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EVOLUTION OF THE NORTH AMERICAN CORDILLERA

William R. Dickinson
Department of Geosciences, University of Arizona, Tucson, Arizona 85721; email: wrdickin@geo.arizona.edu

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■ **Abstract** The Cordilleran orogen of western North America is a segment of the Circum-Pacific orogenic belt where subduction of oceanic lithosphere has been underway along a great circle of the globe since breakup of the supercontinent Pangea began in Triassic time. Early stages of Cordilleran evolution involved Neoproterozoic rifting of the supercontinent Rodinia to trigger miogeoclinal sedimentation along a passive continental margin until Late Devonian time, and overthrusting of oceanic allochthons across the miogeoclinal belt from Late Devonian to Early Triassic time. Subsequent evolution of the Cordilleran arc-trench system was punctuated by tectonic accretion of intraoceanic island arcs that further expanded the Cordilleran continental margin during mid-Mesozoic time, and later produced a Cretaceous batholith belt along the Cordilleran trend. Cenozoic interaction with intra-Pacific seafloor spreading systems fostered transform faulting along the Cordilleran continental margin and promoted incipient rupture of continental crust within the adjacent continental block.

INTRODUCTION

Geologic analysis of the Cordilleran orogen, forming the western mountain system of North America, raises the following questions: 1. When was the Cordilleran system born, and from what antecedents; 2. which rock masses are integral to the Cordilleran continental margin, and how were they formed; 3. which rock masses were incorporated into the Cordilleran realm by tectonic accretion, and what were their origins; and 4. what geologic processes are promoting distension and disruption of the Cordilleran system today?

Figure 1 is a chronostratigraphic diagram of Cordilleran rock assemblages showing their relationships to major phases of Cordilleran evolution. The Cordilleran edge of the Precambrian basement, which forms the Laurentian craton, was first delineated by rifting to form a passive continental margin, along which a thick Neoproterozoic to Devonian miogeoclinal prism of sedimentary strata was deposited. From Late Devonian to Early Triassic time, oceanic allochthons were successively thrust across the miogeoclinal strata as internally deformed tectonic

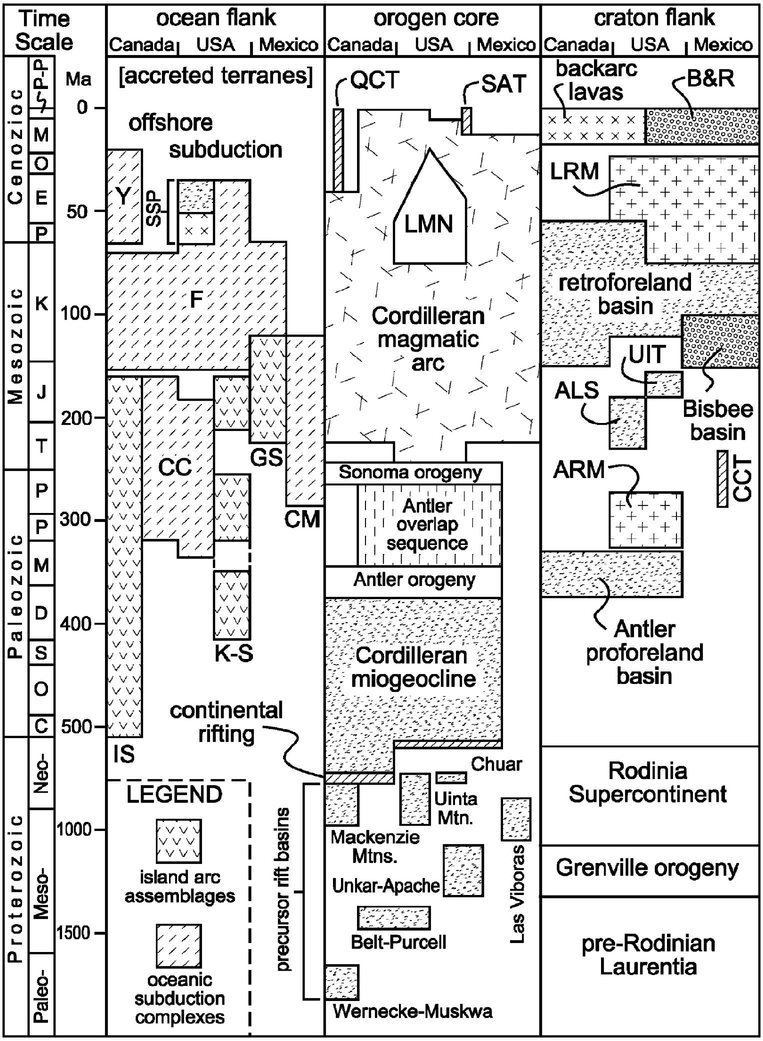

Figure 1 Schematic chronostratigraphic diagram of major Cordilleran rock assemblages (note changes in timescale at 100 Ma and 500 Ma). Canada includes the adjacent panhandle of southeastern Alaska, and Mexico includes the USA-Mexico border region south of the Colorado Plateau. Accreted island-arc assemblages: GS, Guerrero superterrane; IS, Insular superterrane; K-S, accreted arcs of Klamath Mountains and Sierra Nevada foothills. Subduction complexes: CC, Cache Creek; CM, central Mexico; F, Franciscan; Y, Yakutat. Transform faults (*diagonally ruled bars*): CCT, California-Coahuila; QCT, Queen Charlotte; SAT, San Andreas. Other features: ALS, Auld Lang Syne backarc basin; ARM, Ancestral Rocky Mountains province; B&R, Basin and Range taphrogen; LRM, Laramide Rocky Mountains province (LMN, Laramide magmatic null); SSP, accreted Siletzia and overlying forearc basin; UIT, Utah-Idaho trough.

assemblages accreted to the continent. An arc-trench system initiated along the modified continental margin in Triassic time was the tectonic regime that produced Mesozoic-Cenozoic subduction complexes and batholiths most characteristic of the Cordilleran orogen. During the subduction of oceanic lithosphere beneath the Cordilleran margin, Jurassic-Cretaceous accretion of intraoceanic island arcs contributed to the outward growth of the continental block. Beginning in mid-Cenozoic time, impingement of intra-Pacific seafloor spreading systems on the subduction zone at the continental margin gave birth to transform fault systems lying near the edge of the continental block and to associated inland deformation that distended continental crust previously overthickened by Cordilleran orogenesis.

On paleotectonic maps showing the distributions of Cordilleran rock assemblages adapted in part from Dickinson (2000, 2001, 2002) and Dickinson & Lawton (2001a,b; 2003), rock masses are plotted on present geography, with state and province boundaries for orientation, without palinspastic restoration to correct for distortion of rock masses by deformation. Offsets of rock masses across major Cenozoic strike-slip faults are shown, however, and curvatures of tectonic trends by oroclinal bending are indicated by annotations where appropriate.

To aid analysis of accretionary tectonics, the North American Cordillera has been subdivided into nearly 100 formally named tectonostratigraphic terranes (Coney et al. 1980) separated by faulted boundaries of varying tectonic significance and structural style (Silberling et al. 1992). For graphic display at feasible scale, various terranes are combined into generic groupings.

CORDILLERAN OROGEN

The Cordilleran mountain chain of western North America is an integral segment of the Circum-Pacific orogenic belt, which extends along a great circle path for 25,000+ km from the Antarctic Peninsula to beyond Taiwan (Figure 2). The length of the Cordilleran orogen from the Gulf of Alaska to the mouth of the Gulf of California is ~5000 km, or ~20% of the total length of the orogenic belt.

Characteristic geologic features of the Circum-Pacific orogenic belt derive from persistent subduction of oceanic lithosphere at trenches along the flanks of continental margins and offshore island arcs linked spatially to form a nearly continuous chain along the Pacific rim (Figure 2). The rock assemblages of subduction zones where oceanic plates are progressively consumed and of the parallel magmatic arcs built by related igneous activity are the prime signatures of Circum-Pacific orogenesis in the rock record. The oldest rock assemblages of the Cordilleran continental margin that reflect this style of tectonism mark initiation of the Cordilleran orogenic system in mid-Early Triassic time. Older rock assemblages exposed within the mountain chain record preceding tectonic regimes of different character.

Global Orogenic Patterns

In the Philippine-Indonesian region, the Circum-Pacific orogenic belt intersects the Alpine-Himalayan orogenic belt, which is aligned along a different great circle

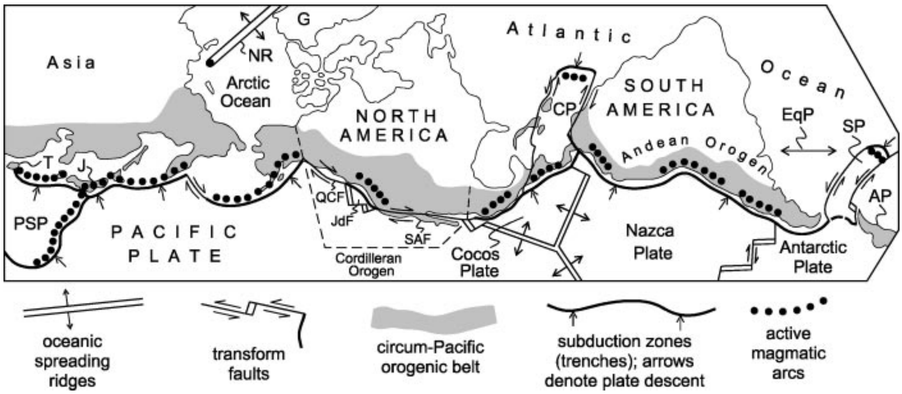

Figure 2 Position of the Cordilleran orogen of western North America along the Circum-Pacific orogenic belt (after Dickinson et al. 1986). Mercator projection with pole at 25° N Lat, 15° E Long (EqP is equatorial plane of projection). AP, Antarctic Peninsula; C, Cascades volcanic chain; CP, Caribbean plate; G, Greenland; J, Japan; JdF, Juan de Fuca plate; NR, Nansen Ridge (northern extremity of Atlantic spreading system); PSP, Philippine Sea plate; QCf, Queen Charlotte fault; SAf, San Andreas fault; SP, Scotia plate; T, Taiwan.

of the globe (Figure 3). Both orogenic belts relate to the breakup of the Permian-Triassic supercontinent of Pangea beginning early in Mesozoic time, but in different ways, as the Atlantic and Indian oceans opened to disrupt Pangea by seafloor spreading. Alpine-Himalayan evolution has involved the successive juxtaposition of disparate continental blocks (e.g., Africa, India, Australia against Eurasia) at suture belts marking the former positions of trenches where intervening ocean basins were closed by plate consumption (Figure 3), but no crustal blocks of comparable size have lodged against the Pacific margin of the Americas.

The ancestral Circum-Pacific orogenic system along the margin of Pangea was born along a great circle path (Le Pichon 1983), rimming an ocean (Panthalassa) that was effectively a paleo-Pacific realm with a Tethyan gulf that projected into the angle between Laurasian and Gondwanan segments of Pangea (Figure 4). The great circle configuration was maintained as expansion of the Atlantic and Indian Oceans led to a modern Pacific only 60% the size of the paleo-Pacific (Le Pichon et al. 1985) by insertion of Australia and its surrounding seas into the Pacific arena to step the Pacific rim eastward for ~5500 km along the Indonesian archipelago (Figure 3). During Circum-Pacific evolution, intra-Pacific seafloor spreading renewed oceanic lithosphere so rapidly that no vestiges of pre-Jurassic paleo-Pacific seafloor remain (Dickinson 1977).

Supercontinent History

The composite supercontinent of Pangea (Figure 4) formed during late Paleozoic time when Gondwana lodged against Laurasia along the Appalachian-Hercynian

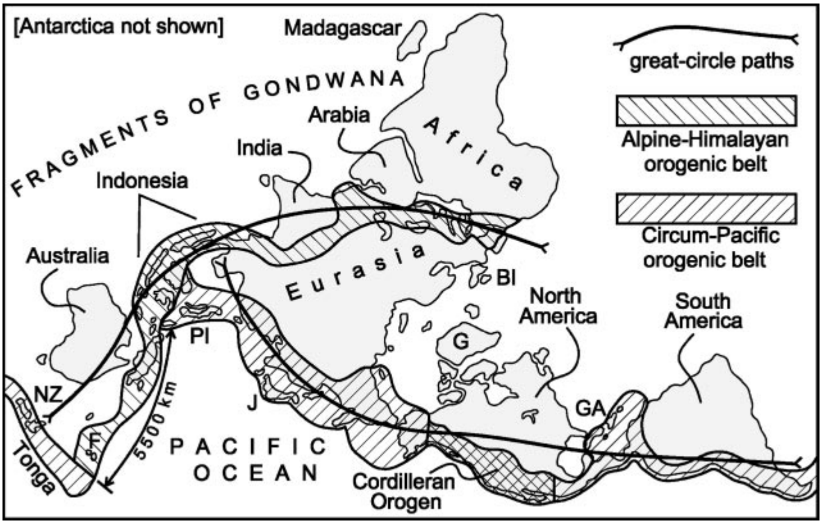

Figure 3 Distribution of continents in relation to the Alpine-Himalayan and Circum-Pacific orogenic belts (Cordilleran orogen: *cross-hatched*) in "circular" projection (after Challand & Rageau 1985). BI, British Isles; F, Fiji; G, Greenland; GA, Greater Antilles; J, Japan; NZ, New Zealand; PI, Philippine Islands.

orogen, a Paleozoic precursor of the modern Alpine-Himalayan system in that both achieved assembly of supercontinents (Pangea and Eurasia) through juxtaposition of previously separate continental blocks. Gondwana was a paleocontinent assembled in Neoproterozoic time (800–550 Ma) by juxtaposition of continental fragments across multiple internal suture belts (Meert & Van der Voo 1997). Laurasia included the ancient continental nuclei of Laurentia (North America) and Baltica (Europe), which had been conjoined in early Paleozoic time and linked in mid-Paleozoic time with Siberia. The Paleozoic precursor of the modern Circum-Pacific system was the Gondwanide orogenic belt, which lay along the Panthalassan (paleo-Pacific) margin of Gondwana, from South America past Antarctica to Australia (Figure 4), where consumption of oceanic lithosphere proceeded without interruption during the progressive assembly of Pangea (Ramos & Aleman 2000, Foster & Gray 2000).

The assembly of Pangea and its subsequent breakup during the assembly of Eurasia over the course of Phanerozoic time finds a Precambrian parallel in the geologic history of an earlier supercontinent, Rodinia, from which continental fragments were at first widely dispersed and then rearranged to form Gondwana and eventually Pangea. Rodinia was aggregated during Grenville orogenesis in Mesoproterozoic time (1325–1050 Ma), and the Cordilleran continental margin first took shape from Neoproterozoic breakup of Rodinia. As yet, however, there is no final consensus on the arrangement of continental cratons within the Rodinian

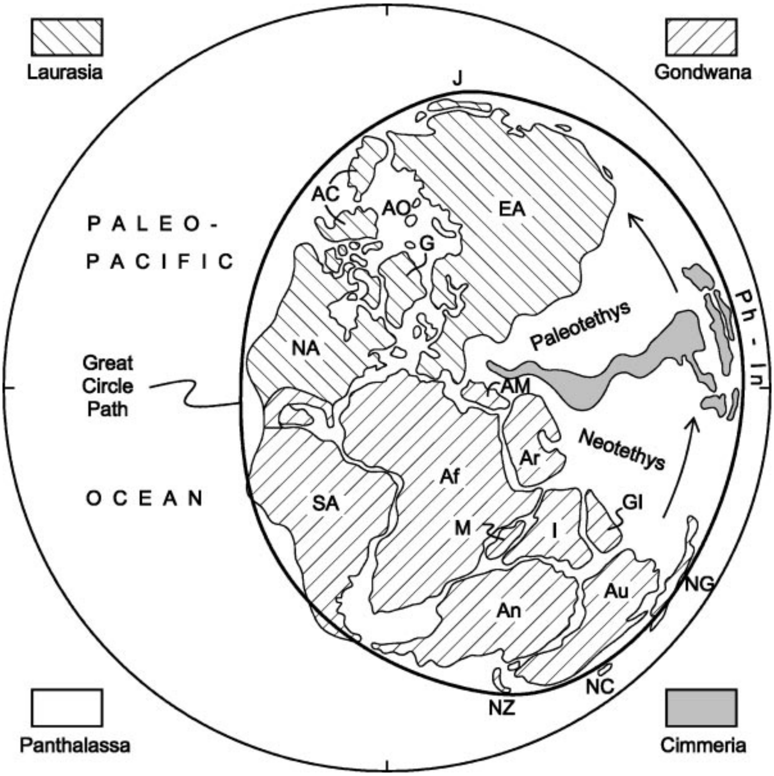

Figure 4 Permian-Triassic configuration of Pangea (Gondwana after Lawver & Scotese 1987) surrounded by Panthalassa (global sea including paleo-Pacific ocean and Tethys gulf) in Lambert equal-area projection (whole Earth). The Arctic Ocean closed by restoring transform slip of Alaska-Chukotka (Patrick & McClelland 1995). Arrows schematically denote the motion of Cimmerian landmasses in transit across the Tethys gulf, originating by rifting off the margin of Gondwana, toward Mesozoic accretion along the southern flank of Eurasia by closure of Paleothys as Neotethys opened. AC, Alaska-Chukotka; Af, Africa; AM, Asia Minor; An, Antarctica; AO, Arctic Ocean (closed); Ar, Arabia; Au, Australia; EA, Eurasia; G, Greenland; GI, Greater India; I, India; J; Japan; M, Madagascar; NA, North America; NC, New Caledonia; NG, New Guinea; NZ, New Zealand; Ph-In, Philippine-Indonesian archipelago; SA, South America.

supercontinent. Continental blocks suggested as conjugate to the Cordilleran rifted margin of Laurentia include Siberia (Sears & Price 2000), Antarctica-Australia (Dalziell 1992), Australia together with an unknown block farther north (Karlstrom et al. 1999), and China (Li et al. 2002). Of the various options, Siberia currently seems the most viable (Sears & Price 2003).

PRECAMBRIAN–PALEOZOIC MIOGEOCLINE

Along the Cordilleran margin of Laurentia, an elongate belt of thick sediment was deposited in Neoproterozoic and lower Paleozoic time as a miogeoclinal prism draped over a passive continental margin formed by rifting during the breakup of Rodinia (Figure 1). The narrow miogeoclinal belt truncates disrupted older Precambrian age provinces of interior Laurentia (Figure 5). The miogeoclinal prism thickened westward from a zero edge along a hinge line at the edge of the Laurentian craton. Miogeoclinal sedimentation continued, unbroken by tectonic disruption, until Late Devonian time, but the timing of its inception was apparently diachronous.

Cordilleran Rifting

North of the trans-Idaho discontinuity (Yates 1968), where the elongate trend of the miogeocline is offset by $>$250 km (Figure 5), basaltic rocks associated with glaciomarine diamictite in basal horizons of the miogeoclinal succession have been dated isotopically at 770–735 Ma (Armstrong et al. 1982, Devlin et al. 1988, Rainbird et al. 1996, Colpron et al. 2002). This time frame provides an age bracket for rifting that initiated deposition of the Windermere Supergroup along a newly formed passive continental margin open to the west in Canada (Ross 1991, MacNaughton et al. 2000). South of the trans-Idaho discontinuity, coeval rifting apparently formed only intracontinental basins in which redbed units such as the Chuar Group (775–735 Ma) of the Grand Canyon were deposited (Timmons et al. 2001), with continental separation delayed in the Death Valley region to the west until after 600 Ma (Prave 1999).

The subparallelism of the trans-Idaho discontinuity and a paleotransform delimiting the southwest margin of Laurentia (Figure 5) suggests that both originated as transform offsets of the nascent Cordilleran margin. Miogeoclinal strata present locally along the trans-Idaho discontinuity form a narrow band exposed only within roof pendants of the Idaho batholith (Lund et al. 2003) and contain intercalated bimodal volcanic rocks (685 Ma), which perhaps reflect prolonged deformation along a marginal offset at the edge of the continental block during the evolution of the rifted Cordilleran margin.

Published subsidence analyses for the Cordilleran miogeocline in both Canada and the USA imply that postrift thermotectonic subsidence of the passive continental margin did not begin until 560–555 Ma in Early Cambrian time (Bond et al. 1983, Armin & Mayer 1983, Bond & Kominz 1984, Levy & Christie-Blick 1991). Replotting the subsidence curves for revised estimates of the beginning of Cambrian time (545 Ma versus 570 Ma) puts onset of thermotectonic subsidence at 525–515 Ma, still within Early Cambrian time on the revised timescale. Projecting subsidence curves backward in time to allow for 1.1–1.2 km of synrift tectonic subsidence ($\sim$2 km of sediment accumulation) in the Great Basin area of the USA (Levy & Christie-Blick 1991) suggests that rifting that led directly to

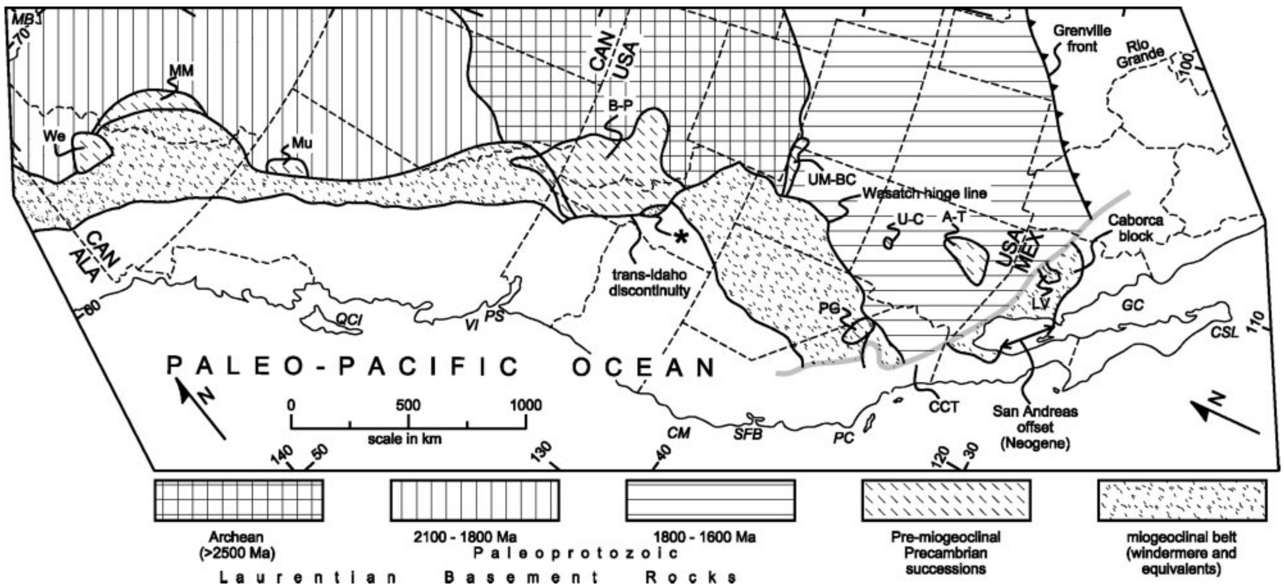

Figure 5 Neoproterozoic-Early Paleozoic Cordilleran miogeocline and premiogeoclinal sedimentary basins along the trend of the North American Cordillera (rock assemblages now present west of the miogeoclinal belt were added to the continental block after mid-Late Devonian time). Asterisk (*) denotes miogeoclinal strata along trans-Idaho discontinuity (Lund et al. 2003). Grenville front is margin of Mesoproterozoic Grenville orogen along which Rodinia was assembled. CCT is Permian-Triassic California-Coahuila transform (Dickinson 2000), which offset the Cordilleran miogeoclinal assemblage of the Caborca block by overprinting an older paleotransform system that delimited the early Paleozoic southwest margin of Laurentia (Dickinson & Lawton 2001a). See text for ages of premiogeoclinal successions. (A-T, Apache-Troy; B-P, Belt-Purcell; LV, Las Víboras; MM, Mackenzie Mountains; Mu, Muskwa; PG, Pahrump Group; U-C, Unkar-Chuar; UM-BC, Uinta Mountain—Big Cottonwood; We, Wernecke). Coastal locales (*italics*): CM, Cape Mendocino; CSL, Cabo San Lucas; GC, Gulf of California; MB, Mackenzie Bay of Arctic Ocean; PC, Point Conception; PS, Puget Sound; QCI, Queen Charlotte Islands; SFB, San Francisco Bay; VI, Vancouver Island.

continental separation south of the trans-Idaho discontinuity occurred during the interval 600–575 Ma (Armin & Mayer 1983). Isotopic dating of synrift volcanic rocks in southern British Columbia at 570 ± 5 Ma (Colpron et al. 2002) documents that active rifting persisted into latest Neoproterozoic time, north as well as south of the trans-Idaho discontinuity.

The indicated time span of 45–65 million years between initial rifting and onset of passive thermotectonic subsidence is comparable to the time span of 55 million years between initial development of Triassic rift basins and the first emplacement of Jurassic oceanic crust along the modern Atlantic continental margin (Manspeizer & Cousminer 1988). Cordilleran unconformities near the Precambrian-Cambrian time boundary (Devlin & Bond 1988, Lickorish & Simony 1995) may stem from reactivation of rift faults or from the influence of eustasy on a rift hinge undergoing flexure from sediment loading of oceanic crust offshore (Fedo & Cooper 2001).

The evidence for two rift events (Colpron et al. 2002), spaced 160–170 million years apart in pre-Windermere and latest Neoproterozoic time, suggests the possibility that two different continental blocks, one west of Canada and one west of USA-Mexico, were once conjugate with Laurentia, but no current Rodinian models are readily compatible with that interpretation. In any case, the onset of thermotectonic subsidence at the same time in both Canada and southward implies that the Windermere passive margin was reactivated at the time of the second rifting event, as suggested by stratal relationships near the USA-Canada border (Devlin 1989).

Precursor Rifts

A number of premiogeoclinal Precambrian sedimentary successions occur along the trend of the Cordillera but lack the longitudinal continuity of the overlying miogeocline (Figure 5). Each was deposited within an intracratonic basin formed by incipient rift extension within Rodinia or before its assembly (Figure 1). From the trans-Idaho discontinuity northward, isotopic dating of basin substratum, intercalated volcanics, and local intrusions establishes age brackets as follows for deep local rift troughs: Wernecke Supergroup, 1820–1710 Ma (Ross et al. 2001); Muskwa Assemblage, 1760–1660 Ma (Ross et al. 2001); Belt-Purcell Supergroup, 1470–1370 Ma (Evans et al. 2000, Luepke & Lyons 2001); Mackenzie Mountains Supergroup, 975–775 Ma (Rainbird et al. 1996); Uinta Mountain Group (and Big Cottonwood Formation), 975?–725? Ma. Farther south, thinner premiogeoclinal successions, deposited either in rift basins or on the craton, include the following: Unkar Group, Apache Group (including Troy Quartzite), and lower Pahrump Group, 1220–1070 Ma (Timmons et al. 2001), with the lowermost Apache Group as old as 1335 Ma (Stewart et al. 2001); Las Víboras Group, 1050?–850? Ma (Stewart et al. 2002); and the Chuar Group, 775–735 Ma (Timmons et al. 2001).

The precursor rift troughs may have acted as subregional guides helping to control the trend of eventual continental separation that initiated miogeoclinal sedimentation. In the Death Valley region, for example, diamictites of the Kingston Peak Formation in the upper Pahrump Group probably include correlatives of both syn-Windermere (∼750 Ma) rift fill and younger (∼600 Ma) synrift deposits at the base of the miogeoclinal succession (Prave 1999). Recent interpretations from deep seismic reflection profiles propose that pre-Windermere Canadian assemblages (Wernecke, Muskwa, Mackenzie Mountains) in the region north of the Belt-Purcell basin are all parts of a composite sediment prism built westward along an evolving passive continental margin that flanked the youngest basement components of the Laurentian craton and persisted over a time span that exceeded a billion years (Snyder et al. 2002). If so, no continental separation by Neoproterozoic rifting was required to form the Windermere continental margin, but the disparate age ranges and outcrop discontinuity of the pre-Windermere successions are difficult to reconcile with the postulate of a pre-Windermere passive continental margin continuous for the length of the Canadian Cordillera.

LATE PALEOZOIC–EARLY TRIASSIC ACCRETION

Between Late Devonian and Early Triassic time, internally deformed allochthons (Figure 6) of oceanic strata were thrust eastward as accretionary prisms across the seaward flank of the miogeoclinal belt when the margin of the Laurentian continental block was drawn into the subduction zones of intraoceanic island arcs that faced the Cordilleran margin and subducted offshore oceanic crust of marginal seas downward to the west (Dickinson 2000). Lithic constituents of the allochthons include pillow basalts, peridotite, and serpentinite of oceanic crust and subjacent mantle, as well as more voluminous argillite, ribbon chert, and turbidites of overlying seafloor sediment profiles. The turbidites of the allochthons include continental slope and rise deposits originally transported off the Laurentian margin and then thrust back toward the craton over the miogeoclinal shelf edge. Exposures where later tectonism and erosion has exhumed the thrust contact show that the allochthons traveled 100+ km across the structurally underlying miogeoclinal assemblage.

Antler-Sonoma Allochthons

Stratigraphic and structural analysis of the overthrust allochthons has documented their emplacement during two discrete episodes of incipient continental subduction termed the Antler and Sonoma orogenies in the USA. The two events were spaced ∼110 million years apart during comparatively brief intervals of time (∼25 million years each) spanning the Devonian-Mississippian and Permian-Triassic time boundaries (Figure 1). The two separate allochthons have been delineated with greatest confidence in the Great Basin of Nevada (USA), where allochthonous but unmetamorphosed oceanic facies of multiple Paleozoic horizons were thrust over autochthonous miogeoclinal facies of the same ages along the Roberts Mountains

(Antler) and Golconda (Sonoma) thrusts (Figure 6). Farther north, the lack of clearcut evidence for the presence of either allochthon along the trend of the trans-Idaho discontinuity (Figures 5 and 6) may stem from engulfment by batholiths, widespread erosion, and burial beneath volcanic cover, but may also reflect anomalous tectonic behavior along that atypical segment of the Cordilleran margin.

Within Canada, post-Triassic internal deformation and tectonic transport of Paleozoic Antler-Sonoma allochthons during Mesozoic arc-continent collision and subsequent retroarc thrusting complicate interpretations of their original character and positions (Hansen 1990, Ghosh 1995). Allochthons of both Antler (Smith & Gehrels 1991; 1992a,b) and Sonoma (Roback et al. 1994, Roback & Walker 1995) age have been identified in the Kootenay structural arc (Figure 6) spanning the USA-Canada border. Farther north, the Sylvester allochthon emplaced above the Cassiar platform (Figure 6) is composed exclusively of post-Devonian rocks (Nelson 1993) and apparently represents only the younger allochthon of Sonoma age. Nearby, however, the widespread and internally complex Yukon-Tanana terrane (Hansen 1988) probably includes both allochthons as well as underthrust miogeoclinal facies (Hansen & Dusel-Bacon 1998). Blueschists of both Devonian (~345 Ma) and Permian (270–260 Ma) ages are present in allochthonous Yukon-Tanana assemblages lying structurally above miogeoclinal strata along the west flank of the Cassiar platform and in the isolated Anvil allochthon (Figure 6) to the east (Erdmer et al. 1998). Juxtaposition of the Anvil allochthon against the Cassiar platform implies ~485 km of post-thrust dextral displacement along the Tintina fault (Figure 6).

Antler-Sonoma Foreland

Tectonic loads of the overthrust Antler-Sonoma allochthons downflexed the Laurentian margin to form an elongate system of markedly asymmetric proforeland sedimentary basins extending across the miogeoclinal belt into the fringe of the interior craton (Lawton 1994, Savoy & Mountjoy 1995). The extent of the Antler foreland basin is defined by an apron of clastic sediment shed toward carbonate platforms of the interior craton, but the Sonoma foreland basin is defined only by the limit of Triassic marine strata (Figure 6). Widespread syndepositional normal faulting of Antler age along the foreland belt in Canada (Gordey et al. 1987) can be interpreted as a response to local extension induced by flexure of the foreland basin floor (Smith et al. 1993).

Proximal sandstone petrofacies along the western fringe of the Antler foreland belt are dominantly quartzolithic, a composition reflective of sediment recycling from the uplifted accretionary prisms of allochthons exposed farther west as sediment sources (Smith et al. 1993). Near the Canada-Alaska border, the ages of detrital zircons in Cambrian and Devonian sandstones suggest derivation of foreland sediment near the northern end of the Cordilleran orogen (Figure 6) from the Paleozoic Innuitian-Ellesmerian orogen of the Canadian Arctic to the northeast (Gehrels et al. 1999).

During the time interval between Antler and Sonoma thrusting along the Cordilleran margin, part of the continental block extending as far west as the Antler foreland basin and thrust front in the USA was disrupted by intracontinental reverse faulting to form yoked basins and uplifts of the Ancestral Rocky Mountains province (Figure 6). The intracontinental deformation, centered on Pennsylvanian time (Figure 1), was related to sequential intercontinental suturing along the Ouachita orogenic belt (Figure 6), where the southern flank of the Laurentian craton was drawn progressively, from east to west, into a subduction zone along the leading edge of Gondwana during the assembly of Pangea (Dickinson & Lawton 2003).

Accreted Island Arcs

Segments of accreted Devonian and Permian island arcs, composed of volcanic and volcaniclastic strata and paired geotectonically with Antler and Sonoma accretionary prisms to the east, are present in the Klamath-Sierran region (Figure 6) of the Cordilleran orogen to the south of volcanic cover in the Pacific Northwest (USA). The Paleozoic Klamath-Sierran arc system evolved as a system of frontal arcs and remnant arcs during slab rollback related to closure of marginal seas between the offshore arc complex and the Cordilleran continental margin (Dickinson 2000).

Farther north in Canada, remnants of comparable Devonian to Permian arc assemblages (Rubin et al. 1990, Brown et al. 1991), underlying the Mesozoic arc assemblages of Quesnellia and Stikinia (Figure 6), are interpreted here as northern analogues of the accreted Klamath-Sierran island arcs. Both east and west of the Cassiar platform (Figure 6), overthrust allochthonous assemblages include both Devonian (365–340 Ma) and Permian (~260 Ma) granitic plutons of arc affinity (Mortensen 1992). Permian island-arc volcanics are closely associated with deformed seafloor volcanics within the internally complex Sylvester allochthon emplaced structurally above the Cassiar platform (Nelson 1993) and in correlative assemblages farther south (Ferri 1997). These occurrences of island-arc remnants within allochthonous Paleozoic assemblages suggest that severe structural telescoping in Canada during superposed mid-Mesozoic arc-continent suturing and later Mesozoic-Cenozoic retroarc thrusting closely juxtaposed island-arc and subduction-zone tectonic elements of Antler-Sonoma age that remain largely separate farther south.

MESOZOIC-CENOZOIC ARC-TRENCH SYSTEM

A Permian-Triassic (284–232 Ma) magmatic arc, built along the edge of Gondwanan crust in eastern Mexico (Dickinson & Lawton 2001a), was sustained by subduction of oceanic crust beneath present-day central Mexico (Figure 6). The northern margin of the subducting Mezcalera plate along the southwestern edge of

Laurentia was defined by the sinistral California-Coahuila transform, which offset the Caborca block of miogeoclinal strata, together with underlying basement and a structurally superposed allochthon of overthrust Paleozoic strata (Figure 6), from southern California into northwestern Mexico (Dickinson 2000, Dickinson & Lawton 2001a). Farther northwest, the transform fault obliquely truncated, along a northwest-southeast trend, island-arc complexes trending northeast-southwest that were accreted to the continental block in the Klamath-Sierran region by Antler-Sonoma orogenesis. Initiation of subduction beneath the truncated continental margin in mid-Early Triassic time (Dickinson 2000) closely preceded the breakup of Pangea, and was the earliest record of Cordilleran orogenesis as an integral facet of the circum-Pacific orogenic belt.

Subsequent evolution of the active Cordilleran continental margin was marked by incremental accretion of subduction complexes at a trench along the continental slope, and by arc magmatism involving both plutonism and volcanism along the edge of the continental block. Multiple imbricate thrust panels of accretionary mélange belts incorporate disrupted stratal successions of seafloor turbidites, argillite, and chert, together with pillow lavas of underlying oceanic crust and structural slices of peridotite and serpentinite derived from oceanic mantle, and with limestone enclaves representing carbonate platforms built on oceanic seamounts. Combined plutonic and volcanic contributions to arc magmatism were emplaced into and erupted through the composite Cordilleran crustal profile along a shifting belt of igneous activity that lay 100–250 km inland from the evolving subduction zone along the continental margin (Armstrong 1988, Armstrong & Ward 1991). The principal record of arc magmatism is a discontinuous alignment of deeply eroded Cretaceous granitic batholiths extending the full length of the Cordilleran orogen. Isotopic studies indicate that the granitic magmas were composed in part of juvenile mantle components and in part of recycled crustal materials (DePaolo 1981, Samson et al. 1991).

Mid-Triassic to Mid-Jurassic Cordilleran Arc

Volcanic assemblages and associated plutons of Upper Triassic to Middle Jurassic age developed within a continuous magmatic arc established along the margin of North America as modified by Antler-Sonoma tectonism. The central segment of the Triassic-Jurassic arc transected miogeoclinal and Laurentian cratonic crust along the truncated continental margin of the southwest USA (Busby-Spera 1988, Schweickert & Lahren 1993), but the arc trend extended southward across the Ouachita suture into Gondwanan crust of eastern Mexico (Dickinson & Lawton 2001a) and northward along the continental margin, as expanded by tectonic accretion, to merge with the Quesnellia or Nicola arc (Mortimer 1987) of the Canadian Cordillera (Figure 7).

The local preservation of forearc basins along the western flank of the nascent Cordilleran arc in both the Canadian Cordillera (Travers 1978) and the USA Pacific Northwest (Dickinson 1979) show that the arc-trench system faced west, subducting seafloor downward beneath North America, even where relations of the arc assemblage to Laurentian basement or miogeoclinal strata are unexposed. Past speculation that the Quesnellia arc was a freestanding intraoceanic structure only accreted to North America by later collapse of an intervening marginal sea or open ocean has been discounted by recent isotopic studies (Unterschutz et al. 2002, Erdmer et al. 2002). The backarc region was flooded in Canada by marine waters, but was occupied in the USA by desert ergs (Figure 7), with the accommodation space for both sedimentary assemblages probably provided by subsidence of the flank of the continental block under the geodynamic influence of a subducted slab in the mantle beneath (Lawton 1994).

West of the Triassic-Jurassic Cordilleran arc assemblage lies a paired subduction complex of mélange and variably deformed thrust panels of oceanic strata forming the Cache Creek terrane and its correlatives in the Canadian Cordillera, the central mélange belt (Baker terrane) of the Pacific Northwest, coeval assemblages in the central Klamath Mountains and the Sierra Nevada foothills of California, and remnants of the Arperos oceanic realm formed on the Mezcalera plate in central Mexico. This nearly continuous alignment of disrupted oceanic materials, conveniently termed the Cache Creek belt (Mortimer 1986), forms a suture zone trapped between the Triassic-Jurassic continental margin and various accreted arc assemblages lying farther west (Figure 7).

The suture belt is probably a compound subduction complex formed of combined tectonic elements added to the flank of North America at a trench lying just offshore from the Triassic-Jurassic Cordilleran arc but also accreted to the flank of intraoceanic arc structures as they approached the Cordilleran margin, with both components representing offscrapings from intervening paleo-Pacific seafloor. Cache Creek blueschists formed by subduction-zone metamorphism have yielded isotopic ages of 230–210 Ma (Late Triassic) in both Canada and the USA (Erdmer et al. 1998), where stratal components of the suture belt range in age from Carboniferous (locally Devonian) to Early or Middle Jurassic (Cordey et al. 1987, Blome and Nestell 1991, Cordey & Schiarizza 1993, Dickinson 2000, Struik et al. 2001, Orchard et al. 2001). In Mexico, where accretion of an intraoceanic arc to the continental margin occurred much later than farther north, only Permian to Early Cretaceous rocks are present within the suture belt of central Mexico (Dickinson & Lawton 2001a).

Mid-Jurassic to Mid-Cretaceous Arc Accretion

In Jurassic-Cretaceous time, tectonic accretion at the Cordilleran subduction zone was punctuated by the arrival of intraoceanic island arcs subducting seafloor downward to the west, rather than to the east, to produce arc-continent collisions (Godfrey & Dilek 2000, Ingersoll 2000, Dickinson 2001). For evaluating accretionary tectonism, a distinction must be drawn (Wright 1982) between incremental accretion within evolving subduction complexes (so-called disrupted terranes), even where far-traveled oceanic components are incorporated, and bulk accretion of tectonic elements transported intact, as integral exotic terranes, to the continental margin. Arc accretion expanded the continental edge by closing the Cache Creek suture and induced the subduction zone and the magmatic arc along the Cordilleran margin to step outward away from the continental interior. Subsequent Cordilleran arc magmatism was widely superimposed on the accreted arc and mélange terranes (van der Heyden 1992, Friedman & Armstrong 1995). The ages of the oldest superimposed plutons of the Cordilleran magmatic arc reflect north-south diachroneity of arc accretion from Middle Jurassic ($\sim$170 Ma) as far south as central California (Schweickert et al. 1999) to Early Cretaceous ($\sim$120 Ma) in Mexico (Dickinson & Lawton 2001a).

CANADIAN TECTONIC ELEMENTS In Canada, two principal accreted tectonic elements, the Stikinia arc and the Insular superterrane, lie west of the Cache Creek suture belt (Figure 7). The Insular superterrane along the present continental fringe includes the Alexander terrane, a Paleozoic arc assemblage of largely pre-Devonian rocks overlain by less deformed Devonian to Permian strata including abundant limestone (Butler et al. 1997), and the Wrangellia terrane, a largely post-Carboniferous succession of Permian arc volcanics and overlying Triassic basalt capped by Upper Triassic limestone (Jones et al. 1977). The two components of the Insular superterrane were amalgamated by Carboniferous time, long before their joint incorporation into the Cordilleran continental margin, for they were both locally intruded by the same pluton (Gardner et al. 1988).

The Stikinia arc farther east is composed dominantly of Upper Triassic to Middle Jurassic volcanic and volcaniclastic rocks, intruded by cogenetic plutons (Marsden & Thorkelson 1992, Mihalynuk et al. 1994, Anderson 1993, Currie & Parrish 1997, MacIntyre et al. 2001). The arc assemblage is flanked on the northeast by a forearc basin (Dickie & Hein 1995, Johannson et al. 1997), lying adjacent to the Cache Creek suture in a position showing that the Stikinia arc faced the Cordilleran margin and subducted seafloor downward to the west. The ages of the youngest arc-forearc strata and the oldest strata in the postaccretion Jurassic-Cretaceous Bowser basin (MacLeod & Hills 1990), resting unconformably on Stikinia (Figure 7), indicate accretion of the northern part of Stikinia by early Middle Jurassic closure of the Cache Creek suture in either Aalenian (Ricketts et al. 1992) or early Bajocian (Thomson et al. 1986, Anderson 1993) time. The youngest strata known from the adjacent segment of the Cache Creek suture belt are Early Jurassic in age (Struik et al. 2001), but farther south in Canada deformed strata of the Cache Creek belt include strata as young as late Middle Jurassic (Callovian) in the Bridge River terrane (Cordey & Schiarizza 1993). The difference in stratal ages along tectonic strike suggest progressive southward closure of the Cache Creek suture from a tectonic hinge point on the north.

STIKINIA-QUESNELLIA OROCLINE Along tectonic strike to the north, the Stikinia arc merges, around the northern limit of the Cache Creek belt, with the northern end of the petrologically and lithologically similar Quesnellia arc along the edge of Triassic-Jurassic North America (Figure 7). This spatial relationship suggests that the Stikinia arc formed originally as a northern extension of the Quesnellia arc, but that oroclinal bending of the Quesnellia-Stikinia arc trend during continued subduction backfolded Stikinia against the Cordilleran margin to juxtapose Stikinia against Quesnellia across the Cache Creek suture, which was thereby enclosed within the tectonic orocline (Nelson & Mihalynuk 1993, Mihalynuk et al. 1994). Paleomagnetic data (May & Butler 1986, Vandall & Palmer 1990) showing no detectable latitudinal movement of Stikinia with respect to North America are compatible with the enclosure interpretation, and isotopic data indicating juvenile crustal origins are similar for Stikinia and Quesnellia arc assemblages (Samson et al. 1989, Smith et al. 1995). Pre-Mesozoic underpinnings of both Stikinia and Quesnellia include Devonian to Permian arc assemblages (Brown et al. 1991, Currie & Parrish 1997), inferred here to have been accreted to Laurentia during Antler-Sonoma events (Figure 6). Both Mesozoic arc assemblages also overlap depositionally upon deformed Paleozoic assemblages (Mortensen 1992, Roback & Walker 1995, Dostal et al. 2001, Acton et al. 2002), interpreted here as overthrust Antler-Sonoma allochthons.

