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The mid-Cretaceous Peninsular Ranges orogeny: a new slant on Cordilleran tectonics? II: northern United States and Canada[1]

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Abstract: The mid-Cretaceous Peninsular Ranges orogeny occurred in the North American Cordillera and affected rocks from Mexico to Alaska. It formed when a marine trough, open for ~35 million years, closed by westerly subduction beneath a 140–100 Ma arc complex. In Part I, we described the features of the orogen in Mexico and California, west to east: back-arc trough, magmatic arc, 140–100 Ma seaway, post-collisional 99–84 Ma granodioritic–tonalitic plutons emplaced into the orogenic hinterland during exhumation, an east-vergent thrust belt, and farther east, a flexural foredeep. In western Nevada, where the Luning–Fencemaker thrust might be a mid-Cretaceous feature, arc and post-collisional plutons occur in proximity. The orogen continues through the Helena salient and Washington Cascades. In British Columbia, rocks of the 130–100 Ma Gambier arc lie west of the exhumed orogenic hinterland and 99–84 Ma post-collisional plutons to collectively indicate westerly subduction. East-dipping reverse faults near Harrison Lake, active from ~100 Ma until ~90 Ma, shed 99–84 Ma debris westward into the Nanaimo back-arc region. Within Insular Alaska, the Early Cretaceous Gravina basinal arc assemblage was deformed at 100 Ma and flanked to the east by a high-grade hinterland cut by post-collisional plutons. In mainland Alaska, the 100 Ma collision of Wrangellia and the Yukon–Tanana–Farewell composite terrane occurred above a southward-dipping subduction zone as shown by the 130–100 Ma Chisana arc sitting on Wrangellia and southward-dipping, northerly vergent thrusts in the Lower Cretaceous Kahiltna basin to the north. The outboard back-arc region was filled with post-collisional detritus of the McHugh complex.

Key words: orogeny, North American Cordillera, arc magmatism, arc-continent collision, slab failure magmatism, Peninsular Ranges orogeny.

Résumé : L'orogenèse des chaînes péninsulaires d'âge crétacé moyen s'est produite dans la cordillère nord-américaine et a touché des roches allant du sud du Mexique à l'Alaska. Elle s'est formée quand une fosse marine, ouverte pendant ~35 millions d'années, s'est refermée par subduction vers l'ouest sous un complexe d'arc de 140–100 Ma. Dans la première partie, nous avons décrit les éléments de l'orogène au Mexique et en Californie qui comprennent, d'ouest et est, une fosse d'arrière-arc, un arc magmatique, un bras de mer de 140–100 Ma, des plutons de granodiorite–tonalite post-collision de 99–84 Ma mis en place dans l'arrière-pays orogénique durant l'exhumation, une ceinture de charriage vers l'est et, plus à l'est, une avant-fosse formée par flexion. Dans l'ouest du Nevada, où le chevauchement de Luning–Fencemaker pourrait être un élément d'âge crétacé moyen, des plutons d'arc et post-collision sont présents à proximité les uns des autres. L'orogène se poursuit par le saillant d'Helena et les montagnes Cascades de l'État de Washington. En Colombie-Britannique, des roches de l'arc de Gambier de 130–100 Ma sont présentes à l'ouest de l'arrière-pays orogénique exhumé et de plutons post-collision de 99–84 Ma, indiquant collectivement une subduction vers l'ouest. Des failles inverses à pendage vers l'est près du lac Harrison, actives de ~100 Ma à ~90 Ma, ont évacué vers l'ouest des débris de 99–84 Ma jusque dans la région de l'arrière-arc de Nanaimo. En Alaska insulaire, l'assemblage d'arc et de bassin de Gravina d'âge crétacé précoce a été déformé à 100 Ma et flanqué à l'est par un arrière-pays de haut degré de métamorphisme recoupé par des plutons post-collision. En Alaska continental, la collision à 100 Ma de la Wrangellie et du terrane composite de Yukon–Tanana–Farewell s'est produite au-dessus d'une zone de subduction à pendage vers le sud, comme l'indique l'arc de Chisana de 130–100 Ma reposant sur la Wrangellie et des chevauchements vers le nord à pendage vers le sud dans le bassin crétacé inférieur de Kahiltna au nord. La région d'arrière-arc externe a été remplie par des détritus post-collision du complexe de McHugh. [Traduit par la Rédaction]

Mots-clés : orogenèse, cordillère nord-américaine, magmatisme d'arc, collision arc-continent, magmatisme de rupture de plaque, orogenèse des chaînes péninsulaires.

Introduction

Ever since the late 1960s, protracted easterly dipping subduction beneath North America has been the standard model to explain the development of the Cordillera (Dickinson 1970). However, in a companion paper (Hildebrand and Whalen 2021, this issue) we demonstrated how the geology of the Peninsular Ranges and Sierra Nevada can be reconciled by a major orogenic event, the Peninsular Ranges orogen, which developed at 100 Ma

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when an east-facing, 130–100 Ma shallow marine arc collided with a west-facing Aptian–Albian passive margin of the North American plate built on the east side of the Bisbee–Arperos seaway. Following closure of the basin by westward-dipping subduction, the oceanic lithosphere, along with much of the rifted margin, broke off from the partially subducted North America plate and fell away into the mantle. This slab failure led to rapid exhumation of the collisional hinterland, which shed post-collisional, 99–84 Ma plutonic and metamorphic debris westward into the back-arc region. The debris was derived in part from a suite of 99–84 Ma mesozonal–catazonal plutons formed when the basaltic–gabbroic portion of the broken and sinking plate melted. Geochemistry and isotopic analyses suggest that pre-collisional arc and post-collisional slab break-off magmas were derived from two different sources at different depths and can be distinguished on a set of discrimination diagrams (for example, Hildebrand et al. 2018).

In Part I (Hildebrand and Whalen 2021, this issue), we showed how 100 Ma easterly vergent thrust faults, interpreted to have formed during the Peninsular Ranges orogeny, were traced north–south through eastern California, Nevada, and Utah. In this paper, we describe the northward continuation of the orogen starting in western Nevada and continuing to south-central Alaska.

Nevada

Rocks and deformation of the Peninsular Ranges orogen are exposed to the north of the Sierra Nevada through western Nevada (Fig. 1), where both arc and slab failure plutons are common, but the subcontinental lithospheric mantle (SCLM) might be different from that beneath the Sierra Nevada (Hildebrand and Whalen 2017). For example, the 93–89 Ma Sahwave intrusive suite is a large post-collisional intrusive complex that is geochemically and petrologically similar in most respects to those in the Sierra Nevada batholith farther south, such as the Tuolumne and Mount Whitney intrusive suites (Supplementary Fig. S1[2]), but has more primitive εNd_T and $^{87}Sr/^{86}Sr_i$ reflecting derivation from younger and less-radiogenic lithosphere (Van Buer and Miller 2010).

In the Santa Rosa Range and Bloody Run Hills of north-central Nevada, both located just west of the Luning–Fencemaker thrust (Fig. 1), Brown et al. (2018) identified plutons of two age groups, 105–101 and 96–93 Ma, and noted that the older plutons had positive εNd_T, ranging from 0.8 to 2.9 and $^{78}Sr/^{86}Sri$ from 0.7045 to 0.7049, whereas plutons of the younger group had εNd_T from –1.5 to −3.2 and $^{78}Sr/^{86}Sr_i$ from 0.7052 to 0.7062. Even though the plutons are close to one another, these data suggest derivation from different sources, with pre-100 Ma intrusions reflecting an arc source and the post-100 Ma plutons indicating assimilation of SCLM. Thus, the lithospheres were juxtaposed at ∼100 Ma just as they were to the south in the Sierra Nevada and Peninsular Ranges. Supporting evidence for the interpretation that the Santa Rosa – Bloody Run Hills now sit atop cratonic lithosphere comes from the 248 ± 1 Ma Koipato volcanics of the Humboldt and adjacent ranges (Fig. 1), which lie to the south and southeast of the Santa Rosa Range in the footwall of the Luning–Fencemaker thrust, and have high initial Sr and negative εNd_T, which Vetz (2011) argued reflected interactions with Paleoproterozoic lithosphere.

The age of the Luning–Fencemaker thrust is poorly constrained because there are no known dikes or other intrusions that cut it and, as it carries mostly Jurassic rocks in its hanging wall, has commonly been assumed to be a Jurassic fault. However, as there are also no known 130–100 Ma arc plutons to the east of the fault, it is possible that the thrust is not Jurassic, but instead a 100 Ma thrust active during the Peninsular Ranges orogeny. Most of the samples collected for $^{40}Ar/^{39}Ar$ ages from Jurassic rocks in the Santa Rosa Range, the Jungo Hills, and the Jackson Range (Fig. 1) yielded plateau ages from 97 to 86 Ma, which are considered to represent regional heating at about 100 Ma (Wyld et al. 2003). Thus, on the basis of arc plutons located just to the west — but not to the east — of the thrust, as well as the $^{40}Ar/^{39}Ar$ plateau ages discussed above, we suggest that the Luning–Fencemaker thrust is a 100 Ma structure, not a Jurassic fault as commonly assumed (Wyld et al. 2003; DeCelles 2004).

Small areas of low-grade sedimentary rocks that fill at least one half graben in the area, are collectively known as the King Lear Formation (Fig. 1) and are dominated by conglomerate and sandstone, but contain an interbedded siliceous ignimbrite dated at 125 ± 1 Ma, as well as a hypabyssal intrusion at 123 ± 1 Ma (Martin et al. 2010). We assume, based on the age, location, and presence of volcanics, that these outcrops represent remnants of early magmatism in the Cinco arc trough.

