to the north-moving Pacific plate (Riddihough, 1977). The independent movement of the Explorer plate

suggests that both the subducted and oceanic portions of the Juan de Fuca plate tore away from the Explorer plate along the Nootka fault, eliminating the effect of slab pull on the Explorer plate and allowing the Juan de Fuca plate to continue to subduct unimpeded. As depicted by Riddihough (1977), the Nootka fault continues beneath North America. We suggest that the subsurface Nootka fault describes a small circle about the Euler pole of the Juan de Fuca plate and separates the subducted slabs of the Explorer and Juan de Fuca plates underneath the interior areas of British Columbia.

The Wells Gray–Clearwater volcanic field was established in British Columbia at ca. 3.5 Ma and was active into Holocene time (Hickson, 1987). This volcanic field is located in southeastern British Columbia, ~250 km inboard of synchronous Garibaldi arc magmatism and is along-strike from the oceanic expression of the

Nootka fault zone. The volcanics are dominantly alkali olivine basalt, with some flows containing mantle xenoliths. Sr and Pb isotopes reflect contamination by a radiogenic crustal component, possibly from the underlying Kootenay terrane (Bevier, 1989). Basalts of the Wells Gray–Clearwater field have been regarded as the far eastern extent of the Anahim volcanic belt (Rogers and Souther, 1983; Hickson and Souther, 1984); however, this relationship was called into question because the age–location trend does not extend into the Wells–Gray Clearwater area, and the Wells Gray–Clearwater field is not along trend with the Anahim belt (Souther, 1986; Hickson, 1987). The Wells Gray volcanics have previously been attributed to thinning crust and the presence of crustal penetrating structures (Gabrielse and Yorath, 1991; Hickson, 1987; Hickson et al., 1995).

We suggest that the subducted extension of the Nootka fault may be the underlying cause of the alkalic composition of the Wells Gray–Clearwater volcanic field. The magmatism may have been largely generated by asthenospheric upwelling facilitated by displacement along the fault. If the fault had a component of vertical tearing, to accommodate possible different dip angles between the Explorer and Juan de Fuca slabs, subslab asthenosphere could flow upward into the mantle wedge. Similarly, if the displacement had a component of extension, a horizontal slab window–like gap would have formed, again providing a pathway for upwelling mantle. In either case, the uprising asthenosphere could have undergone low degrees of decompressional melting and interacted with North American lithosphere to yield within-plate compositions.

Chilcotin volcanism, Anahim belt volcanism, and northern Cordilleran volcanic province magmatism remained active into Holocene time. Bimodal volcanism of tholeiitic affinity began in the Edgumbe volcanic field, Alaska, in Holocene time. Edgumbe volcanic field lavas were derived from mantle melting generated in response to movement along the transform fault separating the Pacific plate from North America, coupled with crustal anatexis, and do not demonstrate a subducted slab component, all of which is consistent with the outcome of the tectonic model (Myers and Marsh, 1981). The Chilcotin Plateau lavas experienced a pulse at 3 Ma (Mathews, 1989), and the northern Cordilleran region experienced a pulse at 2 Ma (Edwards and Russell, 1999). After ca. 3 Ma, the Pemberton volcanic belt shifted westward and the present-day Garibaldi volcanic belt was established (Barr and Chase, 1974; Green et al., 1988). This westward shift of arc volcanism occurred within the portion of the Pemberton belt overlying the subducting Juan de Fuca plate

and may be related to steepening of the Juan de Fuca slab after the break-off of the Explorer plate (Green et al., 1988), providing evidence for different dip angles between the subducting Explorer and Juan de Fuca slabs and supporting the existence of a vertical gap between the two plates at depth, as suggested for the Wells Gray–Clearwater field.

The Salal glacier volcano of the Garibaldi volcanic belt is situated immediately south of the projected subsurface expression of the Nootka transform (Lawrence et al., 1984; Green et al., 1988). This volcano is associated with basaltic and alkalic volcanism, which contrasts with the normal calc-alkaline magmatism of the Garibaldi belt. Lawrence et al. (1984) postulated that this anomalous magmatism is related to the mantle interaction with the subducted edge of the Juan de Fuca plate and/or Explorer plates in the vicinity of the subducted Nootka transform. An outburst of Quaternary-age volcanism occurred in the interior of British Columbia, east of the Garibaldi volcanic arc. The Quaternary Okanagan valley basalts are distinguished from older flows by their valley-filling morphology. These valley basalts have not yet been geochemically characterized.