Most of the contact zone between the Stikinia arc and the Insular superterrane is occupied by a sliver of strongly deformed pre-Mesozoic strata, forming a western arm of the Yukon-Tanana terrane (Gehrels et al. 1991, 1992) including the Taku terrane (Gehrels 2002), which underlie the Mesozoic arc assemblage of Stikinia and are regarded here as a product of Antler-Sonoma orogenesis oroclinally deformed along with Stikinia (Figures 6 and 7). The contact zone was overlapped by thick Upper Jurassic (Oxfordian) to Lower Cretaceous (Albian) strata of the intraarc Gravina basin (McClelland et al. 1992), but underlying metavolcanic rocks that also overlap the contact zone document initial accretion of the Insular superterrane to the western flank of Stikinia by Middle Jurassic (∼175 Ma) time (Gehrels 2001). Mid-Cretaceous thrusting later carried rocks east of the contact zone over the Gravina basin and the Insular superterrane (Gehrels et al. 1990, Rubin & Saleeby 1992).

INSULAR ARC ACCRETION As the Stikinia arc demonstrably faced east, subduction along its western flank could not have drawn the Insular superterrane toward the continental margin. Accordingly, Early to Middle Jurassic arc magmatism (190–165 Ma) within the Insular superterrane, as displayed in the Queen Charlotte Islands (Lewis et al. 1991) and on Vancouver Island (DeBari et al. 1999), is viewed here as evidence for activation of subduction along the eastern flank of the Insular superterrane, to draw the Insular superterrane closer to the back side of the Stikinia arc by subducting intervening seafloor downward to the west. The polarity of the Jurassic arc along the Insular superterrane is seemingly confirmed along tectonic strike to the northwest, beyond the head of the Gulf of Alaska, where Lower to Middle Jurassic plutons intruding the Wrangellia component of the Insular superterrane on the Alaska Peninsula display transverse compositional gradients indicative of a magmatic arc facing the continent (Reed et al. 1983).

Paleomagnetic data suggest that the Alexander terrane lay in the Arctic region near Baltica in mid-Paleozoic time, but that the associated Wrangellia terrane lay near the paleolatitude of the Pacific Northwest by Late Triassic time (Butler et al. 1997). Apparently, the Insular superterrane drifted as an intraoceanic arc structure within the paleo-Pacific Ocean, along paths that cannot be specified with present information, through late Paleozoic and early Mesozoic time before its accretion to the Cordilleran margin along the back side of Stikinia. If the sliver of the Yukon-Tanana terrane along the west flank of Stikinia includes miogeoclinal facies, as seems likely (Gehrels 2000), the oroclinal rotation of Stikinia was apparently initiated by calving of Stikinia off the edge of the Laurentian margin during backarc rifting. The complex plate motions required to achieve accretion of both the Stikinia arc and the Insular superterrane to the Cordilleran margin in the same general time frame (intra-Jurassic) are indeterminate with present information.

PACIFIC NORTHWEST RECONSTRUCTION The longitudinal correlation of premid-Cretaceous tectonic elements southward across the Pacific Northwest from Canada into the USA has long been a challenge (Monger et al. 1982) because of widespread Neogene volcanic cover (Figure 8*A*), the complex kinematics of an intersecting knot of strike-slip faults of latest Cretaceous to Eocene age spanning the USA-Canada border (Figures 7, 9), and structural complexity within the metamorphic cores of mountain ranges near the USA-Canada border where Cretaceous structural telescoping obscured earlier tectonic relationships between older rock masses.

An apparently satisfactory tectonic reconstruction is achieved here (Figure 8*B*) by reversing 105–110 km of Eocene (44–34 Ma) dextral slip on the Fraser River–Straight Creek fault zone and 110–115 km of previous dextral slip on the offset Yalakom–Ross Lake fault system of latest Cretaceous ($<$75 Ma) to Eocene age (Kleinspehn 1985, Umhoefer & Kleinspehn 1995, Umhoefer & Miller 1996, Umhoefer & Schiarizza 1996) and by backrotating the Oregon-Washington Coast Range and the Blue Mountains by 50° each (Figure 8) to recover clockwise tectonic rotations imposed during Eocene time (Heller et al. 1987, Dickinson 2002).

In Figure 8, the southern extension of the Stikinia arc assemblage includes the Cadwallader terrane of southern British Columbia (Rusmore 1987, Rusmore et al. 1988, Umhoefer 1990, Rusmore & Woodsworth 1991) and the Triassic-Jurassic Cascade River–Holden belt (Hopson & Mattinson 1994) in the Cascade Mountains east of the Straight Creek fault. Inland extensions of the Insular superterrane include the Chilliwack, Bowen Lake, and Harrison Lake terranes of southern British Columbia (Friedman et al. 1990, Mahoney et al. 1995), the Swakane Gneiss (Nason terrane) east of the Straight Creek fault in the Cascade Mountains (Mattinson 1972), and the Wallowa–Seven Devils segment of Wrangellia in the Blue Mountains. The Cache Creek suture belt flanking the Triassic-Jurassic continental margin is reconstructed as an alignment of similar lithologic units, including the Cache Creek and Bridge River terranes of southern British Columbia, the Hozameen terrane spanning the USA-Canada border, the Baker terrane of the Blue Mountains, and

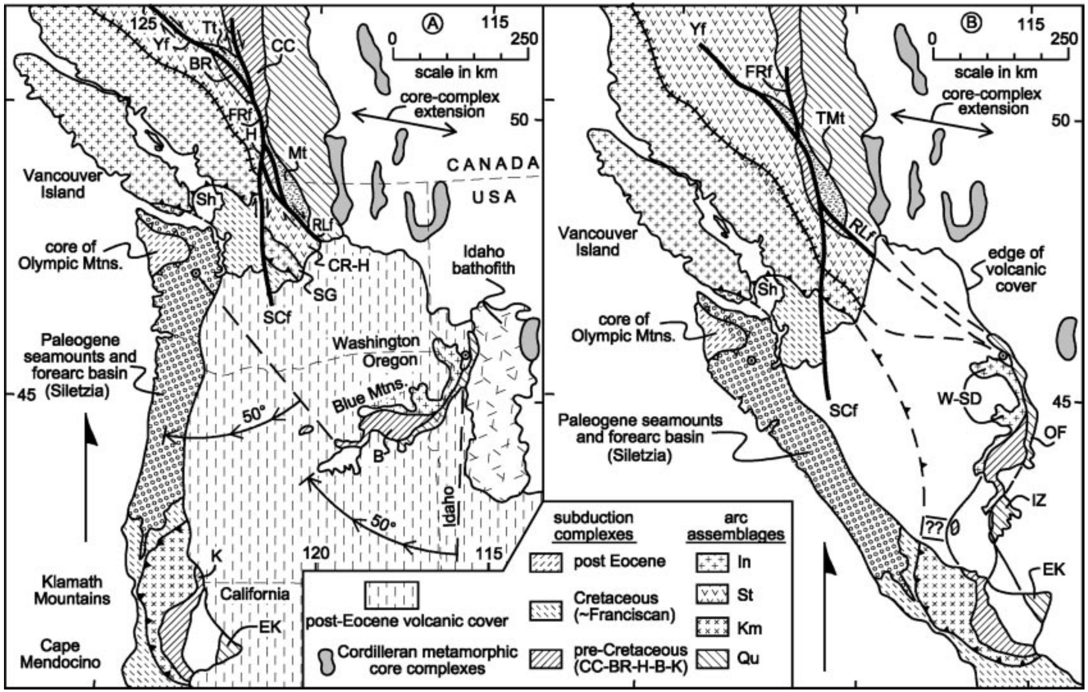

Figure 8 Pre-Oligocene geotectonic features in the Pacific Northwest (USA) and adjacent Canada at present (*A*) and as reconstructed (*B*) before clockwise rotations of the Oregon-Washington Coast Range and Blue Mountains provinces, and before dextral slip on branching faults near the USA-Canada border. Arc assemblages: In, Insular (SG, Swakane Gneiss; W-SD, Wallowa–Seven Devils); Km, accreted western Klamath Mountains arcs; Qu, Quesnellia and related terranes (IZ, Izee forearc basin; EK, eastern Klamath Mountains Mesozoic arc, OF, Olds Ferry terrane or Huntington arc); St, Stikinia (CR-H, Cascade River–Holden belt). Pre-Late Jurassic subduction-complex terranes: B, Baker; BR, Bridge River; CC, Cache Creek; H, Hozameen, K, central Klamath Mountains mélange belt. Other geologic features: Sh, Shuksan thrust system (schematic); TMt, Tyaughton–Methow trough (offset segments: Mt, Methow trough; Tt, Tyaughton trough).

the central mélange belt of the Klamath Mountains (Figure 8*B*). Closer proximity of restored tectonic elements near the USA-Canada border to counterparts in the Blue Mountains could be achieved by additional recovery of the significant Eocene intracontinental extension recorded by Cordilleran core complexes (Figure 8) in southeastern British Columbia (Dickinson 2002).

The Tyaughton-Methow trough (Figure 8) was initiated in Early Jurassic time as a forearc basin flanking the Quesnellia arc (Anderson 1976), but evolved during Late Jurassic and Early Cretaceous time to overlap the accreted Stikinia arc (Garver 1992, Umhoefer et al. 2002). West and south of the Shuksan thrust system (Figure 8), an internally deformed underthrust assemblage, including Upper Jurassic to Lower Cretaceous blueschists and clastic strata (Brown 1987, Brandon et al. 1988, Monger 1991), is presumed to be a northern counterpart of the late Mesozoic Franciscan subduction complex and associated forearc basins of coastal California to the south (Brown & Blake 1987).

USA-MEXICO ARC ACCRETION The Insular superterrane extends as far south as the Blue Mountains (Figure 8) of the Pacific Northwest, where intense mid-Cretaceous crustal telescoping near the Snake River has thrust strata of Wrangellia beneath the Mesozoic continental margin (Lund & Snee 1988). Stratigraphic analysis of Blue Mountains terranes indicates, however, that initial accretion of the Wrangellia component of the Insular superterrane was completed in Middle Jurassic (Bajocian) time (Follo 1992, White et al. 1992, Avé Lallemant 1995), coordinate with accretion farther north in Canada. The oroclinally deformed Stikinia arc apparently does not extend farther south than the Cascades Mountains along the USA-Canada border (Figure 8), and there is no indication that active magmatism was still underway at the southern end of the Insular superterrane when the Wallowa–Seven Devils segment (Figure 8) of Wrangellia was drawn passively into a subduction zone along the continental margin (Dickinson 1979).

Accreted intraoceanic arc assemblages of Jurassic age in the Klamath Mountains and Sierra Nevada foothills of California rest on ophiolitic basement formed near the Triassic-Jurassic time boundary (Dilek 1989, Edelman 1990, Hacker & Ernst 1993, Wright & Wyld 1994, Hacker et al. 1995), as does the Guerrero superterrane (Figure 7) of western Mexico (Dickinson & Lawton 2001a). The accreted Mesozoic arc complexes in the USA and Mexico can perhaps be regarded as southern extensions and descendants of the Jurassic arc along the Insular superterrane where subduction continued southward across paleo-Pacific oceanic crust lying beyond the southern limits of the older Alexander and Wrangellia terranes. In California, severely deformed mélange belts separate accreted arc assemblages on the west from the pre-Jurassic continental margin (Wright 1982, Edelman & Sharp 1989, Edelman et al. 1989b, Dilek et al. 1990, Hacker et al. 1993), but a superimposed magmatic arc built along the Cordilleran continental margin across the accreted tectonic elements by late Middle Jurassic (Callovian) time (Wright & Fahan 1988, Edelman 1990, Edelman et al. 1989a, Harper et al. 1994, Girty et al. 1995) implies arc accretion during early Middle Jurassic (Bajocian) time (170–165 Ma).

In the southwestern USA and Mexico, final closure of the oceanic realm between accreted Mesozoic arcs and the Cordilleran continental margin in Early Cretaceous time promoted slab rollback of the Mezcalera plate to induce crustal extension within the overriding continental block (Dickinson & Lawton 2001a). The resulting border rift belt, including the Bisbee basin and Chihuahua trough, supplanted arc magmatism along the USA-Mexico border region (Figure 7), with Late Jurassic rifting accompanied by bimodal magmatism and followed by Early Cretaceous thermotectonic subsidence (Dickinson & Lawton 2001b). Farther north, the extensional Utah-Idaho trough (Figure 7) of Middle to Late Jurassic age and development of a wide zone of Late Jurassic to Early Cretaceous backarc magmatism (Figure 7) closely followed arc accretion along the California continental margin to the west (Dickinson 2001). Earlier Middle Jurassic thrusting along the Luning-Fencemaker thrust (Wyld 2002), which carried the fill of the Auld Lang Syne backarc basin eastward (Figure 7), coincided closely in timing with arc accretion farther west.

Mid-Cretaceous to Mid-Tertiary Cordilleran Arc

Following Jurassic-Cretaceous arc accretion at the evolving subduction zone along the continental margin, the Cordilleran magmatic arc stepped oceanward to a trend that was largely superimposed upon accreted terranes (Figure 9). Massive Late Cretaceous plutonism, continuing until mid-Eocene time in Canada, formed the major Cordilleran batholith belt along the arc axis. To the west, a parallel belt of Jurassic-Cretaceous forearc basins is prominent along the coastal fringes of the USA and Mexico, and lies immediately inland from exposures of the Jurassic-Cretaceous subduction complex forming the Franciscan superterrane (Figure 9). Farther north in Canada, however, Cenozoic modification of the continental margin by strike slip along the Cenozoic Queen Charlotte transform and its splays has largely disrupted or submerged tectonic elements of the late Mesozoic forearc region.

Past speculation (Cowan et al. 1997), based on paleomagnetic data, that the western part of the Canadian Cordillera, including a large segment of the Cretaceous batholith belt, was transported northward in Cretaceous-Paleocene time from an origin along the continental margin of California or Mexico encounters the insuperable difficulty that no segment of the Cretaceous arc-trench system is missing from California or Mexico (Figure 9). The anomalously shallow paleomagnetic vectors that gave rise to the hypothesis of large lateral displacements can be interpreted instead as the result of widespread pluton tilt coupled with compaction in sedimentary strata (Butler et al. 2001).

Crustal shortening across the Cordilleran orogen gave rise by Late Jurassic time in Canada (Cant & Stockmal 1989) and mid-Early Cretaceous time in the USA (Dickinson 2001) to initiation of a backarc thrust belt that was continuous from the interior flank of the Canadian Cordillera into the Sevier thrust belt (Figure 9). The tectonic load of the thrust sheets downflexed an extensive retroforeland basin with a distal fringe that extended well into the continental interior. Deformation within the Canadian Cordillera produced intraorogen thrusting associated with development of the Skeena foldbelt (Evenchick 1991) accompanied by downflexure of the Sustut basin, and analogous intraorogen deformation formed the Eureka thrust belt in the USA (Figure 9).

Scattered plutons of Late Cretaceous age present in the interior hinterland of the backarc thrust belt, but most prominent in the Omineca region of the Canadian Cordillera (Figure 9), were not an integral facet of the arc magmatism active farther west, but instead were derived largely from sources within underthrust continental crust. Backarc thrusting and the associated retroforeland basin did not extend as far south as the region occupied until mid-Cretaceous time by the Bisbee basin and related rift troughs along the USA-Mexico border. Although somewhat diachronous in timing, backarc rifting (Figure 7) and backarc thrusting (Figure 9) occupied different realms marked by distinct contrasts in geodynamics along the Cordilleran orogen.

In Canada, arc magmatism along the eastern flank of the Coast batholith continued until mid-Eocene time (~45 Ma), as did deformation along the backarc thrust belt. Farther south, however, in both the USA and Mexico, subhorizontal subduction of the Farallon plate during latest Cretaceous through Eocene time altered the progress of both magmatism and tectonism (Dickinson & Snyder 1978). Inland migration and diminution of igneous activity led to a magmatic null through much of the USA Cordillera (Figure 1), and basement-involved crustal shortening produced yoked uplifts and basins of the Laramide Rocky Mountains well inland from the continental margin (Figure 9).

Deformation began ~70 Ma throughout the Laramide Rocky Mountains while thrusting was still underway along the Sevier thrust belt to the west, but its termination was diachronous (Dickinson et al. 1988). The development of Laramide basins and uplifts was complete in the northern part of the Laramide province by mid-Eocene time (~50 Ma), coincident with the terminal phase of deformation along the Sevier thrust belt to the west (DeCelles 1994). Farther south, however, Laramide deformation continued until the end of Eocene time. In Mexico, south of the magmatic null, Laramide basin evolution (Figure 9) during Late Cretaceous and Paleocene time (Dickinson & Lawton 2001b) was accompanied by arc magmatism that migrated inland from the Cretaceous batholith belt near the coast. The time-space pattern of Laramide magmatism and deformation suggests that the shallow angle of plate descent that gave rise to both resulted from subduction of a buoyant oceanic plateau beneath the continental margin (Dickinson et al. 1988).

In the Pacific Northwest, the elongate oceanic seamount chain of Siletzia (Figure 9), which formed during Paleocene-Eocene time (Figure 1) at some unknown distance offshore, was accreted in bulk to the Cordilleran margin early in Eocene time, and was subsequently buried beneath the Eocene forearc basin (Figure 8) of the Oregon-Washington Coast Range (Heller et al. 1987). A subduction complex (Brandon & Vance 1992), composed of premid-Miocene Cenozoic strata underthrust beneath the accreted mass of Siletzia, forms the core of the Olympic Mountains near the USA-Canada border (Figure 9). Incrementally accreted Paleocene-Eocene components of the Franciscan subduction complex are exposed along the coastal fringe of California farther south, but elsewhere most Cenozoic subduction along the Cordilleran margin occurred along an offshore zone still unexposed underwater.

CENOZOIC TAPHROGENY

The Cordilleran arc orogen as a typical segment of the Circum-Pacific orogenic belt reached peak development in Late Cretaceous time. During Tertiary time, arrival at the Cordilleran margin of successive segments of spreading systems bounding the Pacific plate progressively converted segments of the continental margin into transform fault systems along the Pacific plate boundary. As the transform continental margin evolved, subsidiary strike slip and associated crustal extension disrupted the adjacent continental block and gave rise to the rift trough of the Gulf of California, an incipient ocean basin that is expanding obliquely within the transform regime (Figure 10).

North of the Tofino triple junction (Figure 10), subduction associated with waning phases of batholith generation along the coastal fringe of the Canadian Cordillera was supplanted in mid-Eocene time (Hyndman & Hamilton 1993) by dextral slip along the Queen Charlotte transform fault. The change in coastal geodynamics, from convergence to strike slip, was triggered by amalgamation of the offshore Kula and Pacific plates at ~42.5 Ma (Lonsdale 1988). Subsequent Oligocene-Miocene magmatism within the Queen Charlotte Islands was associated with evolution of a slab window (Hamilton & Dostal 2001), and the Neogene Queen Charlotte basin farther east (Figure 10) developed as a pull-apart basin within the transform system (Lewis et al. 1991). The Chatham Strait–Denali fault system, initiated in mid-Eocene time as a branch of the Queen Charlotte transform (Cole et al. 1999), has displaced segments of the Insular superterrane laterally along the continental margin (Figure 10). A discrepancy between 370 km of slip along the Denali fault and 150 km of slip along the linked Chatham Strait fault suggests that 220 km of slip parallel to the continental margin, southward from the elbow where those two fault segments meet, was accommodated by the Coast shear zone along the eastern flank of the Insular superterrane (Gehrels 2000).

Farther south, subduction along the Cordilleran margin continues at the foot of the continental slope along an offshore trend parallel to and coextensive with the active Cascades volcanic arc of the Pacific Northwest (Figure 10). Arc volcanism, which extended southward through the USA in Miocene time (Figure 10), was progressively extinguished south of the Cascades arc by evolution of the San Andreas transform system along the continental margin as the Mendocino triple junction (Figure 10) migrated northward to shorten the Cascades subduction zone. Beginning near the Oligocene-Miocene time boundary, slab-window volcanism evolved in coastal California along a belt parallel to the evolving San Andreas transform (Dickinson 1997). Neogene arc volcanism was extinguished in similar fashion within Baja California when the Rivera triple junction (Figure 10) migrated southward in mid-Miocene time to a position near the mouth of the modern Gulf of California.

Following Laramide events, establishment of mid-Cenozoic arc magmatism along a trend near the USA-Mexico continental margin had been accomplished by the migration of successive volcanic fronts toward the coast (Figure 10) as the slab of oceanic lithosphere subducting beneath the Cordilleran orogen steepened or foundered (Dickinson 2002). Subsequent initiation of the San Andreas transform along the continental margin in Early Miocene time triggered crustal extension within the Basin and Range taphrogen (Figure 10), where multiple fault blocks distended the Cordilleran orogen once the continental block was partly coupled to the Pacific plate. A largely intact remnant of the mid-Cenozoic arc assemblage lies along the Sierra Madre Occidental (Figure 10), where flat-lying volcanic strata form an enclave of largely undistended crust enclosed within the Basin and Range taphrogen. Baja California was calved from mainland Mexico when the San Andreas transform plate boundary south of the USA jumped inland in Late Miocene time to open the Gulf of California by oblique extension.

In the Pacific Northwest (USA), extensive volcanic fields of flood basalt that have erupted behind the Cascades volcanic arc since Early Miocene time mask older rock assemblages over wide areas (Figure 10). The volcanism may have been related to mantle advection induced by deformation of the continental lithosphere after shear was imposed on the continental block by interaction of the Pacific and American plates along the San Andreas transform system at the continental margin (Dickinson 1997). Less voluminous but comparably extensive Middle Miocene and younger volcanic fields of basaltic character in the Canadian Cordillera (Edwards & Russell 2000) may reflect analogous shear coupling of the Pacific and American plates along the nearby Queen Charlotte transform.

SUMMARY PERSPECTIVES

The questions posed in the introduction can be answered as follows:

1. The Cordilleran system, as an integral segment of the circum-Pacific orogenic belt, was established when subduction was initiated between Early and Late Triassic time along a continental margin that had been delineated by Neoproterozoic rifting during the breakup of Rodinia, and later modified in late Paleozoic and earliest Mesozoic time by the emplacement of oceanic allochthons upon the edge of the continental block during the final assembly of Pangea.
2. Rock masses native to the Cordilleran margin include the miogeoclinal prism deposited between Neoproterozoic and Late Devonian time along a passive continental margin, volcanic and plutonic rocks of the Cordilleran magmatic

arc built along an active continental margin from mid-Triassic time to the present, and the sedimentary and volcanic assemblages of basins and lava fields superimposed upon the miogeoclinal succession and the arc assemblage.

3. Accreted tectonic elements include subduction complexes thrust bodily over the miogeoclinal prism in Devonian-Mississsippian and Permian-Triassic time; intraoceanic island arcs sutured to the continental block at those times and also later, between Middle Jurassic and Early Cretaceous time; and subduction complexes accreted incrementally to the continental block at the Cordilleran subduction zone between Late Triassic and mid-Cenozoic time.
4. Postmid-Cenozoic internal distension and incipient dislocation of Cordilleran crust has occurred in response to transform tectonism imposed on the continental margin when intra-Pacific seafloor spreading systems impinged on the Cordilleran trench.

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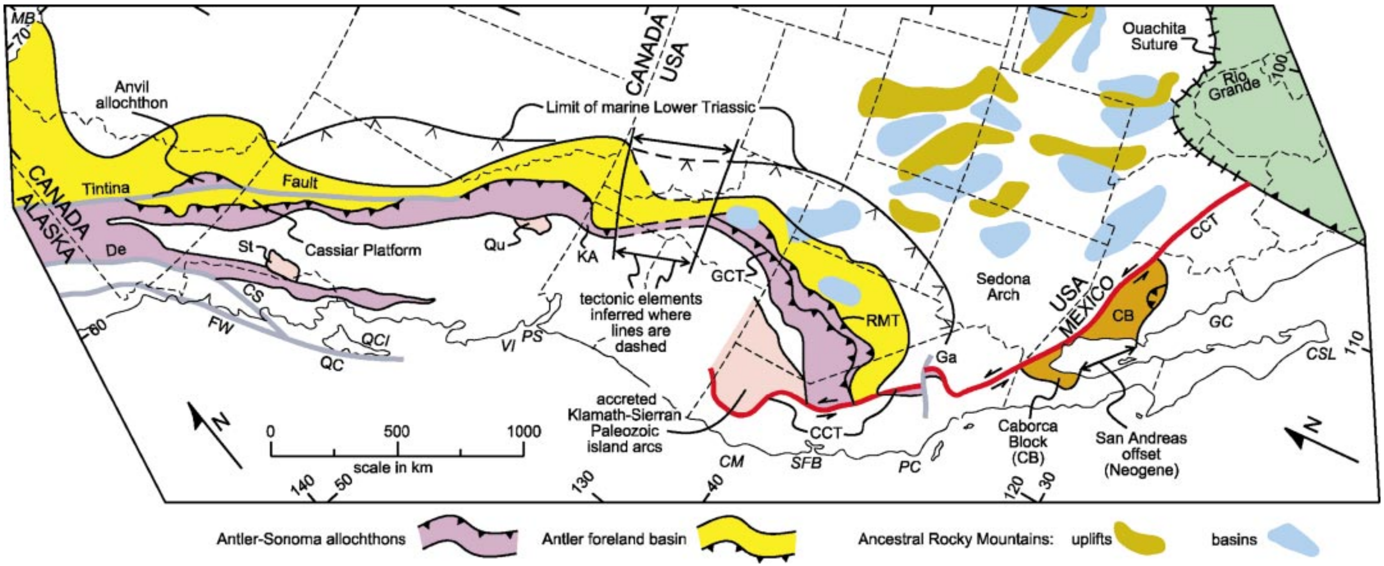

Figure 6 Geotectonic features of the Antler orogen (Late Devonian–Early Mississippian), the Ancestral Rocky Mountains province (Pennsylvanian–Early Permian), and the Sonoma orogen (Late Permian–Early Triassic) of the North American Cordillera (allochthons of Antler and Sonoma age are combined, but note the uncertain continuity of tectonic trends along the trans-Idaho discontinuity of Figure 5). See text for discussion of Kootenay structural arc (KA) and remnants of Paleozoic arc assemblages in Quesnellia (Qu) and Stikinia (St). Key active faults: RMT, Devonian-Mississippian Roberts Mountains thrust; GCT, Permian-Triassic Golconda thrust; CCT, Permian-Triassic California-Coahuila transform. Gondwanan Mexico restored (after Dickinson & Lawton 2001a) to position before mid-Mesozoic opening of the Gulf of Mexico. Tintina and De-CS-FW-QC fault systems are Cenozoic structures. See Figure 5 for geographic legend.

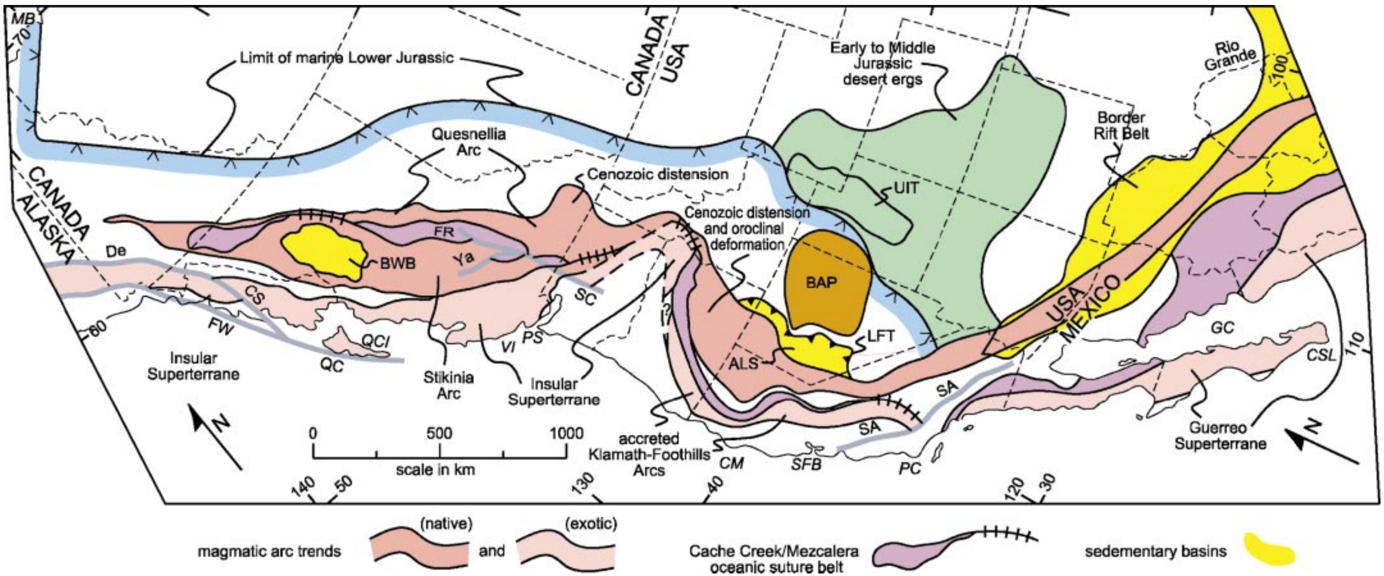

Figure 7 Mid-Early Triassic (~247.5 Ma) to mid-Early Cretaceous (~120 Ma) geotectonic features of the Cordilleran arc-trench system, including intraoceanic arc structures accreted to the Cordilleran continental margin between Middle Jurassic and Early Cretaceous time. Selected geologic features: ALS, Late Triassic to Early Jurassic Auld Lang Syne backarc basin; LFT, Middle Jurassic Luning-Fencemaker backarc thrust system; BAP, zone of diffuse Middle to Late Jurassic backarc plutonism (Nevada-Utah); UIT, Middle to Late Jurassic Utah-Idaho backarc trough; BWB, Late Jurassic to Early Cretaceous Bowser successor basin (superimposed on accreted Stikinia arc). Border rift belt (Late Jurassic to Early Cretaceous) includes Bisbee basin and Chihuahua trough. Cenozoic faults (*gray*): De-CS, Denali-Chatham Strait; FR, Fraser River; FW, Fairweather; QC, Queen Charlotte; RL, Ross Lake; SA, San Andreas; SC, Straight Creek; Ya, Yalakom. See Figure 5 for geographic legend.

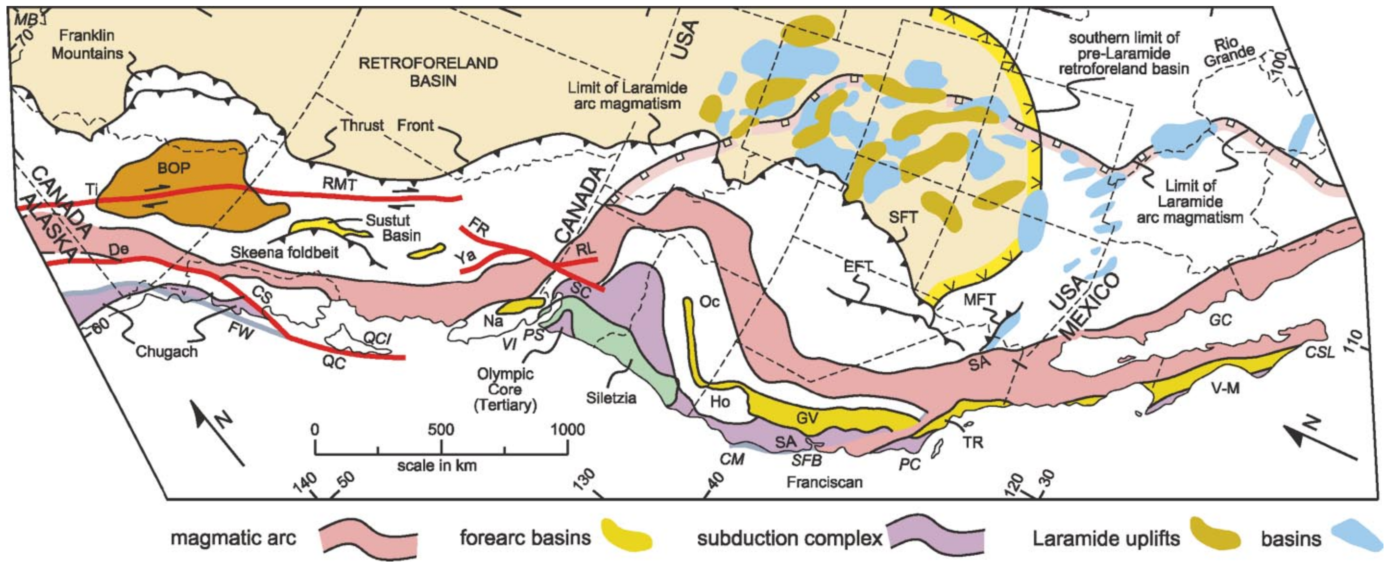

Figure 9 Mid-Early Cretaceous (~120 Ma) to mid-Cenozoic (Eocene/Oligocene boundary) geotectonic features of the Cordilleran arc-trench system, including the Laramide province of intracontinental deformation. The main batholith belt is delineated as the magmatic arc, but subsidiary arc magmatism spread inland for varying distances (BOP, Omineca zone of backarc mid-Cretaceous plutonism). Active Paleogene strike-slip faults (*red lines*): De-CS, Denali-Chatham Strait; FR, Fraser River; QC, Queen Charlotte; RL, Ross Lake; SC, Straight Creek; Ti-RMT, Tintina–Rocky Mountain Trench; Ya, Yalakom. Younger Neogene strike-slip faults (*gray lines*): FW, Fairweather; SA, San Andreas. Key active fold-thrust belts: EFT, Eureka; MFT, Maria; SFT, Sevier. Key forearc basin segments: GV, Great Valley (California); Ho, Hornbrook; Oc, Ochoco; Na, Nanaimo; TR, Transverse Ranges; V-M, Vizcaino-Magdalena. See Figure 5 for geographic legend.

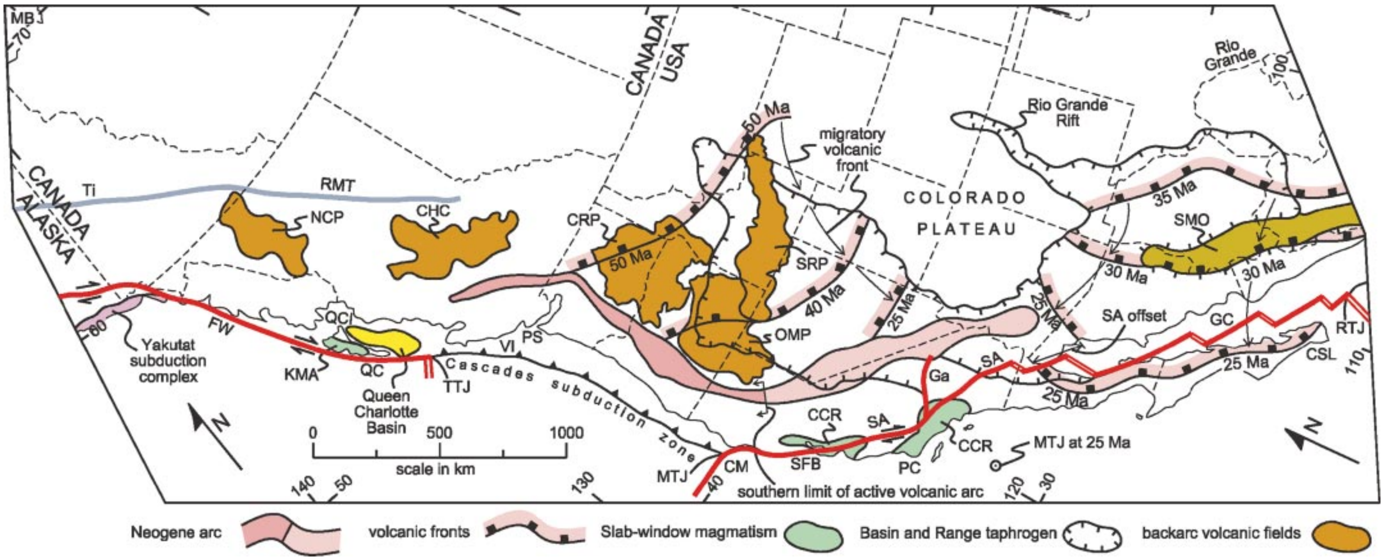

Figure 10 Post-Eocene geotectonic features of the North American Cordillera, including the Basin and Range taphrogen (SMO, Paleogene magmatic arc remnant in the Sierra Madre Occidental). Active Neogene strike-slip faults (*red lines*): FW, Fairweather; Ga, Garlock; QC, Queen Charlotte; SA, San Andreas. Offshore triple plate junctions: MTJ, Mendocino (FFT); RTJ, Rivera (RTF); TTJ, Tofino (RTF). Slab-window magmatism (near the coastal transform systems): CCR, California Coast Ranges (28–0 Ma); KMA, Queen Charlotte Islands (46–17 Ma). Backarc lava fields (erupted inland from main Cordilleran arc trend): CHC, Chilcotin (14–6 Ma); CRP, Columbia River Plateau (17–8 Ma); NCP, Northern Cordilleran Province (8–0 Ma); OMP, Oregon-Modoc Plateau (17–7 Ma); SRP, Snake River Plain (16–0 Ma). See Figure 5 for geographic legend.

Chapter 6: "A synthesis of the Jurassic–Cretaceous tectonic evolution of the central and southeastern Canadian Cordillera: Exploring links across the orogen" *(Evenchick et al.), in* ***SPE433: Whence the Mountains? Inquiries into the Evolution of Orogenic Systems: A Volume in Honor of Raymond A. Price*** *(Sears, Harms, and Evenchick, eds.)*

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A synthesis of the Jurassic–Cretaceous tectonic evolution of the central and southeastern Canadian Cordillera: Exploring links across the orogen

Carol A. Evenchick*
Geological Survey of Canada, 625 Robson Street, Vancouver, British Columbia V6B 5J3, Canada

Margaret E. McMechan*
Geological Survey of Canada, 3303 33rd Street NW, Calgary, Alberta T2L 2A7, Canada

Vicki J. McNicoll*
Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8, Canada

Sharon D. Carr*
Ottawa-Carleton Geoscience Centre, Department of Earth Sciences, Carleton University, Ottawa, Ontario K1S 5B6, Canada

ABSTRACT

Restoration of tectonic elements in the central interior of the Canadian Cordillera southward to their paleogeographic position in the Mesozoic permits comparison of data across the active orogen, recognition of the interplay between coeval lithospheric thickening and basin evolution, and new constraints on models of tectonic evolution. The onset of Middle Jurassic clastic sedimentation in the Bowser basin, on the west side of the Jurassic orogen, occurred in response to accretionary events farther inboard. Shortening and thickening of the crust between the Alberta foreland basin on the east side of the Jurassic orogen and Bowser basin on the west side resulted in an Omineca highland between the two basins and lithospheric loading that influenced their Late Jurassic–Cretaceous sedimentation. The provenance of detritus in these basins, and in the Late Cretaceous Sustut basin on the east side of the Bowser basin, reveals migration of drainage divides in the intervening Omineca highland through time. Synchronous and compatible tectonic events within the basins and evolving accretionary orogen, and in rocks of the Stikine terrane and the western margin of North America, suggest that they were kinematically connected above a lower-crust detachment, beginning in the Middle Jurassic. The Coast belt was part of this wide, dynamically linked bivergent orogen from the mid-Cretaceous to earliest Cenozoic,

*E-mails: Evenchick: cevenchi@nrcan.gc.ca; McMechan: mmcmecha@nrcan.gc.ca; McNicoll: vmcnicol@nrcan.gc.ca; Carr: scarr@earthsci.carleton.ca.

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and the lower-crust detachment rooted near the active plate margin. Nested within the orogen, the east-vergent thin-skinned Skeena fold belt, equivalent in scale to the Rocky Mountain fold-and-thrust belt, was also linked to the detachment system.

Keywords: Canadian Cordillera, tectonic evolution, Bowser basin, Alberta foreland basin, Sustut basin, Omineca belt.

INTRODUCTION

The relationships among basin evolution, fold-and-thrust belt development, and thermotectonic evolution of core zones provide insight into progressive stages of orogenesis. The dynamic link between these realms also provides a basis for geodynamic modeling and feedback for understanding orogen-scale processes. The southeast Canadian Cordillera is an excellent example. The Late Jurassic through earliest Cenozoic inter-relationships among the northeasterly vergent thin-skinned fold-and-thrust belt of the Foreland belt (Fig. 1) on the east side of the Cordilleran orogen, the synorogenic Alberta foreland basin deposits within and east of the Foreland belt, and the internal core zone of the Omineca belt are relatively well understood (Bally et al., 1966; Price and Mountjoy, 1970; Price, 1973, 1981; Brown et al., 1986; McMechan and Thompson, 1989; Fermor and Moffat, 1992; Beaumont et al., 1993). Studies have associated part of their evolution to accretion of terranes in the western Omineca and eastern Intermontane belts during westward underthrusting of the North American plate (e.g., Cant and Stockmal, 1989; Brown et al., 1992a; Price, 1994; Brown and Gibson, 2006; Carr and Simony, 2006). However, analyses commonly extend only as far west as the easternmost of the accreted terranes on the west side of the Omineca belt, and relationships of orogenic processes in the Foreland and Omineca belts to those of the Intermontane and Coast belts farther west have not been explored, nor have their implications for tectonic evolution of the Cordillera as a whole.

New understanding of exhumed mid-crustal rocks exposed in core zones of the Canadian Cordillera (Coast and Omineca belts, Fig. 1) and of depositional and structural histories of the sedimentary basins that flank them, in particular the Bowser, Sustut, and Alberta foreland basins (Fig. 1), is used herein to illustrate links between the major tectonic elements of the orogen. One result is a tectonic reconstruction in which the entire width of the orogen is kinematically linked throughout the Mesozoic–early Cenozoic (Fig. 2). It evolved from a predominantly west-verging "small-cold" (Beaumont et al., 2006) accretionary orogen, ~300 km wide in the Middle Jurassic (Fig. 2A), into an ~1000-km-wide "large-hot" (Beaumont et al., 2006) bivergent orogen in the mid-Cretaceous (Fig. 2B). By the mid-Cretaceous, an unusual geometry developed with two major detachment systems (Figs. 2B and 2C). A lower-crust detachment system extending across the entire orogen rooted near the active plate margin and joined the western magmatic convergent belt (Coast belt) to the eastern front of deformation in the thin-skinned, east-vergent foreland fold-and-thrust belt. A second detachment rooted in the Coast belt rose eastward to relatively high structural levels, forming the basal detachment of the east-vergent Skeena fold-and-thrust belt nested in the interior of the orogen—a fold belt that matches the classic Rocky Mountain fold-and-thrust belt (Foreland belt and eastern Omineca belt) in width and magnitude of horizontal shortening.