Idaho

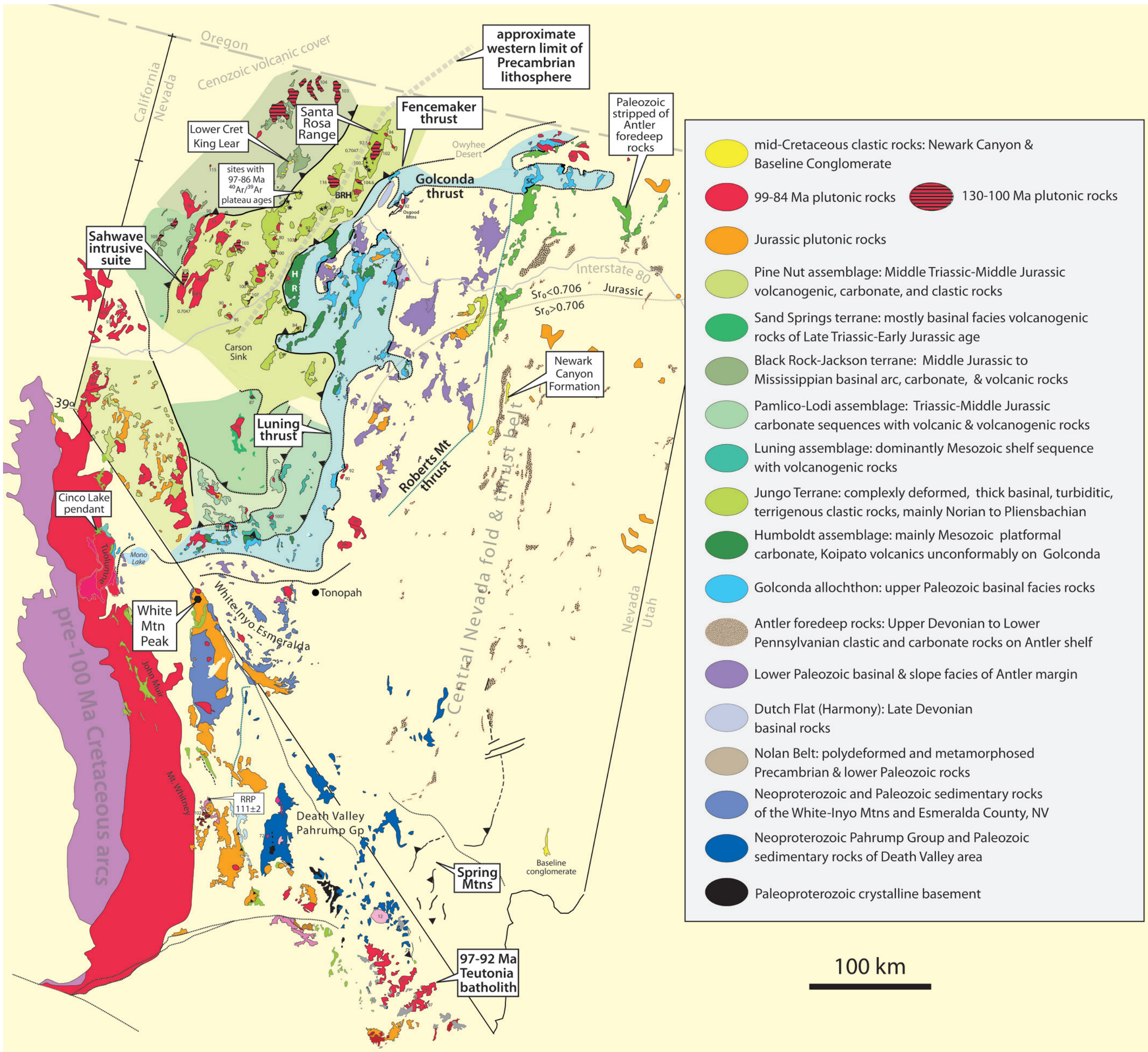
Northward, across the Snake River Plain, several components of the Peninsular Ranges orogen are exposed. Rocks in and adjacent to the Western Idaho shear zone, which lies along the western margin of the Atlanta lobe of the Idaho batholith, are variably deformed, and in places contain deep-seated epidote-bearing, tonalitic to granitic orthogneisses in the age range 118 ± 5 to 105 ± 1.5 Ma, cut by epidote-bearing tonalitic sheets in the age range 92 to 90 Ma (Taubeneck 1971; Hyndman 1983; Manduca et al. 1993; Giorgis et al. 2008). The Western Idaho shear zone (Supplementary Fig. S2[2]) has long been considered as the western margin of cratonic North America based largely on isotopic data (Armstrong et al. 1977; Fleck and Criss 1985, 2007; Criss and Fleck 1987; Fleck 1990; Manduca et al. 1992).

East of the main shear zone, intrusions of the western border zone of the Atlanta lobe of the Idaho batholith, as well as plutons preserved as roof pendants, termed the Early Metaluminous suite, occur within the younger Laramide sector of the batholith and range in age from 98 to 87 Ma (Gaschnig et al. 2010; Kiilsgaard et al. 2001). In addition to the larger bodies, a number of small intrusions in high-grade gneisses of the Sawtooth Range near Stanley, Idaho (Supplementary Fig. S2[2]), yield 95–92 Ma zircon ages (Ma et al. 2017). On the basis of their age, the small bodies appear to be post-collisional bodies of the Peninsular Ranges orogeny, whereas the deformation appears to be related to the Late Cretaceous Laramide orogeny as a post-deformational intrusion was dated to be 77 Ma, about the same age as other post-Laramide intrusions in the area (Hildebrand and Whalen 2017).

Transpressional deformation within the Western Idaho shear zone is older than 90 Ma U–Pb ages of granitic pegmatites that cut the fabric (Giorgis et al. 2008). However, according to Braudy et al. (2017), the shear zone likely initiated after 104 Ma on the basis of 99.5 ± 1.4 Ma and 97.3 ± 0.7 Ma Lu–Hf garnet isochrons, which they interpret to represent the time of peak metamorphism. K/Ar and $^{40}Ar/^{39}Ar$ ages in the area fall in the range 93–85 Ma and partially overlap with the emplacement of the 98–87 Ma metaluminous plutons of the Idaho batholith (Lund and Snee 1988; Manduca et al. 1993; Snee et al. 1995; Giorgis et al. 2008), suggesting that the intrusions were emplaced during exhumation within the hinterland of the Peninsular Ranges orogen. Whole-rock geochemical data of the 98–87 Ma plutonic rocks plotted on our discrimination diagrams classify them as slab failure magmas (Supplementary Fig. S3[2]). Just west of the main mass of the Atlanta lobe (Supplementary Fig. S2[2]), 98–88 Ma foliated epidote–hornblende tonalitic, quartz dioritic, and granodioritic bodies, and gneisses, including the Payette River tonalitic intrusion (91.5 ± 1.1 to 89.7 ± 1.2 Ma), are exposed (Lund and Snee 1988; Manduca et al. 1993; Giorgis et al. 2008; Unruh et al. 2008) and, on the basis of the few existing analyses, appear to be slab failure magmas as well (Supplementary Fig. S3[2]).

[2]Supplementary data are available with the article at https://doi.org/10.1139/-2021-0006.

Fig. 1. Geological map, with Cenozoic cover removed, of eastern California and Nevada showing the Sierra Nevada batholith, Cretaceous plutonic rocks of Nevada, various tectonic terranes of Nevada after Crafford (2007, 2008) and northern Mojave Desert modified from Walker et al. (2002). Death Valley and White–Inyo–Esmeralda sedimentary successions from Stewart (1970). BRH, Bloody Run Hills; HR, Humboldt Range; RRP, 111 ± 2 Ma metavolcanic rocks of the Rugged Rock Pendant from Whitmarsh (1997); SC, Schoonover sequence. Jurassic 0.706 isopleth from Wyld and Wright (2019); location of 97–86 Ma $^{40}Ar/^{39}Ar$ plateau ages in Jurassic rocks west of Fencemaker thrust from Wyld et al. (2003). Ages of plutons in Nevada mainly from N. Van Buer, personal communication, 2018; D. Johns and C. Henry, personal communication, 2020; and Du Bray 2007. [Colour online.]



A variety of orthogneisses occur west of the Payette River tonalite (Supplementary Fig. S2[2]) as well as within the Western Idaho shear zone and the youngest unit, an undeformed biotite–hornblende granodiorite (Rat Creek granodiorite), yields U–Pb ages ranging from 89 to 84 Ma (Braudy et al. 2017). West and north of the Payette River tonalite, two deformed intrusive complexes, the Little Goose Creek and the Hazard Creek, have seven U–Pb ages ranging between 120 and 108 Ma, interpreted as emplacement ages, as well as sparse Jurassic ages, which are compatible with local basement (Patzke 2017). Unfortunately, no modern whole-rock geochemical data have been published to date.

The Potters Pond migmatitic domain (Supplementary Fig. S2[2]) contains meta-arkose metamorphosed after 96.8 ± 5.5 Ma as well as orthogneisses dated at 98.6 ± 0.6, 95.3 ± 0.9, 93.3 ± 0.7, and 92.4 ± 0.7 Ma along with seven gneisses that yielded monazite ages ranging from 98 to 91 Ma (Montz and Kruckenberg 2017). To the southwest of the Snake River Plain, in the northern Owyhee Mountains, several orthogneisses and plutons produced U–Pb zircon ages of 98–86 Ma, all with $^{87}Sr/^{86}Sr_i$ of 0.706 or higher (Benford et al. 2010).

Taken together, the abundance of 100–85 Ma intrusions emplaced during and after deformation in the transpressional Western Idaho shear zone suggests that these intrusions are post-collisional bodies of the Peninsular Ranges orogen. Like the Sierra Nevada and Peninsular Ranges, intrusions younger than 100 Ma show the isotopic interaction with old and enriched continental lithosphere, whereas pre-100 Ma bodies located to the west are isotopically more juvenile (Armstrong et al. 1977; Fleck and Criss 1985, 2007; Criss and Fleck 1987; Manduca et al. 1992).

Foreland basin remnant

East of the Atlanta lobe of the Idaho batholith and west of the Boulder batholith in the Drummond, Montana area (Supplementary Fig. S2[2]), a succession of upper Albian to Santonian sedimentary units over 3500 m thick, likely deposited in a flexural foreland basin, were folded and thrust eastward prior to emplacement of 82 Ma intrusions (Wallace 1987; Wallace et al. 1990). The Blackleaf Formation is the basal unit, comprising three members, Flood, Taft, and Vaughn. Similar named units also occur to the east in the Montana thrust belt near Wolf Creek, but they are very much thinner, and 1–2 million years older, there (Zartman et al. 1995; Singer et al. 2021). In the thick western section, Wallace et al. (1990) pointed out that rocks of the Taft member were uplifted and eroded at about 100 Ma, but that there is no evidence of this in the thinner easterly sections. They (p. 1034) also indicated that the >3000 m thick overlying "sequence of rocks shares no similarities of lithologic succession with rocks of the upper Cretaceous" in the thrust belt north of the Lewis and Clark line and that the dominantly coarse-grained succession was deposited in shallow, brackish water compared with the thin sequence of marine Cenomanian–Santonian black shales deposited farther east (see also Fuentes et al. 2011).