The last frame of the model shows a large slab window extending below northern British Columbia and Yukon. It is bounded to the west by a small amount of subducted Pacific plate that has subducted below the Queen Charlotte transform fault. Geophysical studies of the Queen Charlotte region, however, suggest that the North America and Pacific plates may instead be in a transpressive regime (Mackie et al., 1989; Rohr et al., 2000), suggesting that the subduction component shown by the calculated vectors may instead be taken up as transpression. If this is the case, the slab window would be bounded to the west by the Queen Charlotte fault, and the flange of subducted Pacific crust may have been thermally eroded, or may have torn away from the trench and foundered into the mantle. The model is consistent with the work of Frederiksen et al. (1998), who suggested that low-velocity anomalies observed in the northern Cordillera are due to upwelling mantle in a slab window setting near the subducted edge of the Pacific plate. Furthermore, Preece and Hart (2004) documented adakitic volcanic centers within the younger than 5 Ma Wrangell volcanic belt near the proposed subducted edge of the Pacific plate. These adakites may be attributable to slab-edge melting in this slab window environment (cf. Thorkelson and Breitsprecher, 2005).

According to our tectonic model, the southern boundary of the slab window presently lies just north of Vancouver Island. Inboard, the southern

slab window boundary is a complex zigzag configuration striking 030°NNE. The subducted slab surface expression, which was terminated at the 300 km isobath, is located near the British Columbia–Alberta border, approximately parallel to and 925 km from the trench. The present-day slab window underlies the northern Cordilleran volcanic province, and the subducted slab edge is situated near the northwest extent of the Chilcotin Plateau.

CONCLUSIONS

This paper provides a new plate-tectonic model for the northeastern Pacific Basin and northwestern North America from 53 Ma (middle Eocene) to present. Most of the Eocene to Recent forearc magmatism that occurred within in a semicontinuous belt along the forearc areas of Alaska, British Columbia, Washington, and Oregon is explainable in terms of ridge subduction and slab window tectonics. The model was constructed in 1 m.y. increments, where movement of oceanic plates and ridge–transform sets was governed by vectors from plate-motion studies (Engebretson et al., 1985; Lonsdale, 1988; Norton, 1995). The model subscribes to the theory of interaction of spreading ridges and oceanic transforms with the subduction zone as the principal cause of forearc magmatism. Eight new U–Pb dates for forearc magmatism on Vancouver Island were integrated with previously published dates for Alaska, British Columbia (Queen Charlotte Islands and Vancouver Island), Washington, and Oregon to provide the space-time framework for tectonic reconstruction. The locations of the plate-boundary intersections and associated slab windows were constrained primarily by the distribution and ages of well-dated forearc igneous centers, until ca. 20 Ma, when the configuration and location of ridge–trench intersections became constrained by the magnetic record. The forearc magmatic record is not reconcilable with the Kula–Farallon slab window alone (Haeussler et al., 2003a, and references therein; Breitsprecher et al., 2003), and is instead better explained by concurrent development of slab windows involving ridges among the Kula plate, Farallon plate, Resurrection plate (Haeussler et al., 2003a), and Eshamy plate (this paper).

Despite the forearc emphasis of the model, the positions of plate boundaries and slab windows of the multiridge subduction model also agree with inboard magmatic and structural features of the Cordillera, including the Challis–Kamloops magmatic belt, the Chilcotin Plateau lavas, the Anahim volcanic belt, the northern Cordilleran volcanic province, the Wrangell volcanic belt, and Edgumbe volcanic field.

The model is also in accord with the onset of arc volcanism in the Pemberton and Garibaldi belts and the anomalous Wells Gray–Clearwater volcanic field. Regional tectonic features explained in the context of the slab window model include widespread Eocene-aged exhumation of core complexes, strike-slip faulting throughout northwestern North America, and the formation, transport, and accretion of Eocene terranes, such as the Crescent/Siletz and Yakutat terranes. Altogether, the Cenozoic magmatic history of northwestern North America is explained well by slab window formation and migration.

ACKNOWLEDGMENTS

Funding was provided by grants to D. Thorkelson from the Slab Window Project of the U.S. Geological Survey, Alaska, and the Natural Sciences and Engineering Research Council of Canada. Support was also provided by the Geological Survey of Canada through Bob Anderson. We thank Nick Massey for sharing geochemical data and perspectives, Bohdan Podstawskyj for U-Pb zircon age determinations, Katrin Breitsprecher, Dwight Bradley, Peter Haeussler, Mark T. Brandon, and Stephen Johnston for valuable discussions, and Jacques Houle for advice in the field. Dwight Bradley, Terry Pavlis, and editor Randy Keller provided valuable reviews that led to substantial improvements.

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MANUSCRIPT RECEIVED BY THE SOCIETY 3 MAY 2005
 REVISED MANUSCRIPT RECEIVED 2 NOVEMBER 2005
 MANUSCRIPT ACCEPTED 4 NOVEMBER 2005