In this paper, we include a synthesis of the depositional and structural histories of the two major Mesozoic basins, the Bowser and Sustut basins (Fig. 1), that formed west of the site of initial accretionary orogenesis now preserved in the southern Omineca belt. The Bowser basin is a largely marine basin that formed the Middle Jurassic to Early Cretaceous western continental margin of North America during and following the accretion of its basement, Stikinia (Stikine terrane), to the North American plate (e.g., Evenchick and Thorkelson, 2005, and references therein). The Sustut basin was the site of Late Cretaceous nonmarine, synorogenic clastic sedimentation confined between, and sourced from, the evolving Skeena fold belt on the west and the Omineca highland on the east (Eisbacher, 1974a, 1985; Evenchick, 1991a). Previous work of Ricketts et al. (1992) has shown how early subsidence and sedimentation in the Bowser basin in the latest Early Jurassic or early Middle Jurassic in the central Intermontane belt was related to southwest thrusting of Cache Creek terrane (Fig. 1) over Stikinia, and that these events were contemporaneous with southwest-directed thrust faults in the southern Omineca belt. Eisbacher (1981, 1985) included the depositional and structural history of the basins west of the Omineca belt in a synthesis of depositional patterns across the Cordillera. We build on these works, using data and interpretations from more than 20 yr of research since Eisbacher's analysis, to examine the relationships between Mesozoic depositional and structural events across the orogen. The geological history of the Bowser basin is of fundamental importance to our understanding of the tectonic development of the Cordilleran orogen because it contains the earliest depositional record in response to mid-Mesozoic terrane accretion in the Canadian Cordillera, as well as the record of continued lithospheric response to Jurassic–Cretaceous development of the thickening orogen between the Bowser basin and the Alberta foreland basin; in this regard, the Bowser basin is the western counterpart to the Alberta foreland basin. Also critical to a regional tectonic analysis is the role of the Skeena fold belt, a thin-skinned fold-and-thrust belt that deformed Bowser and Sustut strata at the same time as horizontal shortening occurred at all structural levels in the Coast and Omineca belts and at

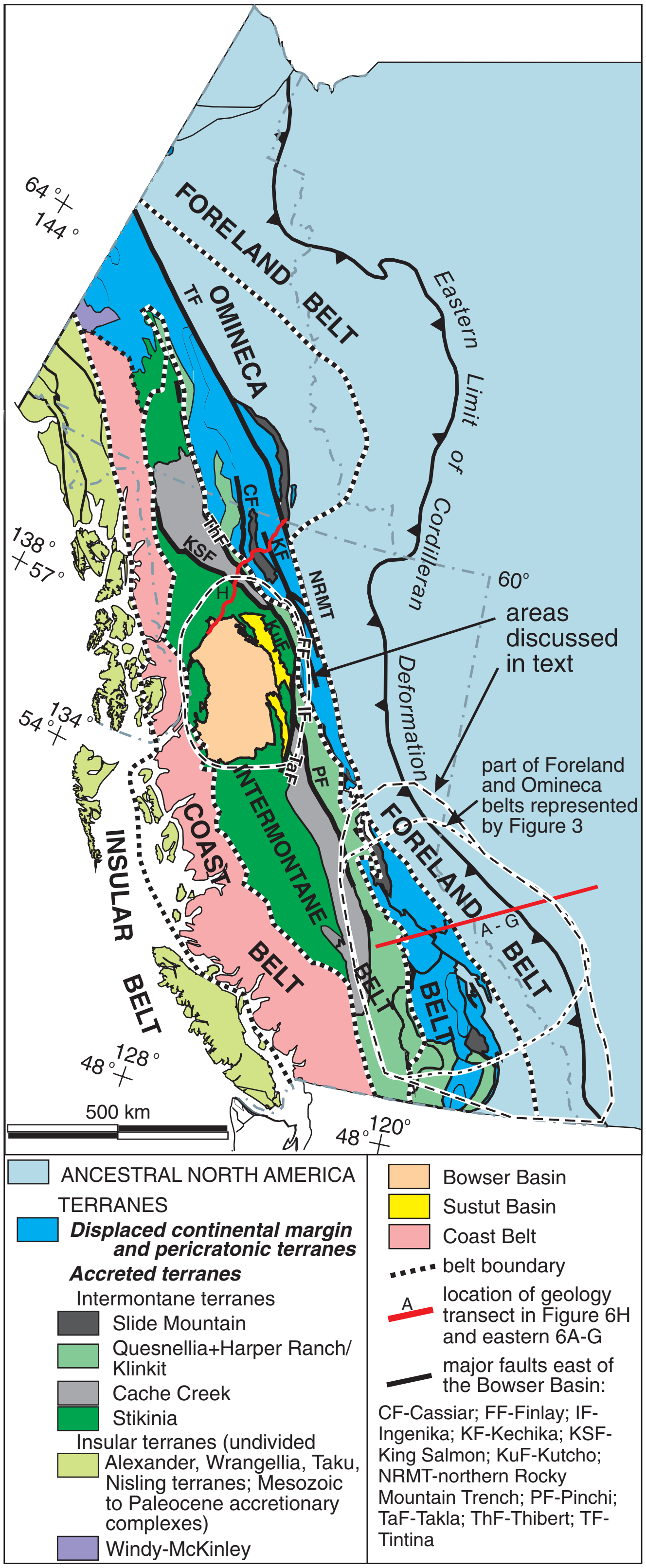


Figure 1. Morphogeological belts and major terranes of the Canadian Cordillera, locations of the Bowser and Sustut basins, major strike-slip faults east of these basins, and outlines of areas discussed in text (modified from Wheeler and McFeely, 1991; Colpron et al., 2006).

upper-crustal levels in the Foreland belt (Evenchick, 1991a). The site of the Bowser basin evolved from a region of marine deposition on the western margin of the continent in the Jurassic, to a region of significant horizontal shortening in the interior of the orogen during the Cretaceous, as well as localized Late Cretaceous sedimentation in the Sustut basin. The Bowser and Sustut basins are thus keystones bridging the eastern and western parts of the Cordilleran orogen.

Data used as a basis for the synthesis presented here include: (1) integrated structural and geochronologic studies in the Omineca belt (O1–O10 in Table 1, and references therein), which reveal the diachronous nature of thermotectonic events at different structural levels, the evolution of structural geometry, and the role of metamorphism and plutonism in orogenic processes; (2) similar types of studies of the Coast belt, which have increased our knowledge of its Mesozoic tectonic evolution, and its association with the Intermontane belt (e.g., Crawford et al., 1987; Rubin et al., 1990; Rusmore et al., 2001); (3) a refined stratigraphic and structural framework for the Alberta foreland basin and Foreland belt (e.g., Mossop and Shetsen, 1994) and a new understanding of foreland basin provenance based on detrital zircon geochronology and isotope geochemistry (e.g., Ross et al., 2005); (4) mapping and interpretation of lithofacies assemblages across the Bowser basin, which illustrate the distribution of depositional environments (e.g., Evenchick et al., 2006); (5) integration of the depositional environments with fossil ages, resulting in the first paleontologically constrained depositional history for the Bowser basin (Evenchick et al., 2001; revised herein); (6) revision of the age of Cretaceous Sustut basin strata (A. Sweet, in Evenchick et al., 2001); (7) detrital zircon geochronology studies of the Bowser and Sustut basins, which refine our knowledge of the evolution of source areas (McNicoll et al., 2005); (8) recognition of at least 160 km of horizontal shortening in the Skeena fold belt (Evenchick, 1991a, 1991b, 2001); and (9) refined estimates of the timing and magnitude of Mesozoic–early Cenozoic dextral transcurrent faults east of the Bowser basin (Gabrielse, 1985; Gabrielse et al., 2006).

This paper begins with a review the geology of the morphogeological belts of the Canadian Cordillera (Fig. 1) with emphasis on Middle Jurassic to early Cenozoic evolution (Fig. 3; Table 1 provides sources of information). The focus is on a transect across the Cordillera that predates transcurrent faulting, wherein the Bowser and Sustut basins are restored to their probable site of formation adjacent to Omineca and Foreland belt rocks in the southeast Canadian Cordillera. We summarize the stratigraphy and depositional and structural histories of the Bowser and Sustut basins (Fig. 3), including new provenance data from detrital zircon studies (Fig. 4). The next section contains descriptions of the events occurring across the orogen in a series of successive "time slices," focusing on timing of basin initiation and sedimentation, deformation, magmatism, and metamorphism. These events are illustrated by paleogeographic maps (Fig. 5) and transorogen cross sections (Fig. 6).

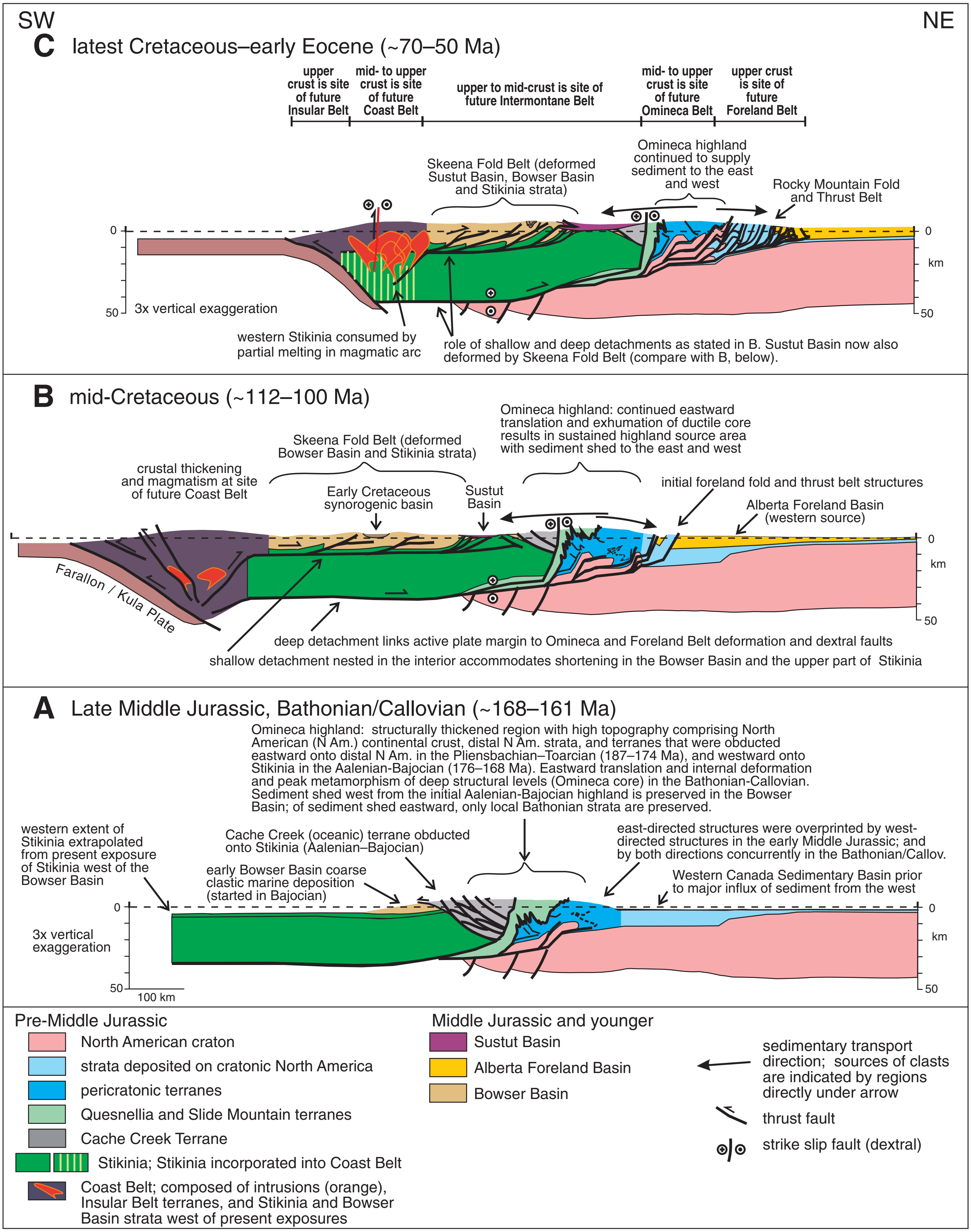

SW
NE
C latest Cretaceous–early Eocene (~70–50 Ma)
upper crust is site of future Insular Belt
mid- to upper crust is site of future Coast Belt
upper to mid-crust is site of future Intermontane Belt
mid- to upper crust is site of future Omineca Belt
upper crust is site of future Foreland Belt
Skeena Fold Belt (deformed Sustut Basin, Bowser Basin and Stikinia strata)
Omineca highland continued to supply sediment to the east and west
Rocky Mountain Fold and Thrust Belt
0
50
km
3x vertical exaggeration
western Stikinia consumed by partial melting in magmatic arc
role of shallow and deep detachments as stated in B. Sustut Basin now also deformed by Skeena Fold Belt (compare with B, below).
B mid-Cretaceous (~112–100 Ma)
Omineca highland: continued eastward translation and exhumation of ductile core results in sustained highland source area with sediment shed to the east and west
Skeena Fold Belt (deformed Bowser Basin and Stikinia strata)
crustal thickening and magmatism at site of future Coast Belt
Early Cretaceous synorogenic basin
Sustut Basin
initial foreland fold and thrust belt structures
Alberta Foreland Basin (western source)
Farallon / Kula Plate
deep detachment links active plate margin to Omineca and Foreland Belt deformation and dextral faults
shallow detachment nested in the interior accommodates shortening in the Bowser Basin and the upper part of Stikinia
A Late Middle Jurassic, Bathonian/Callovian (~168–161 Ma)
Omineca highland: structurally thickened region with high topography comprising North American (N Am.) continental crust, distal N Am. strata, and terranes that were obducted eastward onto distal N Am. in the Pliensbachian–Toarcian (187–174 Ma), and westward onto Stikinia in the Aalenian-Bajocian (176–168 Ma). Eastward translation and internal deformation and peak metamorphism of deep structural levels (Omineca core) in the Bathonian-Callovian. Sediment shed west from the initial Aalenian-Bajocian highland is preserved in the Bowser Basin; of sediment shed eastward, only local Bathonian strata are preserved.
western extent of Stikinia extrapolated from present exposure of Stikinia west of the Bowser Basin
Cache Creek (oceanic) terrane obducted onto Stikinia (Aalenian–Bajocian)
early Bowser Basin coarse clastic marine deposition (started in Bajocian)
east-directed structures were overprinted by west-directed structures in the early Middle Jurassic; and by both directions concurrently in the Bathonian/Callov.
Western Canada Sedimentary Basin prior to major influx of sediment from the west
3x vertical exaggeration
100 km
Pre-Middle Jurassic
North American craton
strata deposited on cratonic North America
pericratonic terranes
Quesnellia and Slide Mountain terranes
Cache Creek Terrane
Stikinia; Stikinia incorporated into Coast Belt
Coast Belt; composed of intrusions (orange), Insular Belt terranes, and Stikinia and Bowser Basin strata west of present exposures
Middle Jurassic and younger
Sustut Basin
Alberta Foreland Basin
Bowser Basin
sedimentary transport direction; sources of clasts are indicated by regions directly under arrow
thrust fault
strike slip fault (dextral)

Figure 2. Generalized cross sections illustrate tectonic elements and key events during the Mesozoic to early Cenozoic evolution of the southern Canadian Cordillera, from a narrow accretionary orogen (A) to a wide orogen with two major detachment systems (B, C). The lower detachment system links structural elements across the entire Cordillera. The upper detachment is the sole fault of the Skeena fold belt nested within the interior of the orogen. The Skeena fold belt is similar in width, magnitude of shortening, and timing of formation to the Rocky Mountain fold-and-thrust belt. Note that zones of penetrative ductile strain are not depicted on the cross sections; please see text and Table 1 for documentation of structural style, particularly in the Omineca and Coast belts.

Our purposes are to identify coeval and tectonically compatible events that illustrate linkages across the orogen and to use these as a basis for discussing its tectonic evolution.

GEOLOGICAL SETTING

The Canadian Cordillera (Fig. 1) is an amalgam of: (1) predominantly sedimentary deposits formed on and adjacent to the continental margin of ancestral North America, referred to herein as the craton margin; (2) terranes made mostly of arc and oceanic rocks that evolved separately early in their histories and subsequently became part of the Cordilleran "collage" (Coney et al., 1980; Gabrielse and Yorath, 1991); and (3) supracrustal rocks deposited in basins during and after terrane accretion. The five morphogeological belts of the Canadian Cordillera reflect both geological history and physiography (Gabrielse et al., 1991; Fig. 1). The major zones of terrane amalgamation and accretion coincide with the metamorphic-plutonic Omineca and Coast belts (Monger et al., 1982). These are flanked by belts of little metamorphosed or relatively low-grade sedimentary, volcanic, and plutonic rocks that form the Foreland, Intermontane, and Insular belts. The Foreland belt contains strata deposited on the ancient cratonal margin of North America, whereas the Intermontane and Insular belts coincide largely with the accreted terranes. In addition, there are syn- and postaccretion successions that were deformed into fold-and-thrust belts of Cretaceous to early Cenozoic age. The principal period of accretion of terranes of the Intermontane belt to North America was in Early to Middle Jurassic (e.g., Gabrielse and Yorath, 1991). The timing of accretion of Insular belt terranes to those farther inboard was either in the Jurassic or earlier (e.g., van der Heyden, 1992; McClelland et al., 1992), or in the Cretaceous (Monger et al., 1982). During and following amalgamation, there was significant orogen-parallel displacement on transcurrent faults within or marginal to the metamorphic-plutonic belts. Faults were dextral in and bordering the Omineca and Coast belts from mid-Cretaceous to early Cenozoic (e.g., Gabrielse, 1991a; Gabrielse et al., 2006). Prior to that, sinistral displacement of uncertain magnitude is inferred for the Early Cretaceous rocks outboard of the Bowser basin (Evenchick, 2001, and references therein).

Foreland Belt in British Columbia and Alberta

The Foreland belt is composed of Proterozoic to early Mesozoic continental-margin strata of the Western Canada sedimentary basin and of Mesozoic synorogenic strata of the Alberta foreland basin, the part of the Western Canada sedimentary basin that had the uplifting Cordillera as its main source (e.g., Stott and Aitken, 1993; Mossop and Shetsen, 1994). In the Foreland belt south of 58°N, west-derived sediments—a signal of emergence of a western source—first appear in the Oxfordian (F1 in Fig. 3; Table 1), except in the most southwestern part of the basin, where there is evidence for a western source in the Bajocian (Stronach, 1984). A major break in foreland basin sedimentation occurred in the Hauterivian to early Aptian, with a long period of pedimentation (Cadomin conglomerates; White and Leckie, 1999) and development of a low-angle unconformity with a stratigraphic separation that increases to the east (F3 in Fig. 3; Table 1). Clast composition and heavy mineral studies indicate that between 53°N and 57°N, western source areas for most of the coarse clastic units of the uppermost Jurassic to the Turonian include volcanic, plutonic, and medium-grade metamorphic rocks (McMechan and Thompson, 1993, and references therein). In contrast, south of 53°N, clast and heavy mineral studies, detrital zircon studies, and isotope geochemistry indicate that the western source areas were limited to low-grade metasedimentary rocks of the Omineca belt or the Foreland belt (McMechan and Thompson, 1993, and references therein; Ross et al., 2005). Exceptions are in Kimmeridgian strata, where detrital mica indicates a metasedimentary source, and in Albian strata, which have clasts of Intermontane belt volcanic and intrusive rocks (McMechan and Thompson, 1993, and references therein; Ross et al., 2005). Significant sediment accumulation and basin subsidence occurred in the western Alberta foreland basin during the Kimmeridgian to Valanginian (section up to 4 km thick; F2, Fig. 3), and over a broader area during the Campanian through Paleocene (section over 5 km thick; F5 in Fig. 3; Table 1).

Contractional thin-skinned structures of the Foreland belt are kinematically linked to structures in the polydeformed and metamorphosed Omineca belt and are essentially structurally continuous with them south of 53°N (McDonough and Simony, 1988; Kubli and Simony, 1994). Deformation of the foreland fold-and-thrust belt generally progressed from west to east and accommodated up to 200 km of horizontal shortening (Bally et al., 1966; Price and Mountjoy, 1970; Price, 1981; Fermor and Moffat, 1992). Structures near the western margin of the Foreland belt formed prior to ca. 108 Ma near 50°N (Larson et al., 2004) and ca. 100–112 Ma near 53°N (F4 in Fig. 3; McDonough and Simony, 1988). Faults in the east deform the youngest preserved strata (upper Paleocene near 53°N; Demchuk, 1990), and contractional deformation continued into the Eocene (Kalkreuth and McMechan, 1996). Most of the shortening of the southern Foreland belt occurred after the Turonian (younger than 89 Ma), concurrent with strike-slip faulting on the northern Rocky Mountain–Tintina fault system (Fig. 1; Price, 1994; Gabrielse et al., 2006).

TABLE 1. SUMMARY OF TIMING OF EVENTS IN THE INTERMONTANE, OMINECA, AND FORELAND BELTS

Intermontane Belt				Omineca Belt				Foreland Belt			
Fig. 3 code	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**
				O10	Penetrative high strain at the deepest structural levels (e.g., Thor-Odin dome; Frenchman Cap, in part); transition from transpressional to transtensional tectonics in Middle Eocene.	Eocene	Parrish et al. (1988); Gibson et al. (1999); Crowley et al. (2000); Hinchey (2005, and references therein)	**F5**	Rapid, coarse clastic sedimentation (100 m/m.y.); dominant Foreland belt source; major period of deformation with deformation front migrating eastward across Front Ranges and Foothills.	Late Campanian–Early Eocene (ca. 77–53 Ma)	Price and Mountjoy (1970); Price (1981); Stott and Aitken (1993); McMechan and Thompson (1993, and references therein); Ross et al. (2005)
I6	Initiation of coarse clastic deposition in Sustut basin (Campanian) and continued deformation in Skeena fold belt to affect youngest Sustut rocks (Maastrichtian). Sources same as in I5, but includes deposition of tuff.	Late Campanian to Early Maastrichtian, (ca. 74–68 Ma)	Eisbacher (1974a); Evenchick and Thorkelson (2005); MicNicoll (2005, personal commun.); McNicoll et al. (2005, 2006)	**O9**	South of 52°N: Penetrative polydeformation and metamorphism (e.g., eastern Selkirk fan, Selkirk allochthon north of Monashee Complex; mid-crustal zone south of Thor-Odin), crystalline nappes carried on ductile shear zones (e.g., Gwillim Creek shear zone in Valhalla Complex); zones of high strain (e.g., Monashee décollement in Frenchman Cap dome).	Late Cretaceous–Paleocene	Carr (1992); Scammell (1993); Crowley and Parrish (1999); Gibson et al. (1999); Johnston et al. (2000, and references therein); Gibson (2003); Kuiper (2003); Williams and Jiang (2005); Carr and Simony (2006); Brown and Gibson (2006, and references therein)				
I5	First deposition of easterly derived metamorphic clasts, clasts of cratonic North American lithologies, and detrital zircons of Proterozoic age (Sustut basin).	late Early Cretaceous (Albian; ca. 112–110 Ma)	Eisbacher (1974a); Evenchick et al. (2001); Evenchick and Thorkelson (2005); V.J. MicNicholl (2005, personal commun.); McNicoll et al. (2005, 2006)	**O8**	Purcell thrust system (post-Bearfoot and related thrusts and pre–Horsethief Creek batholith).	Mid-Cretaceous pre–ca. 93 Ma	Archibald et al. (1983); P.S. Simony (2005, personal commun.)	**F4**	Northeast-directed thrust faulting in westernmost Foreland belt near 53°N; cooling of Yellowjacket gneiss ca. 110 Ma, and 100 Ma deformation thrust faults immediately to east.	110–100 Ma (Albian)	McDonough and Simony (1988, and references therein)
				O7	Emplacement of Malton gneiss basement cored nappes via Bearfoot and related ductile thrusts (post–140–126 Ma as isograds [O6] deflected; pre– or syn–ca. 110 Ma cooling of Yellowjacket gneiss and 100 Ma deformation thrust faults in Foreland belt immediately to east).	late Early Cretaceous	McDonough and Simony (1988, and references therein); Digel et al. (1998); Crowley et al. (2000, and references therein)				

(*continued*)

TABLE 1. SUMMARY OF TIMING OF EVENTS IN THE INTERMONTANE, OMINECA, AND FORELAND BELTS (*continued*)

Intermontane Belt				Omineca Belt				Foreland Belt			
Fig. 3 code	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**
I4	Development of piggyback basin within Skeena fold belt signals end of solely east-derived sedimentation and is probably a record of early Skeena fold belt deformation.	early Early to middle Early Cretaceous (ca. 145–135 Ma)	Evenchick et al. (2001); Evenchick and Thorkelson (2005); V.J. McNicoll (2006, personal commun.)	**O6**	Penetrative polydeformation and metamorphism in the Cariboo, Monashee and Selkirk mountains (51.5–52.5°N latitude). Near line represented by Fig. 3: southeastward imbrication of basement slices at depth with SW-verging folds and faults at higher levels (e.g., Hobson Lake area); fold fans (e.g., Ozalenka fan). South of 52°N and at deep structural levels: generally NE-verging poly-deformation and high-strain zones (e.g., Scammell, 1993).	throughout the Late Jurassic, Early Cretaceous, and Late Cretaceous	Parrish (1995); Currie (1988); Scammell (1993); Digel et al. (1998); Crowley et al. (2000); Reid (2003, and references therein); Gibson (2003)	**F3**	Major period of pedimentation; development of low-angle unconformity with stratigraphic separation increasing eastward except in westernmost Foreland basin.	Hauterivian to Early Aptian (ca. 136–110 Ma)	Stott and Aitken (1993); Stott (1998); White and Leckie (1999)
				O5	Penetrative polydeformation and metamorphism—Allan Creek area, Cariboo Mountains.	ca. 143–126 deformation; ca. 135 Ma metamorphism	Parrish (1995); Currie (1988)				
I3	Transition to nonmarine conditions across central and southern Bowser basin (no record in southwest).	Jurassic-Cretaceous boundary (ca. 145 Ma)	Evenchick et al. (2001); Evenchick and Thorkelson (2005); Smith and Mustard (2006); V.J. McNicoll (2005, personal commun.)	**O4**	SW-verging penetrative deformation and metamorphism—eastern Hobson Lake area, Cariboo Mountains.	ca. 147 Ma syntectonic metamorphism	Reid (2003, and references therein)	**F2**	Rapid, west (Omineca highland) derived sedimentation (to 100 m/m.y.); shallowing- and coarsening-up sequence in Upper Jurassic.	Kimmeridgian-Valanginian (ca. 156–136 Ma)	Poulton (1989); Poulton et al. (1993, 1994b); Stott and Aitken (1993); Stott et al. (1993); Stott (1998)

(*continued*)

TABLE 1. SUMMARY OF TIMING OF EVENTS IN THE INTERMONTANE, OMINECA, AND FORELAND BELTS *(continued)*

Intermontane Belt				Omineca Belt				Foreland Belt			
Fig. 3 code	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**	**Fig. 3 code**	**Event**	**Timing**	**Reference**
I2	Rapid westward migration of facies belts in the Bowser basin.	Late Jurassic (Oxfordian/ Kimmeridgian; ca. 157–151 Ma)	Evenchick et al. (2001); Evenchick and Thorkelson (2005)	**O3**	Onset and/or formation of significant architecture of major structures such as SW-verging folds, fold fans, belts of NE-verging folds & faults. Peak of metamorphism ca. 165–160 Ma (biotite-grade Hobson Lake, Scrip Nappe).	(3a) 173–164 Ma in Kootenay arc & Purcell anticlinorium; >167 Ma Selkirk Fan (3b) 174–162 Ma in Cariboo Mountains	Archibald et al. (1983); Gerasimoff (1988); Struik (1988); Brown et al. (1992b); Parrish (1995); Warren (1997); Colpron et al. (1998); Gibson (2003); Reid (2003, and references therein); Brown and Gibson (2006, and references therein)	**F1**	First regional subsidence and thick W-derived (Omineca highland) sediments; overlies thin, craton-derived, basal transgressive sandstone.	Oxfordian (post–Early Oxfordian (ca. 158–156 Ma)	Poulton (1984, 1989); Poulton et al. (1993, 1994b); Stott (1998)
I1	At least 3 km deposition in Bowser basin focused in northeast trough.	Middle Jurassic (Bathonian to early Oxfordian; ca. 168–158 Ma)	Evenchick et al. (2001); Evenchick and Thorkelson (2005).					**F0**	Time-transgressive erosive disconformity at base of Alberta foreland basin succession.	Callovian-Oxfordian (ca. 164–160 Ma)	Poulton (1984); Poulton et al. (1993, 1994b)
I0	First east-derived sediment from the Cache Creek terrane deposited on Stikinia is the initial coarse clastic deposition in Bowser basin. Rapid exhumation of Cache Creek.	early Middle Jurassic (Aalenian/ Bajocian; ca. 176–168 Ma)	Gabrielse (1991b); Ricketts et al. (1992); Mihalynuk et al. (2004)	**O2**	Southwest-verging thrusts and isoclinal recumbent folds south of 51°N.	onset by 175 Ma	Parrish and Wheeler (1983); Klepacki (1985); Smith et al. (1992); Warren (1997); Colpron et al. (1996, 1998); Gibson (2003); Reid (2003)				
				O1	Obduction of Slide Mountain and Quesnellia terranes onto North America pericratonic terranes (e.g., Eureka, Pundata, Stubbs, Waneta faults and related structures).	ca. <187 Ma, 187–173 Ma	Tipper (1984); Parrish and Wheeler (1983); Murphy et al. (1995, and references therein); Beatty et al. (2006)				

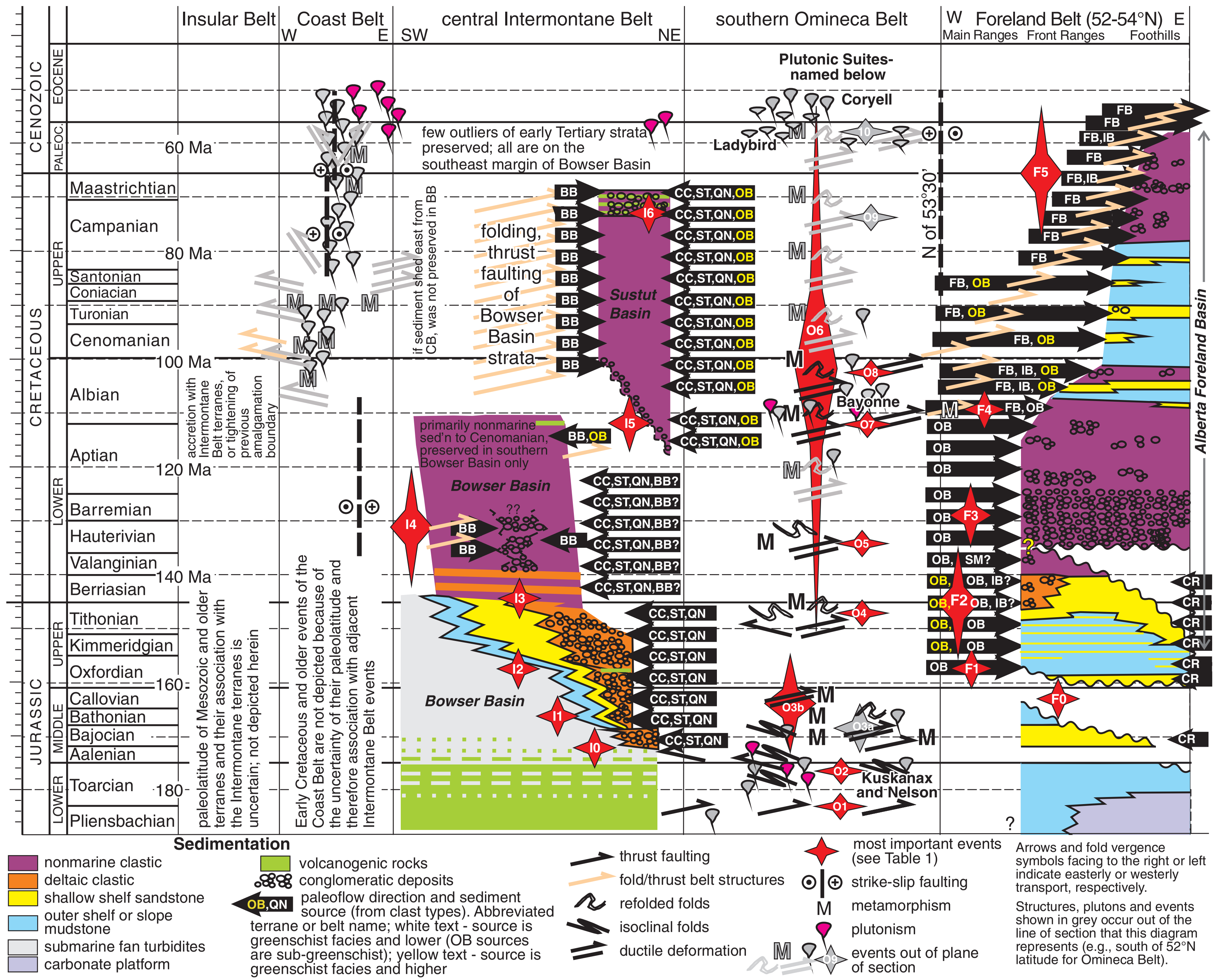
Southern Omineca Belt

The southern Omineca belt is composed of predominantly Proterozoic to Paleozoic supracrustal rocks formed on or near the cratonal margin of North America as well as late Paleozoic and early Mesozoic rocks of the most inboard accreted terranes. These rocks were polydeformed and metamorphosed during and following their Early to Middle Jurassic (ca. 187–174 Ma) accretion (O1 in Table 1), and some structural levels were reactivated and overprinted through the Jurassic, Cretaceous, Paleocene, and early Eocene (Table 1; Fig. 3). Protoliths are: (1) Paleoproterozoic basement exposed in structural culminations of the Monashee Complex and as imbricated thrust slices in the Malton Complex; (2) Mesoproterozoic to Paleozoic supracrustal rocks deposited on the western cratonal margin of North America; (3) lower Paleozoic sedimentary, volcanic, and igneous rocks of the pericratonic Kootenay terrane that formed on or adjacent to the North American margin; (4) Permian volcanic and ultramafic rocks of the oceanic Slide Mountain terrane; (5) Triassic to Lower Jurassic volcanic and sedimentary arc-related rocks of Quesnellia; (6) ca. 175–159 Ma calc-alkaline plutons of the Kuskanax and Nelson Suites; and (7) ca. 110–90 Ma granites of the Bayonne Suite. South of 51°N, Paleocene–Eocene peraluminous leucogranites occur in high-grade rocks exhumed from the mid-crust. The youngest rocks, middle Eocene intrusions of the syenitic Coryell Suite, coincide with the locus of significant Eocene east-west extension.

Although there is a record of Paleozoic interaction of offshore terranes with the western margin of North America (cf. Klepacki, 1985), and there is debate about the tectonic setting and paleogeography of the margin (cf. Thompson et al., 2006), our focus is tectonism related to the Late Triassic to Middle Jurassic obduction of terranes, and subsequent shortening and thickening of the orogen. Obduction of the Quesnellia and Slide Mountain terranes starting at ca. 187 Ma was accommodated by crustal shortening, as documented by folding and northeast-directed thrust faulting at low-grade metamorphic conditions, and this resulted in over 40 km of overlap of these terranes onto North American rocks (O1 in Table 1; Fig. 3). In the Middle Jurassic, shortening and thickening was accommodated by west-verging faults and fold systems, including regional-scale isoclinal folds with limbs tens of kilometers long (e.g., Scrip Nappe; Raeside and Simony, 1983; O2 in Table 1; Fig. 3). In the central Omineca belt, southwest-verging structures continued to form in the west (Schiarizza and Preto, 1987; Reid, 2003; Fig. 3). Elsewhere, upright folds, northeast-verging fold systems, regional fan structures, and northeast-directed thrust faults were superimposed on the southwest-verging nappes (O3 in Table 1; Fig. 3). Deformation was accompanied by regional greenschist-facies metamorphism, with some amphibolite-facies assemblages indicating that the crustal thickness was ~50–55 km (Table 1; Archibald et al., 1983; Colpron et al., 1996; Warren, 1997), or greater, and that this pulse of metamorphism reached its peak in the Middle Jurassic and was followed by rapid exhumation in the Middle Jurassic (O3 in Table 1). We follow the blind "tectonic wedge" model (Price, 1986; Struik, 1988; Murphy, 1989; Colpron et al., 1998) in which basement slices or ramps acted as wedges resulting in the formation of large-scale southwest-vergent structures in the detached and deforming overlying supracrustal rocks. Alternative models to explain the southwest-vergent Middle Jurassic structures, based on development of a retro-wedge geometry above a subduction zone (Brown et al., 1993; Brown and Gibson, 2006), are incompatible with timing and geometry of structures documented within the wedge, and the westward location of the magmatic arc and subduction zone relative to the position of deforming rocks (cf. Colpron et al., 1998). In the Late Jurassic and Early Cretaceous, progressive deformation and metamorphism continued in the Omineca belt (O4 and O5 in Table 1), and the foreland became progressively more involved, thus expanding the deforming highland and locus of thickening northeastward relative to the craton. In the Early Cretaceous, the lower structural levels of west-verging fold systems and related higher-level fold fans in the southern Omineca belt were reactivated and tightened during greenschist- and amphibolite-facies metamorphism, and sheets of basement and cover rocks of the Malton Complex were imbricated by northeast-vergent structures (O6 in Fig. 3). By the end of the Albian, the rocks at the latitude of discussion were being cooled and exhumed. However, more deeply exhumed rocks, now exposed south of 52°N, record zones of ductile moderate to extreme strain, folding, and transposition, indicating that mid-Cretaceous–Eocene strain partitioning at deep structural levels in the Omineca belt was concomitant with northeastward propagation of deformation and thin-skinned shortening in the Foreland belt (O9, O10 in Fig. 3).

Central Intermontane Belt

Most of the central Intermontane belt is underlain by Stikinia (Fig. 1), a terrane composed of volcano-plutonic arc assemblages of Devonian to Permian, Late Triassic, and Early Jurassic to early Middle Jurassic age (e.g., Monger and Nokleberg, 1996, and references therein). These strata exhibit a range of styles of deformation, and metamorphism from subgreenschist to greenschist facies. The Cache Creek terrane, in the eastern Intermontane belt (Fig. 1), is an accretionary complex of rocks formed in oceanic environments, with lesser volcanic arc strata. Rocks of oceanic affinity are Mississippian to Early Jurassic (Toarcian) in age and contain Permian Tethyan fauna that indicate an origin far from the North American craton margin (e.g., Struik et al., 2001). The structural style of Cache Creek terrane is dominated by southwest-vergent fold-and-thrust fault systems, superimposed in places on chaotic disrupted structures, and locally overprinted by northeast-verging fold systems (e.g., Gabrielse, 1991b; Struik et al., 2001; Mihalynuk et al., 2004). In the northern segment (Fig. 1), the Cache Creek terrane structurally overlies Stikinia, and locally Bowser basin strata, along the southwest-directed King Salmon fault (KSF, Fig. 1; e.g., Gabrielse, 1991b). In the southern segment (Fig. 1), the western accretionary boundary of Cache Creek terrane with Stikinia is obscured. In both segments, the eastern boundaries of Cache Creek terrane with Quesnellia are Cretaceous and



Insular Belt
Coast Belt
W
E
central Intermontane Belt
SW
NE
southern Omineca Belt
W
Foreland Belt (52-54°N)
E
Main Ranges
Front Ranges
Foothills
CENOZOIC
EOCENE
PALEOC.
CRETACEOUS
UPPER
LOWER
JURASSIC
UPPER
MIDDLE
LOWER
60 Ma
80 Ma
100 Ma
120 Ma
140 Ma
160
180
Maastrichtian
Campanian
Santonian
Coniacian
Turonian
Cenomanian
Albian
Aptian
Barremian
Hauterivian
Valanginian
Berriasian
Tithonian
Kimmeridgian
Oxfordian
Callovian
Bathonian
Bajocian
Aalenian
Toarcian
Pliensbachian
paleolatitude of Mesozoic and older terranes and their association with the Intermontane terranes is uncertain; not depicted herein
accretion with Intermontane Belt terranes, or tightening of previous amalgamation boundary
Early Cretaceous and older events of the Coast Belt are not depicted because of the uncertainty of their paleolatitude and therefore association with adjacent Intermontane Belt events
few outliers of early Tertiary strata preserved; all are on the southeast margin of Bowser Basin
if sediment shed east from CB, was not preserved in BB
folding, thrust faulting of Bowser Basin strata
Sustut Basin
primarily nonmarine sed'n to Cenomanian, preserved in southern Bowser Basin only
Bowser Basin
Bowser Basin
BB
BB,OB
CC,ST,QN,OB
CC,ST,QN,BB?
CC,ST,QN
CC,ST,QN
I0
I1
I2
I3
I4
I5
I6
Plutonic Suites-named below
Coryell
Ladybird
Bayonne
Kuskanax and Nelson
O1
O2
O3a
O3b
O4
O5
O6
O7
O8
O9
O10
M
N of 53°30'
FB
FB,IB
FB, OB
FB, IB, OB
FB,OB
OB
OB, SM?
OB, IB?
OB,
CR
F0
F1
F2
F3
F4
F5
Alberta Foreland Basin
?
Sedimentation
nonmarine clastic
deltaic clastic
shallow shelf sandstone
outer shelf or slope mudstone
submarine fan turbidites
carbonate platform
volcanogenic rocks
conglomeratic deposits
OB,QN
paleoflow direction and sediment source (from clast types). Abbreviated terrane or belt name; white text - source is greenschist facies and lower (OB sources are sub-greenschist); yellow text - source is greenschist facies and higher
thrust faulting
fold/thrust belt structures
refolded folds
isoclinal folds
ductile deformation
most important events (see Table 1)
strike-slip faulting
metamorphism
plutonism
events out of plane of section
Arrows and fold vergence symbols facing to the right or left indicate easterly or westerly transport, respectively.
Structures, plutons and events shown in grey occur out of the line of section that this diagram represents (e.g., south of 52°N latitude for Omineca Belt).

Figure 3. Chart illustrating late Early Jurassic to early Eocene tectonic events in the Canadian Cordillera, with emphasis on the central Intermontane, southern Omineca, and southern Foreland belts. Events labeled with red stars are discussed in the text; a summary of events and sources of information are provided in Table 1 and in the text. CC—Cache Creek; ST—Stikinia; QN—Quesnellia; BB—Bowser basin; OB—Omineca belt; FB—Foreland belt; SM—Slide Mountain; IB—Intermontane belt; CR—craton. Time scale here and elsewhere in text is after Gradstein et al. (2004).

Cenozoic dextral strike-slip faults such as the Thibert and Pinchi faults, in the north and south, respectively (Fig. 1; Gabrielse, 1985; Struik et al., 2001). The latter is inferred to overprint the Triassic–Jurassic Pinchi suture (Struik et al., 2001). Quesnellia is an arc terrane similar to Stikinia in general age and lithology (e.g., Monger and Nokleberg, 1996; Beatty et al., 2006). Significant differences occur in their pre-Triassic stratigraphy (e.g., Monger and Nokleberg, 1996), and Quesnellia has stratigraphic and structural ties with pericratonic Kootenay terrane, Slide Mountain terrane and the North American continental margin, which demonstrate that it probably formed adjacent to the continent (Beatty et al., 2006, and references therein). In addition, although Quesnellia has considerable north-south extent (Fig. 1), its map area is small relative to Stikinia. On the seismic-reflection profile that crosses most of the terranes in northern British Columbia, Quesnellia and Cache Creek are interpreted to form thin sheets (~2.5 km thick and ~7.5 km thick, respectively), whereas Stikinia is interpreted to be relatively thick (~35 km), comprising the entire crust above the Moho (Cook et al., 2004).

Bowser Basin

Much of northern Stikinia is overlain by strata of the Bowser Lake Group (Fig. 1), which were deposited in the Bowser basin and comprise a widespread upper Middle Jurassic to mid-Cretaceous marine and nonmarine clastic succession at least 6000 m thick (Tipper and Richards, 1976; Eisbacher, 1981; Evenchick and Thorkelson, 2005). Skeena Group, Cretaceous in age, contains rocks similar to coeval facies of the Bowser Lake Group, but includes volcanic successions, and many exposures occur south of its northern limit in the southern Bowser basin (Bassett and Kleinspehn, 1997; Smith and Mustard, 2006). Sand and pebble clasts in Bowser basin strata are dominated in many places by radiolarian chert derived from Cache Creek terrane; these clasts demonstrate the stratigraphic link between Cache Creek terrane and Stikinia and record the final stages of closure of the Cache Creek ocean (Gabrielse, 1991b; Evenchick and Thorkelson, 2005; I0 in Fig. 3).

The Bowser Lake Group is a monotonous assemblage of sandstone, siltstone, and conglomerate lacking laterally continuous stratigraphic markers. Lithofacies assemblages interfinger laterally and repeat vertically on a range of scales (I0 to I3 in Fig. 3). Strata were deposited in submarine fan and interfan, slope, shallow-marine, deltaic, fluvial, and lacustrine environments from southeast, east, and northeasterly sources in an overall regressive basin history (Tipper and Richards, 1976; Evenchick and Thorkelson 2005, and references therein).

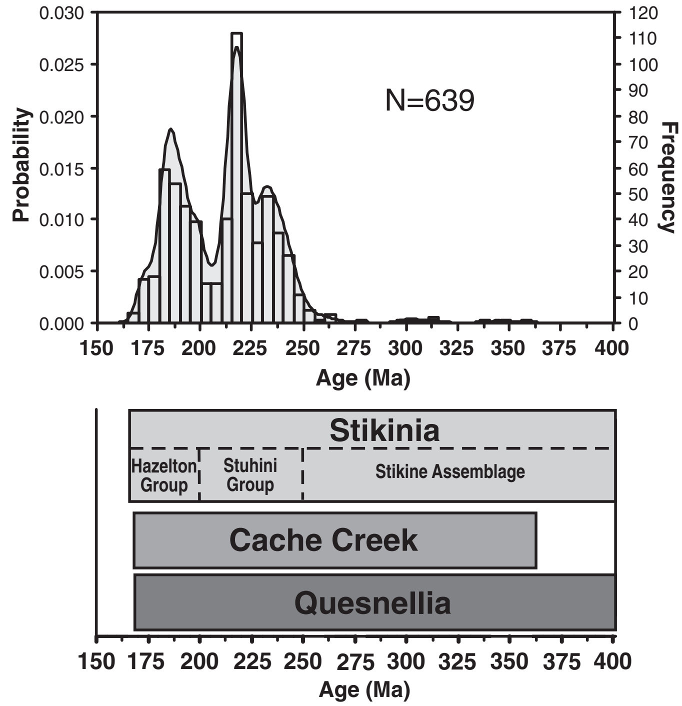


Figure 4. Cumulative probability plot of Bowser basin detrital zircon data for grains older than basin strata (Mississippian to early Middle Jurassic, ca. 360–169 Ma), from all Bowser basin samples (n = 21). Ages of rocks in Stikinia, Cache Creek, and Quesnellia are shown below. Of over 1435 single-grain analyses, 639 have ages older than the depositional age of strata in the basin. This diagram shows only the older Bowser basin data to highlight the ages of pre–Bowser basin sources. Potential source rock ages in Stikinia, Cache Creek and Quesnellia are from: Monger et al. (1991) and references therein; Monger and Nokleberg (1996) and references therein; Mihalynuk et al. (2004) and references therein; and Evenchick and Thorkelson (2005) and references therein.