We infer that the erosion of the Taft member took place as the easterly advancing thrust sheets caused the flexural bulge along the eastern flank of the foreland basin to migrate in front of it. The age of the unconformity, as well as the 3.5 km of Cenomanian–Santonian clastic sediment above it, are consistent with the 100 Ma Peninsular Ranges orogeny, not the older Sevier or younger Laramide deformation.

Detrital zircons were collected from the Vaughn member of the Blackleaf Formation and yielded a peak of 100 Ma (Stroup et al. 2008) consistent with erosional debris expected from the Peninsular Ranges orogen. They also found detrital zircons in rocks of the late Eocene Renova Formation with a distinctive 95 Ma peak dominated by zircons younger than 100 Ma, whereas they found that detrital zircons from the late Oligocene Medicine Lake beds had a 105 Ma peak dominated by detrital zircons in the age range 115–100 Ma. It appears that local drainages were able to access both arc and orogenic hinterland rocks for many millions of years after the collision.

The thrusts in the belt strike north–south and are easterly vergent (Supplementary Fig. S4[2]), whereas the younger Laramide thrusts are oriented northwest–southeast and appear to reflect the buttress effect of the Lewis and Clark lineament, located just to the north (Hildebrand 2015). The thrusts and sedimentary succession are readily interpreted to represent a foreland fold-thrust belt and related flexural foredeep of the Peninsular Ranges orogeny.

Additional sedimentary debris, likely derived from post-collisional plutonic rocks, was shed still farther inboard and is preserved in the Bighorn basin (Supplementary Fig. S2[2]) as rocks of the Mowry Shale, Frontier Formation, and Cody Shale, which yielded youngest detrital peak ages, consistent with their paleontological ages in 13 samples, ranging from 99.4 to 87.7 Ma (May et al. 2013).

Cascades

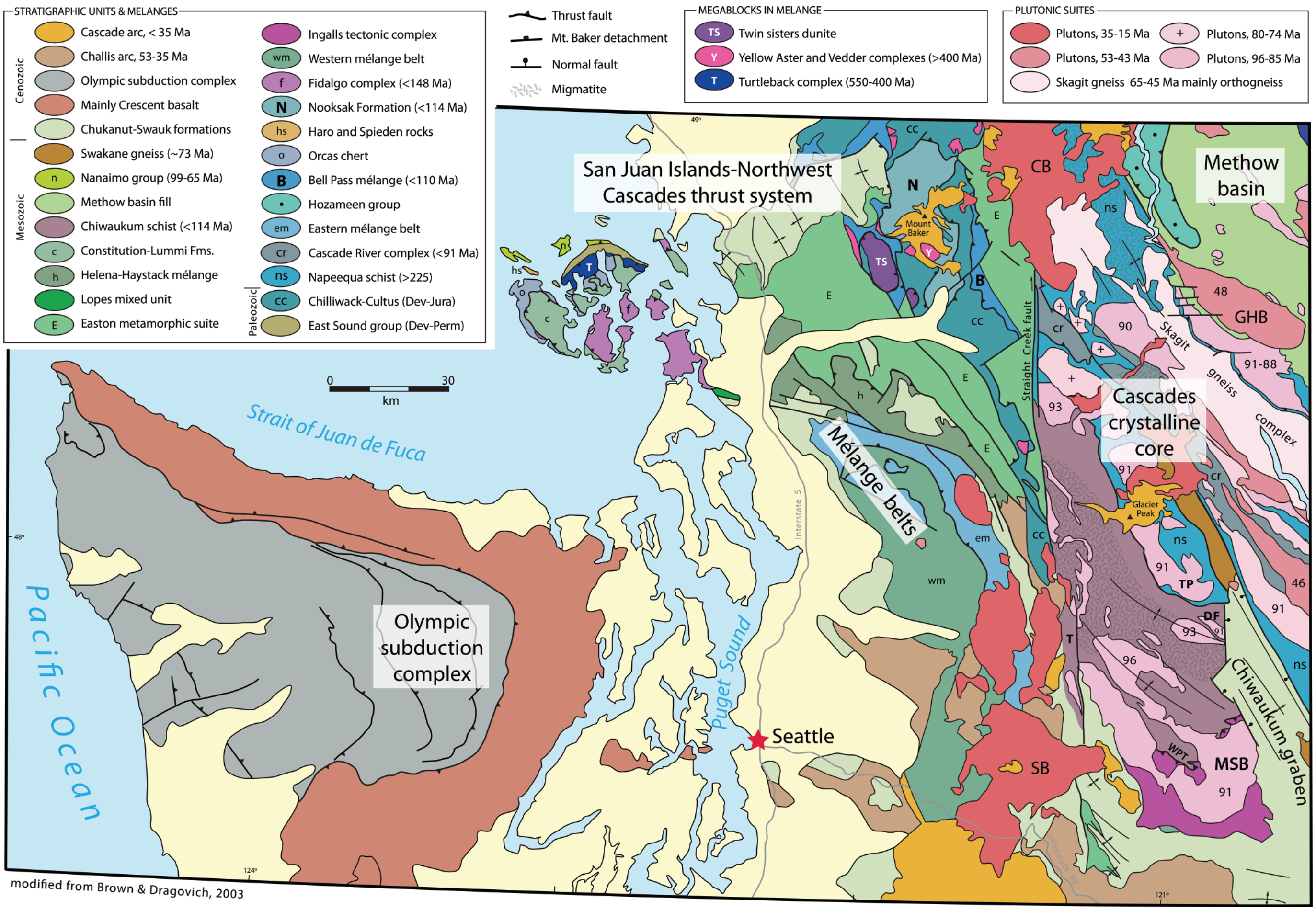
Although younger deformation and metamorphism related to the Laramide orogeny obscure and obliterate some of the older geological development of the Cascades (Miller et al. 2009 and references therein), many components of the Peninsular Ranges orogeny are preserved.

A number of Early Cretaceous metavolcanic and metasedimentary units are exposed in the generally high-grade Cascades crystalline core (Fig. 2). These include the Cascade River schist, which comprises mica schist and biotite paragneiss with lesser amounts of metaconglomerate, metavolcanic rocks, and metaperidotite (Sauer et al. 2017). One sample of the schist yielded detrital zircon populations of 120 Ma, a broad peak centered on 165 Ma, and a maximum depositional age (MDA) of 97 Ma, whereas the other sample produced a prominent peak of 93 Ma, as well as a small peak of 165 Ma, and an MDA of 91 Ma (Sauer et al. 2017). They also examined several samples from the garnet–sillimanite grade Skagit gneiss (Fig. 2) and found most detrital zircons have younger Laramide-age rims, but that the cores are mainly Early Cretaceous, Jurassic, and Triassic. Five of the samples had core MDAs of 121, 115, 112, 108, and 96 Ma, all ±2 Ma. These ages appear to reflect both arc and post-collisional magmatism related to the Peninsular Ranges orogeny.

The high-grade crystalline core was intruded by three groups of plutons with ages similar to other suites within the Cordillera (Fig. 2): plutons attributed herein to the Peninsular Ranges orogen are 96–89 Ma, Laramide bodies are 80–58 Ma, and plutons emplaced during Laramide extensional collapse are 50–45 Ma. The large 96–91 Ma Mount Stuart batholith (Matzel et al. 2006) postdates the early deformational fabric and a thrust fault, known as the Windy Pass thrust (Fig. 2), which placed rocks of the dominantly ophiolitic Ingalls complex over metamorphosed siliciclastic and metavolcanic rocks of the <125 Ma Chiwaukum schist (Miller 1985). Brown and Gehrels (2007) reported a 95 Ma dike that cut the amphibolite-grade Tonga Formation, which is a fault-bounded fragment of the Chiwaukum schist located just east of the Fraser River – Straight Creek fault zone (Fig. 2). We plotted geochemical data from several intrusions (Supplementary Fig. S5[2]), emplaced during regional amphibolite-facies metamorphism at depths of 25–35 km (Shea 2014; Miller et al. 2018) and, as consistent with their post-deformational age of emplacement, they appear to be post-collisional slab failure plutons.

West of the crystalline Cascades core, across the Straight Creek – Fraser River fault and to the west in the San Juan Islands, is an imbricate stack (Fig. 2) of units that were assembled after 110 Ma, and largely before intrusion of the 96–90 Ma plutons (Brandon et al. 1988; Brown and Dragovich 2003; Brown and Gehrels 2007) consistent with the 100 Ma orogenic event. Each thrust slice has a different metamorphic mineral assemblage, but on the whole most show evidence for high-pressure, low-temperature metamorphism (Brown et al. 1981). The lowermost unit of the stack is the Nooksack Formation, which comprises metavolcanic arc and associated metasedimentary clastic rocks with 114 and 153 Ma detrital zircon peaks and an MDA of 114 Ma (Brown and Gehrels 2007). These authors reported that the Nooksack Formation is overlain structurally, but separated by the Bell Pass mélange, from higher thrust slices containing Jurassic, but no Cretaceous, rocks. The mélange contains an incredible variety of clasts in a sandstone–argillite matrix yielding a major 119 Ma detrital zircon peak and a possible MDA of 110 Ma, but two zircons with ages of 105 Ma suggest the MDA might be younger. Individual blocks in the mélange include the 4 km × 10 km slab of Twin Sisters dunite, along with many smaller ultramafic blocks. In addition, blocks or lenses, up to a few kilometres across, of metasedimentary rocks termed the Yellow Aster complex, are dominated by detrital zircons mostly older than 1800 Ma and were strongly metamorphosed and intruded during the Devonian by gabbroic–tonalitic plutons (Brown and Gehrels 2007; Schermer et al. 2018). Other blocks include Lower Cretaceous blueschists, high-pressure, high-temperature Permian amphibolite, Triassic ribbon chert, pillow basalt of oceanic island provenance; and blocks of both underlying and overlying tectonic units. Brown and Gehrels (2007) pointed out the similarity of the

Fig. 2. Geological sketch map of northwestern Washington, illustrating the basic geology of the San Juan Islands – Northwest Cascades thrust system, the western Cascade crystalline core, and surrounding area. CB Chilliwack batholith; DF, Dirty Face pluton; GHB, Golden Horn batholith; MSB, Mount Stuart batholith; N, Nooksack Formation; SB, Snoqualmie batholith; T, Tonga Formation; TP, Ten Peak complex; WPT, Windy Pass thrust. Modified from Brown and Dragovich (2003) with ages of plutons from Miller et al. (2009) and ages of other units from Brown and Gehrels (2007) and Sauer et al. (2017). [Colour online.]