Integration of timing constraints from index fossils with the distribution of major lithofacies assemblages demonstrates the migration of facies boundaries through time (Evenchick et al., 2001; Fig. 3). From Bathonian through early Oxfordian time, the major depocenter was restricted to the north-northeastern part of the basin (I1 in Table 1; Fig. 3) and only a condensed marine section formed at the western side. Sections in the south are also relatively thin and fine grained compared to those in the northeast (Tipper and Richards, 1976). Between mid-Oxfordian and early Kimmeridgian time, there was rapid south and southwest migration of facies proximal to the source over more distal facies (I2 in Fig. 3). New mapping in the central Bowser basin (Evenchick et al., 2006) indicates that the shelf–slope break migrated ~200 km southwestward during this time. The result was a wide, shallow, marine shelf bounded on the southwest by a region of submarine fan deposition at least 80 km wide that, based on its marine fauna, was probably

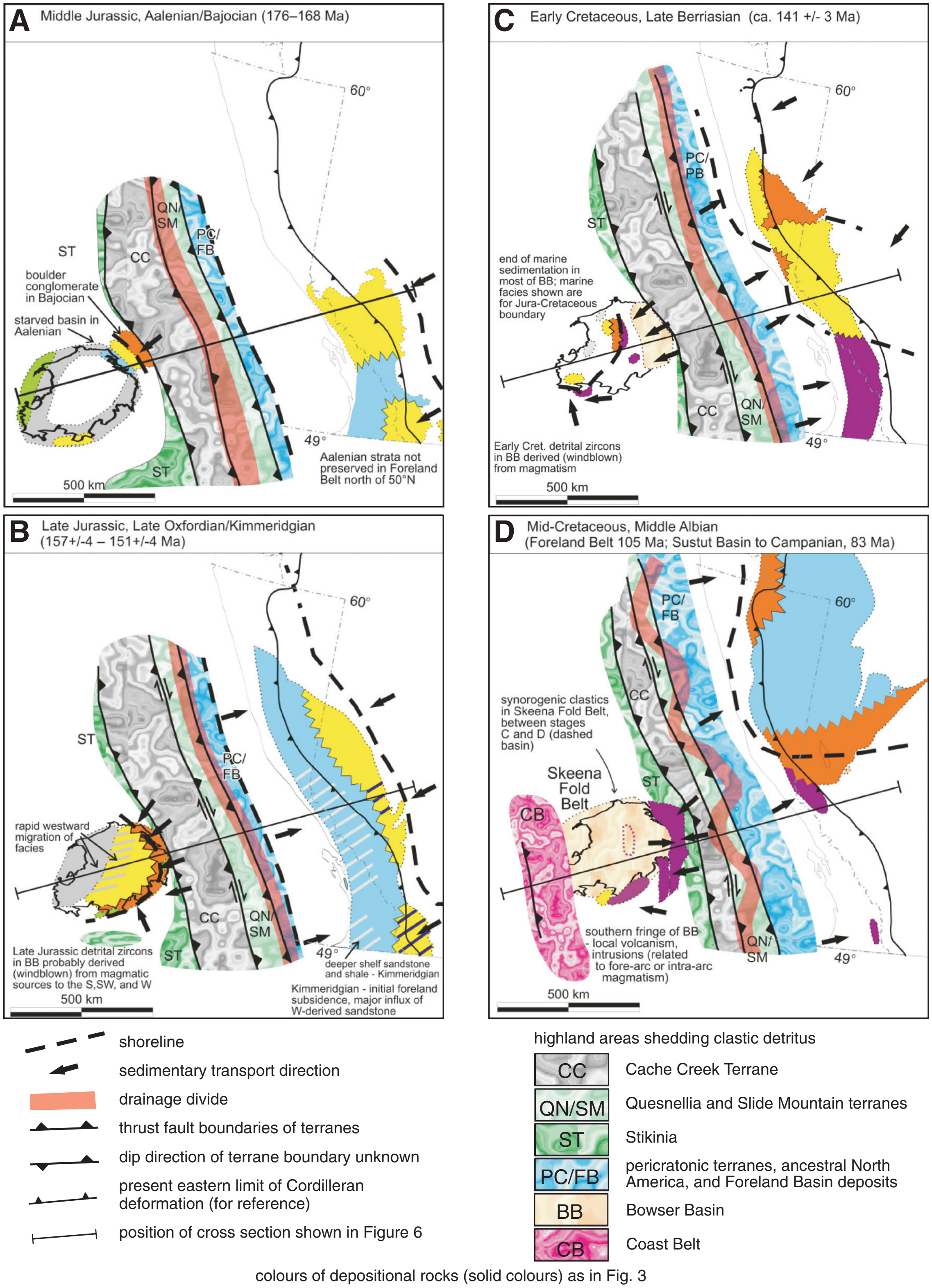


Figure 5 (*on this and following page*). Paleogeographic maps for the early Middle Jurassic through Late Cretaceous showing the inferred position of Bowser basin relative to cratonic North America (as discussed in text; North America is shown in present geographic coordinates), areas and types of sedimentation in the Bowser basin and Foreland belt, inferred source areas and drainage divide, and location of cross sections in Figure 6. The outline of the Bowser basin, with 50% shortening of the Skeena fold belt (SFB) restored, is shown for reference. Sources of information for deposition in the Bowser and Sustut basins are Eisbacher (1974a), Tipper and Richards (1976), Bassett and Kleinspehn (1997), Evenchick et al. (2001; and revisions based on new mapping in Evenchick et al., 2006), and Evenchick and Thorkelson (2005). Sources of information for facies belts in the Alberta foreland basin are Stott (1982, 1998), Hall (1984), Smith (1994), Poulton et al. (1994b), Stott et al. (1993), Leckie and Burden (2001), and McMechan et al. (2006).

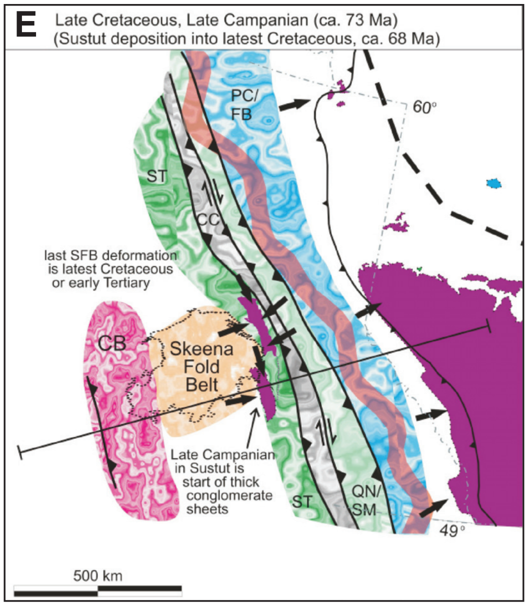


Figure 5 (*continued*).

open to the Pacific Ocean (Poulton et al., 1994a). Facies boundaries in the southern basin also migrated more rapidly than previously (Tipper and Richards, 1976). The shelf–slope break in the central and northern basin remained in about the same position into the latest Jurassic or earliest Cretaceous (I3 in Fig. 3); younger strata are absent in the western basin. In earliest Cretaceous time, deltaic and nonmarine strata were deposited in large parts of the northern basin, and nonmarine strata were deposited in the southern basin (Bassett and Kleinspehn, 1997; Smith and Mustard, 2006; V.J. McNicoll, 2006, personal commun.). By middle Early Cretaceous, these strata included thick conglomeratic braided river and alluvial fan deposits in the north-central Bowser basin (Eisbacher, 1974b; I4 in Fig. 3). Preserved strata of mid-Cretaceous age are restricted to the southern basin and are largely nonmarine clastic, but they include local volcanic centers and minor marine strata (Bassett and Kleinspehn, 1997). The youngest fluvial systems were probably continuous with the oldest Sustut fluvial systems deposited farther northeast. Ages of most Early Cretaceous strata are not narrowly constrained (Bassett and Kleinspehn, 1997; Evenchick and Thorkelson, 2005, and references therein; V.J. McNicoll, 2004, 2006, personal commun.). Deposition of detrital muscovite in the southern basin, likely starting in Aptian or Albian time, signals initiation of metamorphic Omineca belt detritus shed to the west (Bassett and Kleinspehn, 1997), and its initiation was roughly coeval with deposition of muscovite in the Sustut basin farther northeast (Eisbacher, 1974a; Evenchick and Thorkelson, 2005).

New provenance data that refine our understanding of the depositional history of the Bowser basin are provided by detrital zircons analyzed from sandstone samples that have paleontologically well-constrained depositional ages and from diverse ages, map units, and areas of the basin. The results, based on U-Pb sensitive high-resolution ion microprobe (SHRIMP) analyses of over 1435 detrital zircons from 21 samples of Bowser Lake Group, ranging from Bathonian to earliest Cretaceous age, show that the Bowser basin was receiving detritus mainly from sources of Early Triassic age to as young as the depositional age of the rock sampled (Fig. 4; McNicoll et al., 2005). The source regions indicated by paleocurrents, clast types, and facies distribution of Bowser basin strata are Cache Creek terrane, Quesnellia, and Stikinia (e.g., Evenchick and Thorkelson, 2005, and references therein). These are potential sources for the detrital zircons that are older than Bowser basin strata (Fig. 4). However, the age of the youngest zircon population in each rock is indistinguishable from the paleontologically determined depositional age of the rock, and it is interpreted to have originated from wind-blown ash from sources south, southwest, and/or possibly west of the Bowser basin (McNicoll et al., 2005).

Regional relationships indicate that basin subsidence for the Jurassic and earliest Cretaceous Bowser basin was controlled, in part, by flexural subsidence resulting from sediment load and obduction of Cache Creek terrane, and by thermal subsidence resulting from cessation of arc-related magmatism within northern Stikinia (e.g., Eisbacher, 1981; Ricketts et al., 1992; Evenchick and Thorkelson, 2005); the role of dynamic subsidence in Bowser basin evolution is unknown. An outlier of Early Cretaceous braided river and alluvial fan deposits in the northern Bowser basin (I4 in Fig. 3) is interpreted to represent synorogenic deposition within the Skeena fold belt (Evenchick and Thorkelson, 2005).

Sustut Basin

The Sustut Group is composed of more than 2000 m of Late Cretaceous nonmarine clastic strata (Eisbacher, 1974a). Since Eisbacher's (1974a) description and interpretation of the group, new constraints on the ages of units (A. Sweet, in Evenchick and Thorkelson, 2005), and additional stratigraphic and structural relationships, have clarified its tectonic significance (Evenchick and Thorkelson, 2005).

The lower of two formations, the Tango Creek Formation, is up to 1400 m thick and overlies Triassic to Late Jurassic units, including Bowser Lake Group, with angular unconformity (Eisbacher, 1974a). It is dominated by sandstone, siltstone, and mudstone, and in the upper part, by mudstone, calcareous siltstone, and calcareous sandstone. Paleocurrents, which in the lowest part are to the south and southwest, and high quartz clast content of sandstone are both consistent with the interpretation of derivation from a northeastern, Omineca belt source (Eisbacher, 1974a). Quartzite characteristic of Lower Cambrian miogeoclinal strata occurs as pebbles; these were also likely derived from the Omineca belt. In the middle and upper part of the formation, paleoflow to the northeast was accompanied by an increase in chert content in sandstone, interpreted as Cache Creek clasts recycled from Bowser Lake Group strata. The base of the formation is diachronous, ranging from Barremian–early Albian, to Coniacian–Campanian, and the upper age limit is late Campanian (I5 in Fig. 3).

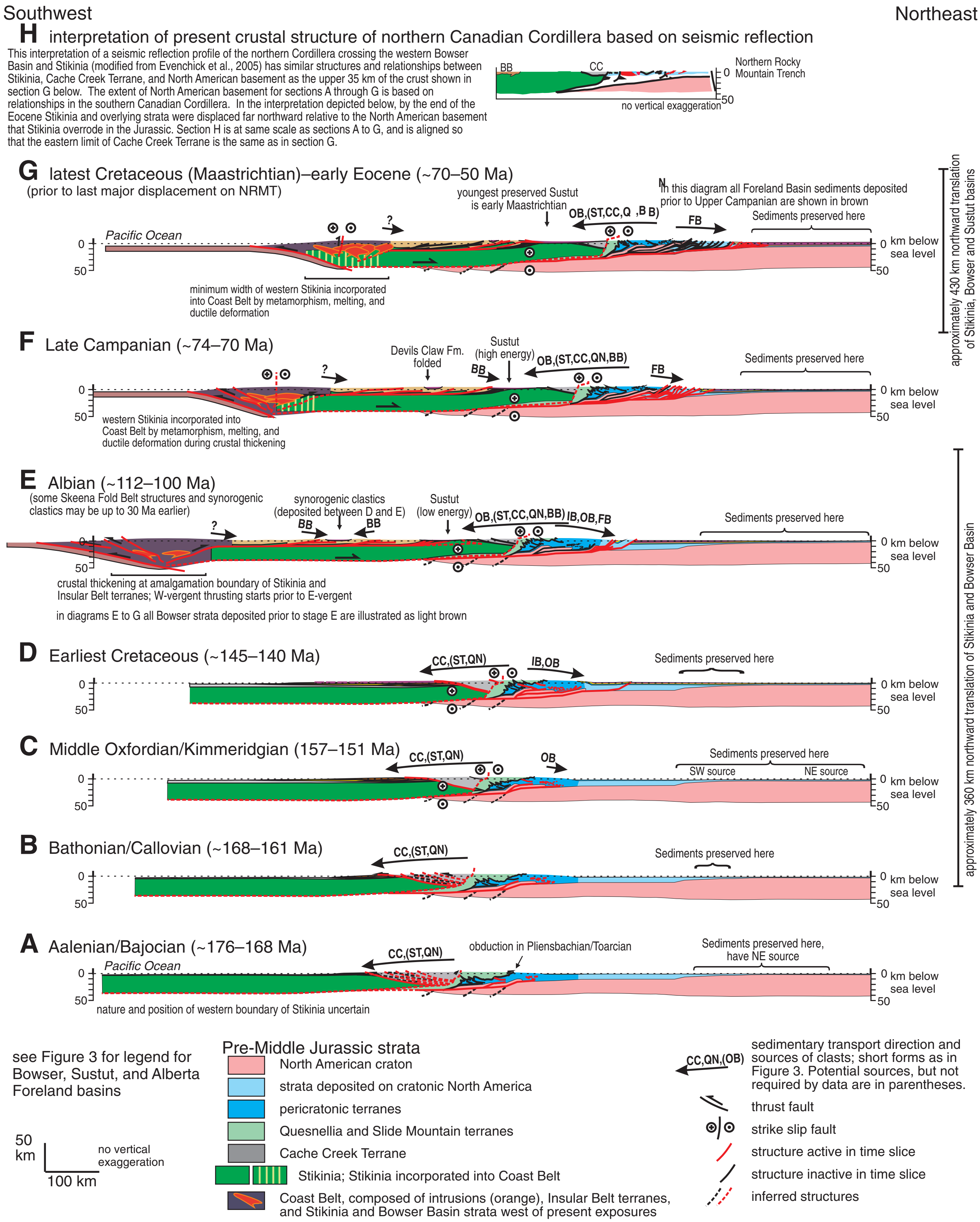


Figure 6 (*on this and following page*). Cross sections illustrating major depositional, structural, metamorphic, and plutonic events in the region discussed in the text. Abbreviations are as in Figure 3. Note that zones of penetrative ductile strain are not depicted on the cross sections; please see text and Table 1 for documentation of structural style, particularly in the Omineca and Coast belts. The set with no vertical exaggeration shows structural relationships. The set with 3× vertical exaggeration is included to show the depositional units. The location of cross sections A, C, D, and E are given in Figure 5.

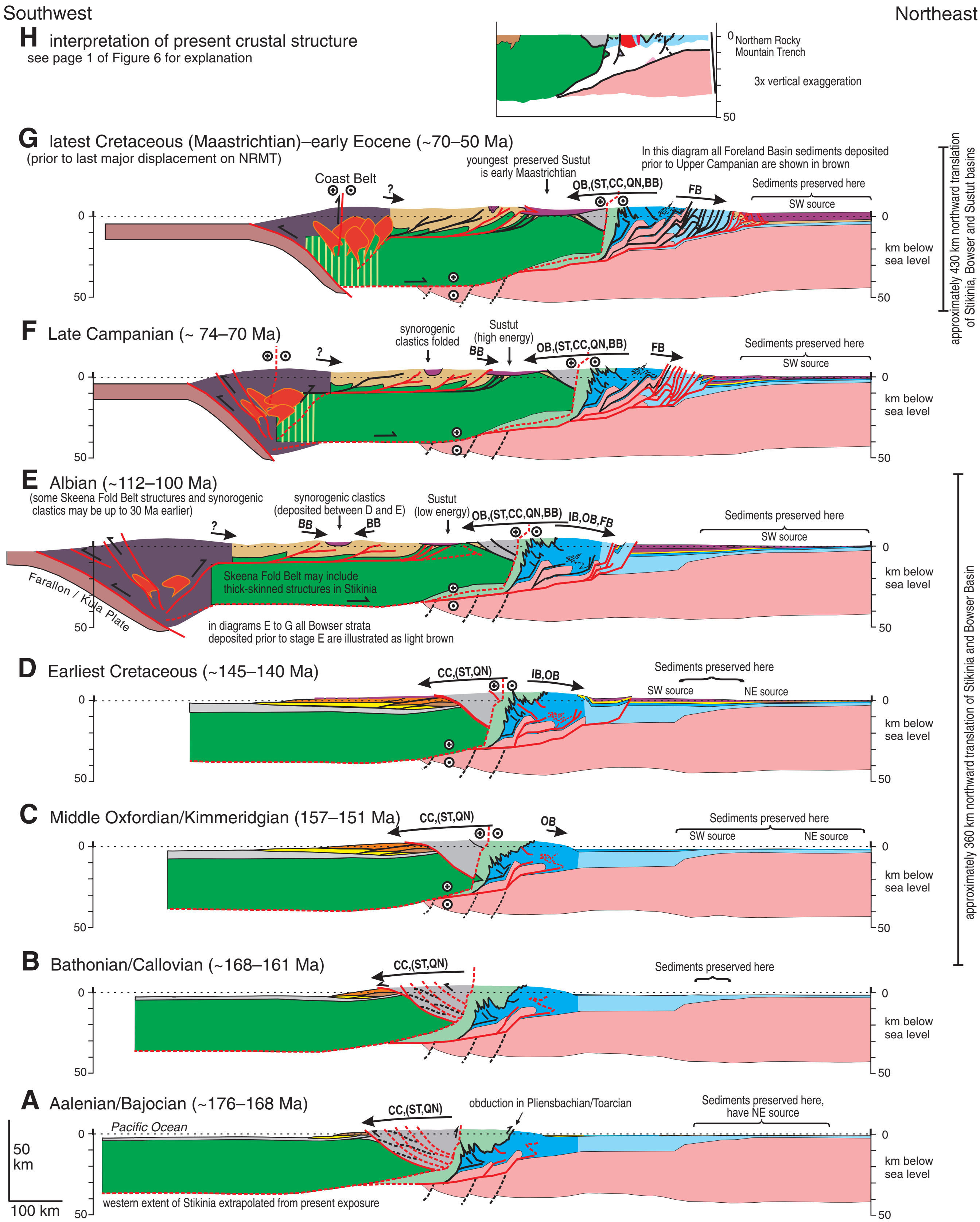


Figure 6 (*continued*).

The Tango Creek Formation is abruptly and conformably overlain by the Brothers Peak Formation, which is characterized by polymict conglomerate, sandstone, and felsic tuff (Eisbacher, 1974a). A basal conglomeratic succession is commonly more than 50 m thick. Paleocurrents were primarily southeast, longitudinally down the basin, with input from the north, east, and west (Eisbacher, 1974a). Strata are late Campanian to late early Maastrichtian (A. Sweet, in Evenchick et al., 2001). Two felsic tuff layers in the formation have been dated at ca. 75 and ca. 71 Ma (V.J. McNicoll, 2004, personal commun.). The dramatic change upward from the mudstone and siltstone of the upper Tango Creek Formation to the conglomerate-rich base of the Brothers Peak Formation (I6 in Fig. 3) indicates a marked increase in energy of Sustut fluvial systems.

Paleocurrent and clast types show that initial Sustut basin deposits had an eastern Omineca belt source that continued throughout the depositional history; however, early in this history, development of the Skeena fold belt provided an additional, southwest, source of sediment (Eisbacher, 1974a). Detrital zircon data show that the Sustut basin shared similar Triassic to Early Cretaceous sources with the Bowser basin, and/or zircons were recycled from the Bowser basin (McNicoll et al., 2005). In contrast, the Sustut basin also received Archean, Paleoproterozoic, Mesoproterozoic, Paleozoic, and Late Cretaceous zircons that are clearly distinct from Bowser basin sources (McNicoll et al., 2005) and are consistent with provenance studies of the Sustut Group that illustrate an Omineca belt source in the late Early Cretaceous (Eisbacher, 1974a).

Skeena Fold Belt

The Skeena fold belt is a regional fold-and-thrust belt that is best expressed in the thinly bedded clastic rocks of the Bowser and Sustut basins, but also involves Stikinia, as shown by folded contacts, structural culminations of Stikinian rocks within the fold belt, and klippen of early Mesozoic Stikinian rocks on Cretaceous Sustut basin strata. Details of the geometry, magnitude of shortening, and timing of the fold belt have been documented by Evenchick (1991a, 1991b, 2001) and Evenchick and Thorkelson (2005). The fold belt has accommodated a minimum of 44% (160 km) northeasterly shortening, it locally terminates to the northeast in a triangle zone within the Sustut basin, and is inferred to root to the west in the Coast belt.

Folds of a range of scales, from several hundred meters to a kilometer or more in wavelength, are the most obvious structures, and northwest-trending fold trains occupy most of the fold belt. Most verge northeast, and they vary from upright to overturned. Thrust faults in Bowser strata are apparent where they juxtapose the Bowser Lake Group against other map units, but they are difficult to recognize in most of the basin unless hanging-wall or footwall cutoffs are exposed. Thrust faults and/or detachment zones are required by the style of folding and are inferred to be largely bedding-parallel, blind thrusts (Evenchick, 1991b).

The age of the fold belt is constrained by regional stratigraphic relationships. The youngest folded marine Bowser basin strata, deposited at the Jurassic–Cretaceous boundary, lack a western source or other indications that the fold belt had evolved significantly. Sustut basin strata unconformably overlie contractional structures that involve Bowser basin and Stikinia strata, illustrating contractional deformation and erosion of Stikinian and Bowser strata in the northeast prior to the Albian (I5 in Fig. 3). Uppermost Sustut Group strata are the youngest deformed rocks. Western Skeena fold belt structures are overlain by flat-lying Pliocene volcanic rocks (Evenchick and Thorkelson, 2005). These relationships demonstrate that the fold belt was initiated in the Early Cretaceous with at least some deformation prior to the Albian, and that it ended in latest Cretaceous or early Cenozoic time.

Direct constraints on the magnitude of shortening for specific structures and/or time periods are sparse. A ca. 84 Ma post-tectonic pluton near the central Bowser basin intruded Late Jurassic strata and constrains deformation there to between Late Jurassic and Campanian (Evenchick and McNicoll, 1993). Synorogenic clastic rocks include an Early Cretaceous piggyback basin in the north-central part of the fold belt, which records Early Cretaceous (I4 in Fig. 3) subaerial erosion of topographic highs in the central and eastern fold belt. The Sustut Group itself is synorogenic and records a western source (Skeena fold belt) from Albian through early Maastrichtian time. Angular unconformities within the lowest Sustut strata are indications of Albian or Cenomanian tectonism in the northeastern fold belt. The dramatic change within the Sustut Group from low- to high-energy fluvial systems at the base of the Brothers Peak Formation (I6 in Fig. 3) indicates increased relief in the late Campanian.

Coast and Insular Belts

The Coast belt is composed mainly of Middle Jurassic to early Cenozoic plutonic rock and lesser amounts of greenschist-to amphibolite-, and locally granulite-facies metamorphic rock (e.g., Gabrielse et al., 1991). In the northern Coast belt (Fig. 1), Cretaceous continental arc magmatism, interpreted to be associated with accretion of terranes and subduction of Farallon, Kula, and possibly Resurrection plates, migrated eastward, resulting in a western 105–90 Ma arc and an eastern 80–50 Ma arc, separated by the Coast shear zone (e.g., Crawford et al., 2005, and references therein). Contractional deformation of these arcs resulted in significant crustal thickening in the mid-Cretaceous, continuing to ca. 60 Ma; arc igneous activity ended at ca. 50 Ma with the beginning of a period of extension, and the eastern side of the arc was exhumed on bounding shear zones (e.g., Andronicos et al., 2003; Crawford et al., 2005). Farther west, the Insular belt consists of little metamorphosed Late Proterozoic to early Mesozoic volcanic arc terranes (mainly Wrangellia and Alexander) that are stratigraphically distinct from the more inboard terranes (e.g., Monger and Nokleberg, 1996, and references therein). Additional components of the Insular belt are minor late Mesozoic and Cenozoic accretionary complexes (e.g., Gabrielse et al., 1991) and Proterozoic to Paleozoic metamorphosed continental-margin sequences (e.g., Gehrels and Boghossian, 2000, and references therein). The western boundaries of Stikinia and the Bowser basin are within the Coast belt, but their relationships with Insular belt terranes are obscured by the large volume of intrusive rocks, medium- to high-grade metamorphism, and high-strain zones; thus, the timing of amalgamation of terranes of the Intermontane and Insular belts is controversial. It was either in the Jurassic or earlier (e.g., van der Heyden, 1992; McClelland et al., 1992) or the mid-Cretaceous (e.g., Monger et al., 1982). Uncertainty in the accretion history and in magnitude of postaccretion orogen-parallel translation leads to uncertainties about the Middle Jurassic to early Cenozoic paleogeography of these terranes. For these reasons we discuss only Cretaceous and early Cenozoic features of significant strike length, and only in general terms. Examples are: Cretaceous northeast-vergent contractional structures on the east side of the Coast belt that merge with contractional structures in the Intermontane belt (e.g., Evenchick, 1991a, 1991b; Journeay and Friedman, 1993; Rusmore and Woodsworth, 1994); mid-Cretaceous crustal thickening and magmatism (e.g., Crawford et al., 1987, 2005); and west-vergent contractional structures on the west side of the Coast belt (e.g., Rubin et al., 1990; Journeay and Friedman, 1993).

MESOZOIC TECTONIC EVOLUTION

Paleogeographic Reconstructions—Restoration of Jurassic to Eocene Orogen-Parallel Faults

Discussion of the Mesozoic evolution of the Bowser basin and Skeena fold belt in the context of tectonism in the Omineca and Foreland belts requires restoration of displacement on orogen-parallel strike-slip faults within and bordering the Omineca belt (Fig. 1). The discrepancies between geological and paleomagnetic estimates of dextral motion in the Late Cretaceous and Eocene have been reviewed and discussed by Gabrielse et al. (2006). The amount of Late Cretaceous to Eocene dextral displacement on the array of faults including the Tintina, Northern Rocky Mountain Trench fault, and related splays (Fig. 1) is estimated to be 490 km, with 430 km primarily in the Eocene, and ~60 km on the Northern Rocky Mountain Trench fault in early Late Cretaceous (Gabrielse et al., 2006). Additional strike-slip faults of this age are the Kechika-Spinel (80 km; a splay of the Northern Rocky Mountain Trench fault), and the Pinchi fault. The magnitudes and specific ages of older displacement on these faults, and of older faults, are less well constrained. Offset of lower Paleozoic facies boundaries suggest a total of ~700 km displacement (Gabrielse et al., 2006) on the Northern Rocky Mountain Trench fault; if 490 km of this was in the Late Cretaceous to Eocene, earlier displacement must have been ~210 km. Other Late Cretaceous or older fault systems to consider are (see Fig. 1): (1) the pre-Eocene part of the Kechika-Sifton fault, a splay of Northern Rocky Mountain Trench (~90 km); (2) the Kechika-Thudaka-Finlay-Ingenika-Takla system (~110 km); and (3) the Cassiar-Kutcho-Thibert system, which must be younger than the early Middle Jurassic (or younger) contractional structures and Early Jurassic intrusions that they displace. Faults of the latter two fault systems that cut the Cassiar batholith are considered to be synchronous with intrusion of the batholith at ca. 95–110 Ma (Gabrielse et al., 2006).

Figure 5A reconstructs the paleogeography of the early Middle Jurassic to account for the faults listed previously. Their continuation to the south as discrete structures is problematic (see Gabrielse et al., 2006), which makes an accurate paleogeographic reconstruction challenging. Accordingly, we restore Stikinia and the Bowser basin southward to approximately the latitude indicated by strike-slip faults in the north (~800 km), yet outboard enough to allow for shortening in the Foreland and Omineca belts (~300 km), stacking of terranes in the Omineca belt, and Eocene extension. Following Gabrielse (1985), we depict the Cache Creek terrane and Quesnellia in Figure 5A as continuous belts, with Cache Creek in thrust fault contact with Stikinia. In doing so, we assume that most of the later dextral fault displacement occurred within and/or between Cache Creek terrane and Quesnellia and is now obscured by Paleogene and younger strata of the southern Intermontane belt. Stikinia and the Bowser basin are restored with clockwise rotation from the present orientation, which in part reflects the restoration of known faults and allows for rotation about the Euler pole during northward translation (Price and Carmichael, 1986). To depict the minimum original width of Stikinia and the Bowser and Sustut basins, horizontal shortening of the Skeena fold belt, estimated at ~50%, is restored from across the Bowser basin, from more than half of the Sustut basin and most of Stikinia.

This reconstruction puts the Bowser basin adjacent to the site of the southern Canadian Cordillera at the time of final closure of the Cache Creek ocean in the Aalenian-Bajocian, just prior to the onset of major clastic deposition. In the absence of constraints on net displacements on strike-slip faults for specific periods, we assume, from Gabrielse et al. (2006), that dextral displacements were (1) ~300 km between Middle Jurassic and mid-Cretaceous time, with ~100 km on the fault systems west of the Northern Rocky Mountain Trench fault and 200 km on the Northern Rocky Mountain Trench fault, and (2) ~500 km on the Northern Rocky Mountain Trench–Tintina faults and related splays in the Late Cretaceous to late Eocene.

Relationship of Structures in Neighboring Intermontane, Omineca, and Foreland Belts, and Development of Basins and Fold-and-Thrust Belts Flanking the Omineca Belt

In the following sections, the major deformation, metamorphic, magmatic, and depositional events in the north-central and southeastern Canadian Cordillera are summarized for a series of time periods from the Middle Jurassic to Early Cretaceous, with emphasis on the regions that became the Intermontane, Omineca, and Foreland belts. The morphogeologic belt terminology applies to the Cordillera today; however, we use the terms as a convenient way to refer to regions that eventually became the belts. Cross sections (Fig. 6) for each time period show the tectonic evolution and relationships between events in neighboring belts, such as links between highland sources and basins of deposition. Although the cross sections are drawn with the east side fixed relative to eastward translation of terranes and structures, events described are the result of the North American plate moving westward relative to the hot-spot reference frame, with its western convergent plate boundary interacting with, and accumulating, parts of microcontinents, volcanic arcs, and intervening ocean basins of the Pacific Ocean—the future terranes of the Cordillera (e.g., Engebretson et al., 1985; Coney and Evenchick, 1994). Deposition in the Bowser and Alberta Foreland basins, shown in Figure 5, and development of mid-crustal structures in the Omineca belt and those in the upper crust in the Foreland belt, shown in Figure 6, are relatively well constrained because they are now exposed, are confidently inferred by plunge projection of structures, or have been drilled by hydrocarbon exploration wells. The geometry of the boundary between the Cache Creek terrane and Quesnellia in Figure 6, however, is highly uncertain because the only contacts known are the dextral faults that may obscure possible earlier low-angle faults. Other geometries that satisfy the constraint of the Cache Creek terrane forming a major part of the source region for clasts deposited in the Bowser basin are possible. The thickness of Stikinia in Figure 6 is interpreted from seismic sections (Cook et al., 2004), but its thickness through time is poorly constrained. It could have been significantly thinner in the Jurassic and later thickened by a combination of thin- and thick-skinned contractional structures during formation of the Skeena fold belt, with inversion of early Mesozoic extensional structures. The nature and position of the original western boundary of Stikinia are poorly constrained, and thus only the palinspastically restored width of exposed Stikinia strata is represented Figure 6.

Early Jurassic–Pliensbachian/Toarcian (ca. 190–176 Ma)

Prior to the Middle Jurassic, Stikinia was separated from cratonic North America by a region of marginal or pericratonic terranes, Quesnellia, and the last vestiges of the Cache Creek ocean, which was closing probably as a result of southwest subduction beneath Stikinia and northeast subduction beneath Quesnellia (e.g., Mortimer, 1987; Marsden and Thorkelson, 1992). Upper parts of Cache Creek terrane were delaminated to form the accretionary complex that now sits in thrust contact above Stikinia. Deposition in the Cache Creek ocean ended in the latest Early Jurassic (Toarcian; e.g., Struik et al., 2001), and northern Stikinia evolved from a region of widespread subduction-related volcanism in the Pliensbachian and earlier, to a largely marine clastic environment with minor volcanism by the Aalenian-Bajocian (Fig. 3; e.g., Marsden and Thorkelson, 1992; Anderson, 1993). The structural overlap of Quesnellia over North American basement at this time is documented by changes in the geochemistry of volcanic-arc rocks (Ghosh and Lambert, 1995) and the geochemistry and composition of detritus in volcaniclastic sediments of the arc (Petersen et al., 2004). Early Jurassic I-type intrusions of the Kuskanax and Nelson Suites indicate the presence of a subduction zone, and their isotope geochemistry indicates primitive signatures contaminated with that of continental North America (Armstrong, 1988; Ghosh and Lambert, 1995). In northern Quesnellia, there was increasing cratonal influence on granitic rocks of Triassic to mid-Cretaceous age (Gabrielse, 1998). Quesnellia and Slide Mountain terranes were obducted onto the North American pericratonic terranes and imbricated and thrust eastward along the Eureka, Pundata, Stubbs, and related faults between ca. 187 and 173 Ma (O1 in Fig. 3; Murphy et al., 1995, and references therein). Obduction was closely followed by the onset of southwest-directed thrusts and isoclinal recumbent folding, which continued in the Middle Jurassic (O2 in Fig. 3). Dating of east-vergent structures in eastern Quesnellia in the east-central Intermontane belt indicates that the initial stages of obduction there occurred at ca. 186 Ma (Nixon et al., 1993). In the Foreland belt, a Sinemurian to middle Toarcian cherty carbonate platform south of 54°N changed westward and northward into a narrow belt of shale and carbonate sandstone and then into a westward thinning unit of phosphatic mudstone and limestone (Asgar-Deen et al., 2004). All facies in the Foreland belt were overlain by a thin, euxinic, black shale deposited during the worldwide Toarcian (anoxic black shale) "transgressive" event (Poulton et al., 1994b).

Early Middle Jurassic–Aalenian/Bajocian (ca. 176–168 Ma)

An early phase of Bowser basin deposition began with subsidence in the northeast, marked by a starved phase in the Aalenian, and followed in the Bajocian by deposition of subaerially eroded Cache Creek strata in the northeasternmost part of the basin (I0 in Fig. 3; Fig. 6A). These events are interpreted to be a result of southwest thrusting of Cache Creek strata onto Stikinia (Gabrielse, 1991b; Ricketts et al., 1992). Stacking of Quesnellia and pericratonic terranes onto or against Cache Creek probably facilitated crustal thickening in the source area (Fig. 6A). Some of the first Cache Creek chert clasts deposited in the Bowser basin are close in age to the youngest blueschist (173.7 ± 0.8 Ma) resulting from Cache Creek subduction, a relationship interpreted by Mihalynuk et al. (2004) to indicate rapid exhumation of northern Cache Creek strata at ca. 174–171 Ma. In the northern part of the south segment of Cache Creek terrane, westward obduction of Cache Creek onto Stikinia occurred between 190 and 165 Ma, and west-vergent structures were overprinted by east-vergent ones (Struik et al., 2001).

In the Omineca belt, the formation of large southwest-verging, recumbent isoclinal folds and polyphase deformation record progressive crustal thickening and low- to medium-grade regional metamorphism (O1–O3 in Table 1; Fig. 3). By ca. 173–168 Ma, northeast-verging fold and fault systems were superimposed on southwest-verging fold systems, south of 52°N, as indicated by the relationships of syn- and post-tectonic ca. 173–168 Ma plutons of the Kuskanax and Nelson plutonic suites in the Selkirk and Purcell Mountains (Armstrong, 1988; Parrish and Wheeler, 1983; Colpron et al., 1998; Gibson, 2003; O2 in Table 1; Fig. 3).

North of 52°N, polyphase southwest-verging fold systems in the Cariboo Mountains and thrust faults at higher structural levels were coeval with northeastward thrusting on shear zones and penetrative deformation at depth (O3b in Table 1; Fig. 3; Struik, 1988). Following Price (1986), Struik (1988), and Murphy (1989), we use a tectonic wedge model with southwest-directed back thrusting and folding of the cover above northeast-directed detachments and thrusting of the craton to explain the geometry and structural evolution of the internally deforming and thickening crust (Fig. 6A). In the Foreland belt, 30–90 m of shelf mudstone and northeasterly derived sandstone were deposited in the Bajocian and possibly Aalenian (Fig. 3; Hall, 1984).

The western Omineca belt and eastern Intermontane belt collectively defined a growing region of structurally thickened Cache Creek, Quesnellia, and pericratonic terranes, the growth of which was concomitant with the onset of westward deposition of Cache Creek detritus into the Bowser basin. Changes in structural vergence are present in both Intermontane (Cache Creek) and Omineca belt rocks and may indicate the onset of decoupling of supracrustal rocks from the westward underthrusting of North America. Coincident timing of southwest-verging structures in the Omineca belt with southwest thrusting of Cache Creek on Stikinia has been used to suggest a link between the southern Omineca belt and Stikinia at this time (Ricketts et al., 1992). The size and types of clasts that were deposited in the Bowser basin record deposition from the west side of the Omineca highland, in high-energy conditions, during rapid exhumation of Cache Creek strata, and they provide further evidence for linkage between these realms. The structurally thickened crust and emerging highland had little effect on the sedimentary record preserved in the Foreland belt over 350 km to the east (Figs. 3, 5A, 6A), suggesting that the loading occurred on weak lithosphere too far to the west to have elastically depressed the thick craton to the east.

Late Middle Jurassic–Bathonian/Callovian (ca. 168–161 Ma)

A major increase in deposition of chert-rich detritus in the northeastern Bowser basin began in the Bathonian, with up to 3000 m of strata deposited in base-of-slope to deltaic environments, and expansion of the extent of Cache Creek chert clasts ~130 km farther southwest than in the Bajocian (I1 in Table 1; Figs. 3 and 6B). Rapid denudation of northern Stikinia or the Cache Creek terrane is indicated by deposition of ca. 161 Ma dacite boulders in the early Callovian slope assemblage of the Bowser basin (Ricketts and Parrish, 1992). The development of fan deltas in the northern part of the basin demonstrates high sedimentation rates (Ricketts and Evenchick, 1991, 2007). Sections of this age elsewhere in the basin are considerably finer grained and thinner and lack the spectacular submarine channel deposits present in the north. This scenario continued into the early Oxfordian, with ~30 km of south and southwest migration of facies boundaries. Southern Bowser basin facies migrated northerly away from a westerly trending arch of Stikinia that defined the south margin of the basin (Tipper and Richards, 1976).

In the Omineca belt, the crust was 50–55 km thick or greater (O3 in Table 1), and it likely formed a broad highland that was internally deforming as it was being translated toward the craton (Table 1; Fig. 6B). Southwest-verging polyphase folding was ongoing in the Cariboo and Monashee Mountains (Reid, 2003; O3b in Fig. 3), but elsewhere, upright folds and northeastward-verging fold systems and faults dominated after ca. 168–167 Ma (Warren, 1997; Colpron et al., 1998; Gibson, 2003; O3a in Fig. 3). The peak of regional metamorphism occurred at ca. 165–160 Ma, although at higher structural levels, it had started to wane (Gerasimoff, 1988; Warren, 1997; Parrish, 1995). Plutons of ca. 167–159 Ma age, within the Nelson Suite, crosscut large-scale belts of folds (e.g., Scrip nappe, early Selkirk fan, Dogtooth structure, and Kootenay arc), indicating that the architecture of these belts had largely formed by the end of the Middle Jurassic (Warren, 1997, and references therein). In the southwest Foreland belt, an 80-m-thick Bathonian section appears to provide the first record of increased subsidence due to tectonic loading (Poulton et al., 1993), and the local presence of cherty quartz-arenites provides the first evidence of a western source area (Stronach, 1984). Elsewhere in the Foreland belt, the first preserved record of west-derived sediments occurs in the Upper Jurassic (Bally et al., 1966; Poulton, 1984; Poulton et al., 1993; Fig. 3). A period of pre-Oxfordian uplift removed much of the Middle Jurassic record in the eastern Foreland belt.

In summary, links across the orogen may be inferred from evidence for continued growth of the Omineca highland, which loaded the lithosphere and affected the sedimentation patterns of adjacent basins as it was translated inboard from distal transitional crust that was significantly thinned during Proterozoic and early Paleozoic rifting, onto thicker, more rigid transitional crust. This is expressed in the northeastern Bowser basin by substantially thicker and more widespread deposits sourced from the Cache Creek terrane, and in the Foreland belt by the first, but limited, westerly derived foreland basin sediments. A time-transgressive Callovian–Oxfordian erosive disconformity developed across the Foreland belt (F0 in Fig. 3) and adjacent craton, marking the northeastward migration of the forebulge (Poulton, 1984). Detrital zircon analyses from Bowser basin strata indicate that in addition to wind-blown ash, the source areas were Triassic to Middle Jurassic in age (McNicoll et al., 2005). These data, combined with paleocurrents, facies distribution, and clast types in the Bowser basin indicate that the Cache Creek, and Stikine, and/or Quesnel terranes (Fig. 4) formed the upper structural levels of at least the western Omineca highland, and that no cratonic North American detritus was being shed to the west (Fig. 6B).

Late Jurassic–Middle Oxfordian/Kimmeridgian (ca. 158–151 Ma)

Starting in the middle Oxfordian, the Bowser basin changed from a relatively narrow northeastern depocenter of base-of-slope to deltaic deposits, to widespread deposition when the shelf–slope break migrated ~200 km west (I2 in Table 1; Figs. 3, 5B, and 6C). The result was accumulation of up to 2 km of shelf deposits in the central basin, and over 4 km of submarine fan deposits in the western basin. Deltas migrated westward to occupy much of the northeastern basin. Facies boundaries in the southern basin also migrated rapidly, and pebbly deposits became widespread there for the first time, accompanied by local volcanic flows in the Oxfordian (Tipper and Richards, 1976). Regionally, the dominant pebble and sand type remained Cache Creek radiolarian chert (e.g., Evenchick and Thorkelson, 2005, and references therein), but at the basin's southern margin, clasts reflect a local source of Stikinia strata from the arch to the south.

In the Omineca belt between 52°N and 53°N, a zone of medium-grade metamorphism that trends approximately north-northwest across the Cariboo, Monashee, and Selkirk Mountains represents the deepest exposed levels of rocks at this latitude. These rocks preserve evidence of penetrative shortening throughout the Late Jurassic and Early Cretaceous accompanied by metamorphism, folding, shearing, and reactivation of structures (Ferguson, 1994; Currie, 1988; Digel et al., 1998; Crowley et al., 2000; Reid, 2003; Ghent and Simony, 2005, and references therein). Deformation progressed northeastward and carried the accreted terranes farther onto the craton (Fig. 6C). South of 52°N, post-tectonic plutonism and cooling indicate rapid exhumation and quenching at higher structural levels in the Late Jurassic (Warren, 1997, and references therein), providing a source for sediments deposited in the Alberta foreland basin. Pronounced flexural subsidence and development of a two-sided foredeep trough in the Alberta foreland basin began in the Oxfordian. Immense quantities of west-derived silty mud were transported into the trough along its western side at the same time as thin, craton-derived, basal, transgressive sandstone was deposited along its eastern side (Poulton, 1984; Stott, 1998; F1 in Fig. 3). Pronounced subsidence continued through the Kimmeridgian, and sands containing detrital chert, stretched quartz, and mica derived from sedimentary or metasedimentary rocks were deposited into the trough from the west (Hamblin and Walker, 1979). These are the first indications of an Omineca belt source for the Alberta foreland basin; however, isotope geochemistry studies of these strata have not detected juvenile material from the accreted terranes (Ross et al., 2005; F2 in Fig. 3; Fig. 6C).

We conclude that links across the orogen in the Oxfordian can be inferred from events within the Omineca belt and significant changes in regions bordering it. The Omineca highland was likely maintained by continued internal structural thickening during northeastward translation of the Omineca belt core relative to the North American craton. To the west, the Bowser basin experienced a dramatic westward migration of facies boundaries, indicating sustained topography in the source areas, and to the east, the Foreland belt experienced its first significant flexural subsidence and deposition. The latter resulted from the elastic response of the first widespread loading of the North American craton by thickened North American supracrustal rocks and the western thickened lithosphere of pericratonic terranes, Quesnellia, Cache Creek, and possibly Stikinia. Detrital zircons in Bowser basin strata indicate that the sources were still within the accreted terranes (Cache Creek, and Stikinia and/or Quesnellia) that were incorporated into the western Omineca highland, whereas sources for Alberta foreland basin deposition included mica and quartz from exhumed metamorphic rocks of continental origin then exposed at the surface of the eastern Omineca highland. Accordingly, the drainage divide in the Omineca highland is shown on Figures 5B and 6C as lying in the eastern part of the carapace of accreted terranes such that sediments were transported from the exhumed terranes west to the Bowser basin, and Alberta foreland basin drainages had access only to the eastern Omineca highland.

Early Early Cretaceous (ca. 145–135 Ma)

In the early Early Cretaceous, deposition in the Bowser basin changed from widespread marine and marginal marine, to nonmarine, including floodplain deposition marginal to deltas, and low-energy fluvial systems (I4 in Table 1; Figs. 3 and 6D). The only widespread strata preserved of this age are in the north-central part of the basin. Other strata deposited in this stage, or prior to the Albian, include the synorogenic coarse clastic deposits in the north-central Bowser basin and nonmarine strata in the southern basin (Evenchick and Thorkelson, 2005; Bassett and Kleinspehn, 1997; V.J. McNicoll, 2006, personal commun.; Smith and Mustard, 2006).

In the Omineca belt, Early Cretaceous zones of penetrative deformation and metamorphism occurred at mid-crustal levels within the existing edifice (Currie, 1988; Digel et al., 1998; Crowley et al., 2000; Reid, 2003; O4–O6 in Fig. 3), while imbrication of the basement and northeastward translation of the belt occurred on shear zones at depth, near the base of the edifice (Fig. 6D). Middle Jurassic plutons, such as the Hobson Lake and Fang plutons, which crosscut early southwest-verging structures, record Early Cretaceous quenching to low temperatures as they were progressively exhumed to higher structural levels. The interplay between exhumation of older structures and renewed metamorphism and penetrative deformation at depth is consistent with a model of progressive shortening of the Omineca belt during northeastward translation.

In the Foreland belt, west-derived sands and conglomerates flooded into the Alberta foreland basin, locally filling the western foreland trough with over 1.5 km of lower Lower Cretaceous sediments (Stott, 1998; Fig. 3). Sedimentation changed from marine to nonmarine in the late Tithonian south of 52°N, and during the late Berriasian to early Valanginian further north (52°N–55°N). In the Berriasian, a west-sourced delta system with rivers carrying volcanic and metamorphic clasts and abundant radiolarian chert entered the basin near 54°N, indicating that the depositional system had sources in the Slide Mountain terrane and metamorphic parts of the Omineca belt (McMechan et al., 2006). Relatively rapid Late Jurassic to Early Cretaceous sedimentation (Valanginian; up to 100 m/m.y.) was followed by pedimentation and conglomeratic sedimentation in the Hauterivian and Barremian (F3 in Fig. 3). There was substantial erosion of lower Lower Cretaceous and Upper Jurassic sediments in the eastern part of the Alberta foreland basin, and the conglomerates were derived solely from sedimentary sources (Gibson, 1985, and references therein; Ross et al., 2005). The westernmost Alberta foreland basin sediments were being deformed and eroded by the end of the Valanginian (Fig. 6D). Near the line of section in Figure 6D, northeast-directed thrusting likely reached the western Foreland belt in the early Early Cretaceous with the initiation of the Malton Gneiss basement slice (O7 in Table 1).

From the material presented here, we conclude that the early Early Cretaceous was a time of significant change across the Cordillera and that events in adjacent regions were kinematically linked within the developing orogen. The basins flanking the Omineca highland became regions of exclusively nonmarine deposition, with >1.5 km of sediment deposited in the western Alberta foreland basin, and probably >2 km of sediment deposited in the Bowser basin. A major middle Early Cretaceous unconformity in the Alberta foreland basin developed within all but the westernmost areas. This period also records the first significant deformation of western Alberta foreland basin deposits, and possibly the initial deformation in the Skeena fold belt. The age of the latter is not narrowly constrained, but the presence of locally derived early to middle Early Cretaceous coarse synorogenic clastic rocks in the north-central Bowser basin suggests that deformation started in this period, or shortly thereafter. Detrital zircons in Early Cretaceous Bowser basin strata have similar source ages as the older strata, indicating that the drainage divide remained in the western Omineca–eastern Intermontane components of the Omineca highland, and detritus of only Cache Creek, Quesnellia, and Stikinia was carried westward (Fig. 5C and 6D).

Albian (ca. 112–100 Ma)

By the mid-Cretaceous, the Insular belt terranes were accreted to western Stikinia, and a major phase of ductile deformation, metamorphism, and magmatism began in the Coast belt (e.g., Monger et al., 1982; Crawford et al., 1987, 2005; Fig. 6E). In the west-central Coast belt, greenschist metamorphism prior to ca. 98 Ma was followed by burial locally to 8 kbar (800 MPa; 30 km), and regional west-directed thrusting started at ca. 100 Ma (e.g., Crawford et al., 1987; Rubin et al., 1990). Farther east in the central Coast belt, crustal thickening by thrust faulting and emplacement of tabular syntectonic plutons resulted in 6 kbar (600 MPa) metamorphism by ca. 90 Ma (Crawford et al., 1987). Rocks in the central Coast Belt, west of the southern Bowser basin, presently overlie 30 km of crust (Morozov et al., 1998; Hammer et al., 2000), suggesting that in the mid-Cretaceous, part of the central Coast belt crust was up to 60 km thick (e.g., Crawford et al., 1987). Ductile east-directed thrusting involving Stikinia on the east side of the central and southern Coast belt began at ca. 90 Ma, about the same time as the major crustal thickening described previously (e.g., Rusmore and Woodsworth, 1994; Rusmore et al., 2000; Crawford et al., 2005).

Shortening of Bowser basin strata in the Skeena fold belt started before the Albian, and there is no record of Albian or younger Bowser basin deposition except at the southern margin of the basin (Evenchick, 1991a; Evenchick and Thorkelson, 2005; Bassett and Kleinspehn, 1997; I5 in Table 1; Fig. 3). Widespread fluvial deposition in the Sustut basin began in the Albian with an eastern (Omineca belt) source of detrital micas and clasts distinctive of the early Paleozoic Cordilleran margin (Eisbacher, 1981), including recycled detrital zircons of Archean, Paleoproterozoic, Mesoproterozoic, and Paleozoic ages initially derived from the North America craton (McNicoll et al., 2005; I5 in Fig. 3; Fig. 6E). Omineca belt sources may be represented in the southern Bowser basin earlier than Albian (Bassett and Kleinspehn, 1997). The depositional overlap of Skeena fold belt structures by basal Sustut Group strata illustrates the extent of contractional deformation, which reached far to the northeast part of the Bowser basin prior to Albian time (Eisbacher, 1981; Evenchick, 1991a). Clasts of chert recycled from the Bowser Lake Group into the Sustut basin in this period suggest that the Skeena fold belt formed highlands southwest of the Sustut basin (5e in Fig. 3; Fig. 5D; Eisbacher, 1981).

In the Omineca belt, the general structural style established in the Early Cretaceous prevailed, whereby zones of penetrative deformation in the mid-crust and movement on deep-seated shear zones accommodated deformation and translation across the belt (O7 in Table 1; Fig. 6E) and transferred shortening across the orogen from the plate margin to the active foreland. Emplacement and stacking of imbricated basement-cored nappes of the Malton Complex occurred in the Albian, as bracketed by the ca. 140–120 Ma isograds south of the complex, which were deflected during emplacement of the complex, and the ca. 110–100 Ma cooling dates for the complex (O7 in Table 1). The ca. 105–90 Ma Cassiar batholith, now located in the northern Omineca belt, was probably emplaced in the southern Omineca belt and translated northward. Geochemistry indicates derivation mainly from melting of continental crust (Driver et al., 2000), which was associated with, and facilitated by, movement on mid-Cretaceous transcurrent faults within a transpressive environment (Gabrielse et al., 2006).

In the Foreland belt, renewed subsidence of the Alberta foreland basin occurred during the Albian. The greatest subsidence was associated with extensional faulting near 56°N above the ancestral Peace River Arch, following a pattern established in the Aptian (Stott, 1993). Several regional transgressive-regressive cycles in the north caused an alternation of marine, coastal, and nonmarine environments, whereas nonmarine deposition and disconformities developed in the south (Fig. 5D; Stott, 1993; Smith, 1994). Rock fragments and heavy minerals in Albian strata indicate a mixed sedimentary, metamorphic, and volcanic/intrusive source between 53°N and 56°N, and a dominant volcanic and intrusive Quesnellian source south of 53°N (McMechan and Thompson, 1993; Ross et al., 2005). Exhumation and cooling of metasedimentary rocks and basement slices above northeast-directed thrust faults occurred in the Albian near 53°N (F4 in Fig. 3; McDonough and Simony, 1988).

We conclude that the Albian was a milestone in tectonic development across the southern Canadian Cordillera, including the Coast belt. A fundamental change in detrital sources for regions west of the Omineca highlands was marked by a flood of clasts derived from deeper levels of the Omineca belt, including the first clasts of exhumed metamorphic rocks that had originally been deposited on the Paleozoic margin of cratonic North America (Fig. 6E). Either exhumation and erosion of the Omineca belt reduced the carapace of accreted terranes on the west side of the highland, or increased surface uplift of the Omineca belt shifted the drainage divide eastward. After an initial flood of clastics into the Sustut basin from the east (Omineca belt), deposition in the basin was focused in a northwest-trending trough, confined between the Omineca highland and highlands of the Skeena fold belt (Fig. 6E). Widespread sedimentation in the Alberta foreland basin driven by flexural and dynamic processes (Beaumont et al., 1993) produced a westward-thickening wedge, locally up to 1.7 km thick, that extended 1000 km eastward onto the craton. In most of the southern Alberta foreland basin, sediments were derived from the Omineca and Foreland highlands. In contrast, south of 51°N, the Albian was the only time when rivers flowed unimpeded from a region underlain by Quesnellia strata in the western Omineca belt into the foreland basin with little input from central Omineca or Foreland belt strata (Leckie and Krystinik, 1995). The structural architecture formed in the Jurassic to Early Cretaceous within the Omineca and eastern Intermontane belts was carried northeastward by deep-seated zones of deformation, and in the Foreland belt deformation occurred mainly on northeast-directed thrust faults (Fig. 6E). The net result of mid-Cretaceous tectonism in the Coast belt was significant crustal thickening by stacking of thick crustal slabs. Involvement of Stikinia in the east-directed ductile thrust system on the east side of the Coast belt is the basis for the interpretation that the high-level structures of the Skeena fold belt, also involving Stikinia, root in the Coast belt (Fig. 6E; Evenchick, 1991a). We speculate that the crustal thickening in the Coast belt may have provided a western source of sediment deposited over deformed Bowser basin strata of the western Skeena fold belt. Strata of this age are not preserved, but apatite fission-track and vitrinite reflectance data suggest that since the latest Cretaceous–early Cenozoic, 4.4–7 km of section has been eroded from the northwest Bowser basin–Skeena fold belt (O'Sullivan et al., 2005).

Late Campanian/Maastrichtian–Early Eocene (ca. 74–50 Ma)

The contractional ductile deformation of the Coast belt that began in the mid-Cretaceous continued into the earliest Cenozoic with emplacement of large volumes of magma and an eastward migration of magmatism; deformation included the development of large recumbent nappes in the core of the Coast belt (younger than ca. 85 Ma), with the result that thick crust was created by latest Cretaceous time (e.g., Crawford et al., 1987, 2005; Fig. 6G). These events occurred in an environment of dextral transpression that lasted into earliest Cenozoic time (Rusmore et al., 2001; Andronicos et al., 2003). Exhumation began during the Cretaceous during contraction (e.g., Crawford et al., 1987). Extension, pluton emplacement, and rapid (2 mm/yr) exhumation followed in the Paleocene and early Eocene (Hollister, 1982; Andronicos et al., 2003, and references therein).

The last major depositional change in the Sustut basin was in the late Campanian, when the relatively low-energy fluvial systems of the Tango Creek Formation were succeeded by high-energy systems that deposited sheets of conglomerate of the Brothers Peak Formation (Eisbacher, 1981; I6 in Table 1; Figs. 3 and 6F). Deformation of uppermost Sustut strata constrains the youngest Skeena fold belt deformation to post-Maastrichtian. Preliminary apatite fission-track thermochronology results (O'Sullivan et al., 2005) and thermal maturity data (Stasiuk et al., 2005) suggest that a few kilometers of strata have been eroded from above present exposures of Sustut and northeastern Bowser strata, and 4.4–7 km has been eroded from northwestern Bowser strata (O'Sullivan et al., in 2005) since the latest Cretaceous–early Cenozoic; some of this "missing" section may have been mid-Cretaceous age, as described earlier, but some, or all, may have been late Maastrichtian to early Eocene age, in part supplied from the west during the rapid exhumation of the Coast belt.

Continued internal deformation and northeastward translation of the Omineca belt (O9 in Table 1; Fig. 6F) is manifested by out-of-sequence structures, such as the pre–ca. 93 Ma Purcell thrust (Archibald et al., 1983; P.S. Simony, 2005, personal commun.), by crystalline thrust nappes, such as the Gwillim Creek shear zone in Valhalla complex, which carried metamorphic rocks inboard and ramped them onto cold basement (Carr and Simony, 2006), and by belts of penetrative deformation within mid-crustal rocks, such as those exposed in the eastern Selkirk Mountains and in the Monashee Mountains north, west, and south of the Monashee Complex (Johnston et al., 2000; Gibson, 2003; Hinchey, 2005; Williams and Jiang, 2005; Brown and Gibson, 2006). Dextral displacement primarily in the Eocene, on faults within and transecting the western Omineca highlands, amounted to ~430 km (Gabrielse et al., 2006) and resulted in the final northward movement of the Bowser-Sustut region and western Omineca highlands (including Cassiar batholith) relative to cratonic North America.

A major change from dominantly marine shale with pulses of westerly derived deltaic sand to dominantly nonmarine coarse clastics occurred during the Santonian in the western Alberta foreland basin south of 51°N (Stott, 1963; Leahy and Lerbekmo, 1995; Payenberg et al., 2002); a similar change occurred at the base of the late Campanian elsewhere in the western foreland basin (Dawson et al., 1994; 4e in Fig. 3). At the eastern margin of the Foreland belt, rapid, nonmarine sedimentation probably continued into the early Eocene, with up to 4 km of late Campanian to Paleocene strata preserved (Stott and Aitken, 1993; F5 in Fig. 3). An additional 2–2.5 km accumulation of Paleocene and Eocene strata, no longer preserved, is inferred for the eastern edge of the Foreland belt, south of 54°N, from coal reflectance data (Nurkowski, 1984; Kalkreuth and McMechan, 1996). Most Santonian to Paleocene sediments were derived from the Foreland highlands and from volcanic airfall (Ross et al., 2005). Two exceptions occur near 53°N (Fig. 3), where local conglomerates, one late Maastrichtian and the other middle Paleocene, contain a few andesitic pebbles (Jerzykiewicz, 1985), which indicate that the drainage divide locally extended into the Intermontane belt. The leading edge of the thrust system progressed eastward from the central Foreland belt to its eastern edge during the late Campanian to early Eocene (Fig. 6G), doubling its width and resulting in exhumation, erosion, and cannibalization of the Alberta foreland basin wedge (Price and Mountjoy, 1970). Motion on major thrust faults in the Eastern Front Ranges (e.g., Lewis, McConnell) deformed late Campanian strata, and major thrust faults in the Foothills (e.g., Bighorn, Brazeau) deformed Paleocene strata (Price, 1981; McMechan and Thompson, 1993). Alberta foreland basin subsidence ended with the cessation of contractional deformation in the early Eocene (Fig. 6G).

The late Campanian to earliest Eocene was the last period of trans-Cordilleran horizontal shortening and sedimentation that may be attributed to kinematic links across the orogen. It was characterized by northeastward translation of the Omineca belt on mid-crustal structures at the same time as significant nonmarine deposition in basins on the east and west sides of the Omineca highland, thickening and horizontal shortening at high crustal levels in the Skeena fold belt and Foreland belt, and crustal thickening and eastward migration of magmatism in the Coast belt. The start of this period marks one of the fundamental changes in sedimentation in the basins flanking the Omineca highlands. In the Sustut basin, the change was from low- to high-energy fluvial deposition, and in the Alberta foreland basin, it changed from deltaic marine pulses to entirely nonmarine deposition. Confinement of Sustut Group deposition to a linear northwest-trending trough with southeast paleocurrents suggests that the Skeena fold belt and Omineca highland continued to form topographic barriers bounding the Sustut basin. During the late Campanian to early Maastrichtian, ~1.5 km of strata accumulated in the Sustut basin, while up to 2 km accumulated in the Alberta foreland basin. Deposition in the Alberta foreland basin continued into the early Eocene, forming a late Campanian to Eocene westward-thickening wedge over 6 km thick at the eastern margin of the Foreland belt. Deposition of strata, no longer preserved, across the Sustut basin and/or a broader and younger basin and originating in part from exhumation of the Coast belt, may have continued into the latest Cretaceous or earliest Cenozoic.

LINKS ACROSS THE OROGEN AND THEIR SIGNIFICANCE

The previous section demonstrated the kinematic and dynamic links across the orogen through time, illustrated by Figure 3 and tectonic interpretations in Figure 6. To summarize:

(1) Middle Jurassic obduction of Quesnellia and Slide Mountain terranes onto pericratonic terranes and distal North America was followed closely by southwest obduction of Cache Creek onto Stikinia, which resulted in initiation of flexural subsidence in the Bowser basin and deposition of coarse Cache Creek detritus in the Bowser basin (I0, I1, O1, O2 in Table 1);

(2) Ongoing deformation, crustal thickening, and/or exhumation in the Omineca belt from the Late Jurassic to Paleocene occurred while sediment was shed from the Omineca highland eastward into the Alberta foreland basin, and westward into the Bowser and Sustut basins (I2–I6, O3–O9, F1–F5 in Table 1; expanded on herein);

(3) Mid-Cretaceous initiation of thin-skinned shortening in the Foreland belt, major crustal thickening in the Coast belt, thin-skinned shortening in the Skeena fold belt, initiation of the Sustut basin, and continued eastward translation of the exhuming core of the Omineca belt above a basal detachment system (I5, O6, O7, O8, F4 in Table 1; Crawford et al., 1987)—in summary, shortening across the width of the Cordillera—was accommodated at different structural levels;

(4) Approximately synchronous pulses of synorogenic coarse clastics were deposited in the Sustut basin and Alberta foreland basin in the late Campanian–early Maastrichtian, during a period of major structural thickening and denudation (I6, O9, F5 in Table 1); and

(5) Horizontal shortening across all belts lasted into the latest Cretaceous to early Cenozoic (I6, O9, O10, F5 in Table 1; e.g., Crawford et al., 1987).

Such links are permissible, despite uncertainties in timing and magnitude of superimposed transcurrent faults, because fundamental aspects of each belt have considerable strike length. Although the ~300-km-long Bowser basin is the most restricted tectonic element discussed, it was the major center of sedimentation west of the evolving orogen in the Middle and Late Jurassic, and it was probably localized adjacent to the region of maximum crustal thickening in the Omineca belt.

From these examples of synchronous and compatible tectonic events in adjacent belts, we suggest that the Intermontane, Omineca, and Foreland belts were kinematically connected from the Middle Jurassic to early Cenozoic, and that the Insular and Coast belts were included in this kinematic connection from mid-Cretaceous to early Cenozoic. Both the Bowser and Alberta Foreland basins received sediment from a persistent source in the Omineca highland, which was composed of the eastern accreted terranes, pericratonic terranes, and supracrustal rocks of western North American. This source lasted for ~120 m.y., from earliest Bowser basin deposition to the final Alberta foreland basin deposition. Throughout the evolution of the orogen, changes in source rocks for each basin reflect changes in the level of exhumation in the intervening highland, which progressively exposed deeper rocks, and migration of the drainage divide. The relationship between basin evolution and lithospheric loading for the Foreland and Omineca belts has been well established (e.g., Beaumont et al., 1993). This relationship is less direct for the Bowser basin because much of northern Stikinia was already the site of marine deposition, although limited in thickness, prior to the Bajocian/Bathonian. Subsidence was caused, at least in part, by cooling of Stikinia, which probably continued through at least the early phases of coarse Bowser basin deposition. Facies relationships indicate that the northeastern part of the basin was the primary site of deposition during the late Middle and early Late Jurassic, presumably in response to lithospheric loading by the growing accretionary orogen to the east and by accumulating sediment. The large scale of both the Foreland and Bowser basins and widespread contractional structures are incompatible with the alternative explanation of basin formation in a transtensional regime. Therefore, the basins are most likely related, in differing degrees, to subsidence due to flexure and progressive loading of the lithosphere by the thickened crust between them and by deposition of sediments from the same persistent source, the elevation of which was maintained by the processes that resulted in continued exhumation of the Omineca belt.

Shortening of the orogen and inferred linkages across the orogen were likely accommodated by a lower-crustal detachment as illustrated in Figure 6. Support for this interpretation is provided by interpretation of SNORCLE line 2a, which displays a low-angle boundary between Stikinia and North American rocks that rises eastward from the lower crust to middle crust (Fig. 6H; Cook et al., 2004; Evenchick et al., 2005). This boundary may be part of the lower-crustal detachment illustrated in Figure 6. The deformation and structural architecture of mid- and upper-crustal rocks in the western Foreland and eastern Omineca belts were established in the Middle Jurassic to Early Cretaceous (Carr and Simony, 2006, and references therein); therefore, linkage of mid-Cretaceous to Eocene structures in the Foreland and Omineca belts must have occurred via a detachment that passed beneath the western Foreland belt. This interpretation is consistent with geophysical data that are, in part, controlled by outcrop and drill-hole information (Cook et al., 1988, 1992; Cook and van der Velden, 1995).

We suggest that the southern Canadian Cordillera evolved from a relatively narrow, doubly vergent, “small-cold” orogen in the Jurassic with the core centered in the Omineca belt (Figs. 6A–6D) to a much broader doubly vergent, “large-hot” orogen in the mid-Cretaceous (terminology of Koons, 1990; Beaumont et al., 2006). The mid- and Late Cretaceous orogen (Figs. 6E–6G) may be viewed as a wide doubly vergent orogen with the predominantly east-directed structures in the Omineca and Foreland belts on the eastern, retro-wedge side (terminology of Willett et al., 1993), and west-directed structures of the accretionary orogen of the Coast belt on the pro-wedge side, at the active oblique subduction margin, all linked by a basal detachment. At this time, the Jurassic core of the orogen was mainly translated eastward and exhumed as part of the retro-wedge, which included the active thrust front in the eastern Foreland belt. However, within this first-order, large-scale geometry, there was a detachment beneath the east-directed upper-crustal Skeena fold belt, which soled westward into mid- and lower-crustal ductile structures in the eastern Coast belt. Thus, there were two coeval cratonward-verging upper-crust fold-and-thrust belts, equally large in cross section, at the same latitude in the Cordilleran orogen: the Rocky Mountain fold-and-thrust belt and the Skeena fold belt.

The Mesozoic history of Stikinia as an arc and (or) back arc may have facilitated development of the lower-crust detachment. Hyndman et al. (2005) argued that back arcs or recent back arcs are hot as a result of transfer of convective heat below thin lithosphere, and that the high temperature results in weak lower crust, facilitating development of lower-crust detachments, which separate crustal elements above from underlying lithosphere in the manner of “orogenic float” (Oldow et al., 1990). The decay in temperature after the source is removed is slow enough that former back-arc regions may remain weak long after cessation of arc activity (Hyndman et al., 2005). From Hyndman et al.’s (2005) calculations of temperature decay, and Monger and Nokleberg’s (1996) review of the evolution of arc development in the Cordillera, Stikinia probably remained relatively hot for the period discussed herein. To illustrate their model, Hyndman et al. (2005) explained the relationship between the modern collision of the Yakutat block in the Gulf of Alaska and shortening in the Mackenzie Mountains at the front of the thrust belt as being facilitated by a detachment in a weak lower crust. We suggest that the connectivity of tectonic elements across the Cordillera in the Cretaceous is an ancient example of this phenomenon.

Recognition of a deep detachment across the orogen may provide a broader context for understanding structural relationships. For example, Rusmore et al. (2001) posed the problem of accommodation in the Coast belt of the large horizontal displacement inferred from plate motions. They concluded that reverse motion on the Paleocene Coast shear zone represents the orthogonal component of oblique convergence, but we question the magnitude of shortening that may be accommodated by this structure, and instead suggest that a lower-crust detachment transferred a component of shortening eastward. This is also an effective way to accommodate the regional transpression inferred by Rusmore et al. (2001) who associated dextral faults in the Coast belt with dextral faults in the western Omineca belt–Northern Rocky Mountain Trench.

A question that arises from consideration of the Cordillera as one kinematically connected orogen is the perplexing thickness of Stikinia, which is interpreted from seismic-reflection data as being ~35 km thick (e.g., Cook et al., 2004). Was this Paleozoic–early Mesozoic arc terrane, possibly built partly on rifted fragments of continental margin, always thick, or was it substantially thinner prior to its accretion to North America and then thickened during Cretaceous Cordilleran-wide contraction?

In this interpretation, the mid-orogen Skeena fold belt was carried piggyback above a lower-crust detachment connected to the Rocky Mountain fold-and-thrust belt. This scenario is unusual in modern or ancient orogens. A factor that may have contributed to the geometry is the mechanical effect of the stratigraphy within the Intermontane belt. The Bowser succession, formed just prior to deformation, is a thinly bedded succession with substantial mechanical heterogeneity, and therefore it was relatively weak. In contrast, Stikinia is composed of units of limited lateral continuity, such as volcanic edifices and surrounding sedimentary units with rapid facies changes, and associated plutons. Pre-Triassic strata underwent at least two phases of deformation and local low-grade metamorphism, and Early Jurassic intrusions pierce all sub-Bowser stratigraphy. Compared to the Bowser succession, Stikinia contains few laterally continuous horizontal weak layers. An exception is the thinly layered early Middle Jurassic clastic succession, immediately below the Bowser Lake Group, which, along with underlying layered Early Jurassic volcanic successions, was intimately involved in thin-skinned deformation. We infer that the relative strength of Stikinia, combined with a relatively weak lower crust, as discussed previously, localized a deep detachment, whereas the mechanical heterogeneity of the Bowser succession and immediately underlying strata of Stikinia facilitated an upper-crustal detachment leading to the Skeena fold belt in uppermost Stikinia, Bowser, and Sustut strata.

CONCLUSIONS

Synthesis of the Jurassic and Cretaceous depositional and tectonic histories of the central Intermontane belt, and the southern Omineca and Foreland belts, when considered in paleogeographic context, reveals sedimentation and structural linkages across the orogen and highlights the tectonic interplay among crustal thickening, basin formation, and topographic evolution. We conclude that coeval and tectonically compatible events in regions considered to have been in close proximity require kinematic linkage across the entire orogen. In our model for the mid- and Late Cretaceous, a detachment in the lower crust extended from the active plate boundary and Coast belt eastward below the Intermontane belt and then rose into the middle crust in the Omineca belt and, ultimately, to the upper crust at the front of the orogen in the Rocky Mountain fold-and-thrust belt. It was also connected to dextral strike-slip faults, facilitating regional transpression partitioned across the orogen in response to oblique plate convergence. In the Intermontane belt, this lower-crust detachment carried Stikinia as well as the Skeena fold belt, an upper-crust and craton-verging fold-and-thrust belt nested in the interior of the orogen. The basal detachment of the Skeena fold belt rooted in ductile structures on the east side of the Coast belt. Development of this nested fold belt was in part a consequence of the mechanical stratigraphy of Stikinia and overlying basins. Consideration of the orogen as a whole, in paleogeographic context, should lead to improved tectonic models of Cordilleran evolution.

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The Cordilleran Ribbon Continent of North America

Stephen T. Johnston

School of Earth & Ocean Sciences, University of Victoria, Victoria, British Columbia, Canada V8W 3P6; email: stj@uvic.ca

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Laramide orogeny, orocline, accretionary orogenesis, plate tectonics, paleomagnetism, Wilson cycle

Abstract

The North American Cordilleran Orogen is the result of a two-stage process: (*a*) Triassic-Jurassic accretion within Panthalassa forming SAYBIA, a composite ribbon continent, and (*b*) Late Cretaceous collision of SAYBIA with North America. This model requires that a large portion of the continental foreland of the orogen is exotic. The exotic continental component of SAYBIA, Cassiar Platform, is distinguished from the autochthon on the basis of its (*a*) Triassic Eurasian fauna; (*b*) involvement in a major Late Triassic-Early Jurassic orogenic event; and (*c*) young, in part Grenvillian basement and mantle. A mid-Cretaceous magmatic arc records west-dipping subduction beneath the east-margin of SAYBIA. The related accretionary prism consists of imbricated shale, chert, and deep-water limestones (the Medial Basin) and overlies an isotopically juvenile mantle domain. Carbonatite complexes delineate the cryptic suture separating SAYBIA and the autochthon. Paleomagnetic and paleobotanical data place SAYBIA 2000 km to the south relative to the autochthon at 80 Ma. Late Cretaceous thrust belt development records transpression between the north-moving ribbon continent and the autochthon. Pinning against the Okhotsk-Chukotka arc in Siberia buckled SAYBIA, giving rise to the Alaskan promontory.

INTRODUCTION

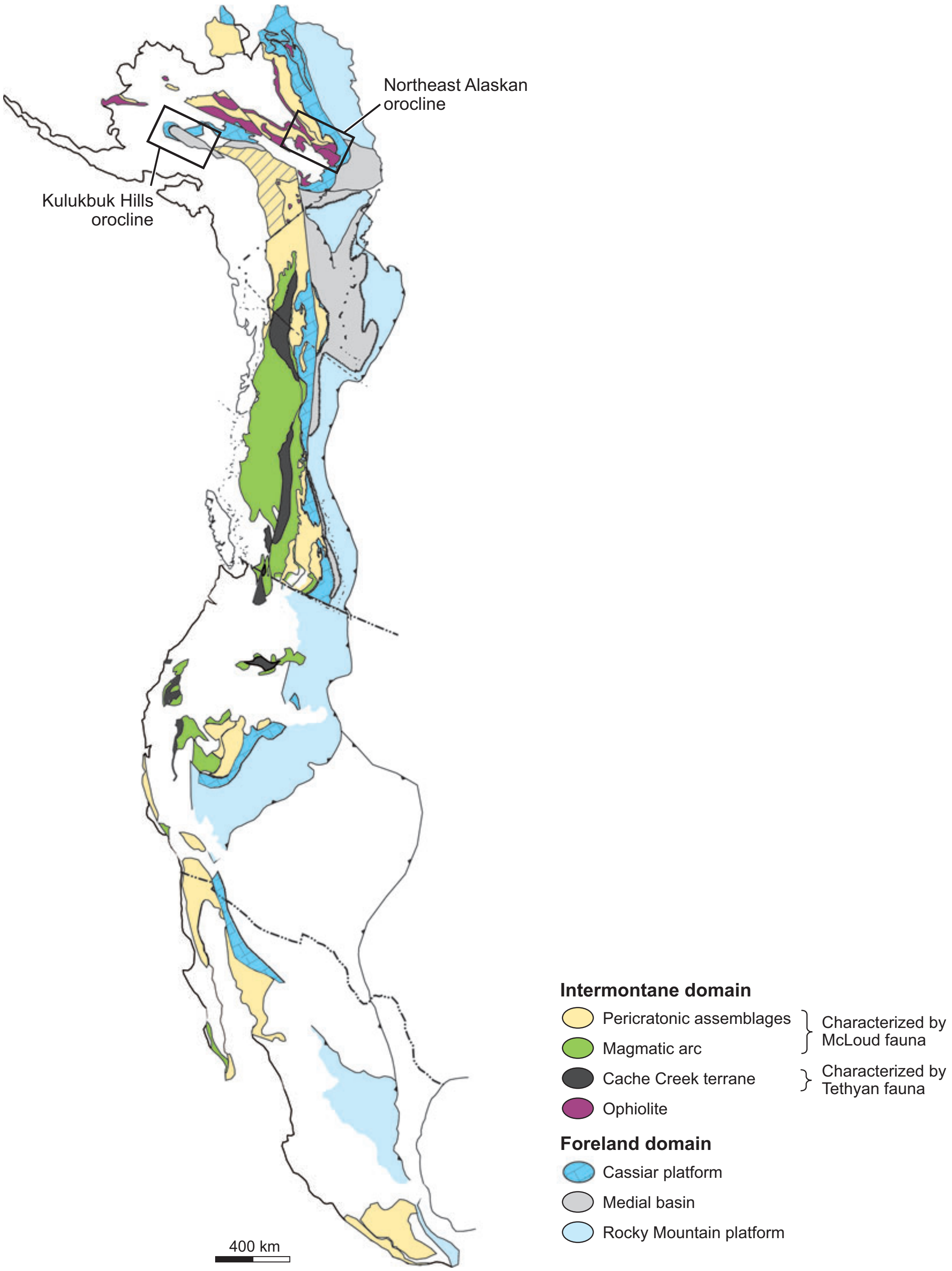
Between the autochthonous, undeformed strata of the plains to the east and the active convergent margin to the west lies the north-south trending Cordilleran Orogen of western North America (**Figure 1**). Within the confines of the orogen lies a boundary between deformed crust of North American affinity (para-autochthonous strata) and accreted, exotic crust. Determining the location, geometry, nature, and evolution of the boundary between exotic and para-autochthonous North American crust has been the subject of intense debate (Cook & Erdmer 2005, Johnston 2001), and is the focus of this paper.

Understanding the boundary between North American and accreted crust is fundamental to understanding the processes responsible for orogenesis and the growth of continents. For example, orogens recording a complete Wilson cycle, including a terminal continental collision, are commonly interpreted to result in significant continental growth (e.g. Bird & Dewey 1970, Hoffman 1980). The Cordilleran Orogen of western North American, however, is thought to represent an incomplete Wilson cycle in that it appears to have developed in the absence of a terminal continental collision. Instead, the Cordillera is interpreted as an accretionary orogen, and its evolution is explained as the result of the incremental, thin-skinned addition of terranes to the continental margin above a landward-dipping subduction zone (Monger 1997). Continental growth is not a requirement of accretionary orogenesis, and subduction erosion of the continent may even result in a net loss of continental mass.

If the Cordillera is, therefore, strictly attributable to accretionary processes, the bulk of the orogen may consist of little-disturbed North American crust underpinned by North American mantle (Cook et al. 2004, Snyder et al. 2002). I start by reviewing the basic character of the Canadian portion of the orogen, and assess a primary assumption in Cordilleran studies: that all continental assemblages are of North American affinity. I demonstrate that a large portion of the continental foreland of the orogen is exotic with respect to the autochthon and forms part of a composite ribbon continent, previously referred to as SAYBIA (Johnston 2001), which extends along strike to the northwest into Alaska and south into the conterminous United States of America. I finish by presenting a model of Cordilleran orogenesis as a product of a two stage process: (*a*) the accretionary construction of a composite ribbon continent, SAYBIA, followed by (*b*) collision of SAYBIA with North America. This model marks a return to a more Wilson cycle–style interpretation of the Cordillera, as it involves a continental collision and implies that North America has grown significantly westward during orogenesis. A key requirement of this model is the presence of a cryptic suture within the orogenic foreland, a region that has been extensively mapped

Figure 1

The Cordilleran Orogen of western North America. Yellow striped region in Alaska, shown here as a portion of the pericratonic belt (*yellow*), has recently been reinterpreted as being part of the Medial shale basin of the Foreland belt (Dusel-Bacon et al. 2006).


Northeast Alaskan orocline
Kulukbuk Hills orocline
400 km
Intermontane domain
Pericratonic assemblages
Magmatic arc
Cache Creek terrane
Ophiolite
Characterized by McLoud fauna
Characterized by Tethyan fauna
Foreland domain
Cassiar platform
Medial basin
Rocky Mountain platform

and studied. Determining the nature and location of the cryptic suture constitutes the primary test of this model.

GENERAL GEOLOGY

In the most general of terms, the Cordilleran Orogen is divisible into an eastern foreland domain characterized by sedimentary strata of continental affinity, a central intermontane domain consisting of oceanic assemblages, and a western insular domain of mixed oceanic and continental assemblages (**Figure 2**). The boundary region between the foreland and intermontane domains is referred to as the Omineca, the diagnostic component of which is the Omineca magmatic belt (OMB), and is commonly considered to consist largely of crust and mantle that extends west from and is a continuation of foreland domain crust and mantle.

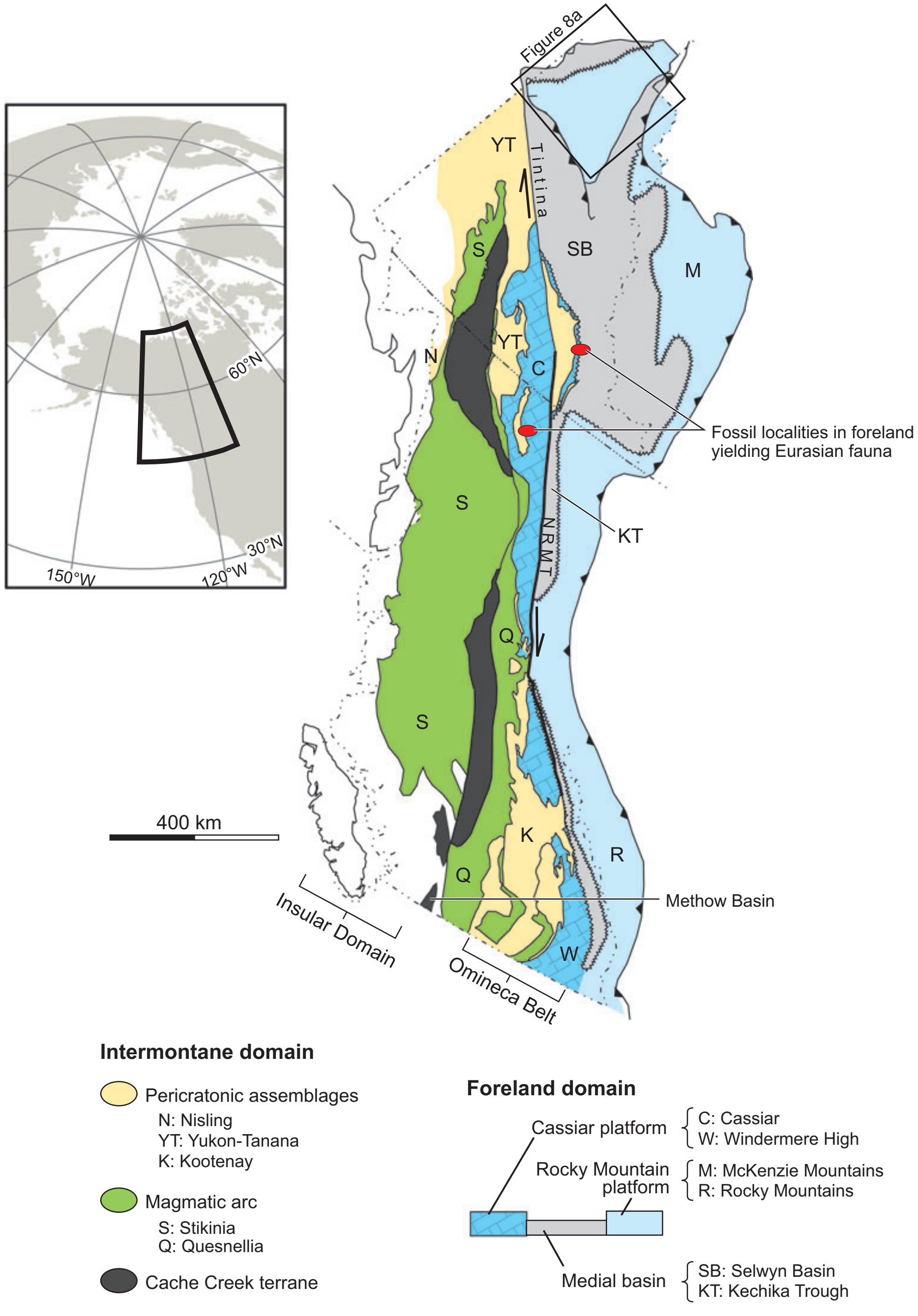
Crust of the insular domain, to the west of the intermontane domain, is exotic (Nokleberg et al. 2005) and was added to the North American margin sometime between Lower Jurassic and Lower Cretaceous time (McClelland & Mattinson 2000). I therefore focus on the eastern portion of the orogen, that region straddled by the foreland and intermontane domains, for it is within this region that the boundary between the ancient west margin of North America and accreted crust is located. Geological relationships limit accretion to having occurred between the Triassic and Upper Cretaceous. Hence, pre-Triassic tectonism and post-Cretaceous extension and magmatism in the southern Canadian Cordillera are little discussed.

Thorough reviews of the geology of the Canadian Cordillera are provided elsewhere and as a **Supplemental Appendix** (follow the **Supplemental Material** link from the Annual Reviews home page at **http://www.annualreviews.org**). The three salient points for this discussion are as follows:

1. Paleozoic to Middle Jurassic strata of the foreland domain are divisible into an easterly shallow water continental platform (the Rocky Mountain Platform); a medial basinal domain of shale, chert, and deep-water limestone (the Medial Basin); and a westerly shallow water platform (Cassiar Platform).
2. The intermontane domain is characterized by a mid-Paleozoic to mid-Mesozoic arc (Stikinia-Quesnellia) and a related accretionary complex (Cache Creek terrane) that includes offscraped seamounts that originated in the Tethyan domain.

Figure 2

Geological map of the Canadian Cordillera showing divisions of the foreland and intermontane domains, location shown in inset at upper left. The jagged line indicates a mapped facies boundary separating shallow water platformal sequences (M, McKenzie Mountains; R, Rocky Mountains; C, Cassiar; W, Windermere High) from basinal strata (SB, Selwyn Basin; K, Kechika Trough). The Tintina–Northern Rocky Mountain Trench (NRMT) fault is the locus of >400 km of Eocene dextral displacement (Gabrielse et al. 2006).



60°N
30°N
150°W
120°W
Figure 8a
YT
Tintina
SB
M
S
N
YT
C
Fossil localities in foreland
yielding Eurasian fauna
S
KT
NRMT
Q
S
400 km
K
R
Q
Methow Basin
W
Insular Domain
Omineca Belt
Intermontane domain
Pericratonic assemblages
N: Nisling
YT: Yukon-Tanana
K: Kootenay
Magmatic arc
S: Stikinia
Q: Quesnellia
Cache Creek terrane
Foreland domain
Cassiar platform
C: Cassiar
W: Windermere High
Rocky Mountain platform
M: McKenzie Mountains
R: Rocky Mountains
Medial basin
SB: Selwyn Basin
KT: Kechika Trough

3. Pericratonic assemblages and structurally interleaved ophiolite separate the intermontane and foreland domains, and provide a record of Paleozoic Andean-type arc and related marginal basins.

CASSIAR PLATFORM: EXOTIC?

Despite the foreland position of the Cassiar Platform, a number of lines of evidence imply an exotic non–North American origin for the shallow-water continental platform. These include faunal and geological data inconsistent with autochthoneity, and paleomagnetic and paleobotanical data that require significant mobility of the platform relative to the autochthon into the Late Cretaceous.