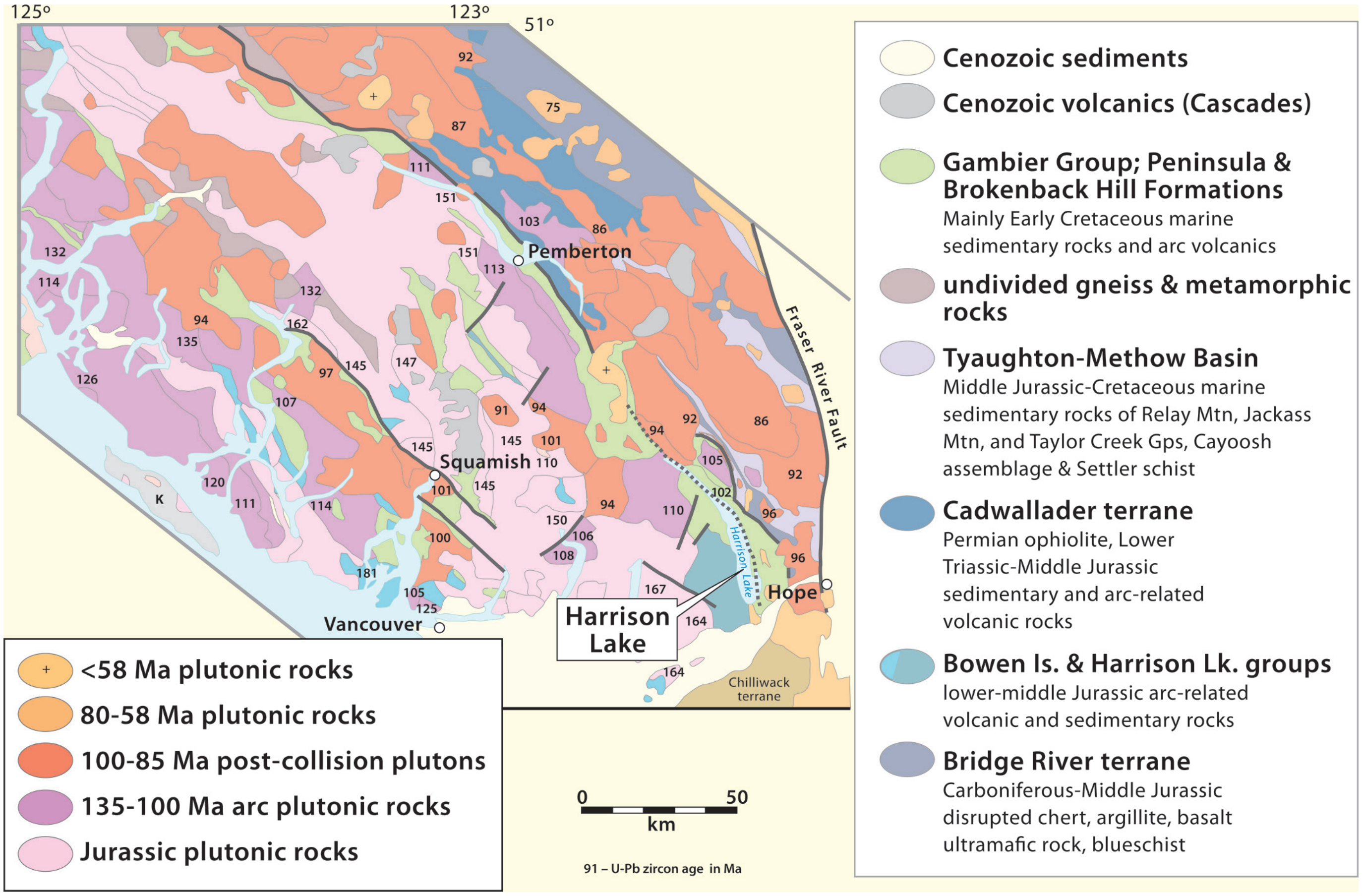
zircon age profiles in the blocks of Yellow Aster complex to many rocks of western North America, such as the older nappes in the Shoo Fly complex and the Antelope Mt. Quartzite of the Yreka terrane in the Klamaths. However, Schermer et al. (2018) suggested they are more similar to Yukon–Tanana and Alexander terranes. Because the mélange contains a wide variety of blocks from varied settings, including both continental and oceanic, as well as high-pressure, low-temperature metamorphism, we suggest that the mélange is part of the suture zone between the lower North American plate and the Peninsular Ranges arc-bearing composite terrane.

Coast Mountains batholith of British Columbia

The geology of the Coast Mountains batholith of British Columbia (Roddick and Hutchinson 1974; Roddick 1983) is similar in many aspects to that of the Peninsular Ranges and Sierra Nevada. The Coast Mountains batholith is traditionally interpreted as a manifestation of eastward subduction, initially beneath the Insular terrane, but that magmatism migrated eastward beneath Stikinia with time (Gehrels et al. 2009). The geology of the batholith is extremely complex, because not only are rocks of the Peninsular Ranges orogen present, but rocks deformed and intruded during the Campanian–Paleocene Laramide orogen (Rusmore and Woodsworth 1991), as well as an Eocene deformational and magmatic belt (Erdmer and Mortensen 1993; Johnston and Canil 2007) largely occupy the same north–south structural corridor (Hildebrand 2015; Hildebrand and Whalen 2017).

On the mainland, the westernmost parts of the Coast Mountains batholith (Fig. 3) comprise Triassic basement of Wrangellia covered by dominantly greenschist-grade metasedimentary and metavolcanic rocks of the Early to Middle Jurassic Bowen Island and Harrison Lake groups, all cut by a variety of plutons (Friedman et al. 1990; Mahoney et al. 1995). Rocks of the Bowen Island – Harrison Lake groups are overlain, along a deeply incised unconformity, by metavolcanic and metasedimentary rocks of the Gambier Group, which is divided into two formations: (1) a basal assemblage, known as the Peninsula Formation, dominantly composed of Berriasian–Valanginian conglomerate, *Buchia* coquina, and arkose, overlain by (2) rocks of the late Valanginian to Albian Brokenback Hill Formation (Fig. 3), which is a succession of submarine to subaerial volcanic and volcaniclastic rocks ranging continuously in composition from basalt to rhyolite, as well as a variety of sandstone, siltstone, black shale, and volcanic cobbly to bouldery conglomerate, containing clasts as large as 1–2 m across (Crickmay 1925; Arthur et al. 1993; Lynch 1992, 1995). The Brokenback Hill formation contains Albian fauna, rhyolite dated at 112.0 ± 0.3 Ma by U–Pb on zircon (Lynch 1992, 1995) and MDAs, determined from detrital zircon grains, of 116–114 Ma (Dorsey 2018). The Peninsular Formation was shown by Dorsey (2018) to be younger than 138 Ma.

Fig. 3. Geological sketch map of the southern portion of the Coast plutonic complex, British Columbia, modified from Friedman and Armstrong (1995), with additional ages from Gibson and Monger (2014). Note that abundant 135–100 Ma plutons, as well as arc volcanic and sedimentary rocks of the dominantly marine Gambier Group, sitting on Late Jurassic basement in pink, all lie west of the Harrison Lake structural break, whereas ~101–86 Ma post-collisional plutons, shown in brick red, span the entire area. [Colour online.]



Farther north, Mahoney et al. (2009) described and grouped the Lower Cretaceous plutons in the Bella Coola region into two suites: the 141–131 Ma Firvale Suite, which consists of hornblende–biotite granodiorite to granite, and the 123–110 Ma Desire suite, a heterogeneous suite of pyroxene–hornblende- and biotite-bearing diorites and granodiorites, commonly deformed by shear zones characterized by a strong foliation and (or) locally developed mylonitic fabrics.

Early work on the structure of the southern Coast plutonic complex suggested that ~100 Ma faults along the eastern boundary of the Lower Cretaceous arc rocks in the Harrison Lake region (Fig. 3), were low-angle thrust faults that placed high-grade metamorphic rocks westward over low-grade arc rocks (Journeay and Friedman 1993), an interpretation that led to the general notion that subduction beneath the belt was eastward. However, more recent work suggests that the faults are much steeper and have over 10 km of reverse displacement (Friedman et al. 1992; Brown et al. 2000; Gibson and Monger 2014) and that the higher-grade rocks were exhumed westward and placed atop the western plutonic terrane (Fig. 4), which consists dominantly of pre-collisional Lower Cretaceous arc plutons and greenschist-grade arc volcanic and associated epiclastic rocks as discussed above.