FAUNAL PROVINCIALITY OF CASSIAR PLATFORM AND THE INTERMONTANE DOMAIN

In Yukon, Upper Triassic strata from near the Cassiar Platform–Medial Basin boundary includes *Epigondolella* and *Paragondolella*—conodont species that are Eurasian. In North America, these species are only known from the exotic Wrangellia terrane of the insular domain (Orchard 2006). Eurasian fauna similarly characterizes the intermontane domain. Permian strata of Stikinia-Quesnellia are characterized by schwagerinid fusulinids and additional fauna that are similar to those found in the McCloud Limestone of northern California and Nevada (Carter et al. 1992) (**Figures 1** and **2**). The fauna of the McCloud Belt terranes (Miller 1987) is distinct from the fauna of coeval North American strata. The degree of separation required to produce this faunal provincialism is assumed to be >1000 km and may be much greater (Stevens et al. 1990). A more distal origin for Stikinia-Quesnellia, consistent with the constraints provided by the Tethyan Cache Creek seamounts (Johnston & Borel 2007), is suggested by Devonian and Lower Carboniferous strata that are characterized by conodonts of exclusively Eurasian derivation (Orchard 2000), and by Permian and Triassic strata that, although characterized by a mixed faunal assemblage, includes corals, conodonts, and radiolarian that are otherwise unknown outside of Eurasia (Reid & Tempelman-Kluit 1987, Stanley & Senowbari-Daryan 1999). The slices of ophiolitic crust tectonically interleaved with the pericratonic assemblages are similarly characterized by Permian McCloud fauna, and Triassic mixed fauna, in part, of Tethyan affinity (Dusel-Bacon & Harris 2003, Nelson 1993). Hence, Eurasian fauna characterizes Devonian through Triassic strata of the intermontane domain, as well as the Cassiar Platform, distinguishing them from coeval North American strata of the Rocky Mountain Platform.

CONTRASTING TRIASSIC–JURASSIC GEOLOGICAL EVOLUTION

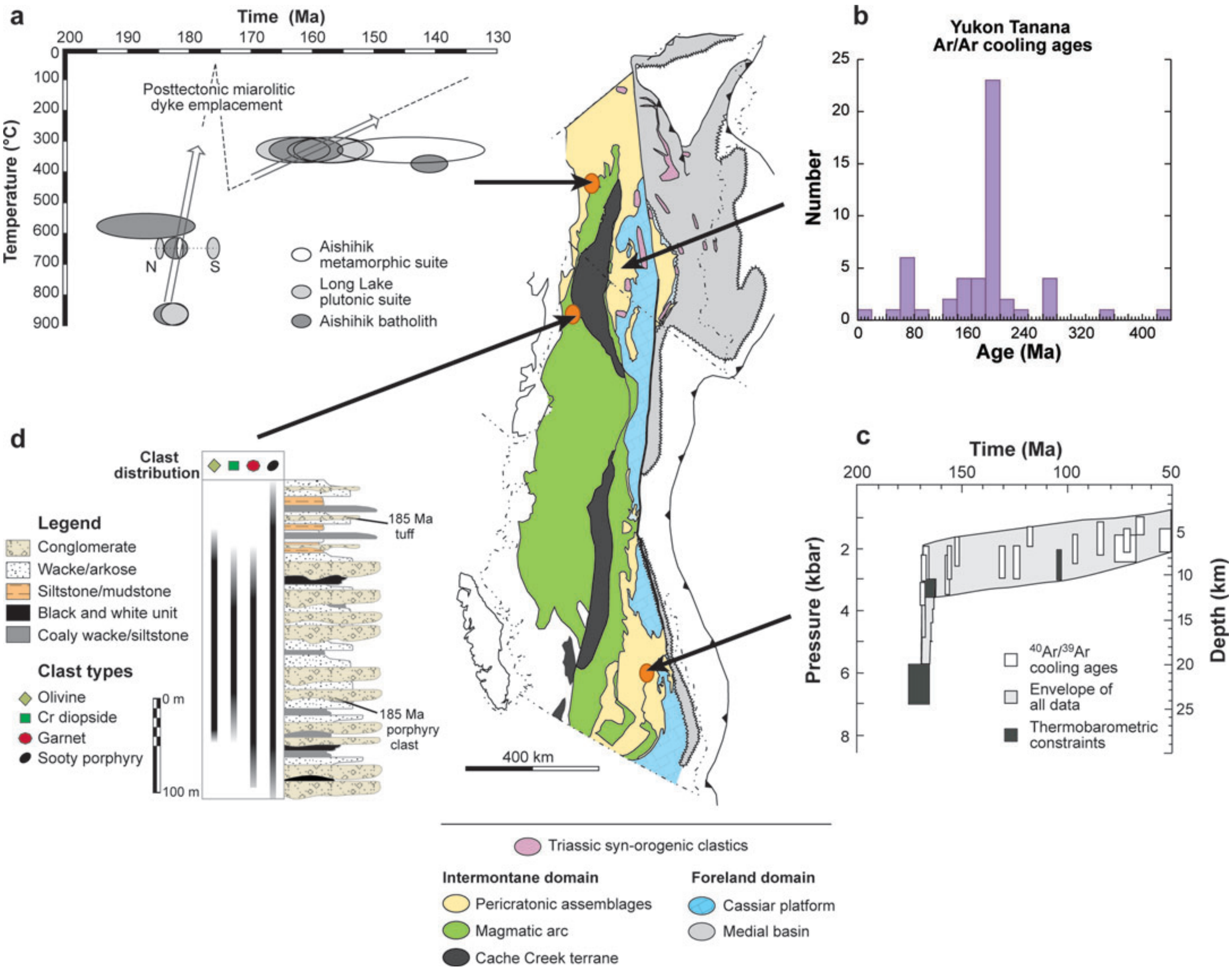
Stikinia-Quesnellia and the pericratonic assemblages were involved in a major Late Triassic collisional event that was not recorded on the autochthon. In the southern

Cordillera, rocks of Quesnellia overthrust the pericratonic assemblages along a major east-verging thrust fault that is plugged by Early Jurassic intrusions (Murphy et al. 1995). Rapid uplift and exhumation of deeply buried pericratonic strata is recorded by Early to Middle Jurassic cooling ages on continental margin assemblages that had been metamorphosed at pressures of greater than 7 kbar (Colpron et al. 1996) (**Figure 3**). In Yukon, Stikinia tectonically overlies the pericratonic assemblages (Johnston & Canil 2007). Early Jurassic posttectonic plutons and shallow-level miarolitic dyke swarms intrude and stitch together Stikinia and pericratonic assemblages (Johnston et al. 1996a) that had been previously metamorphosed at pressures of 8 to 12 kbars (Johnston & Erdmer 1995) (**Figure 3**). Collision is recorded in Stikinia-Quesnellia by the Late Triassic termination of arc magmatism and Early Jurassic molasse deposition.

Collisional orogenesis was thick-skinned, involving exhumation of lowermost mantle lithosphere. Pleinsbachian (~185 Ma) molasse shed off the collision zone includes clasts of ultrahigh-pressure garnet peridotite and eclogite unroofed from depths of 100 to 150 km (MacKenzie et al. 2005) (**Figure 3**). Micaceaous Triassic flysch was deposited across the pericratonic assemblages and their tectonically interleaved slices of ophiolite, the Cassiar Platform, and parts of the Medial Basin. These Triassic strata were derived from the west, are locally conglomeratic, and are interpreted as syn-orogenic sediments (Colpron et al. 2006; Colpron et al. 2007) (**Figure 3**). Detrital zircons of demonstrable pericratonic assemblage origin characterize Triassic strata overlying Medial Basin strata, consistent with interpretation as an overlap assemblage (Beranek & Mortensen 2006, Beranek & Mortensen 2007).

In contrast, passive margin sedimentation continued to characterize the Rocky Mountain Platform through at least the Middle Jurassic, in the south, and until the Cretaceous in the north (Gordey et al. 1992) (see **Supplemental Figure 1**). Phosphorite and, in deeper water strata, chert accumulation imply that the passive margin faced west toward an open ocean basin characterized by upwelling of large-scale deep water currents (Poulton 1984, Poulton & Aitken 1989). Westerly derived flysch and molasse did not inundate the southern Rocky Mountain Platform until the Upper Jurassic (155–152 Ma), and even then, most of the siliciclastic sediment appears to have been derived from autochthonous source terranes (Ross et al. 2005). Clastic sediments derived from erosion of isotopically juvenile source terranes, such as the oceanic arcs and ophiolite of the intermontane domain, do not appear in the foreland basin until 120 Ma (Ross et al. 2005).

The thick-skinned Late Triassic orogeny that involved the intermontane domain terranes, the Cassiar Platform, and the Medial Basin did not load and cause isostatic flexure of the lithosphere on which the Rocky Mountain Platform was located. Even in the southern Canadian Cordillera, where the orogenic welt is closest to the Rocky Mountain Platform (only 200 km to the west after palinspastic restoration of younger thrust faults), there is no evidence of any Triassic–Early Jurassic loading of the North American lithosphere and no orogenic sediments shed east off of the thickened crustal welt that characterized the orogen. Neither did the orogen impede or inhibit continued oceanic upwelling and related phosphatic sediment deposition along the North


a
Time (Ma)
Temperature (°C)
Posttectonic miarolitic dyke emplacement
N
S
Aishihik metamorphic suite
Long Lake plutonic suite
Aishihik batholith
b
Yukon Tanana Ar/Ar cooling ages
Number
Age (Ma)
c
Time (Ma)
Pressure (kbar)
Depth (km)
$^{40}Ar/^{39}Ar$ cooling ages
Envelope of all data
Thermobarometric constraints
d
Clast distribution
185 Ma tuff
185 Ma porphyry clast
Legend
Conglomerate
Wacke/arkose
Siltstone/mudstone
Black and white unit
Coaly wacke/siltstone
Clast types
Olivine
Cr diopside
Garnet
Sooty porphyry
0 m
100 m
400 km
Triassic syn-orogenic clastics
Intermontane domain
Pericratonic assemblages
Magmatic arc
Cache Creek terrane
Foreland domain
Cassiar platform
Medial basin

American passive margin. The implication is that the Triassic orogeny involving the intermontane domain terranes and the Cassiar Platform took place far removed from the Rocky Mountain Platform and involved plates separate from the North American plate.

CRUSTAL BASEMENT AND MANTLE PROVINCIALITY

The basement to the Rocky Mountain Platform consists of 1.84 Ga and older crust and mantle of the Canadian Shield (Ross 2000). This contrasts with the basement underpinning the Medial Basin, the Cassiar Platform, and the intermontane domains, which is younger. In the northern Cordillera, the Coates Lake Diatreme intrudes Proterozoic sedimentary rocks near the mapped eastern boundary of the Medial Basin (**Figure 4**). Lower crustal granitic xenoliths in the diatreme yield crystallization ages of 1.1 Ga (Jefferson & Parrish 1989, Mortensen & Colpron 1998). 1.0 to 1.1 Ga xenocrystic zircons characterize Paleozoic diatremes that intrude in or near the east margin of the Medial Basin in the southern Cordillera (Parrish & Reichenbach 1991) (**Figure 4**). Metabasite xenoliths in intrusive breccias in the Wernecke Mountains (Yukon) were recrystallized at 1.15 Ga (Milidragovic et al. 2007). 1.0 to 1.2 Ga crystalline rocks are unknown in the cratonic basement to the east and indicate that the depositional basement to Medial Basin and Cassiar Platform is a distinct, in part, Grenvillian-aged basement.

Precambrian crystalline rocks crop out within the pericratonic assemblages in British Columbia, including, from south to north, the Priest River Complex (Idaho), the Monashee Complex, the Malton Complex, and the Sifton Range (**Figure 4**). The basement complexes are exposed within structural culminations and have been interpreted as para-autochtonous extensions of the cratonic North American basement to the east (Parrish 1992). However, the Priest River Complex, which lies west of the 3.3 Ga to 2.6 Ga Medicine Hat province of the craton, crystallized at 2.55 to 2.65 Ga, is intruded by 1.59 Ga felsic plutons and cannot be readily correlated with any cratonic basement to the east (Doughty et al. 1998) (**Figure 4**). Despite the abundance of Archean crust and mantle abutting the east margin of the Canadian Cordillera, no Archean crust has yet been documented within the orogen. Granite and felsic cobbles in a conglomerate in southern Quesnellia yield crystallization ages of 1.03 to 1.04 Ga, and were likely derived from erosion of exposed basement (Erdmer et al.

Figure 3

Map of the Cordillera showing region affected by Triassic orogeny. Only the Rocky Mountain Platform (not colored here) was unaffected. Examples of constraints on timing of crustal thickening and subsequent exhumation and cooling include (*a*) cooling curve from Nisling pericratonic assemblage showing Early Jurassic cooling and exhumation (Johnston et al. 1996a), (*b*) Ar-Ar cooling ages for Yukon-Tanana terrane that peak at 190 Ma (Breitsprecher & Mortensen 2004); (*c*) exhumation curve for Kootenay pericratonic assemblage showing unroofing by 180 Ma (Colpron et al. 1996); and (*d*) the stratigraphic record from Stikinia of unroofing of ultra-high-pressure rocks at 185 Ma (Canil et al. 2006). Also indicated is the distribution of westerly derived, Triassic, syn-orogenic clastic sequences (Murphy et al. 2006).

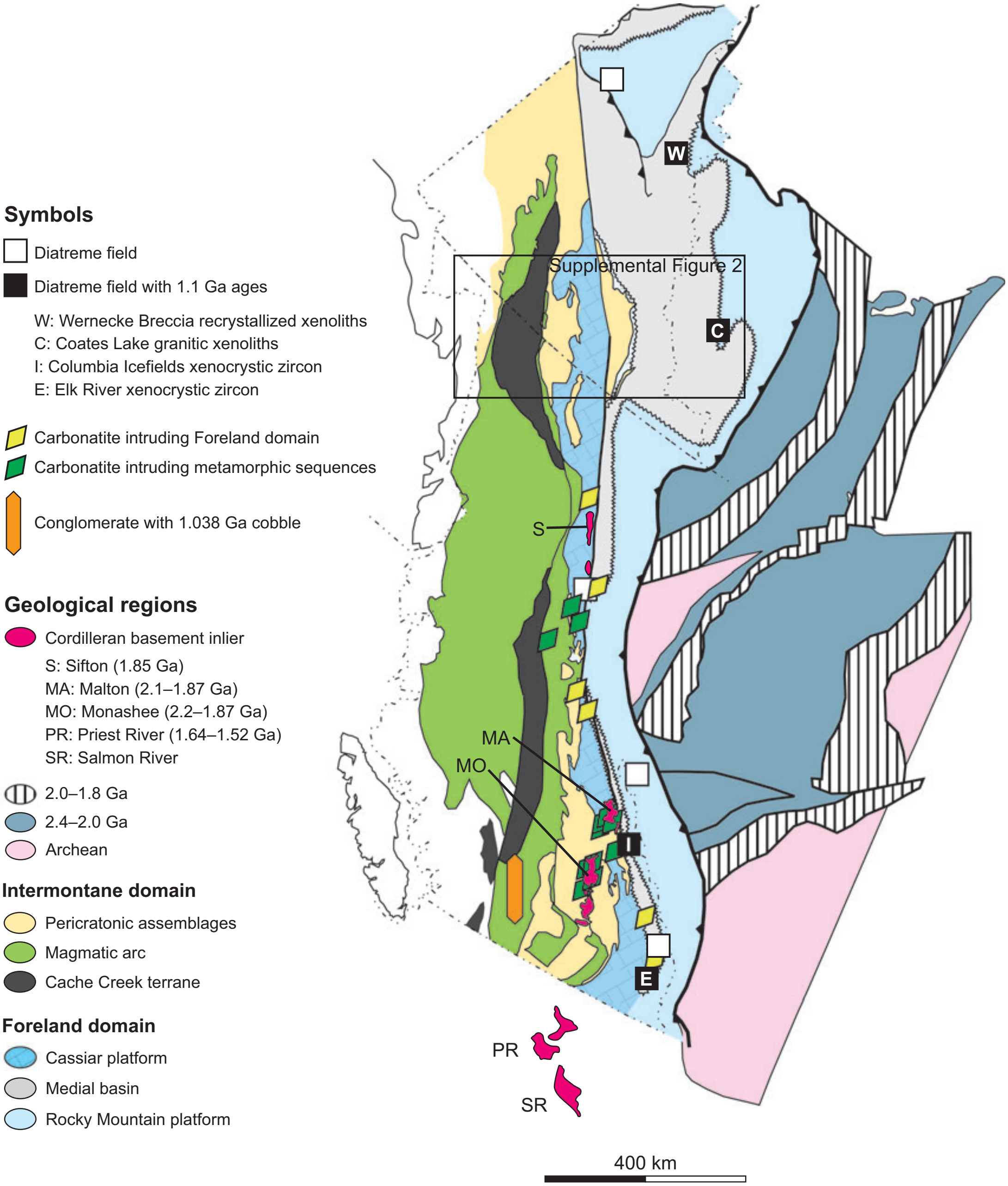


Figure 4

Diatremes (*squares*) and carbonatites (*diamonds*) of the Cordillera. Inliers of Precambrian basement within the Cordillera are indicated (*magenta*), as are the domains of the autochthonous Precambrian basement. Box shows location of **Supplemental Figure 2.**

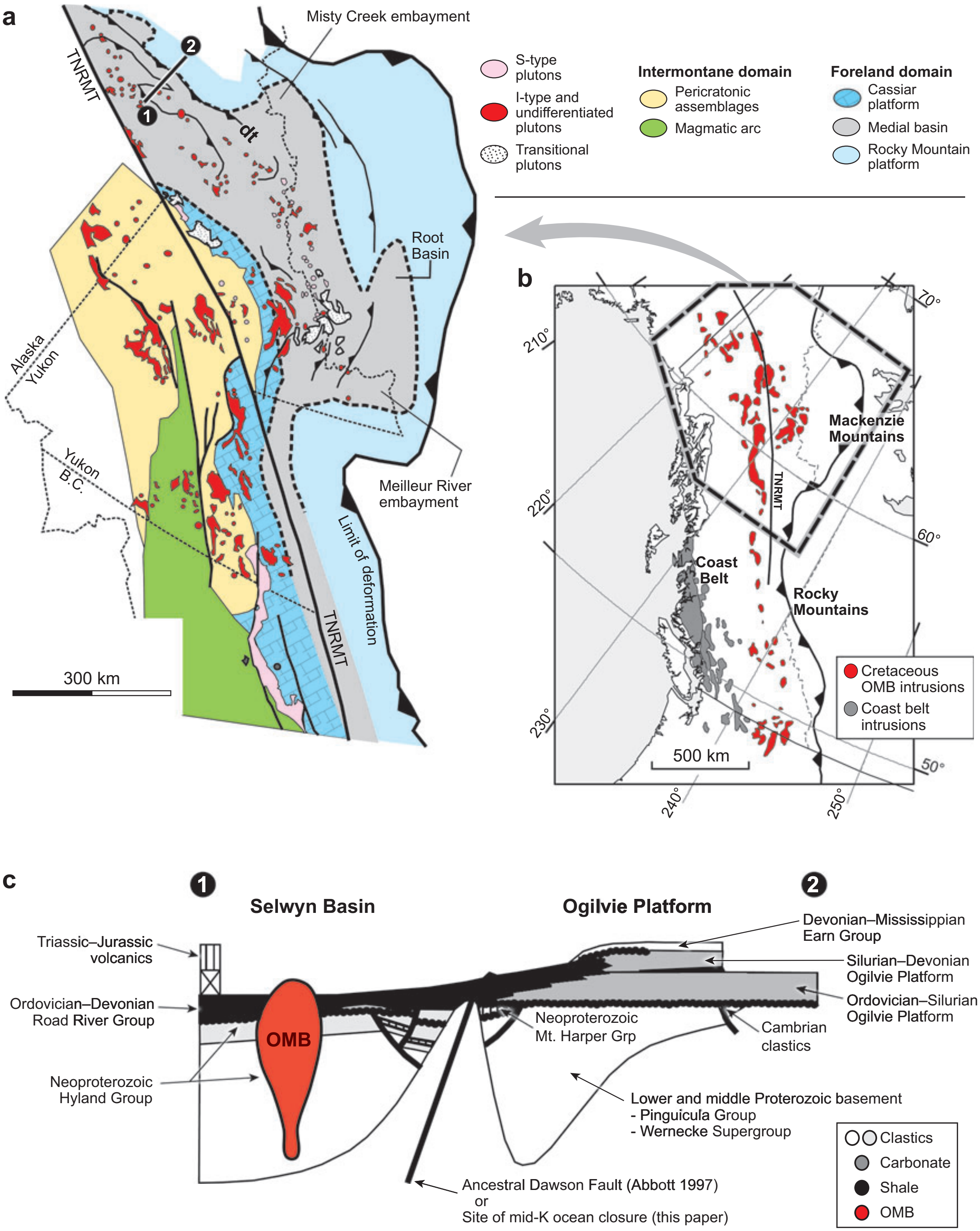
2002) (**Figure 4**) >700 Ma younger than any known basement within the adjacent autochthon.

North American mantle can similarly be distinguished from the mantle underpinning the Cassiar Platform and intermontane domain. Mantle xenoliths in Cretaceous and younger kimberlite pipes intruding the authochthon range in age from Archean to 1.8 Ga (LeCheminant et al. 1996). Alkalic magmatic rocks derived from melting of the cratonic mantle lithosphere are isotopically radiogenic. For example, phlogopite separated from the Lac de Gras kimberlites of the Slave craton yields initial Sr values ranging from 0.704 to 0.706 (Creaser et al. 2004). In contrast, Re-Os isotopic studies of mantle xenoliths sampled by young alkali basalts erupting through Stikinia-Quesnellia and the pericratonic assemblages all yield similar Os model ages of 1.1 Ga, interpreted as the age of melt extraction and lithospheric mantle formation (Peslier et al. 2000) (**Supplemental Figure 1**). Isotopic studies of the alkali basalts indicate that the mantle source region beneath the Medial Basin is less radiogenic (iSr = 0.7034) than either the cratonic mantle to the east or the mantle underpinning the Cassiar Platform to the west (Abraham et al. 2001) (**Supplemental Figure 1**).

MID-CRETACEOUS ARC MAGMATISM OF THE INTERMONTANE DOMAIN, CASSIAR PLATFORM, AND THE MEDIAL BASIN

A defining feature of the Canadian Cordillera is the mid-Cretaceous Omineca Magmatic Belt (OMB). Here I focus on the northern OMB (**Figure 5**) for which there is a significant geochemical and geochronological database. The northern OMB is a belt of I- and S-type plutons (**Figure 5**) that intrude the intermontane domain, Cassiar Platform and the Medial Basin. Resulting contact metamorphism is restricted to narrow aureoles (Gordey & Anderson 1993, Pigage & Anderson 1985, Smith & Erdmer 1990). The magmatic belt youngs to the northeast (**Figure 6**) (Breitsprecher & Mortensen 2004, Hart et al. 2004a, Mortensen et al. 2000). In the west, plutons intruded from 115 to 100 Ma, with small volume plutons as old as 124 Ma. Successively younger orogen-parallel intrusive bands to the northeast terminate in a set of 92 +/− 1 Ma plutons. The southwest to northeast age progression is accompanied by changes in lithology, chemistry, and structure (Hart et al. 2004a, Mortensen et al. 2000). The western plutons are midcrustal, concordant, foliated, metaluminous, calc-alkaline hornblende-biotite granodiorite sills with titanite and magnetite. Discordant, shallow-level plugs of granite are minor. The intrusions are spatially associated with and were syn-kinematic with steep, dextral-transpressive faults (Johnston 1999). Initial Sr is ~0.707 (n = 12) and geochemical data indicate enrichment in large-ion lithophile elements (LILE) and negative Nb anomalies (Selby et al. 1999).

Younger plutons to the northeast are peraluminous and felsic, foliated to massive hornblende-biotite granodiorite and muscovite-biotite granite (**Figure 5**). Accessory ilmenite and monazite indicate reduced magmas (Hart et al. 2004a). Magmatism was syn- to postkinematic; different magmatic phases are juxtaposed along brittle-ductile high-strain zones (Gordey & Anderson 1993), with predominantly shallow-level plutons lying elongate parallel to and spatially associated with northeast-verging dextral


a
Misty Creek embayment
TNRMT
dt
Root Basin
Alaska
Yukon
Yukon
B.C.
Meilleur River embayment
Limit of deformation
TNRMT
300 km
S-type plutons
I-type and undifferentiated plutons
Transitional plutons
Intermontane domain
Pericratonic assemblages
Magmatic arc
Foreland domain
Cassiar platform
Medial basin
Rocky Mountain platform
b
70°
210°
220°
60°
230°
50°
240°
250°
Mackenzie Mountains
TNRMT
Coast Belt
Rocky Mountains
Cretaceous OMB intrusions
Coast belt intrusions
500 km
c
Selwyn Basin
Ogilvie Platform
Triassic–Jurassic volcanics
Ordovician–Devonian Road River Group
Neoproterozoic Hyland Group
OMB
Devonian–Mississippian Earn Group
Silurian–Devonian Ogilvie Platform
Ordovician–Silurian Ogilvie Platform
Neoproterozoic Mt. Harper Grp
Cambrian clastics
Lower and middle Proterozoic basement
- Pinguicula Group
- Wernecke Supergroup
Ancestral Dawson Fault (Abbott 1997)
or
Site of mid-K ocean closure (this paper)
Clastics
Carbonate
Shale
OMB

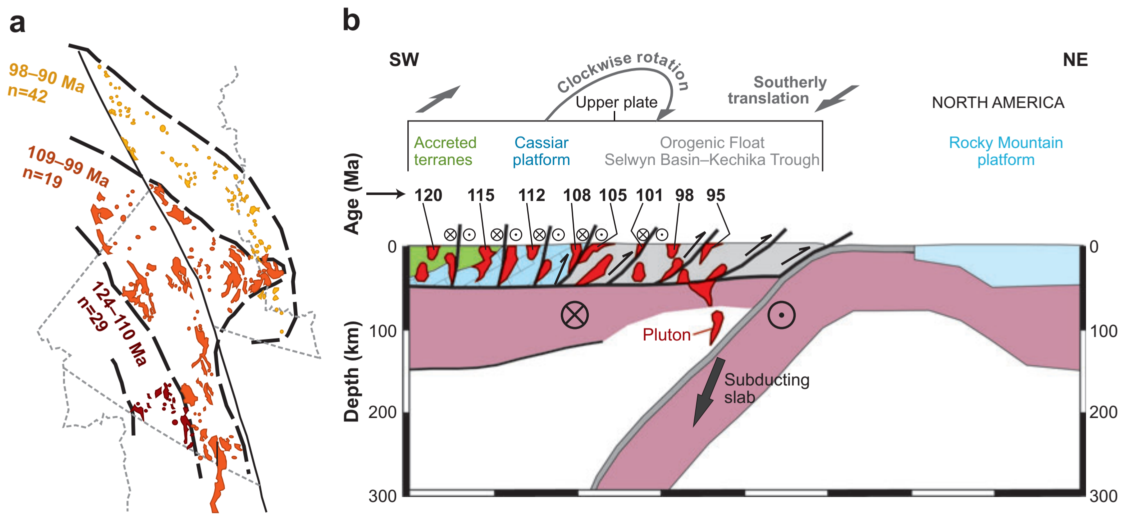


Figure 6

(*a*) Northern OMB plutons contoured (*long dash lines*) by age (Breitsprecher & Mortensen 2004). After Mortensen et al. (2000) and Hart et al. (2004b). n, number of age determinations. (*b*) Schematic cross section showing tectonic setting of OMB arc (plutons in *red*) at ~95 Ma (Oldow et al. 1990). Oblique subduction is resolved into NE-verging thrust faults and dextral strike-slip that displace trenchward crust south (*toward reader*). The addition of orogenic float along thrusts results in younging of arc plutons to the NE (*large arrow* indicates motion of subducting slab). Arrows at top indicate southward displacements and clockwise rotations attributable to dextral strike-slip fault.

transpressive thrust faults (Murphy 1997) (**Figure 5**). The thrust faults root west beneath and are interpreted to share a common basal detachment with coeval dextral strike-slip faults. Hornblende and biotite-bearing intrusions have initial Sr of 0.706 to 0.710, LILE enrichment, and negative Nb anomalies; two mica granites have initial Sr of 0.709 to >0.730 (n = 15) (Driver et al. 2000). The most northeasterly plutons intruded from 94 to 90 Ma, have circular map-patterns and are bimodal metaluminous to alkaline, quartz monzonite to syenite, with rare gabbro and lamprophyre (Anderson 1987).

Figure 5

(*a*) The northern Omineca magmatic belt (OMB) showing distribution of S- (*pink*), I-type and undifferentiated (*red*), and (*stipple*) transitional plutons that are intermediate between I- and S-types (Gordey & Anderson 1993, Hart et al. 2004a, Mortensen et al. 2000). Four-hundred and twenty-five kilometers of Eocene displacement along the Tintina–Northern Rocky Mountain Trench (TNRMT) fault has been restored. Teeth on mid-Cretaceous thrust faults, including the Dawson Thrust (dt), point into the hangingwalls, opposite their sense of vergence; strike-slip faults are dextral. (*b*) Inset location map shows Cordilleran distribution of mid-Cretaceous OMB (*red*) and Coast belt (*gray*) intrusions. (*c*) Cross section at lower right (line 1–2 in *panel a*) shows palinspastically restored stratigraphic section across hangingwall (*left*, intruded by OMB) and footwall (*right*) sequences of Dawson Thrust (Abbott 1997).

Hornblende-biotite granodiorites characterized by LILE enrichment and negative Nb anomalies, like those of the OMB, are commonly explained as arc magmas (Hamilton 1995). An arc-interpretation has, however, been previously ruled out because of the broad width and lack of a nearby, temporally associated subduction complex. Previous explanations of OMB magmatism have included back-arc magmatism behind the coastal arc developed on the Insular Belt Wrangellia-Alexander terranes (Hart et al. 2004a, Kidwell et al. 2005, Mair et al. 2006) and melting of a thickened crustal welt (Armstrong 1988, Driver et al. 2000, Monger et al. 1982). These models have difficulty explaining the volume, lithology, and structural setting of the northern OMB. Continental back-arcs are typically characterized by small volumes of alkalic magmatism intruded during extension, modern flat slabs are amagmatic (Gutscher et al. 2000), and thermal modeling indicates that slab flattening yields no significant magmatism (English et al. 2003). Crustal welts yield irregularly developed, amphibolite-grade crystalline terranes with spatially associated plutons that postdate crustal thickening and lack a mantle component. Although Late Triassic orogeny thickened the crust in the OMB region, this crustal welt was unroofed by 180 Ma (Colpron et al. 1996, Johnston et al. 1996a). Neither back-arc magmatism nor melting of a crustal welt explains the systematic southwest- to northeast-younging and related geochemical changes of the OMB. I suggest a model involving the structural addition of orogenic float to the upper plate of a convergent margin (Oldow et al. 1990).

Orogenic float (Oldow et al. 1990), when added to the upper plate at a convergent margin, is bound by landward-dipping thrust faults that young oceanward and root into a basal decollement. Oblique subduction is resolved into margin-normal thrusts and coeval margin-parallel strike-slip faults. Assuming a fixed trench, the addition of orogenic float to the upper plate is accommodated by the displacement of upper plate crust, including previously intruded arc plutons, away from the trench (Oldow et al. 1990). Because arcs are produced at a relatively fixed distance from the trench (Hamilton 1995), the addition of orogenic float gives rise to a broad zone of arc plutons that young oceanward (**Figure 6**).

In the orogenic float model, the northern OMB is an arc built initially on an upper plate consisting of the intermontane domain and Cassiar Platform, and the Medial Basin an accretionary prism of orogenic float. The width and northeast-younging of the arc is attributable to displacement of upper plate crust to the southwest away from a fixed trench, accommodating accretion of orogenic float derived from a subducting lower plate. The distribution of arc plutons and accretionary prism strata implies west-dipping subduction of oceanic lithosphere that lay east of Cassiar platform, consistent with the northeast younging direction of the arc, and the northeast vergence of the thrust faults bounding the orogenic float (**Figure 6**). Strata of the Rocky Mountain platform form the footwall to the imbricated Medial Basin strata, implying that it was the entry of buoyant North American continental lithosphere into the trench that terminated subduction at 92 Ma. Because the abyssal strata of the Medial Basin are allochthonous orogenic float, eastward salients of the basinal strata (**Figure 5**) are probably in part structural artefacts (discussed below).

The eastward transition from I-type, oxidized magnetite-bearing plutons to reduced ilmenite-bearing, I- and S-type plutons reflects arc migration onto the accreting orogenic float (**Figure 6**). Underplating of argillaceous float by gabbro would melt the float, explaining the radiogenic S-type plutons. The radiogenic character of some I-type plutons has been used to argue that all of the intrusions are the result of crustal melting. Subduction of transitional lithosphere and fluxing of the mantle wedge by continental components derived from the heavily sediment-laden downgoing plate may, however, explain the evolved, radiogenic mantle-derived magmatism.

Interpretation of the northern OMB as an east-facing arc implies that the Medial Basin is host to a cryptic suture. The imbricated, cleaved, disrupted and deformed chert, argillite, and deep-water limestones of the Medial Basin "accretionary prism," limits the cryptic suture to being beneath or along its east margin. It remains unclear if any of the voluminous suite of ophiolitic rocks mapped in Alaska (Patton et al. 1994) originated within the forearc of the OMB arc. Along strike to the south, the OMB in northern Idaho is characterized by the Salmon River suture in which ophiolitic rocks were obducted, sheared, and overprinted by OMB intrusions, all between 130 and 80 Ma (McClelland et al. 2000). These relationships are consistent with the Salmon River ophiolite being a preserved remnant of the suture separating the OMB arc from North America.

PALEOMAGNETIC AND PALEOBOTANICAL DATA

Cretaceous bedded sedimentary and volcanic rocks that unconformably overlie the intermontane domain consistently yield anomalously shallow paleomagnetic inclinations relative to cratonic North America (Enkin et al. 2003, 2006; Irving et al. 1996, Wynne et al. 1998) (**Figure 7**). These paleomagnetic data imply that the intermontane domain crust lay >2000 km to the south relative to the autochthon between 90 Ma and 70 Ma (Enkin 2006). Paleomagnetic studies of plutons provide less consistent results, probably owing to the difficulty in constraining paleohorizontal and the age of magnetic remanance. Although some studies of plutons have been interpreted as showing little evidence for a southerly origin (McCausland et al. 2006, Symons et al. 2005), the bulk of pluton studies yield results consistent with results obtained from layered supracrustal rocks (Enkin 2006, Irving & Wynne 1992).

In Yukon, 70 Ma volcanic flows and interlayered sedimentary rocks of the Carmacks Group unconformably overlie the pericratonic assemblages and Stikinia-Quesnellia (**Figure 7**). Paleomagnetic studies of the Carmacks Group indicate deposition 1950 ± 600 km to the south relative to cratonic North America (Enkin et al. 2006, Johnston et al. 1996b, Marquis & Globerman 1988, Wynne et al. 1998) (**Figure 7**). The most easterly exposure of the Carmacks Group, at Solitary Mountain, lies just 5 km west of the fault zone along which the pericratonic assemblages and the Cassiar Platform are juxtaposed (Colpron et al. 2005). A 105 Ma batholith, one of the mid-Cretaceous OMB plutons, plugs the fault zone. The contact aureole for the batholith extends unbroken across the fault zone into the pericratonic assemblages, limiting fault juxtaposition of the intermontane domain and the adjacent Cassiar Platform to having occurred prior to 105 Ma, long before deposition of the Carmacks Group.

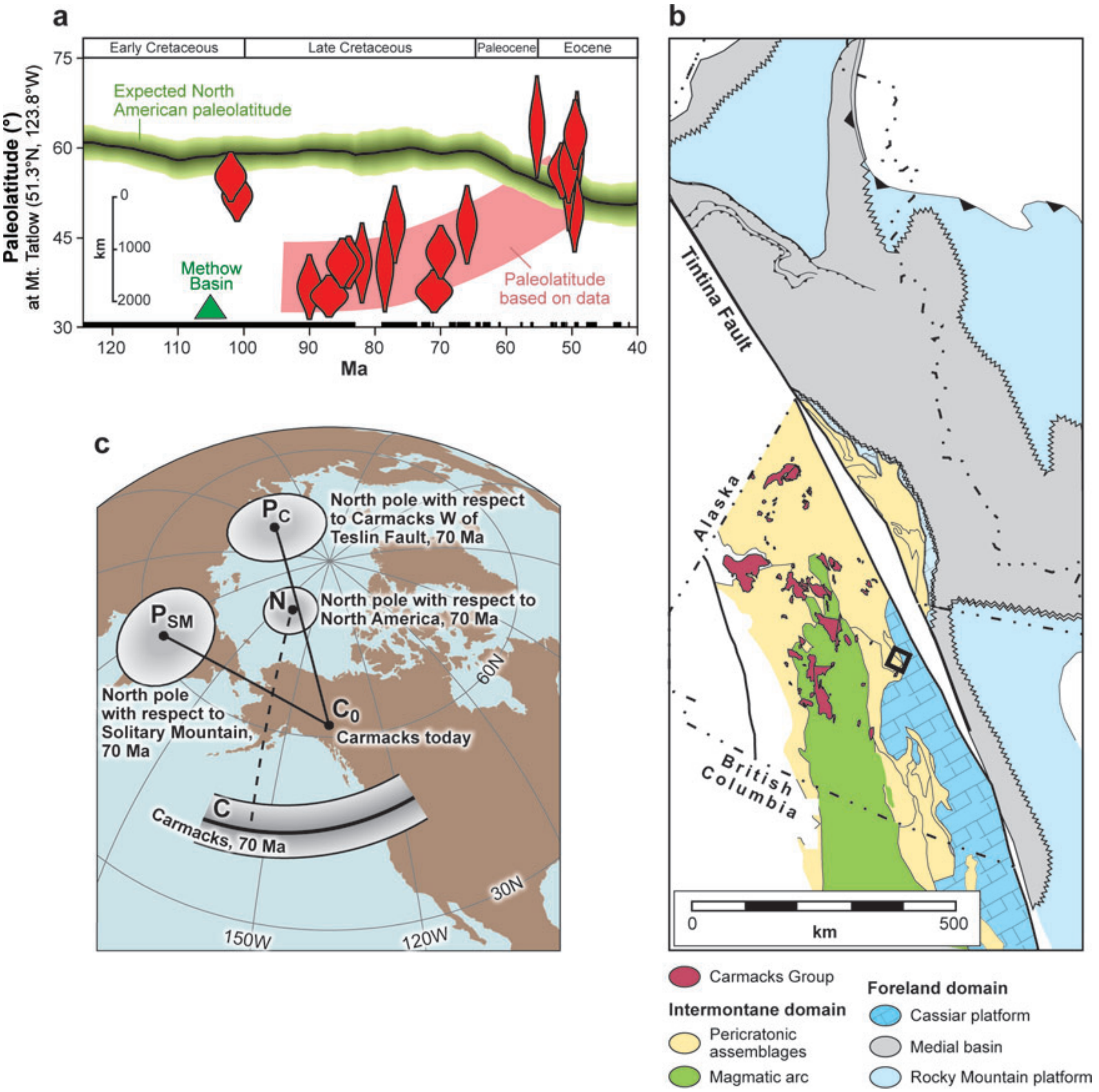


Figure 7

(*a*) Paleolatitudes relative to the expected North American paleolatitude (*green line*) for Cretaceous and Paleocene bedded rocks based on paleomagnetic data (*red diamonds*; see Enkin 2006 for primary data and locations) and on paleobotanical data (Miller et al. 2006) (*green triangle*, the Methow Basin). Global polarity record (black: normal; white: reversed) shown at bottom. (*b*) Geology map, with 425 km dextral motion along Tintina fault restored, showing distribution of Carmacks Group (*maroon*) (Gladwin & Johnston 2006, Wynne et al. 1998). Black box indicates location of Solitary Mountain. (*c*) Paleomagnetic results for the Carmacks Group showing far-sided poles relative to the cratonic North American pole location. Carmacks Group at Solitary Mountain yields a different azimuth indicating rotation relative to Carmacks Group to the west (Enkin et al. 2006).

Intrusions of the mid-Cretaceous OMB are, as indicated above, widespread in Yukon, pinning together Stikinia-Quesnellia, the pericratonic assemblages, Cassiar Platform, and much of the Medial Basin (**Figure 5**) (Woodsworth et al. 1992). The plutons extend east almost to the eastern margin of the Medial Basin (Gordey & Anderson 1993). Hence, the paleomagnetic results for the Carmacks Group apply to the Cassiar Platform and much of the Medial Basin (Gladwin & Johnston 2006). The implications of these findings are that (*a*) the Cassiar Platform did not become fixed to autochthonous North America until the Eocene; (*b*) that it, together with Medial Basin strata to the east and intermontane domain strata to the west, resided ~2000 km to the south relative to cratonic North America at 70 Ma; and (*c*) that the structures along which Late Cretaceous northward motion were accommodated lie in the easternmost Medial Basin, along the eastern boundary of the Medial Basin, or in the Rocky Mountain Platform to the east.

Layered sedimentary rocks overlying the intermontane domain locally contain fossil leaves that can be used to estimate mean annual temperature (MAT). Because the MAT is primarily a function of latitude (Miller et al. 2006), paleobiographic MAT analyses can be used to constrain paleolatitude. The mid-Cretaceous (110 to 100 Ma) Winthrop Formation, part of the Methow Basin of the southwestern intermontane domain (**Figure 2**), contains fossil angiosperm leaves. Leaf-margin analysis of the angiosperm fossils indicates a subtropical to tropical growth environment, consistent with a latitude of 38.4° and implying deposition 2200 km to the south (Miller et al. 2006) (**Figure 7**), consistent with the paleomagnetic data. Because the intermontane domain can be tightly tied to the pericratonic assemblages and the Cassiar Platform by the Early Jurassic, the southerly latitude indicated by the paleobiographic analysis provides further confirmation of the mobility of the Cassiar Platform and western Medial Basin into the Late Cretaceous.

Paleomagnetic studies of older rocks in the Cordillera similarly imply significant mobility of the intermontane domain relative to cratonic North America. Ophiolitic sequences interleaved with the pericratonic assemblages include Permian abyssal seafloor sedimentary rocks. A primary paleomagnetic remanance in the sedimentary rocks yields anomalously shallow inclinations relative to cratonic North America, implying deposition >2000 km to the south (Richards et al. 1993). Triassic and Jurassic paleomagnetic results from Stikinia-Quesnellia yield paleolatitudes that are broadly similar to cratonic values but that require large and variable rotations, and for which the hemisphere of origin is ambiguous (Irving & Wynne 1992).

BENDS OF FAULTS AND FACIES BOUNDARIES

The east margin of the Medial Basin is characterized by numerous eastward salients, including the Meilleur River and Misty Creek embayments (**Figure 2**). The distribution of shallow- and deep-water facies is commonly interpreted as primary features reflecting the geometry of the original rifted west margin of the continent (Cecile et al. 1997). However, two structural observations are inconsistent with the distribution of shallow- and deep-water facies being a primary feature: Cretaceous faults commonly parallel and mimic the older facies boundary and interpreted Paleozoic rift

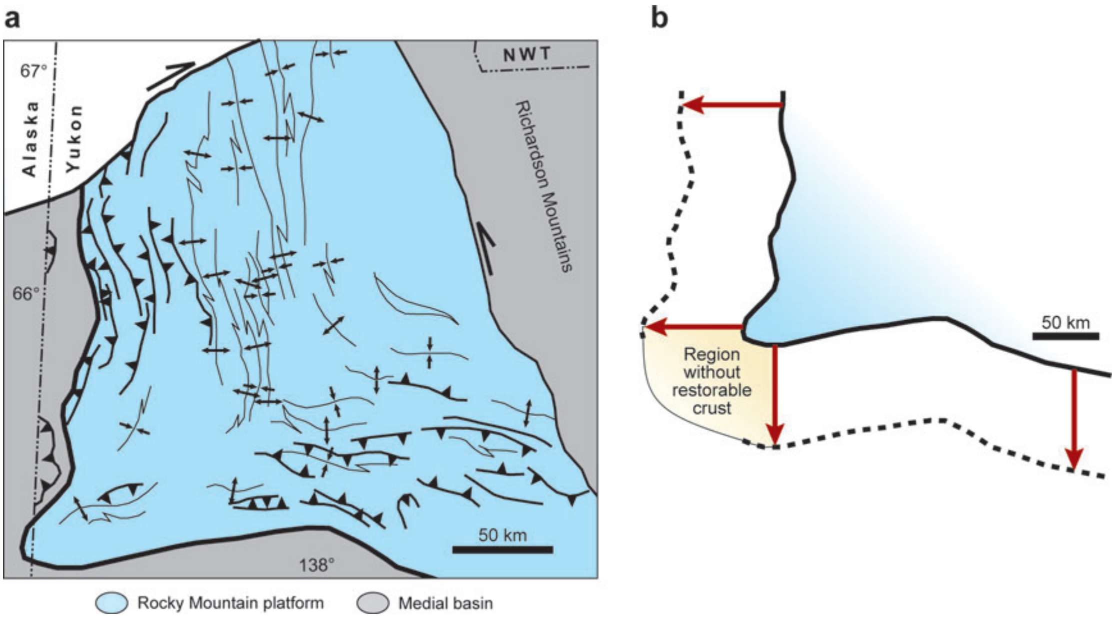


Figure 8

(*a*) Structural map of the Ogilvie Mountains (location on **Figure 2**) showing distribution of shallow and deep water facies and of Cretaceous thrust faults (McMechan et al. 1992, Norris 1985). (*b*) Simplified palinspastic restoration of facies boundary, assuming 75 km shortening along the south and west limbs of the mountain belt.

structures are continuous around bends of 180°. I provide two examples, the Ogilvie deflection and the Misty Creek embayment.

The Ogilvie Mountains are a foreland-verging fold and thrust belt of mid-Cretaceous age that is spatially coincident with the boundary between basinal shales to the south and the more northerly Ogilvie or Yukon Platform (**Figures 3** and **8**). To the west, the south margin of the platform describes a 90° deflection from an east-west trending feature to a north-south orientation. The change in the orientation of the facies boundary is paralleled by a change in the orientation of the fold and thrust belt, a feature referred to as the Ogilvie deflection (Norris 1972). The east-west trending fold and thrust belt verges north, and is continuous through the Ogilvie deflection into east-verging faults of the north-south trending portion of the belt.

Palinspastic restoration of the thrust sheets results in a significant room problem; the restored east-west trending southern and north-south trending western portions thrust belts are restored a considerable distance away from one another around the hinge region, implying that what is now the hinge region lacked crust prior to the development of the fold and thrust belt (**Figure 8**). This room problem is significant—the "hole" in the palinspastic restoration covers an area of almost 10,000 km^2. The problem can also be thought of as a line-length problem: Propagation of thrust sheets continuous around a preexisting bend toward the foreland should have resulted in enormous strike-parallel contraction within the hinge region, for which there is no

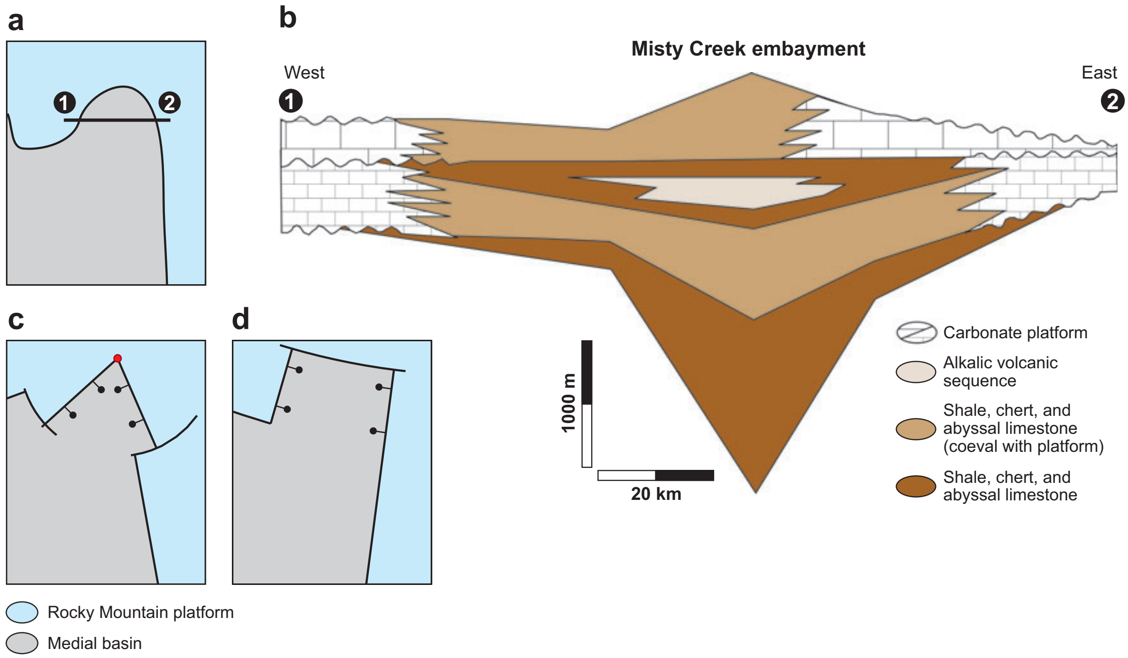


Figure 9

(*a*) Simplified geology map of Misty Creek embayment (location indicated on **Figure 5**). (*b*) Cross section 1–2 showing symmetric "steer's head" geometry of lithofacies across basin (Cecile et al. 1997). (*c*, *d*) Simplified models showing how steer's head rift (*black dots with stems on down-dropped blocks*) has to end either in a pole of rotation (*red dot*) or a transform fault, respectively.

evidence (**Figure 8**). Bending of an originally linear fold and thrust belt removes the room/line length problem and implies that the facies boundary geometry is the result of strain and is not primary.

Cross sections drawn perpendicular to the long axis of the north-northwest-trending Misty Creek embayment show that it is symmetric, with a "steer's head" profile (White & McKenzie 1988) consisting of a central "head" of shale and deep-water limestone with fringing "horns" of shallow-water carbonates (Cecile 1982) (**Figure 9**). Based on the observed steer's head symmetry, a model of symmetrical rifting was employed to explain the origin and geometry of the embayment (Cecile et al. 1997). A prediction of a model of parallel, symmetrical opposing rifts is that they must terminate along strike in either an Euler pole of rotation, or (given a distant Euler pole) a transform fault (**Figure 9**). Instead, the sedimentary facies and bounding structures are continuous through 180° around the entire embayment; sections drawn perpendicular to the facies boundary are everywhere identical, a geometry that cannot be produced through any simple rift model. Bending of an originally linear rift margin best explains the geometry of the Misty Creek embayment.

Parallelism of Cretaceous thrust faults around major bends in facies boundaries, and continuity rift margins around bends of 180°, indicate that the geometry of the platform to basin facies boundary is not a primary feature of the orogen, but is the result of Cretaceous bending, about vertical axes of rotation, of originally more linear rift margins and fold and thrust belts.

Rift-Related Carbonatites and Alkaline Igneous Complexes

The margins of the Medial Basin are characterized by Lower to Middle Paleozoic alkaline igneous and carbonatite complexes along the length of the orogen (Pell 1994) (**Figure 4**). The intrusive complexes are locally spatially associated with coeval alkaline and potassic basalts (Goodfellow et al. 1995). Deformation of the complexes is indicated by the presence of a weak to moderately well-developed foliation, local mylonitization, folding, boudinage, and truncation by thrust faults (Johnston & Pyle 2005, Pell 1994). The presence of syn-magmatic extensional structures (Johnston & Pyle 2005) and proximity to the margins of the Medial Basin imply that the complexes lie along and mark ancient rift zones.

Global study of deformed alkaline igneous and carbonatite complexes indicate that they commonly lie along and characterize suture zones; >90% of deformed African nepheline syenite and carbonatite complexes lie along and mark known and inferred sutures (Burke et al. 2003). Because alkaline igneous complexes and carbonatite intrusions characterize intracontinental rifts (Bailey 1977), their location within collisional orogens can be explained in terms of a Wilson cycle model, with magmatism occurring along and marking the rifted margin of a continent, and deformation occurring during ocean closure and subsequent collision (Burke et al. 2003). The distribution of Cordilleran carbonatites, alkaline igneous complexes, and coeval alkaline basalts within the Medial Basin is, therefore, consistent with the Medial Basin being a cryptic suture separating a west-facing autochthonous continental margin (the Rocky Mountain Platform) from a more westerly, east-facing exotic continental margin represented by the Cassiar Platform.

Late Cretaceous Fold and Thrust Belt Formation

The Canadian Cordillera is characterized along its entire length by a Late Cretaceous, foreland-verging fold and thrust belt (**Figures 1** and **2**), of which the Rocky Mountains of southwest Alberta are the most spectacular and well-known manifestation. Shortening accommodated by the thrust belt decreases northward from a maximum of >250 km in southern Alberta (Price & Sears 2000) to less than 50 km in the north (McMechan et al. 1992). Cooling ages, paleomagnetic studies, direct dating of fault rocks, and the age of deformed and overlapping undeformed strata limit deformation to having occurred between the Campanian (~80 Ma) and the Early Eocene (~50 Ma) (Enkin et al. 2000, McMechan et al. 1992, Price 1981, van der Pluijm et al. 2006).

Explaining the origin of the fold and thrust belt is problematic. Most workers assume fold and thrust belt formation postdates previous accretion of the intermontane and insular domains (Monger et al. 1982). Hence the fold and thrust belt is assumed to have formed 1000 to 1500 km inboard from the active Late Cretaceous west margin of the continent and in the absence of any collisional event (English & Johnston 2004). Noncollisional explanations of the fold and thrust belt appeal to interaction with the slab subducting beneath the west margin of the continent. A compressive stress regime is inferred to have resulted from rapid relative convergence between the North American and the oceanic plate to the west (Hyndman 1972), possibly coupled with the presence of a thermally weakened back-arc region (Hyndman et al. 2005). Alternative explanations of compression involve flat slab subduction of the oceanic plate (Dickinson 2004), possibly due to the presence of an oceanic plateau on the subducting plate (Murphy et al. 2003) or because of subduction of a spreading ridge (Bird 1988). A model of transcurrent deformation links fold and thrust belt formation to 430 km of dextral strike-slip motion along the Northern Rocky Mountain–Tintina trench fault (**Figure 2**) via an inboard transfer of displacement (Price & Carmichael 1986).

Flat slab models, no matter what the cause, do not explain shortening along the entire length of the orogen. The Late Cretaceous fold and thrust belt runs the length of the continent (**Figure 1**); hence, appeals to subducted oceanic plateau and spreading ridges, although possibly explaining some local variation in structural style, do not provide a framework for explaining the entire thrust belt. Linking fold and thrust belt formation to the dextral strike-slip Northern Rocky Mountain–Tintina trench fault suffers the same dilemma in that it provides a local explanation for a continent-scale problem. In addition, displacement on the strike-slip fault occurred in the Eocene (Gabrielse et al. 2006) and would, therefore, have been coeval with only the final increments of thrust belt shortening recorded along the McConnell and related thrust faults (van der Pluijm et al. 2006). Appeals to high relative convergence are inconsistent with there having been at least two and possibly three different oceanic plates west of North America, each moving in different directions. Neither does it explain why fold and thrust belt formation occurred within the continent, well removed from the margin.

As discussed above, paleomagnetic and paleobiographic data from Cretaceous layered sedimentary and volcanic rocks imply >2000 km of northward motion of the intermontane domain, Cassiar Platform, and much of the Medial Basin. The Late Cretaceous fold and thrust belt is, therefore, coeval with the timing of northward motion, and is located along the eastern boundary of the region that moved north. The simplest explanation of the fold and thrust belt is, therefore, that it lies along and marks the boundary between the crust that moved north and the autochthon, and that it is a product of transpression during northward motion (Johnston 2001). In this model, the Northern Rocky Mountain–Tintina trench fault records the last component of northward motion and the McConnell thrust system the last component of convergence between the Cassiar Platform–Medial Basin and the autochthon.

THE CORDILLERAN COMPOSITE RIBBON CONTINENT

The Cassiar Platform is exotic with respect to autochthonous North America and has the following characteristics: (*a*) It has Triassic Eurasian fauna; (*b*) it was involved in a major Late Triassic orogenic event, whereas the coeval ancient west margin of North

America remained a passive margin facing west toward a broad open ocean; (*c*) it is underlain by a basement and lithospheric mantle that, having formed at least in part at 1.1 Ga, is 700 Ma younger than the youngest portions of the basement and lithospheric mantle underpinning the autochthon; (*d*) it is separated from autochthonous mantle by isotopically juvenile and oceanic-like mantle beneath the Medial Basin; (*e*) it is bound to the east by a belt of carbonatites, which likely delineate a cryptic suture; (*f*) it is characterized by an east-facing mid-Cretaceous magmatic arc that records west-dipping subduction beneath its eastern margin; and (*g*) it lay 2000 km to the south relative to the autochthon as recently as 80 Ma as indicated by paleomagnetic and paleobotanical data. The geometry of the boundaries separating platformal and basinal facies within the foreland domain is the product of Cretaceous tectonism and is not a reflection of the shape of the ancient rifted margin of the continent.

If the Cassiar Platform is not North American, from where does it hail? As discussed by Johnston & Borel (Johnston & Borel 2007; see **Supplemental Appendix**), the Cache Creek terrane places tight constraints on the location of the Cassiar Platform from the Permian through the Jurassic. Tethyan fauna and DUPAL-anomaly basalts characterize off-scraped Cache Creek seamounts, constraining the seamounts to having originated in the Tethyan Sea sensu stricto. The seamounts, Stikinia-Quesnellia, the pericratonic terranes, and the Cassiar Platform are pinned together by the end of the Triassic. Hence, the Cache Creek seamounts constrain the paleogeographic location of the Cassiar Platform, placing it in central Panthalassa, >4000 km west of the autochthon at 180 Ma (the most easterly possible point that could have been reached by the seamounts assuming that they migrated eastward out of the Tethys Sea at 11 cm year^{-1}) (**Figure 10**). Continental crust within the Cordilleran Orogen is, therefore, divisible into an eastern autochthonous platform and a western allochthonous platform, separated by a cryptic suture located within or along the margins of the Medial Basin (**Figures 1** and **2**).

The Cassiar Platform and intermontane domain terranes constitute a composite ribbon continent that can be followed northwest into Alaska (Johnston 2001), where it is continuous through two major oroclines, the Kulkbuk Hills orocline in the southwest (Bradley et al. 2003) and the Northeast Alaskan orocline in the northeast (Patton & Tailleur 1977) (**Figure 1**). The Farewell terrane is continuous around the hinge of the Kulukbuk Hills orocline, consists of an east-facing carbonate platform and a more easterly basinal facies of shale and chert, which are correlative with the Cassiar Platform and Medial Basin, respectively. Precambrian basement to the Farewell terrane carbonate platform consists of metasedimentary rocks and metabasite intruded by rhyolites that yield zircon crystallization ages as old as 979 Ma (Bradley et al. 2003), consistent with the evidence for a Grenvillian basement beneath the ribbon continent in the Canadian Cordillera. Lower Paleozoic Farewell terrane strata are characterized by distinct Siberian fauna, including trilobites, conodonts, and bachiopods (Blodgett et al. 2002, Dumoulin et al. 2002). Silurian aphrosalpingid sponges are only known elsewhere from the Alexander terrane of the insular domain and the Urals (Soja & Antoshkina 1997). Based on the faunal provinciality, Bradley et al. (2003) concluded that the Farewell terrane lay far removed from autochthonous North America, probably throughout the Paleozoic, consistent with an exotic origin for the

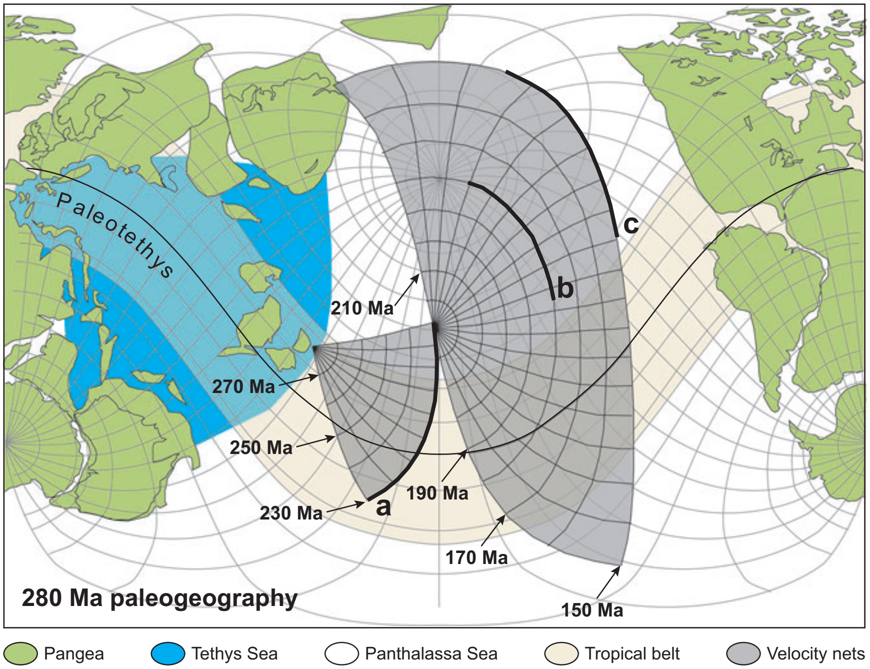


Figure 10

Paleogeographic map of Earth at 280 Ma (Stampfli & Borel 2002). The Tethys Sea (*blue*) separates the Laurasian (*north*) and Gondwanan (*south*) components of the supercontinent Pangea (*green*). The tropical belt is indicated through the uncolored superocean, Panthalassa, and the Tethyan Sea. Two velocity nets, one constructed for the period 280 Ma to 230 Ma and a second for the period 230 Ma to 150 Ma are shown. The velocity nets define the potential translation paths for the Cache Creek seamounts (assumptions outlined in text). Bold lines indicate the limits for the location of (*a*) seamount accretion to Quesnellia-Stikinia at 230 Ma (a point on this curve in the northernmost tropics is then used as the point of origin for the 230–150 Ma velocity net); (*b*) the intermontane domain terranes and Cassiar Platform at 180 Ma upon cessation of exhumation subsequent to Triassic orogenesis; and (*c*) these same terranes at 150 Ma, the time of drowning of the passive margin of western North America, and the first influx of westerly derived orogenic sediments onto the autochthon (Johnston & Borel 2007).

Cassiar Platform. Correlative strata in the western Brooks Range yield a paleomagnetic remanance acquired during mineralization at 330 Ma that places the Cassiar Platform >3000 km to the south and having since rotated 50° to 70° counterclockwise (Lewchuk et al. 2004).

In northernmost Alaska, the autochthon is largely buried beneath a thick Cretaceous foreland basin siliciclastic sequence, but is recognizable from aeromagnetic images (**Figure 11**). The boundary between autochthonous and exotic platform margins lies within the Brooks Range, but does not appear to be separated by a basinal facies equivalent to the Medial Basin (**Figures 1** and **12**). It may be that the excess of basinal facies rocks forming the Selwyn Basin were derived from this portion of

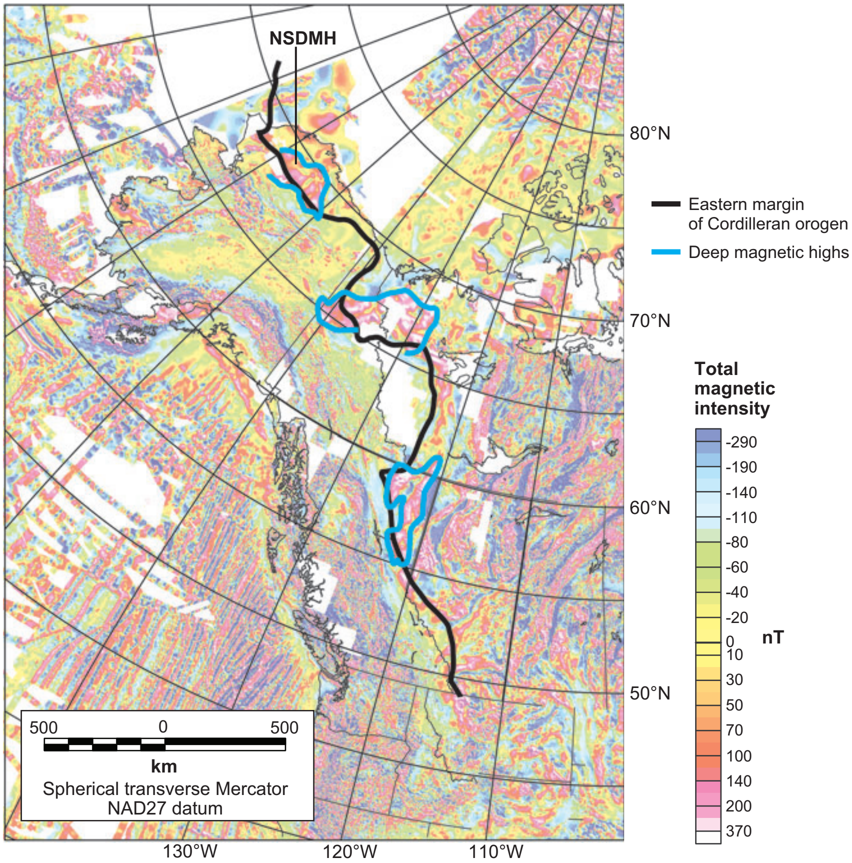


Figure 11

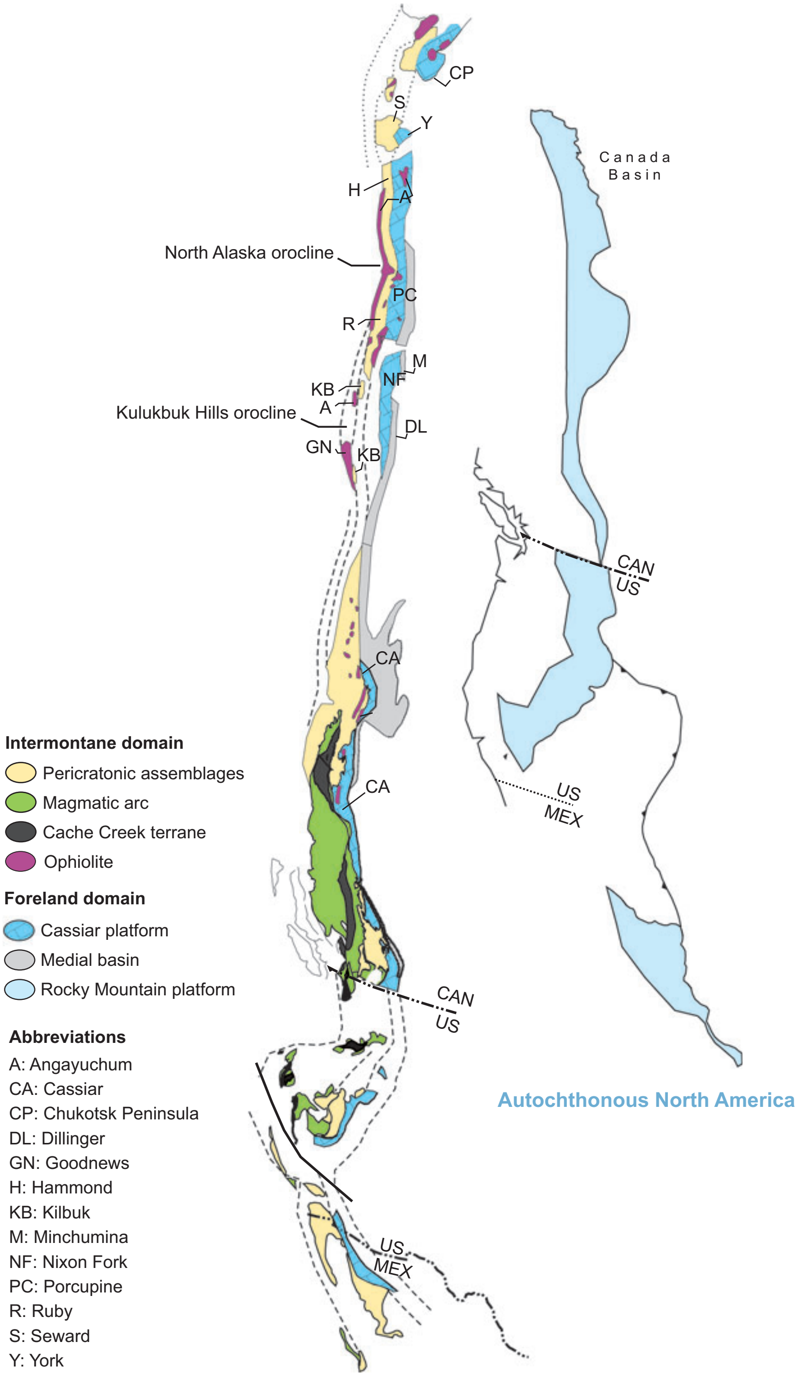
Magnetic anomaly map of northwestern North America (Saltus & Hudson 2007). The black line shows the eastern margin of the Cordilleran Orogen. Blue lines outline major deep magnetic highs that characterize the autochthon and include the NSDMH (north slope deep magnetic high).

the margin during mid-Cretaceous margin parallel motion in response to oblique subduction beneath the east-margin of the ribbon continent.

The ribbon continent is continuous to the south into the coterminous United States (**Figures 1** and **12**). Accretionary assemblages characterized by Tethyan

Figure 12

Palinspastic restoration of the Cordilleran composite ribbon continent to its geometry prior to buckling giving rise to the Kulukbuk Hills and Northeast Alaskan oroclines. There has been no attempt to restore bending and faulting of the southern coterminous U.S.A. portion of the ribbon continent. The ribbon continent is shown as being separated from the autochthon, but it may have been in close proximity to at least parts of the autochthon.



Restored Cordilleran ribbon continent

fauna are correlative with the Cache Creek terrane and can be followed south into California (**Figure 1**). Volcanic and sedimentary sequences that bound the accretionary assemblages to the east are characterized by McCloud fauna and are correlative with Stikinia-Quesnellia. East of the McCloud fauna terranes are ophiolitic (e.g., the Golcanda allochthon) and pericratonic assemblages (e.g., the Roberts Mountains allochthon), that together are correlative with the pericratonic assemblages of the eastern intermontane domain. Identifying strata correlative with the Cassiar Platform and Medial Basin remains, however, controversial. Nonetheless, the Tethyan and McCloud faunal belts provide a template for identifying the southern continuation of the Ribbon continent as far south as Mexico.

The geometry of the ribbon continent in Alaska is the result of oroclinal bending of the originally more linear ribbon continent (Johnston 2001). Bending postdates mid-Cretaceous magmatism, as parallel belts of 120 to 100 Ma magnetite- and ilmenite-bearing plutons are continuous around the Alaskan oroclines (Hart et al. 2004a). Post-Eocene strata unconformably overlie the oroclines, providing a minimum age constraint for oroclinal buckling. Northward subduction of the largely oceanic plate that bore the ribbon continent into a subduction zone marginal to and dipping beneath the Okhotsk-Chukotka arc of eastern Siberia resulted in northward motion of the continental ribbon (**Figure 13**). Buckling resulted from pinning of the leading edge of the ribbon continent into the Siberian upper plate, while the trailing portions of the ribbon remained within and moving north as part of the subducting lower plate. As the buckling ribbon continent was transferred from the lower to upper plate it became progressively overprinted by upper plate arc magmatism of the Okhotsk-Chukotka arc in Siberia and the Late Cretaceous to Eocene Kluane arc in Alaska. Buckling of the smaller Bowers-Shirshov-Kamchatka ribbon was coeval with buckling of the Cordillera ribbon continent and indicates that the two ribbons lay within and were translated north within the same plate (**Figure 13**). It seems likely that the combined buoyancy of the buckling Cordilleran and Bowers-Shirshov-Kamchatka ribbons eventually resulted in the failure of the Okhotsk-Chukotka subduction zone, at which point the oceanic plate bearing the ribbon continents broke behind the two "terrane wrecks" initiating the Aleutian subduction zone (**Figure 13**), and giving rise to the major change in motion of the oceanic plates of the Pacific basin at 45 Ma. Palinspastic restoration of the Alaskan oroclines restores the southern U.S. portion of the ribbon continent well to the south (**Figure 12**).

The nature of the boundary between the ribbon continent and the autochthon in the Late Cretaceous immediately prior to northward displacement and orocline formation remains poorly constrained. The ribbon continent has previously been depicted as being separated from the autochthon by a basin (Johnston 2001) (**Figure 12**). There is, however, little evidence to support the presence of a broad open oceanic basin separating the ribbon continent from the autochthon in the Late Cretaceous. A more likely scenario is that the ribbon continent adjoined the autochthon along a major transcurrent fault boundary (**Figure 13**). Minor transpression, together with the buckling of the ribbon continent in the north, would have

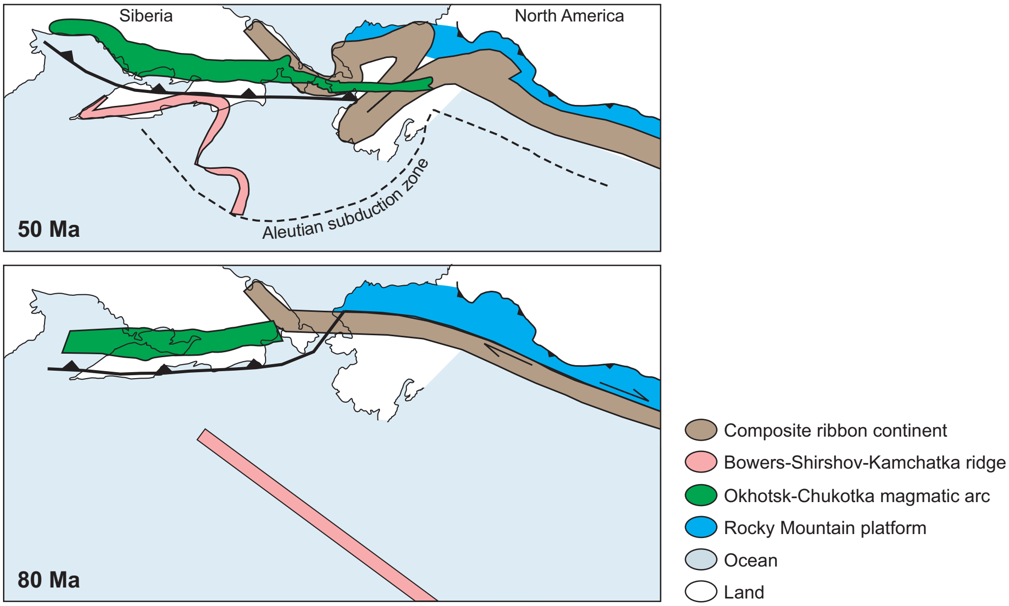


Figure 13

Schematic maps at 80 and 50 Ma showing buckling of the composite ribbon continent (*brown*) and the Bowers-Shirshov-Kamchatka ridge (*pink*) in response to subduction of oceanic lithosphere to the north giving rise to the Okhotsk-Chukotka magmatic arc (*green*). Northward motion is accommodated by a dextral transform fault that is gradually overprinted by foreland-verging thrust belts owing to a component of transpression. As the composite ribbon continent buckles and is transferred from the lower to upper plate, it becomes progressively overprinted by arc magmatism (Kluane arc in Alaska). Eventual failure of the Okhotsk-Chukotka subduction zone leads to the initiation of the Aleutian subduction zone (*dashed line*) oceanward of the buckled ribbon continent.

resulted in the original transcurrent boundary faults being carried inboard onto the autochthon where they were reactivated as thrust faults (Johnston 2001).

Two distinct phases of Cordilleran orogenesis can be distinguished (Johnston & Borel 2007). A Triassic to Early Jurassic accretionary phase involved the amalgamation of seamounts (Cache Creek), oceanic arcs (Stikinia-Quesnellia), pericratonic assemblages, and continental lithosphere (Cassiar Platform) (**Figure 10**). Accretionary orogenesis spanned at least 50 Ma (230 to 180 Ma) and produced a composite ribbon continent previously referred to as SAYBIA (Johnston 2001). The Upper Jurassic (150–155 Ma) drowning of the North American passive margin and the coeval deposition of the first orogenic clastic sediments on the autochthon records the initiation of the collisional second phase of Cordilleran orogenesis. The Late Cretaceous–Eocene Rocky Mountain fold and thrust belt records the terminal phase of collision, and involved transpression between the northward-translating composite ribbon continent and the autochthon (**Figure 13**). Cordilleran orogenesis was, therefore, far more akin to a complete Wilson cycle than has been previously recognized. Proterozoic and younger rifting of the west margin of Laurentia established a passive margin and led to the formation of an adjacent ocean basin. Closure of that basin in the Upper Jurassic led to collision and tectonic burial of the passive margin beneath a colliding continent. What distinguishes the Cordillera Wilson cycle from the classic Wilson cycle is that (*a*) the colliding continent was a composite ribbon continent constructed through an earlier accretionary orogenic phase, (*b*) collision was prolonged (150 to 50 Ma) and involved a final stage (80 to 50 Ma) in which the major motion was margin parallel (>2000 km of margin parallel displacement versus only hundreds of kilometers of margin normal convergence), (*c*) final collision involved oroclinal buckling of the accreting ribbon continent, and (*d*) the high aspect ratio of the colliding ribbon continent (long and narrow) prevented collision from being terminal—subduction continued beneath the new west margin of the continent.

DISCLOSURE STATEMENT

The author is not aware of any biases that might be perceived as affecting the objectivity of this review.

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ARTICLE

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Intra-oceanic subduction shaped the assembly of Cordilleran North America

Karin Sigloch[1] & Mitchell G. Mihalynuk[2]

The western quarter of North America consists of accreted terranes—crustal blocks added over the past 200 million years—but the reason for this is unclear. The widely accepted explanation posits that the oceanic Farallon plate acted as a conveyor belt, sweeping terranes into the continental margin while subducting under it. Here we show that this hypothesis, which fails to explain many terrane complexities, is also inconsistent with new tomographic images of lower-mantle slabs, and with their locations relative to plate reconstructions. We offer a reinterpretation of North American palaeogeography and test it quantitatively: collision events are clearly recorded by slab geometry, and can be time calibrated and reconciled with plate reconstructions and surface geology. The seas west of Cretaceous North America must have resembled today's western Pacific, strung with island arcs. All proto-Pacific plates initially subducted into almost stationary, intra-oceanic trenches, and accumulated below as massive vertical slab walls. Above the slabs, long-lived volcanic archipelagos and subduction complexes grew. Crustal accretion occurred when North America overrode the archipelagos, causing major episodes of Cordilleran mountain building.

Continents grow through subduction magmatism and collision of arcs and other buoyant crustal fragments at their margins. Poorly understood, such collisions are of broad scientific interest because they cause rapid geographic changes, affecting climate, ocean circulation, biota and the formation of economically important metal deposits. North America was enlarged by a sequence of massive terrane collisions relatively recently (between 200 million years (Myr) ago and 50 Myr ago), which created the mountainous Cordillera of the American West[1].

Reconciling geological records on land with those of the ocean basins has proved difficult. Magnetic stripes on the sea floor are the basis of all quantitative plate tectonic reconstructions, and well-preserved Atlantic spreading records indicate that North America has moved westward continuously since the breakup of Pangaea (about 185 Myr ago), away from Africa and Europe[2]. In contrast, more than half of the seafloor records of proto-Pacific (Panthalassa) ocean spreading are missing. The Pacific plate records the existence of another major oceanic plate to its northeast since at least 180 Myr ago: the Farallon plate. This plate is usually assumed to have filled the eastern Panthalassa basin, extending to the western margin of North America and subducting under it, although this is not required by the magnetic seafloor data. Hence, the Farallon plate, invoked as the causative agent in almost all major land geological events since late Jurassic times[2,3], should also have transported the terranes to the continental margin.

However, dozens of terranes have accreted to North America since 200 Myr ago[1], but not to the Andean margin of South America, supposedly a closely analogous setting. The terranes are mostly Triassic to Cretaceous island arc–subduction assemblages, but include three more heterogeneous superterranes: Intermontane (IMS)[4,5], Insular (INS)[4] and Guerrero (GUE)[6], which are essentially microcontinents. Their exact origins remain mysterious, but the terranes formed at various latitudes and times, as inferred from palaeomagnetic observations and fossil faunas, and pre-assembled at others. This implies the temporary existence of additional oceanic plates in the northeastern proto-Pacific Ocean[7,8], which are missing from quantitative plate reconstructions.

The mantle retains a memory of ancient plate configurations, in the form of subducted slabs, which body-wave tomography images *in situ* as domains of faster-than-average seismic wave velocities. Beneath North America these slab relics are massive, almost vertical walls extending from 800 to 2,000 km in depth, and typically 400–600 km wide (Fig. 1, Supplementary Fig. 1). The largest wall runs from northwest Canada to the eastern USA and on to Central America, and has been called the Farallon slab, one of the most massive features in global tomographies[9–12]. We argue that in fact most of this slab wall is not Farallon and subdivide it into the Angayucham (ANG), Mezcalera (MEZ) and the Southern Farallon (SF) components (Fig. 1b). This reinterpretation is based on our most recent tomography model[13], which utilizes dense USArray data[14] in addition to global network data, using a cutting-edge waveform inversion method: multiple-frequency P-wave tomography[13,15].

Besides putting the known eastern slab walls[9–12] in sharper focus, we discovered[15] that another, more westerly slab wall, the Cascadia Root (CR in Fig. 1b, depth 700 km to more than 1,800 km), connects upward continuously to the present-day Cascadia trench, into which the last remnant of the Farallon (Juan de Fuca) plate is subducting. This makes CR a Farallon slab by definition, and prompted us to re-evaluate whether MEZ/ANG/SF do really constitute subducted Farallon Ocean floor, which to our knowledge has never been questioned.

For plate reconstructions, the crucial question is whether (and how) these lower-mantle slab walls have moved laterally since they were deposited beneath their corresponding volcanic arcs (past versus present x–y positions in an absolute reference frame). Here we quantitatively test the hypothesis that they have not moved appreciably, that is, there has been only vertical sinking within our observational uncertainties (a few hundred kilometres laterally). Motivated primarily by imaged slab geometries, this null hypothesis also seems sensible in light of the Cenozoic subduction record, where absolute trench motion on average contributed only 10–30% of total plate convergence[16,17].

[1]Department of Earth and Environmental Sciences, Ludwig-Maximilians-Universität, Theresienstrasse 41, 80333 Munich, Germany. [2]British Columbia Geological Survey, PO Box 9333 Stn Prov Govt, Victoria, British Columbia V8W 9N3, Canada.

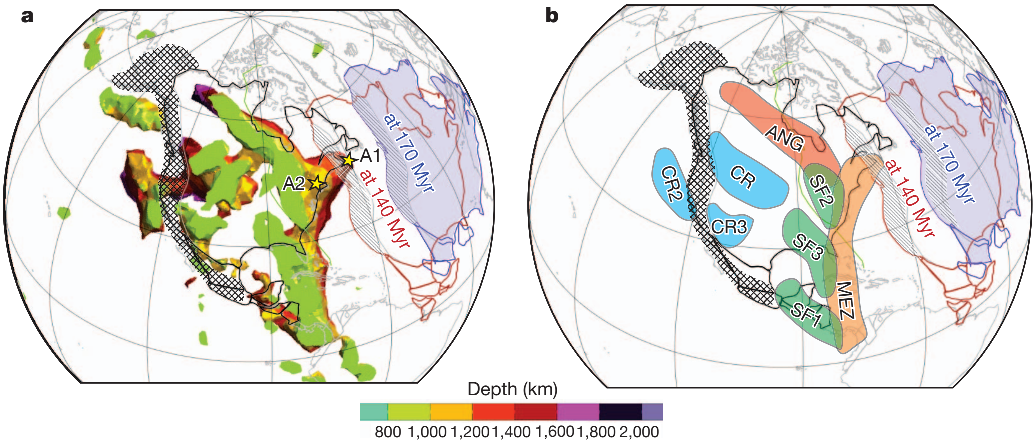

Figure 1 | Slabs under North America and continental motion over time. **a**, Subducted slabs at and below 900 km depth. P-wave tomography model[13] rendered as three-dimensional (3D) isosurface contours, which enclose faster-than-average structure (threshold $dv_P/v_P = 0.25\%$, where v_P is the P-wave velocity). Colour signifies depth and changes every 200 km; the scene is illuminated to convey 3D perspective. At a sinking rate of about 10 mm yr^{-1}, this slab assemblage should have been deposited from about 200 Myr ago to 90 Myr ago. Reconstructed continent positions at 140 Myr ago are shown in a hotspot reference frame[21] and at 170 Myr ago in a hotspot/palaeomagnetic hybrid frame[22]. The hatched area represents location uncertainty for continental margin during Jurassic/Cretaceous times; the cross-hatched area shows terranes that accreted during Cretaceous and early Tertiary times. **b**, Interpretative legend. The slab walls divide into four groups: Cascadia/Northern Farallon slabs (blue) and Southern Farallon slabs (green), owing to eastward subduction; Angayucham (ANG, red) and Mezcalera (MEZ, orange) slabs, owing to westward subduction. Before 140 Myr ago, sizeable ocean basins separated North America from the ANG/MEZ trenches.

Slabs and arcs at stationary trenches

Figure 2a and b shows how a steep, widened slab wall could be piled up by nearly vertical sinking beneath a long-lived, stationary trench and volcanic arc. An Andean-style west-coast trench could not have been stationary because North America moved westward as the Atlantic Ocean spread. This contradiction is resolved by westward intra-oceanic subduction before the arrival of the continent, followed by a polarity switch of subduction to its current eastward motion into a continental-margin trench (Fig. 2c). Such a scenario implies that the imaged lower-mantle slabs MEZ/ANG/SF are Jurassic to Cretaceous in age, allowing the collision of North America with their subduction zones to cause the Cretaceous terrane accretions.

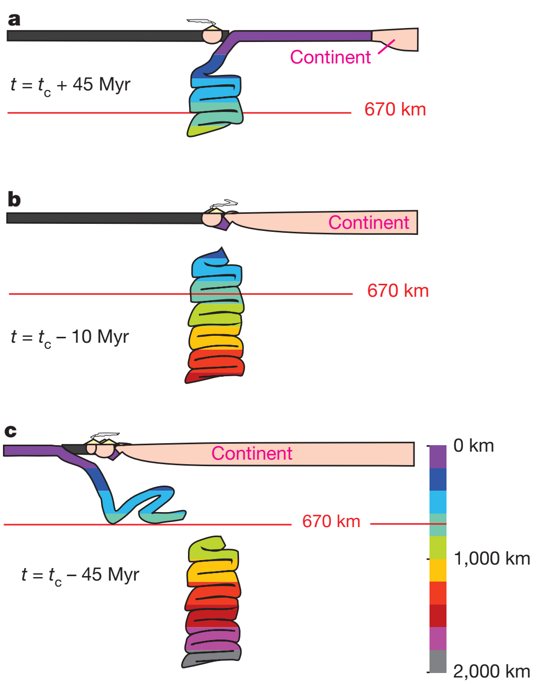

Figure 2 | Schematic cross-section and evolution of a terrane station. t_c denotes the time of arc-continent collision. Motions are shown in a lower-mantle reference frame. **a**, Well before the collision, both trench and arc are active. Slab buckling is due to the viscosity contrast around 670 km, but the backlog reaches into the upper mantle. **b**, Around t_c and up to about 10 Myr later, the continent overrides the trench and accretes its arc terranes, while the slab breaks. **c**, Well after the collision, the slab wall continues to sink. Seaward, a new Andean-style subduction has developed. Anchored in the lower mantle, the slab wall is sinking vertically at a steady-state rate of approximately 10 mm yr^{-1} in all three panels.

To the extent that slabs sink vertically, they record palaeo-arc and trench locations in an absolute sense. Thus, vertical sinking permits quantitative predictions of the location and timing of continent-trench collisions when tomography and absolute plate reconstructions are combined. These predictions can be tested against the docking times of arc terranes inferred from land geology. Abrupt upward truncations of the slab walls, which are well resolved tomographically (Supplementary Fig. 2), correspond to the shutdown of the overlying trench-arc systems, and hence to docking times (Fig. 2).

If the trench remains stationary, a vertical slab pile is deposited beneath it. If the trench moves (but every parcel of slab sinks vertically), the imaged slab will dip towards older trench locations, assuming no dramatic lateral variations in sinking rate. The observed lower-mantle walls are widened to 400–600 km laterally, that is, 4–6 times the thickness of oceanic lithosphere—this is not artificial blur, but the actual reason for their robust tomographic visibility[18]. Thickening is probably achieved by slab folding above the 670-km viscosity jump, deviations from vertical sinking being due mainly to the folding process itself (Fig. 2). In convection models, slab folding occurs preferentially beneath the kind of stationary trenches postulated here[18–20]. Massive, thickened slabs like these can be expected to be the drivers of 'mantle wind', rather than blowing in it: that is, if anything sinks vertically, it should be these slabs.

Such slab walls indicate that their overlying trenches remained in the same absolute locations for a long time, with arc and accretionary complex growth stationed above these locations. Observation of massive slab walls leads us to think of their associated, intra-oceanic trenches as 'terrane stations' where new crustal material is gathered to await transfer to a continental margin. Terrane stations above ANG and MEZ were not conveyed eastward into a continental Farallon trench. Rather, North America migrated westward, collided, and accreted the ANG and MEZ terrane stations. Hence slab walls tie the now-displaced terranes to a laterally unchanged subsurface, constraining absolute locations and temporal evolution of oceanic trenches more than a hundred million years after their demise.

Figure 1 shows the reconstructed western margin of North America[21], together with schematic outlines of the lower-mantle slab walls from tomography[13]. North America's relative westward motion is well constrained by the Atlantic spreading record[2,22], independently of any absolute reference frame. Comparison to the basic geometry of ANG and MEZ suggests that these two slabs did not subduct beneath the continental margin: (1) the slabs are vertical from 800 to 2,000 km depth, indicating a stationary trench, whereas the margin moved westward continuously. (2) The outlines of ANG and MEZ, especially the pronounced eastward promontory of MEZ, do not match the outlines of the continental margin (Fig. 1). If continental subduction had controlled slab deposition, then slab curvature should reflect the curvature of the continent. (3) West of ANG/MEZ, the slab is smeared out laterally in the upper 800 km (Supplementary Fig. 1), as might be expected from a trench dragged along by a migrating continent: direct observational evidence for a switch in subduction mode after override, from stationary oceanic to migrating continental.