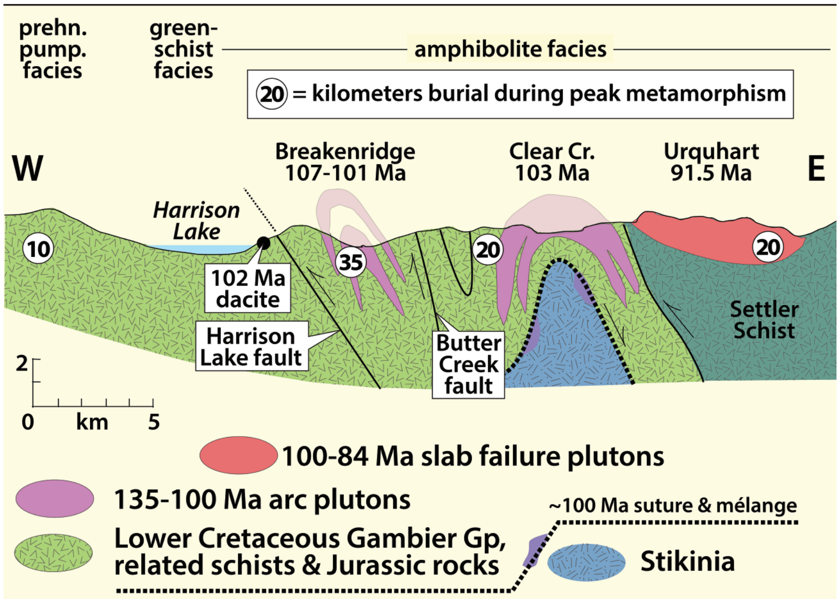
These findings suggest that the arc was on the western block prior to collision at 100 Ma, and that the upper plate was more regionally extensive than previously thought. If so, then the reverse faults are east side up and, therefore, the collisional suture related to basinal closure, if preserved, should lie to the east, not the west.

The rocks of the Harrison Lake region (Fig. 5) constrain the age of deformation to be about 100 Ma as a 102 Ma metadacite, the 107–101 Ma Breakenridge intrusive sheets, and the 103 Ma Clear Creek pluton are all deformed, whereas the post-deformational Hornet Creek, Spuzzum, and Ascent Creek bodies are all 97–96 Ma (Brown and Walker 1993; Gibson and Monger 2014). The collision in this area represents the collision between the Peninsular Ranges composite terrane, composed of parts of the Insular and Intermontane superterranes, which had previously collided during the Jurassic (Monger et al. 1982), but were partly separated when the Lower Cretaceous trough opened.

Friedman et al. (1990) showed that tight isoclinal folds along the steeply dipping Butter Creek fault (Figs. 4 and 5), which clearly postdate the 100 Ma isoclinal folding, transposition of bedding, and thrust faults that imbricate the metavolcanic and metasedimentary rocks, as well as Breakenridge gneiss, were cut by the 91.5 ± 2 Ma Mason pluton. This implies that the majority of the exhumation within the hinterland belt took place in less than 10 million years.

Farther north, within the eastern part of the batholith, rocks of Yukon–Tanana and Stikine terranes are exposed and are generally high-grade metasedimentary migmatites and orthogneiss, typically at upper amphibolite to granulite grade with sillimanite growth after kyanite or staurolite (Hutchison 1970; Stowell and Crawford 2000; Hollister and Andronicos 2000; Rusmore et al.

Fig. 4. Cross section from west to east through the Harrison Lake fault belt illustrating the structural relief across faults. Modified from Monger and Brown (2016). [Colour online.]



2000). Arc rocks, generally at relatively low metamorphic grade, lie to the west of the high-grade hinterland belt (Gibson and Monger 2014) and indicate that the Yukon–Tanana composite terrane was part of the lower plate in the collision and that subduction was westward beneath the arc (Fig. 6).

The post-collisional Ecstall plutonic suite

Both the hinterland and arc terranes of the Coast plutonic complex were intruded by a suite of post-collisional plutons in the age range 99–85 Ma (Brown and McClelland 2000; Brown et al. 2000; Gehrels et al. 2009; Mahoney et al. 2009; Girardi et al. 2012), just as typically occur farther south (Fig. 7). The intrusive suite consists of large homogeneous bodies of tonalite with lesser quantities of granodiorite and quartz diorite, commonly with megascopically visible euhedra of titanite and epidote (Gehrels et al. 2009). Where the bodies intruded the hinterland belt, they were emplaced into high-grade rocks during their exhumation (Crawford et al. 1987; Himmelberg et al. 2004). Mahoney et al. (2009) dated one intrusion by U–Pb zircon to be 86.8 ± 0.3 Ma and included it in his Big Snow suite.

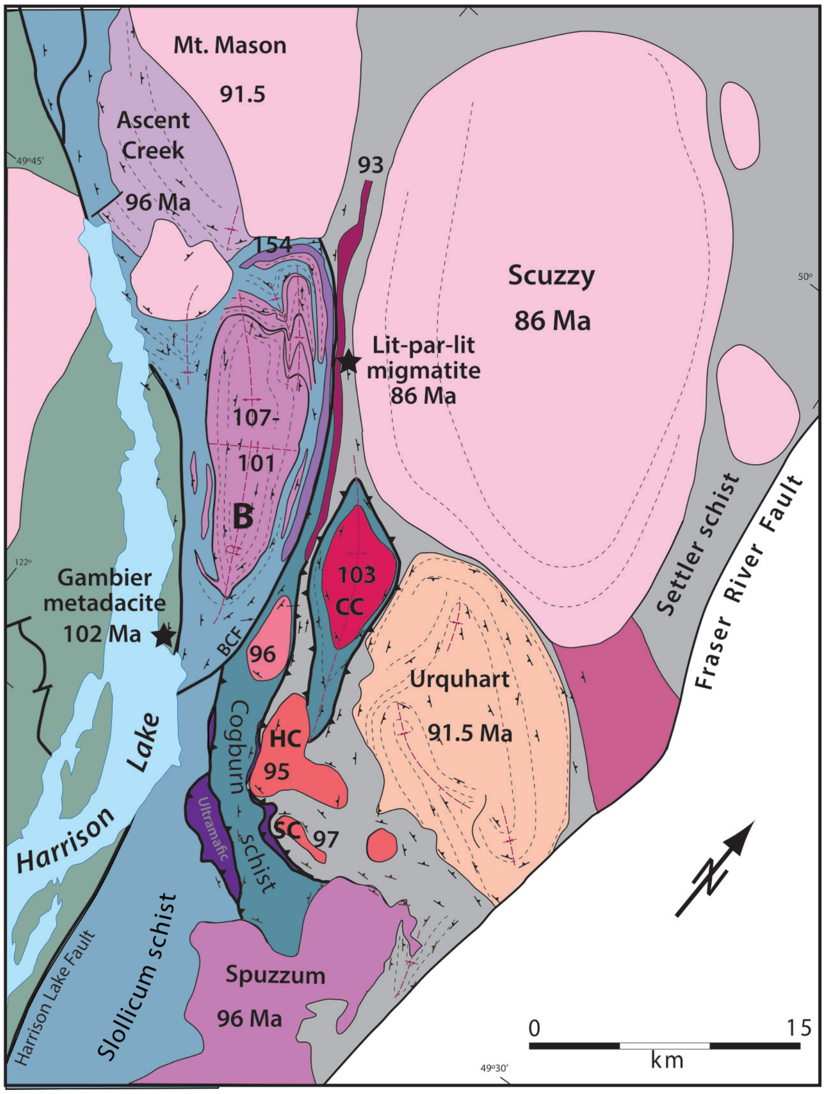
The western part of the Coast Plutonic complex continues northward along the coast where some 50 km south of the USA–Canada international border (Fig. 7), the 20 km by >80 km, epidote-bearing, 98 ± 4 Ma hornblende dioritic–granodioritic Ecstall pluton intruded deformed and high-grade wall rocks at depths of 25–30 km and has a $^{40}Ar/^{39}Ar$ hornblende age of 90 ± 3 Ma, which indicates rapid uplift and exhumation (Woodsworth et al. 1983; Brownlee and Renne 2010) consistent with the Harrison Lake region farther south. Other U–Pb zircon data yield an age of 91 ± 0.5 Ma for the pluton (Butler et al. 2002). This age conflicts with $^{40}Ar/^{39}Ar$ hornblende ages; however, Brownlee and Renne (2010) showed that the Ar systematics were severely disturbed by younger intrusions to the east. The U–Pb ages determined by Butler et al (2002) also young eastward, so additional study is warranted. Nevertheless, the available data indicate that the pluton is no younger than 91 Ma.

The plutonic wall rocks are kyanite-bearing, metamorphosed at 30 km depth, and deformed into tight isoclines and by cleavage, both of which are transected by a more steeply dipping foliation containing steep lineations adjacent to east-dipping, westerly vergent shear zones dated at about 90 Ma by $^{40}Ar/^{39}Ar$ (Crawford et al. 1987). Crawford and Hollister (1982) pointed out that the metamorphic grade increases structurally upward from west to east, and varies from chlorite in the west to kyanite+muscovite+melt (8.1 kbar) in the east, indicating westward transport of higher over lower-grade metamorphic units. The structures responsible for this transport appear to be the typical reverse faults found along much of the western Peninsular Ranges orogen and are similar in age to those around Harrison Lake.

As the post-collisional suite was previously unnamed in the Coast plutonic complex, we name the 100–84 Ma post-collisional suite the Ecstall suite after the deep-seated Ecstall pluton. We plotted available geochemistry from the post-collisional 100–85 Ma intrusions on our discrimination diagrams and, consistent with our model, they plot as slab failure plutons (Supplementary Fig. S6[2]) as do the contemporaneous plutons farther south.

Girardi et al. (2012), who studied the geochemistry of a sector within the batholith, suggested that it is unlikely that the post 100 Ma plutons fractionated or interacted with other rocks in a significant way above the stability limit for plagioclase in these composition rocks, which, on the basis of experiments with amphibolite and eclogite, is about 100 km (Rapp et al. 1991). Overall, the geochemical data from rocks studied by Girardi et al. (2012) are similar to those of post-collisional plutons farther south, in that compatible elements vary systematically with SiO_2 but that with incompatible elements there is no correlation. The same patterns are found in the Tuolumne plutonic suite (Hildebrand et al. 2018), and indicate that as magmas rose, they modified their composition by fractional melting and assimilation of the SCLM. However, because the SCLM beneath the Coast batholith was relatively juvenile and non-radiogenic, the Nd and Sr isotopes are more juvenile, or arc-like, than plutons that interacted with old and enriched SCLM. Coast Range plutons have $\delta^{18}O$ values ranging from 7.2% to 10%, which indicates that the magmas were derived from, or interacted with, rocks that had been weathered near the surface (Wetmore and Ducea 2011). Those researchers pointed out that the high $\delta^{18}O_{quartz}$ coupled with the primitive Sr, Pb, and Nd indicate that the source of the magma must have been mainly mafic, volumetrically large, had primitive radiogenic isotopes, and were altered by low temperature meteoric or

Fig. 5. Plutons of the Coast Range plutonic complex east of Harrison Lake bracket the time of deformation to be about 100 Ma. Modified after Brown and McClelland (2000) and Brown et al. (2000). Ages from Gibson and Monger (2014). Rocks of the Slollicum schist are similar lithologically to those of the Gambier Group west of the Harrison Lake fault, but are generally at higher grade. They were shown by Dorsey (2018) to be the same age with maximum depositional ages of 120 and 112 Ma. Rocks of the Cogburn Group are a greenschist–amphibolite mélange of oceanic rocks, whereas Settler schist is composed of amphibolite-grade pelitic–psammitic metasedimentary rocks. Both are poorly dated. The Butter Creek fault (BCF) is cut by the 91.5 Mt Mason pluton; thus, the main exhumation of the hinterland took place in less than about 10 million years. B, Breakenridge pluton; CC, Clear Creek pluton; HC, Hat Creek pluton; SC, Settler Creek pluton. [Colour online.]