Quantitative prediction of arc accretion

Continent motions in Figs 1 and 3 are tied to an absolute hotspot reference frame[21] and rendered by the palaeo-geographic information system GPlates[23,24]. Like the vertically sinking slab walls, vertically rising plumes are thought not to have significant lateral motions relative to the lowermost mantle[25] (smaller deviations are correctable[21]), so that the hotspot and 'slab wall' reference frames are equivalent (no relative motion). In this merged reference frame, the lateral overlay of North America's reconstructed western margin with a slab wall amounts to a spatiotemporal prediction of trench override and terrane accretion.

For example, Fig. 1a shows the North American margin overriding point A1, eastern promontory of the MEZ arc, sometime before 140 Myr, probably around 150 Myr ago. Slab sinking velocities can be inferred: the shallow end of MEZ beneath A1 is seen to have sunk to a depth of approximately 1,500 km, implying an averaged sinking rate of 10 mm yr^{-1}. The MEZ promontory shallows to the southwest. This is unrelated to trench polarity (the slab did not dip northeastward), but rather reflects differential sinking times (subduction at A1 was choked off earlier than at A2). Sinking rates could be estimated from any well-resolved point on the upward truncation of a slab wall, but we choose five points A1–A5 that are associated with supporting evidence from land geology (Table 1 and Fig. 3). Predicted override ages are Jurassic–Cretaceous (146–55 Myr ago), becoming progressively younger westward, and truncation depths shallow to the west as expected. Sinking rate estimates range between 9 and 12 mm yr^{-1} (± 2 mm yr^{-1}), consistent with findings of 12 ± 3 mm yr^{-1} globally[12]. Figure 3 renders the override sequence in four time slices, each showing only slabs that should already have been deposited at the time, assuming the sinking rate was 10 mm yr^{-1} (the average of A1–A5 in Table 1).

The CR must be a Farallon slab because the Farallon (Juan de Fuca) plate is still subducting into it today[15]. Pacific seafloor records indicate continuous Farallon spreading since about 180 Myr ago[3,26], so that at a sinking rate of 10 mm yr^{-1}, the over-1,800-km-deep CR accounts for the entire lifetime of the (northern) Farallon plate. The presence of this CR slab implies that the equally deep and thickened ANG slab further east cannot represent Farallon lithosphere, as has been assumed[9,12,13,27,28]. Rather, the ANG slab must have dipped in the opposite direction (southwestward) in order to have sourced sufficient plate material from an ocean basin that lay to the northeast, the consumption of which accommodated the westward drift of North America. This scenario for transporting North America away from the former Pangaea provides an alternative to westward rollback of a continental Farallon trench.

Hence our inferred trench/plate evolution in Fig. 3 differs fundamentally from the commonly accepted scenario of MEZ/ANG as products of east-dipping, Farallon-beneath-continent subduction. Westward subduction of the ANG and MEZ slabs, probably since early Jurassic times, consumed the ocean that bounded North America on the west (the ANG and MEZ basins in Fig. 3). Both basins were consumed in a zipper-like fashion: MEZ closed from north to south, ANG from south to north, as North America gradually overrode and shut down the ANG/MEZ arcs between 150 and 50 Myr ago. Further west, the early Farallon Ocean subducted into two east–west-oriented Cascadia slabs (CR/CR2), but established additional segments SF1/SF2 after a clockwise rotation at about 147 Myr ago[22,26]. Thus at northerly latitudes, two long-lived terrane stations of opposite polarity coexisted in ANG and CR, a variant not considered in Fig. 2. Upon override, the east-dipping Farallon trenches started rolling back with North America. Moderate complexities in the Pacific–Farallon spreading record[3,26] probably reflect the transitions of individual trench segments from intra-oceanic to Andean-style.

Supporting evidence from land geology

We now use the terrane-station property of oceanic trench/arc systems to test archipelago override predictions made by tomography and plate reconstructions. The collision of North America with buoyant arc crust should coincide with observed deformation and accretion events.

Figure 3b shows inferred terrane locations before override: each active trench/arc system may include a subduction complex or exotic fragments. Using geological relationships in the present-day Cordillera[1], we can match most hypothesized terranes with actual ones. ANG terranes (red) now make up the interior of Alaska, in fault contact with the Angayucham and related ophiolites. Those studying Alaska have long inferred a southwest-dipping subduction[29].

Green terranes west of A1 represent the Franciscan subduction complex of present-day California. Two superterranes from earlier subduction had already loosely accreted to North America before archipelago override began: the IMS closest to the continent[4,5], and the GUE to the south[6], whereas the INS superterrane[4] probably provided the subduction nucleation for the MEZ arc.

To provide for independent validation, our calibration points for sinking rates were chosen at tectonic events that are sharply defined in time and space (A1–A5 in Table 1). Events B1–B5, which are interleaved with the A1–A5 events, represent widespread Cordilleran orogenic and accretion episodes, demonstrating explanatory power on a continental scale. Four stages of override are distinguished, as follows.

Stage 1 (see Fig. 3a, b) is the beginning of the override of the east-verging MEZ promontory. Deformation was initially localized to the Pacific Northwest, as predicted by our model. Incipient deformation of the hinterland generated molasse that flooded the continental platform about 157 Myr ago between 45 and 55° N (ref. 30). A flip in subduction direction at about 165 Myr ago[31] is recorded by the transition from proto-Franciscan formation (for example, the Red Ant formation[32], shown as orange terranes southwest of A1 in Fig. 3b) to Franciscan formation (green), marking an early subduction handover from MEZ to SF2.

Stage 2 (Figs 3c, d) is the time of margin-wide orogenies as North America collided with an increasingly wide swath of MEZ/SF. This caused the Sevier and Canadian Rocky Mountains orogenesis since around 125 Myr ago. Inboard parts of IMS were partly constructed on top of stable North American crust in southern California[33] and had largely collapsed by 110 Myr ago, increasingly shedding zircons onto stable North America[34], and vice versa; whereas IMS and the active Sierra Nevada arc shed zircons into the Franciscan trench[33] (SF2). Intrusion of Omenica magmatic belts successively eastward into northern IMS and adjacent displaced North American strata[8] (B1, about 124–90 Myr ago) can be attributed to prolonged override of the MEZ promontory.

Stage 3 (Fig. 3e, f) is when North America entered the Farallon hemisphere. As ANG collided obliquely, its terranes (red; now interior Alaska) were accreted along the Canadian margin. Override of A3

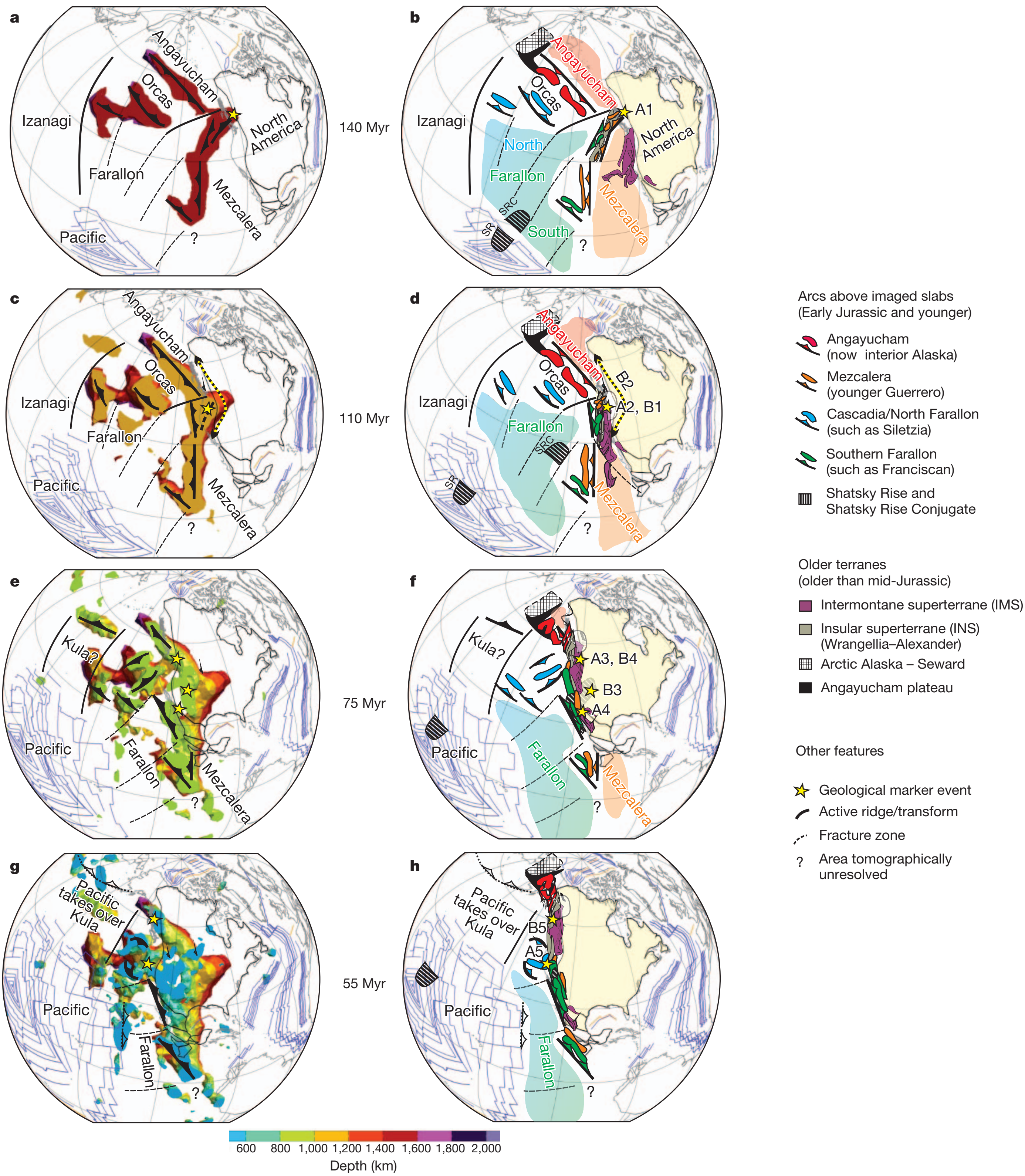

Figure 3 | Sequence of trench overrides and terrane accretions. The left column (**a**, **c**, **e**, **g**) shows time-depth slices at $t = 140$, 110, 75 and 55 Myr ago; the tomography model and plate reconstructions are rendered as in Fig. 1. Each slice shows only material that should have been deposited by that time, that is, slab at and below a depth of $v \times t$, where $v = 10\,\mathrm{mm\,yr^{-1}}$ is the assumed sinking rate. All slabs are coloured according to their current depths, but mentally their upper truncations should be migrated up to the surface, representing the inferred active arc locations at each time. The 140-Myr-ago slice renders slab below 1,400 km depth; the 110-Myr-ago slice renders slab below 1,100 km; the 75-Myr-ago slice below 750 km; and the 55-Myr-ago slice below 550 km. Blue lines are preserved seafloor isochrons. The associated maps in the right column (**b**, **d**, **f**, **h**) are interpretative cartoons showing the evolution of inferred trench and terrane geometries. Yellow stars mark the tectonic events of Table 1. SR, Shatsky Rise; SRC, Shatsky Rise Conjugate.

Table 1 | Sequence of archipelago override

Event	Geometric/kinematic event description	Matched geological validation event	Reconstructed time, *t* (Myr ago)	Slab depth, *d* (km)	Slab sinking velocity, *v* (mm yr^{-1})
A1	Start override of MEZ promontory. Overridden segment is replaced by incipient South Farallon trench SF2.	Initiation of Rocky Mountain deformation, recorded by synorogenic clastic wedge (160–155 Myr ago). Initiation of Franciscan subduction complex/South Farallon (165–155 Myr ago).	146 ± 24	1,500 ± 100	10 ± 2
B1	Gradual override of MEZ promontory by North America (Pacific Northwest).	Omenica magmatic belts in Pacific Northwest (124–90 Myr ago).	-	-	-
A2	End override of MEZ promontory (shallowest point of slab wall).	-	111 ± 8	1,050 ± 50	9 ± 1
B2	Widening collision of North America margin with archipelago (MEZ/ANG/SF arcs).	Margin-wide strong deformation: Sevier and Canadian Rocky Mountains (since about 125 Myr ago).	-	-	-
A3	Override of ANG arc, followed by slab window.	Carmacks volcanic episode due to slab window (72–69 Myr ago).	74 ± 7	850 ± 50	12 ± 1
A4	South Farallon trench steps westward after accretion of Shatsky Rise Conjugate plateau.	Sonora volcanism due to slab window: Tarahumara ignimbrite province (85 ± 5 Myr ago).	88 ± 3	800 ± 50	9 ± 1
B3	Strong transpressive coupling of Farallon plate to superterranes as buoyant Shatsky Rise	Laramide orogeny, basement uplift more than 1,000 km inland (85–55 Myr ago).	-	-	-
B4	Conjugate subducts.	Northward shuffle of INS, IMS, and ANG terranes along margin (85–55 Myr ago).	-	-	-
A5	Override of CR arc by Pacific Northwest USA	Last terrane accretions: Siletzia, Pacific Rim (55–50 Myr ago).	55 ± 7	600 ± 30	12 ± 2
B5	Final override of westernmost ANG.	Explosive end of Coast Mountain arc volcanism (55–50 Myr ago).	-	-	-

Column 2 describes tectonic events in terms of geometrically predicted slab–margin interactions. Column 3 describes matching events from the land geological record. For well-localized events (A1–A5), we estimate slab depth *d*, time *t* since last subduction (margin override), and slab sinking rate *v*, from the tomographic and plate models. Calculations are explained in the Methods; for uncertainty analysis see the Supplementary Information. Geologically observed timing does not enter the calculations, since the role of the geological events is to validate the geometrically inferred results. Events B1–B5 represent additional, first-order tectonic episodes explained by the scenario of archipelago override (these are not localized enough spatiotemporally to estimate *d, t* and *v*).

was accompanied by a strong pulse of intermediate to basaltic volcanism, the Carmacks formation at around 72–69 Myr ago[35]. Such high-temperature, mainly primitive volcanism arises from juxtaposition of hot, sub-slab asthenosphere after a slab, broken through arc collision (Fig. 2b), has started to sink[36].

The South Farallon trench was migrating westward with North America, still building the Fransciscan formation. Around A4, slab geometry indicates an outboard step from SF2 to SF3. This coincides with a strong regional pulse of ignimbrite volcanism at 90 Myr ago (Tarahumara formation, up to 4 km thick and 400 km inland[37,38]) as the southern California/Sonora margin traverses the intermittent slab gap.

The location and timing of this trench step-back coincide almost perfectly with the inferred arrival of a buoyant oceanic Farallon plateau at the North America margin, the conjugate half of the Shatsky Rise (Fig. 3 includes its reconstruction by ref. 39). Plateau collision is a proposed mechanism[39,40] for choking subduction and causing the basement uplifts of the Laramide orogeny at around 85–55 Myr ago (B3). We suggest that the event explains another first-order observation of Cordilleran geology (B4): the INS/IMS and ANG terrane packages, but not Franciscan and GUE, were rapidly shuffled northward along the margin between 85 and 55 Myr ago, by many hundreds to over 2,000 km (ref. 41). The convergence vector of the Farallon plate did have a large northward component at the time[2], but terrane transport additionally requires strong coupling to the Farallon plate and decoupling from North America. A buoyant Farallon plateau, compressed against INS/IMS and unable to subduct, could have achieved such coupling until the trench re-established itself further west.

Stage 4 (Fig. 3g, h) is the end of archipelago override. North America overrides point A5 at 55 ± 7 Myr ago, in excellent agreement with last observed terrane accretions in the Pacific Northwest (Siletzia, Metchosin and Pacific Rim terranes at about 55–50 Myr ago[42–44]). The trench stepped west (clear upward truncation at A5) as the terranes accreted, converting intra-oceanic CR into today's continental Cascadia subduction. Also at about 55–50 Myr ago, ANG was terminally overridden (B5), accompanied by explosive volcanism as the Coast Mountain arc of British Columbia shut down.

By then, slab complexity in the upper mantle rivalled today's western Pacific—not surprising given the numerous forced reorganizations. The simple depth–age relationship suggested by Fig. 3 need not apply to flat-lying Cenozoic slabs[13] or to isolated deeper fragments like Kula or CR3, because all sinking estimates were calibrated on slab walls. We do not attempt to interpret upper-mantle structure and the Cenozoic land record here, but our archipelago model provides a new framework for doing so.

Better constraints on surface and mantle

Long-lived, stationary oceanic trenches explain two problematic observations, previously thought to be unconnected: the near-vertical geometries of the super-slabs under North America (without invoking exceptional mantle rheologies or *ad hoc* shifts in absolute reference frame), and the long series of arc terrane accretions during Cretaceous times[1] (not explained by the South American analogy).

An archipelago offshore Mesozoic North America had previously been suggested on the basis of land geology[7,8], but lacked the absolute spatial constraints provided by seismic tomography and our vertical sinking/terrane stations concept. Now-displaced terranes can be tied to their original, seismically imaged trench locations, but also to reconstructed continent locations, in an absolute reference frame. The hypothesis that continental collision with stationed terranes caused the various episodes of Cordilleran mountain building becomes testable.

Before 150 Myr ago, North America was clearly located too far east for MEZ/ANG to pass for continental trenches, even when longitudinal uncertainties of absolute reference frames[45] are factored in (Fig. 1 and Supplementary Fig. 3, and Supplementary Tables 1 and 2). The apparently 'wrong' geometry and locations of the vertical slab walls—under the hypothesis of subduction of the Farallon plate beneath North America—had been recognized[12,27,46,47] with two kinds of solutions suggested. A longitudinal shift of the global lithospheric shell relative to the lower mantle[12], specifically a Cretaceous westward excursion that tapered down as the Atlantic opened, could have held North America stationary above MEZ/ANG. Alternatively, upper-mantle slab, spread out laterally by a west-coast trench, must somehow have aggregated into steep piles when transitioning into the lower mantle. This requires lateral displacements by over 1,000 km of huge volumes of slab[27,46]. Some convection simulations have produced such behaviour[46,48], whereas others suggest essentially vertical sinking[47], as do newer observations[12,13,18].

We showed that, within observational uncertainty limits, predicted and observed geological events are consistent. This validates simple vertical sinking and seems to explain all North American observations, including accreted terranes, but is incompatible with the widely accepted continental Farallon trench since 175 Myr ago or before. The observed proportionality between slab depth and time since override (consistent sinking rates across all three slab walls) is not required for our argument, but rather increases confidence in its correctness.

Vertical slab sinking provides much tighter constraints on palaeogeographic reconstructions than arbitrarily movable slabs. This scenario follows the principle of parsimony, so that future investigations should start with it, and seek observations requiring a departure from it. The equivalence of hotspot and "slab wall" absolute reference frames, which is implied by vertical sinking, is of great interest because slabs reach back farther in time than hotspot tracks (200 Myr ago or more[12,49] versus about 130 Myr ago[21,22]), and they constrain absolute palaeo-longitude, which palaeomagnetic data alone cannot[50]. However, to quantitatively realize a global subduction reference frame[12], it will be necessary to re-examine whether trenches commonly assumed to have been continental were not actually intra-oceanic.

METHODS SUMMARY

We postulate that subduction into all slab walls imaged tomographically beneath North America[13] originated before the arrival of North America's western margin. Hence we must demonstrate sufficiently old slab ages, and cessation or flipping of subduction when the continent overrode the slabs. Plate reconstructions predict the timing of margin arrival above a slab, but only if slab[13] and plate reconstructions[12,21,22,50–52] can be linked to the same absolute reference frame. Hence override predictions are correct to the extent that slab walls sank vertically, meaning that their x–y locations since subduction are unchanged in a linked hotspot reference frame. Uncertainty is best quantified from slab wall geometry itself: deviation from vertical sinking probably did not exceed a wall's half-width (200–300 km), else such steep geometries could not have built up over a long time.

Uncertainties about absolute locations of North America's palaeo-margin arise from imperfections in plate reconstructions. Owing to terrane accretions and removals, there is also uncertainty about the shape and westward extent of North America's margin, compared to its present-day outlines. We discuss individual uncertainties in the Supplementary Information, and propagate them into cumulative uncertainties for the times at which North America's palaeo-margin overrode selected points A1–A5 on the palaeo-trenches. With relative uncertainties of only 10–15% (Table 1 and Supplementary Tables 1 and 2), this yields the old (Jura–Cretaceous) slab ages required to support intra-oceanic subduction.

Spatiotemporal predictions of trench override are verified by terrane observations: override should coincide with observable collision events, because buoyant island arcs or plateaus are overridden. Uncertainties on terrane observations are difficult to quantify, but particularly characteristic events (A1–A5) can nonetheless be singled out and used successfully for validation (Table 1). Clear upward truncations of all slab walls offer direct observational evidence for continental override of oceanic trenches, and are used to calculate slab-wall sinking rates.

Full Methods and any associated references are available in the online version of the paper.

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Supplementary Information is available in the online version of the paper.

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Author Contributions K.S. generated the tomographic model, and integrated it with quantitative plate tectonic reconstructions in GPlates. M.G.M. provided the geological background and made the terrane maps of Fig. 3. Both authors contributed equally to developing the tectonic arguments and to the writing.

Author Information Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Readers are welcome to comment on the online version of the paper. Correspondence and requests for materials should be addressed to K.S. (sigloch@geophysik.uni-muenchen.de) or M.G.M. (Mitch.Mihalynuk@gov.bc.ca).

METHODS

The MEZ/ANG slab walls have been among the most robust features in global-scale body-wave tomography, starting with the work of Grand[9–13,28]. The deep end of the CR slab was already visible in some of the earlier studies. Its continuous upward connection to present-day Cascadia subduction was pointed out by ref. 15, hence its identification as a Farallon slab. The model on which we base our discussion here[13] is an inversion of P-wave observations recorded by North American broadband stations, using a cutting-edge waveform inversion technique (multi-frequency tomography) on a global, adaptive grid. Method discussion and formal resolution tests are presented in ref. 13. The higher resolution compared to global tomography models is largely due to densely spaced stations from the USArray experiment in the western half of the US, and waveforms recorded 2005–2008, which were not included in any of the above global models. Our calibration points for sinking rate, especially A3–A5, lie within the mantle subvolume that considerably benefits from resolution improvements afforded by the new USArray data.

We postulate that subduction into all slab walls imaged tomographically beneath North America[13] originated before the arrival of North America's western margin. Hence we must demonstrate sufficiently old slab ages, and cessation or flipping of subduction when the continent overrode the slabs. Plate reconstructions predict the timing of margin arrival above a slab, but only if slab and plate reconstructions[12,21,22,50–52] can be linked to the same absolute reference frame. (Technically we accomplish this with the free community palaeo-geographic information system GPlates[23,24] and a compilation of digitally published reference frames[45].) Hence override predictions are correct to the extent that slab walls sank vertically, meaning that their x–y locations since subduction are unchanged in a linked hotspot reference frame. Uncertainty is best quantified from slab wall geometry itself: deviation from vertical sinking probably did not exceed a wall's half-width (200–300 km), else such steep geometries could not have built up over a long time.

Uncertainties about absolute locations of North America's palaeo-margin arise from imperfections in plate reconstructions. Owing to terrane accretions and removals, there is also uncertainty about the shape and westward extent of North America's margin, compared to its present-day outlines. We discuss individual uncertainties in the Supplementary Information, and propagate them into cumulative uncertainties for the times at which North America's palaeo-margin overrode selected points A1–A5 on the palaeo-trenches. With relative uncertainties of only 10–15% (Table 1 and Supplementary Tables 1 and 2), this yields the old (Jura–Cretaceous) slab ages required to support intra-oceanic subduction.

Spatiotemporal predictions of trench override are verified by terrane observations as follows. Override should coincide with observable collision events, because buoyant island arcs or plateaus are overridden. Uncertainties on terrane observations are difficult to quantify, but particularly characteristic events (A1–A5) can nonetheless be singled out and used successfully for validation (Table 1). Sinking-rate calculations provide an additional plausibility check. Consistent results of 9–12 mm yr^{-1} ($\pm$1–2 mm yr^{-1}) across all three slab walls show that this type of feature seems to sink rather predictably and evenly. If this is the case, then conversely the override prediction times that lead to the rate estimates should be adequate.

Sinking at about 10 mm yr^{-1} is considered a lower-mantle sinking rate, but we obtain it as an average over both upper and lower mantle. This is explicable if a wall were to have sunk in steady state: upon subduction of its youngest end (Fig. 2b), its lower part would already have entered the viscous lower mantle and would have been setting the speed limit from below, which would have acted on the entire pile. Such a wall, widened to 4–6 times the lithospheric thickness (400–600 km), and sinking at about 10 mm yr^{-1}, generates the same material throughput as the typical 40–60 mm yr^{-1} of unbuckled plate convergence in the uppermost mantle[16,20]—a plausibility check that confirms the continuity of upper and lower mantle fluxes.

Sinking rates are obtained by dividing the imaged depth of a wall's shallowest end (depth reached since the end of subduction) by predicted time since the trench override. Besides timing uncertainty, a spatial uncertainty about the slab's true depth enters, as discussed quantitatively in Supplementary Fig. 2. These uncertainties are comparatively small, because upward truncations of the slab walls are sharply imaged. This is also a striking visual feature (see, for example, Fig. 3g or Supplementary Fig. 2): along their lengths, MEZ and ANG show abrupt upward truncations in red to orange, yellow and green depth levels. This provides direct observational evidence for westward subduction—after trench override, there was no slab left to subduct. In contrast, the less-complete upward truncation of eastward-dipping CR around A5 (cyan colour level, a more localized slab window) indicates 'only' a larger terrane accretion event and trench step-back when the margin transitioned from intra-oceanic to Andean-style. Upward truncations shallow to the west (A1 at red level, A2 at yellow level, A3/A4 at green level, and A5 at cyan level), reflecting more and more recent ages for termination of subduction.

In summary, the observational uncertainties that we discuss in the Supplementary Information are:

(1) Uncertainties in plate reconstructions at the surface. When exactly did the North American palaeo-margin overlie a given point in the reference frame, for example, A1-A5? This includes relative reconstruction uncertainties (essentially ambiguities about the Atlantic Ocean opening; they are small, and neglected in our calculations); uncertainties about absolute reference frame (these are considerable; we attempt to quantify them by comparing different reference frames[12,21,22,45,50–52] (Supplementary Fig. 3); and uncertainties about the true westward extent of the palaeo-margin over time. The latter uncertainty is most difficult to quantify, since it requires knowledge about accreted terrane locations, which shifted over time. Our best guess of the uncertain area's extent is hatched in Fig. 1a and b and Supplementary Fig. 3a and b. Margin uncertainty is the biggest contributor to reconstruction uncertainties (typically 6–7 Myr), except for the oldest point A1, where absolute reference frame uncertainty dominates (Supplementary Table 2).

(2) Uncertainties in palaeo-trench locations, relative to the slab. Were trench points A1–A5 centred on the imaged slab walls, systematically offset to one side, or oscillating? We assume that the trenches ran centred, for lack of evidence to the contrary. A constant offset would hardly change the calculations, producing a strong correlation with margin uncertainty rather than independent uncertainties. Periodic trench advance and retreat—as buckling folds are being laid down—cannot be excluded, but we are unaware of observational evidence. Unless these oscillations significantly exceeded the half-width of the slab wall of 200–300 km (unlikely, given the tall slab piles), they would not dominate over margin uncertainty.

(3) Uncertainties about present-day slab depth (pertinent only to sinking-rate estimates). Evaluation of tomographic blur yields significantly smaller relative uncertainties than reconstruction errors (Supplementary Table 1). The two types of relative errors enter symmetrically into cumulative uncertainty on sinking rate.

Regarding our hypothesis of intra-oceanic trenches, the most important, qualitative assessment of uncertainty is that all the plate reconstructions considered[21,22,45,50–52] (with one exception[12] discussed below) agree that two ocean basins should have existed between North America and the MEZ/ANG slabs before 140 Myr ago (Fig. 1a and Supplementary Fig. 3a). These oceans were considerably wider than the uncertainty in margin extent, implying intra-oceanic subduction origins for all imaged slab walls, provided they are older than 140 Myr. That this is the case is shown by good agreement between predicted and geologically observed collision events (within the moderate error bars of Table 1). An extrapolation of sinking rates of about 10 mm yr^{-1} to the lower ends of the slab walls at over 1,800 km depth implies that subduction originated at least 180 Myr ago.

The one reference frame[12] that does not predict wide ANG/MEZ ocean basins in the late Jurassic/early Cretaceous was explicitly designed to minimize their extents, by imposing a constraint that keeps North America's western margin stationary above the MEZ/ANG slabs (that is, an Andean-style margin is enforced a priori; SUB on Supplementary Fig. 3b). This is accomplished by introducing an additional degree of freedom, an *ad hoc*, otherwise non-observable, westward shift of the lithospheric shell relative to the lower mantle. In solving for this longitudinal shift, the method considers a global slab inventory, but for the times discussed here, the influence of the MEZ/ANG slabs dominates. Hence, this reference frame does not lend itself to evaluating the existence of the ANG/MEZ oceans.

Supplementary Information for Sigloch & Mihalynuk 2013

“Intra-oceanic subduction shaped the assembly of Cordilleran North America”

- **Three supplementary figures (S1, S2, S3)**
- **Discussion of data uncertainties and error propagation**

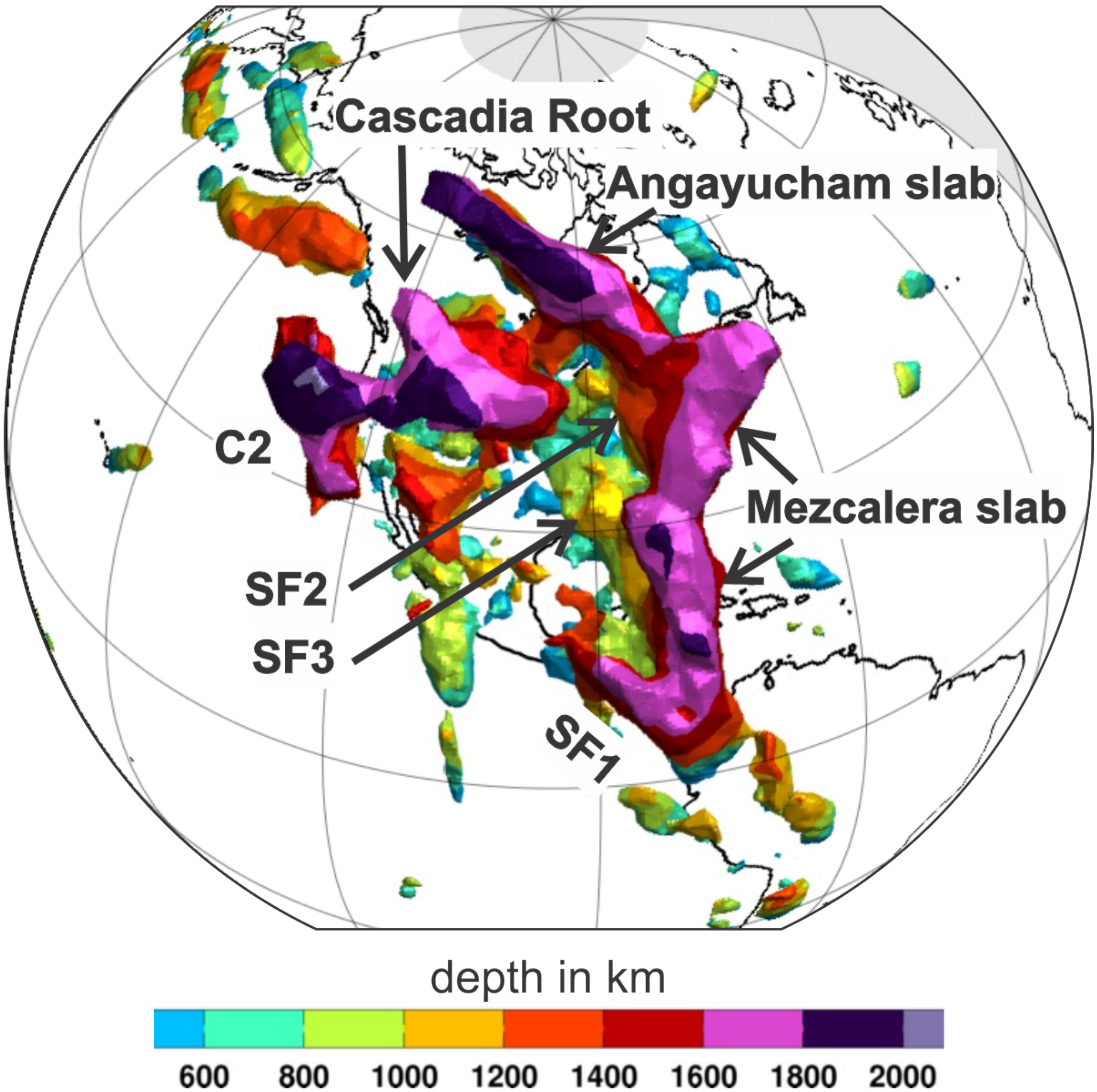
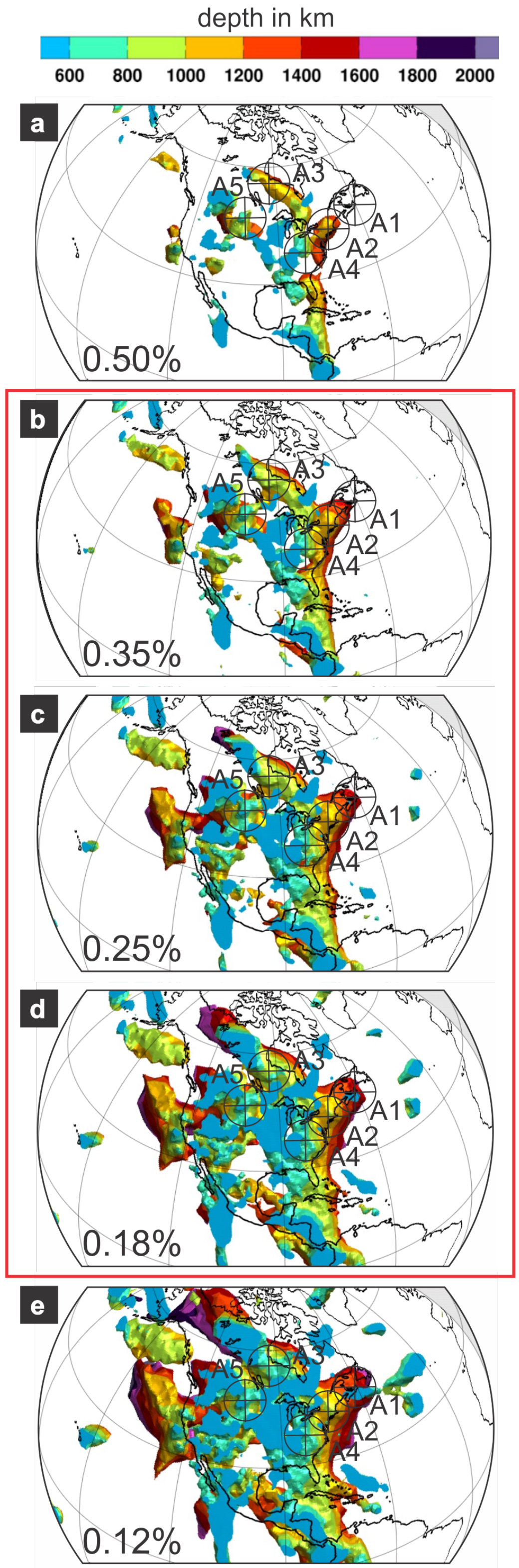

Cascadia Root
Angayucham slab
C2
Mezcalera slab
SF2
SF3
SF1
depth in km
600
800
1000
1200
1400
1600
1800
2000

Figure S1



depth in km
600
800
1000
1200
1400
1600
1800
2000
a
A3
A5
A1
A2
A4
0.50%
b
A3
A5
A1
A2
A4
0.35%
c
A3
A5
A1
A2
A4
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d
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A4
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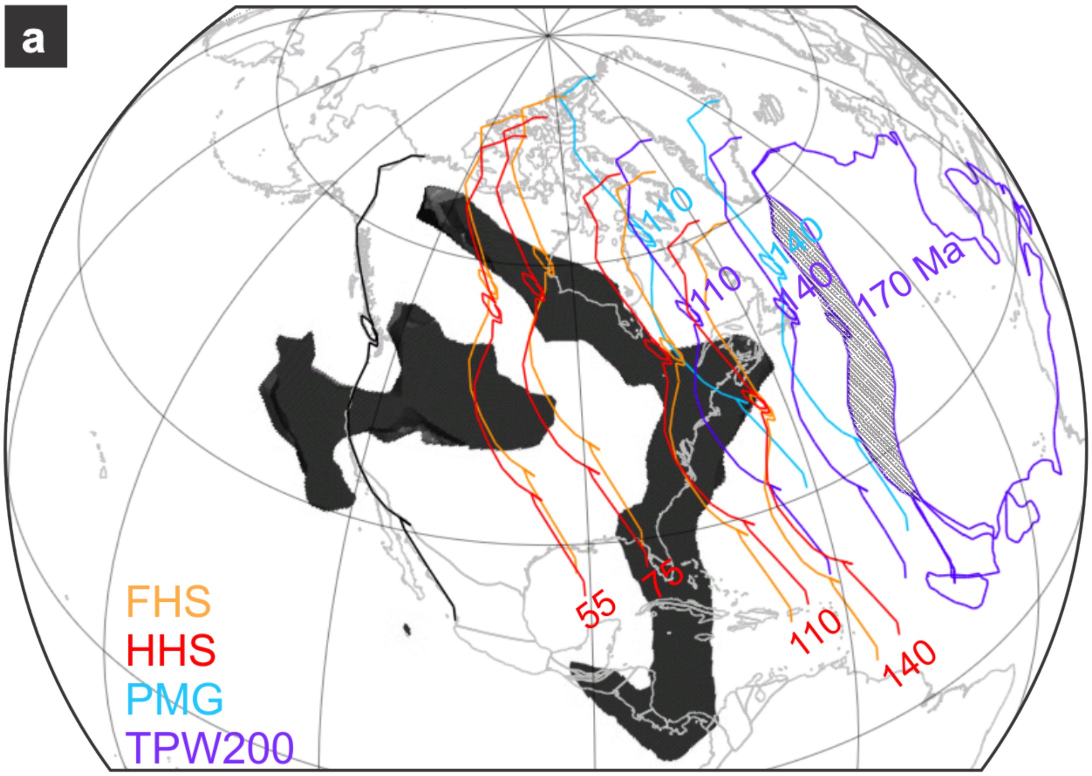
Figure S2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140
110
140
170 Ma
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75
110
140
FHS
HHS
PMG
TPW200

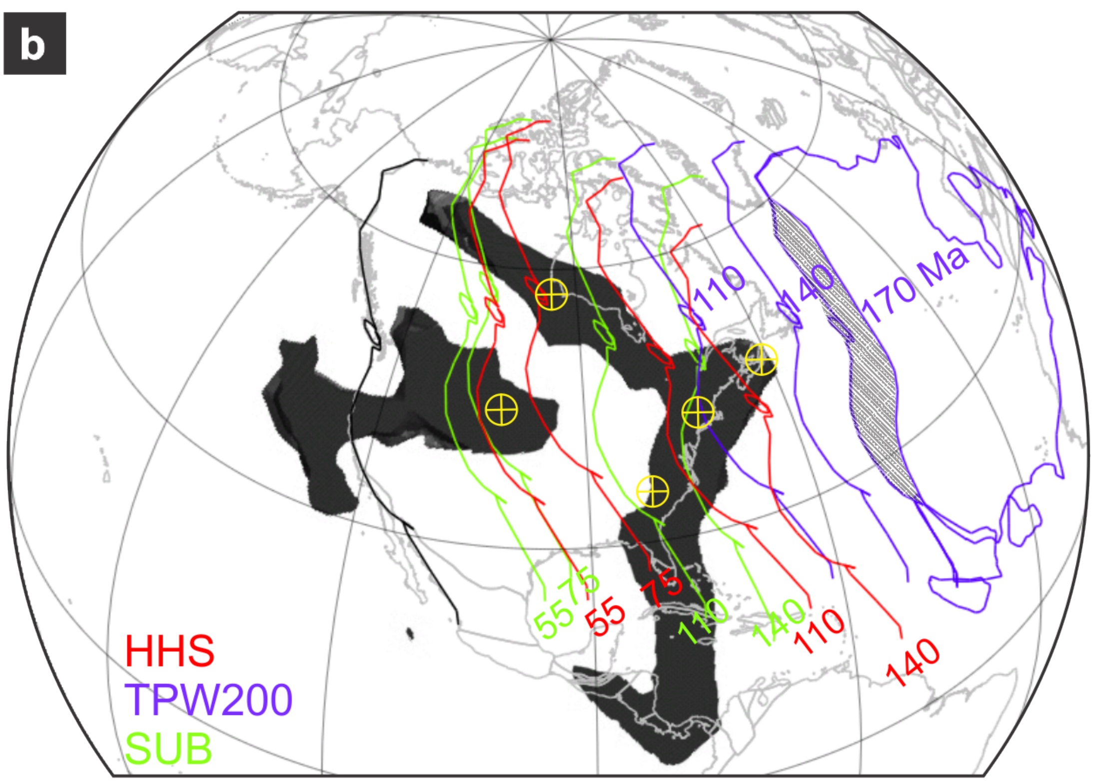
b
110
140
170 Ma
55
75
55
75
110
140
110
140
HHS
TPW200
SUB

Figure S3

Legends of Supplementary Figures S1-S3

Figure S1. Vertical slab walls ─ inside-out view of seismically fast structure under North America. This 3-D rendering contours fast structure in the P-velocity tomography model by Sigloch 2011 [13], at the same isosurface threshold of dVp/Vp=0.25% as used in figures 1 and 3. It is an "inside-out" view, which moves the deepest structure to the foreground (pink/purple shades at 1800-2000 km), and the shallowest levels to the background (blue, 600-400 km). The perspective corresponds to that of an imaginary observer sitting at the center of the earth, looking upward (except that such a viewpoint would also flip east and west, omitted here). Structure shallower than 400 km is not rendered, since it may represent fast cratonic lithosphere rather than subducted slabs. This view shows most clearly the almost vertical geometry of the deep slab walls below ~800 km depth, the segmentation of the Mezcalera and Angayucham walls, and their clear spatial separation from the crescent-shaped Cascadia Root and slab C2 further west. By contrast, material in the transition zone is smeared out laterally (yellow, green, blue shades). The vertical walls carry the geometric signature of intra-oceanic trenches, which can and do remain stationary over long periods, whereas the shallower slabs were deposited into a trench dragged along by the migrating continent.

Figure S2. Estimation of uncertainties in slab depth. Panels a to e render the same P-wave model [13] at different contouring thresholds dVp/Vp: +0.50%, +0.35%, +0.25%, +0.18%, and +0.12% (the incremental factor is always $\sqrt{2}$). Only fast structure at and below 600 km is shown. The preferred threshold is dVp/Vp=0.25%, used in panel c and in all other figures, and assumed to yield the most probable values for slab depth d. Panels b and d are used to obtain one-sigma error bounds σ_d. The five crosshairs mark calibration points A1-A5. Unit on color bars is depth in km.

Figure S3. Comparison of North America´s westward migration in different absolute reference frames. a: Colored lines show reconstructed western margin over time, in four absolute reference frames. Orange: fixed hotspot FHS [51]; red: hybrid moving hotspot HHS [21]; cyan: paleomagnetic, PMG [50]; purple: true polar wander, TPW/TPW200 [22,52]. Colored labels give time in Ma for their corresponding reference frames; the times used correspond to the four time slices of figure 3. The PMG and TPW frames differ from HHS only prior to 100 Ma, and are shown only at these earlier times. Irregular, dark grey contours are slab outlines at and below 1500 km (identical to those in figure 1a), which should have been subducted by 140 Ma. Present-day landmasses are shown in light grey. The hatched area along the NA margin at 170 Ma represents the assumed uncertainty about true westward extent of the margin (for all times).

b: Same as panel a, except that the subduction reference frame[12], SUB, is compared to HHS and TPW/TPW200. Yellow crosshairs mark calibration points A1-A5. The construction of SUB is fundamentally different from the other frames: it allows for shifts in paleo-longitude if this maximizes the superposition of reconstructed trench lines with imaged slabs. Since the Cretaceous Farallon trench was *assumed* to run along the continental margin, and since it is mapped onto the MEZ/ANG slabs, this frame enforces the Andean-style subduction scenario. This results in a very westerly position of Cretaceous North America that minimizes the extents of the Mezcalera and Angayucham Oceans.