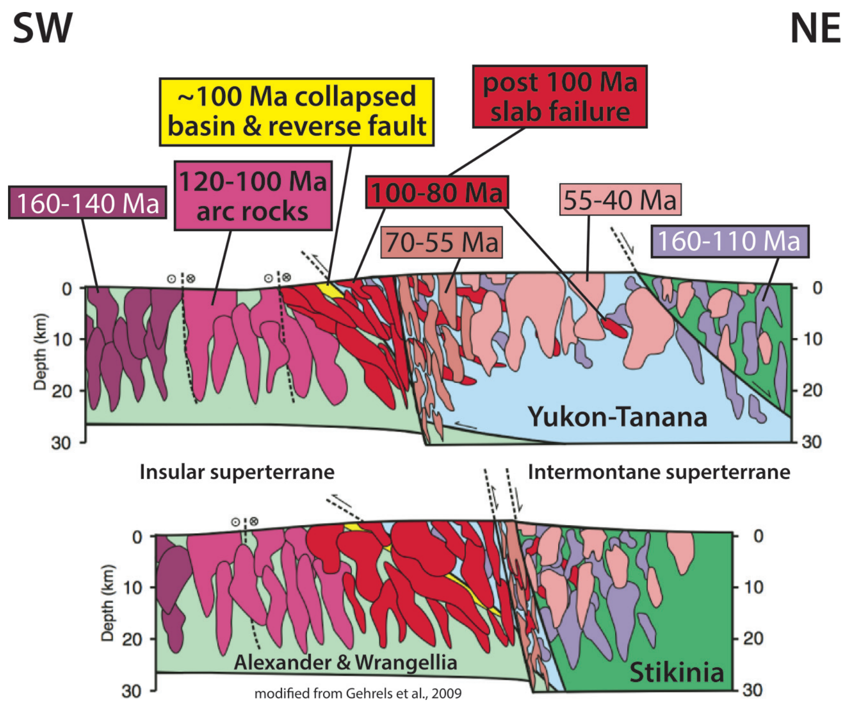
sea water. Just as with the Sierra Nevada and Peninsular Ranges batholiths, the upper basaltic–gabbroic part of oceanic crust is a potential source of these magmas. When compared with post-collisional plutons of the Sierra Nevada and Peninsular Ranges, Sr and Nd isotopes are less evolved, but Sierran plutons commonly have lower, mantle-like $\delta^{18}O$ values.

Nanaimo and Queen Charlotte groups, British Columbia

To the west of the orogen on eastern Vancouver Island, a Late Cretaceous sedimentary succession, known as the Nanaimo Group, sits unconformably upon rocks of Wrangellia (Monger et al. 1982; Mustard 1994; Mustard et al. 1995). Much like their equivalents to the south, this succession is typically considered to represent a fore-arc deposit situated west of the arc represented by the Coast Ranges batholith. However, the rocks of this succession have pronounced detrital zircon peaks (Fig. 8) between 100 and 85 Ma (Matthews et al. 2017), ages that are too young to be arc related, but do match post-collisional plutonic ages, so are interpreted here to represent debris shed from the exhumed hinterland belt.

Other researchers have noted the far-sided paleopoles for the Nanaimo Group and considered that it must have traveled from the south (Enkin et al. 2001). However, the lack of exposure between Vancouver Island and the mainland means there is no recognized fault along which this northerly rotation might have taken place. But, as discussed above, a trough with similar 100–85 Ma detrital zircons occurs all along the western side of the Peninsular Ranges orogen, obviating the requirement for Vancouver Island to have moved relative to the mainland and Coast batholith to provide the necessary-age zircons. In addition, paleomagnetic results from more interior locations, such as those obtained from rocks of the 95–85 Ma Silverquick and Powell Creek formations (Enkin et al. 2006*a*) and

Fig. 6. Cross sections through the Coast plutonic complex, modified from Gehrels et al. (2009) illustrating that the 120–100 Ma Lower Cretaceous arc plutons lie west of the collapsed Gambier basin, shown in yellow, whereas post-collisional rocks in red span the contact of the 100 Ma collision zone. These relations document that subduction was westward-dipping on the east side of the Insular superterrane prior to collision with the Intermontane terrane at 100 Ma. During the Peninsular Ranges orogeny, the leading edge of the Intermontane superterrane was pulled beneath the arc located on the Insular superterrane. Following slab failure, the area was riddled with post-collisional plutons and the collision zone exhumed to expose high-grade metamorphic rocks of the lower plate. [Colour online.]



Methow block (Granirer 1985), approximate those from rocks of the Nanaimo basin (Enkin et al. 2001). Irving et al. (1995, 1996) suggested a model involving northward migration along the cryptic Intra-Quesnellia fault; whereas Hildebrand (2015) used piercing points to show how 1300 km of northerly migration could be restored along faults well inboard of those traditionally investigated for such displacements, similar to models of Enkin et al. (2006*b*) and Gladwin and Johnston (2006).

Haida Gwaii lies offshore from the region just south of the international border (Fig. 7). There, Dorsey (2019) sampled sandstones and pebbly conglomerates from the Cretaceous Queen Charlotte Group (Haggart 2004) for detrital zircons. His zircon analyses document the progressive deroofing of the arc and post-collisional hinterland belt (Fig. 7) of the Peninsular Ranges orogen at this latitude.

Trapped units

Within the northern Cordillera, the lack of recognition of a 100 Ma collision has led to difficulties in interpreting the setting of various terranes. Recognizing the rifting and spreading to form the early Cretaceous seaway and its subsequent 100 Ma demise might resolve some of these problems.

Several terranes are caught between North America and the Peninsular Ranges composite terrane. These terranes include relic parts of the Insular and Intermontane superterranes, because as discussed earlier, the two superterranes had initially collided during the Jurassic, but were subsequently dismembered during the Early Cretaceous. In this scenario, fragments of the previous Intermontane terrane could have been rifted to constitute part of the western block and, similarly, fragments of the Insular terrane could have been transferred to the eastern side of the basin. Here we briefly discuss two terranes: Spences Bridge terrane and Methow basin, both of which span the international border.

Thorkelson (1986) suggested that at least one piece of the Early Cretaceous arc, located just east of the Fraser River – Pasayten fault system in southern British Columbia (Supplementary Fig. S7[2]), and termed the Spences Bridge Group, faced westward during magmatism. Overall, the Spences Bridge Group is a late Albian two-part volcanic succession that rests unconformably on several rock units of the Intermontane superterrane and was considered to have formed by eastward subduction prior to, and after, a mid-Cretaceous collision with the Insular terrane (Thorkelson and Smith 1989).

Volcanic rocks of the group were divided into two successions: a lower unit of calc- alkaline basaltic to rhyolitic arc-type volcanic and intercalated sedimentary rocks containing late Albian fossils, overlain by more localized andesitic lava flows of a broad shield volcano interpreted to be post-collisional and cut by plutons with K–Ar ages of 98–97 Ma (Thorkelson and Smith 1989). They determined that the lower lavas have $^{87/86}Sr_i$ ranging from 0.70316 to 0.7040 and εNd_T from 5.0 to 7.8, whereas the upper andesitic rocks are more primitive with $^{87/86}Sr_i$ as low as 0.70298 and εNd_T as high as 8.8. We plotted their trace element analyses on our discrimination diagrams. Although some elements were affected by alteration, least mobile elements (Ta/Yb and La/Sm vs. Sm/Yb) support their model of arc magmatism overlain by younger andesitic magmatism derived from a different, likely deeper, mantle source. We envision that the Albian arc suite represents part of the arc built on a fragment of the Intermontane superterrane within the Peninsular Ranges composite terrane west of the seaway and that the upper suite represents early post-collisional magmas derived from beneath the collision zone during slab break-off. In a different approach to explain the location of the Spences Bridge Group, Lynch (1995) developed an arc–arc collision model with both westerly and easterly subduction.

The lack of recognition of at least one Jurassic collision followed by a 100 Ma mid-Cretaceous collision led to a complex

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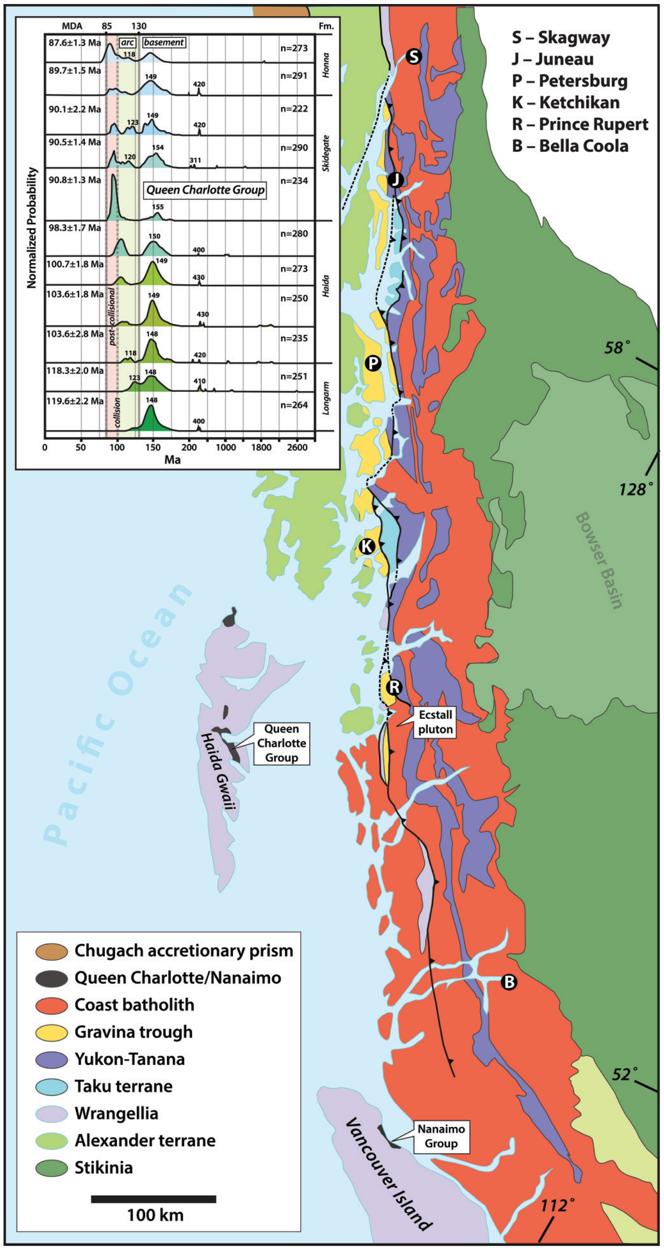


Fig. 7. Geological sketch map of part of the Coast plutonic complex, illustrating the various terranes, the Gravina trough assemblage and detrital zircon suites from Cretaceous sedimentary rocks of the Queen Charlotte Group on Haida Gwaii (Dorsey 2019) illustrating a progressive deroofing sequence ranging from Jurassic basement upwards through 130–100 Ma arc and 99–88 Ma post-collisional magmatism. See Supplementary Fig. S8[2] for detailed map and detrital zircon results for Gravina trough and related rocks. [Colour online.]

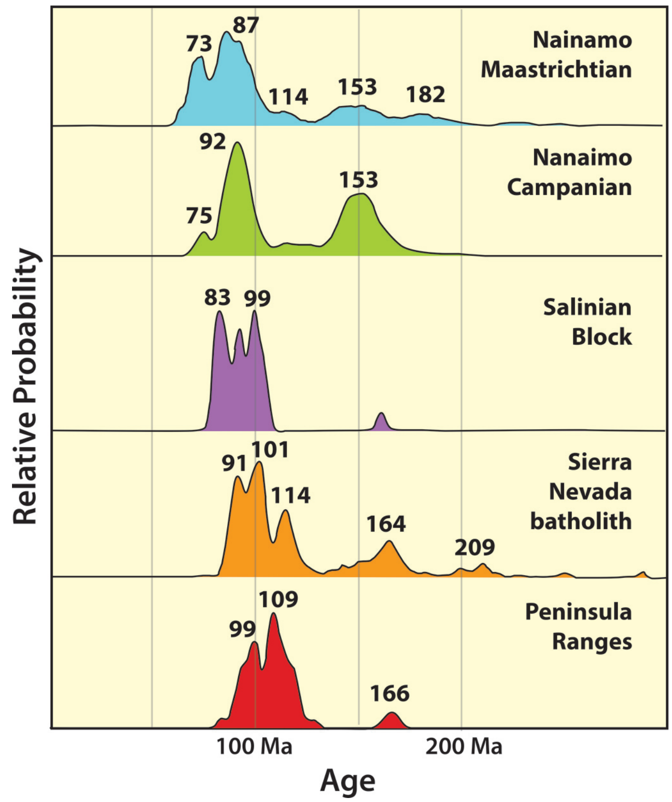


Fig. 8. Detrital zircon probability diagrams for rocks of the Nanaimo Group, illustrating the similarity of dterital ages within rocks of the Salinian block, Sierra Nevada, and Peninsular Ranges (modified from Matthews et al. 2017). Note the consistent mid-upper Jurassic peaks reflecting basement, as well as the obvious 130–100 Ma arc and 100–83 Ma post-collisional peaks. Salinia, as well as rocks of the Sierra Nevada and Peninsular Ranges, were relatively unscathed by younger Laramide deformation so their outboard regions do not contain 75 ± 5 Ma detrital zircons (see fig. 8 in Hildebrand and Whalen 2021). [Colour online.]

model using large-magnitude sinistral strike-slip faults to relocate the Methow basin, interpreted as a fore-arc basin, into a retro-arc position by younger dextral strike-slip faults to explain its present location on the east side of the Coast plutonic complex between the Insular and Intermontane terranes (Monger et al. 1994; Gehrels et al. 2009; Yokelson et al. 2015). The basin is a fault-bounded block (Haugerud et al. 1996) lying to the west of the Spences Bridge Group, and separated from it by the east-vergent Pasayten thrust (Umhoefer and Miller 1996). Existing data suggest that the Methow basin preserves a remnant of the Early Cretaceous arc trough. Metasedimentary rocks dominate the trough, are dominantly siliciclastic and immature with a high modal plagioclase content, and were deposited from the middle Albian to the Cenomanian atop upper Paleozoic to lower Mesozoic cherts and greenstones of the Hozameen and Bridge River groups (Kleinspehn 1985). Upper parts of the sedimentary succession were intruded by a 97.5 Ma sill (Dragovich et al. 1997) and by 96–88 Ma plutons (Haugerud et al. 1996). Detrital zircons from both northern and southern outcrop belts (Surpless et al. 2014) yield two main age peaks: an Oxfordian peak and an Early Cretaceous peak (Fig. 9). Similar profiles from several locations typically exhibit a general magmatic lull between about 140 and 125 Ma. Geochemical analyses of mudrocks from the block (Surpless et al. 2014) plot in the arc field on our discrimination plots, consistent with their presumed proximity to Lower Cretaceous arc material as well as 160 Ma Jurassic arc rocks. The uppermost unit, the Goat Wall Formation, comprises andesitic flows and siliceous ignimbrites intercalated with volcanogenic sedimentary rocks containing detrital zircons as young as 105 Ma (Surpless et al. 2014). Based on the ages and compositions of the Methow rocks, the block is interpreted to be part of the Early Cretaceous arc and seaway.

Zircons older than about 200 Ma (Fig. 9) are interpreted to have been derived from rocks well to the south, near the Gondwana–

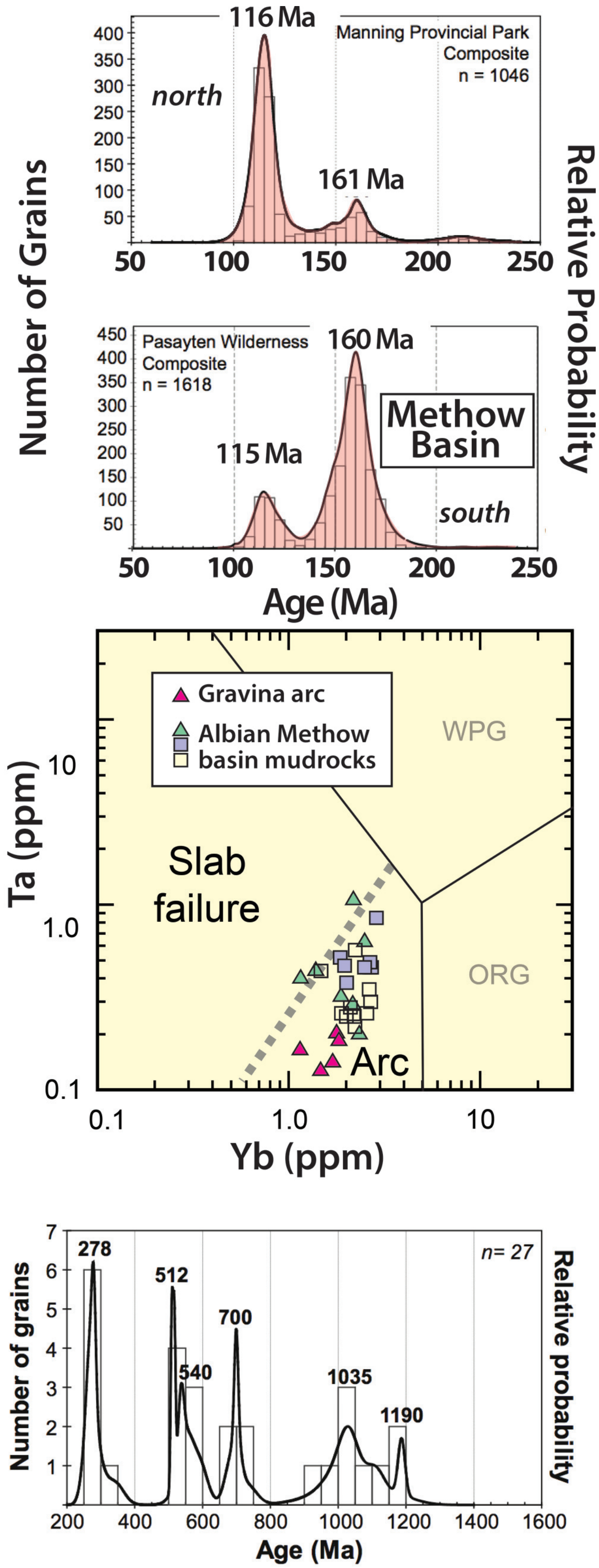


Fig. 9. Detrital zircon histograms showing relative probability vs. age for sedimentary rocks of the Methow block, modified from Surpless et al. (2014). These plots show the typical Jurassic basement peaks as well as the 130–100 Ma arc peaks. The Ta vs. Yb discrimination diagram shows that fine-grained sedimentary rocks within the basin reflect arc debris, which fits with their 130–100 Ma age. ORG, ocean-ridge granite; WPG, within-plate granite. Gravina arc samples from Rubin and Saleeby (1991). The lower plot has few total zircons, but their peaks suggest a southern Gondwana–Ouachita source. [Colour online.]

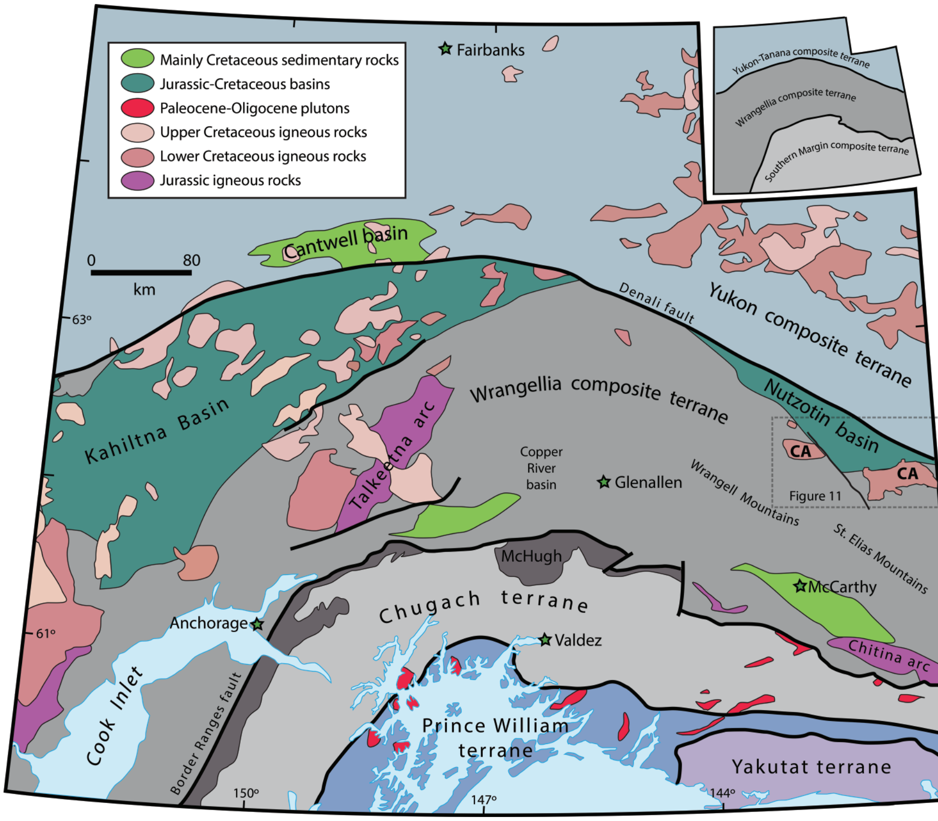
Laurussian collision zone as they contain a 278 Ma Permian peak, Cambrian peaks of 540 and 512 Ma, a 700 Ma Gondwana peak, and two Grenville age peaks of 1035 and 1190 Ma: there are no Paleoproterozoic or Archean zircons (Surpless et al. 2014). If the 700 Ma detrital zircons were derived from Gondwana, then the Methow block probably migrated northward while the basin remained open.

Gravina trough

Late Jurassic to Early Cretaceous volcaniclastic and mafic to intermediate volcanic rocks known as the Gravina succession (Fig. 7 and Supplementary Fig. S8[2]), generally interpreted to be remnants of a magmatic arc, crop out in Insular Alaska and on the mainland to the east (Berg et al. 1972; Rubin and Saleeby 1991; McClelland et al. 1992; Ricketts 2019). In the west, rocks of the Gravina succession unconformably overlie rocks of the Insular terrane, whereas to the east they sit structurally upon Permo-Triassic rocks of the Taku terrane. On the basis of Nd–Sr isotopic and detrital zircon characteristics, the Taku terrane is interpreted as part of Yukon–Tanana terrane, implying that an older suture must lie between the Insular and Intermontane superterranes in that region (Kapp and Gehrels 1998; Gehrels 2002; Giesler et al. 2016) and to the south (Crawford et al. 1987). Overall, metamorphic grade and deformation increase from west to east, reaching their peak in the high-grade gneiss complex east of the Great Tonalite Sill complex within the Coastal batholith (Brew et al. 1989).

Cohen and Lundberg (1993) demonstrated that rocks of the Gravina basin are dominantly volcaniclastic wackes and argued that they were derived from an arc terrane, which is commonly inferred to have been located to the west. More recent studies of the sedimentary rocks in the main outcrop belt, referred to as the western facies, and those in the eastern belt, showed that each belt contains different detrital zircon assemblages (Supplementary Fig. S8[2]). The western facies contains three main populations of 417–411, 165–140, and 120–105 Ma detrital zircons, whereas rocks of the eastern facies yield a pronounced Jurassic peak of 156 Ma with smaller groups of 380–310, 560–520, 1310–920, and 1955–1755 Ma, but none of Cretaceous age (Kapp and Gehrels 1998; Yokelson et al. 2015). The 120–105 Ma zircons in the western facies were likely derived from plutons of the Muir–Chichagof suite of biotite–hornblende granodioritic, tonalitic, and gabbroic plutons, located mainly to the west, along with the Jurassic 165–145 Ma Chilkut–Chichagof plutons (Brew and Morrell 1983), which are the likely source for the Jurassic detrital zircons in the western parts of the trough. Several 110–105 Ma Alaska-type mafic–ultramafic bodies crop out along the belt and attest to mafic arc magmatism (Himmelberg and Loney 1995; Rubin and Saleeby 2000). A dioritic pluton, the Jualin diorite (Supplementary Fig. S8[2]), is unconformably overlain by conglomerate holding boulders of the diorite, as well as a variety of metavolcanic and metasedimentary clasts (Redman 1984). The pluton was dated by U–Pb on zircon to be 105 ± 1 Ma (Kapp and Gehrels 1998), so provides a maximum age for the deposition of the upper units of the sedimentary succession within the Gravina basin, as well as the subsequent deformation and metamorphism.

Fig. 10. Geological sketch map of south-central Alaska illustrating the distribution of some of the geological features discussed in the text. CA, Chisana arc. Modified from Trop and Ridgway (2007). Note location of Fig. 11. [Colour online.]



Haeussler (1992) studied the structural development of the basin and found evidence for an early extensional phase characterized by normal faults, followed by younger isoclinal folds, thrust faults, and at least one 12 km wide shear zone. The eastern margin of the basin is tectonic with east-dipping reverse, or thrust, faults placing high-grade rocks of the Coast batholith and slices of Taku terrane over rocks of the Gravina Basin to the west (Rubin et al. 1990). Eastern facies rocks are typically at higher metamorphic grade, as they commonly contain kyanite or sillimanite and are more deformed than western facies rocks. They are dominated by Jurassic detrital zircons: Lower Cretaceous zircons are absent (Yokelson et al. 2015). The dramatic grade jump from west to east suggests that the westerly vergent faults are not thrusts but are reverse faults.

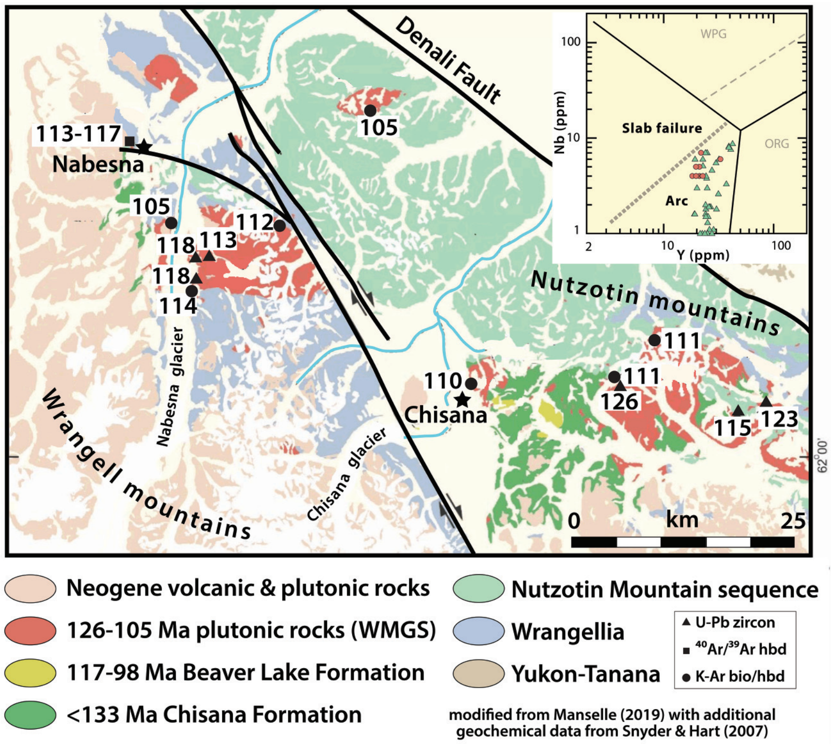
Following deposition of the Gravina sequence, the rocks were metamorphosed, deformed, and then intruded by a swarm of post-deformational 95–90 Ma plutons (Haeussler 1992; Himmelberg et al. 2004; Gehrels et al. 2009). To the east, many 99–89 Ma plutons were emplaced in the high-grade hinterland of the orogen (Gehrels et al. 2009; Girardi et al. 2012). The plutons are generally plagioclase–seriate bodies of biotite–hornblende granodiorite and tonalite (Douglass et al. 1989; Himmelberg et al. 2004). Their emplacement age is poorly constrained by modern standards but appears to be between 102 and 88 Ma. The largest is the post-tectonic Moth Bay pluton (Supplementary Fig. S8[2]) (Cook et al. 1991; Saleeby 2000; Rubin and Saleeby 1987, 2000) from which a sample yielded U–Pb analyses of highly discordant zircon fractions with an age of 102 +3/−2 Ma, and a $^{40}Ar/^{39}Ar$ date of 96 ± 1 Ma (Sutter and Crawford 1985). Other plutons were dated as 90.5 and 88.5 Ma with K–Ar ages of 93–80 Ma (Saleeby 2000). The Moth Bay pluton postdates deformation and metamorphism of the schists and is a biotite–epidote–hornblende tonalite emplaced at about 9 kbar on the basis of geobarometry in kyanite–sillimanite-bearing schists in the aureole and Al-content in hornblende from the pluton (Cook et al. 1991). These data collectively demonstrate that the Gravina basin was inverted, thickened, cut by plutons, and exhumed — all in <10 million years. Sparse trace element geochemistry is only available for a few samples from the Moth Bay body and they plot as post-collisional rocks (Supplementary Fig. S8[2]).

Wrangellia – North America collision in Alaska

To the north, the Peninsular Ranges orogen continues around the Gulf of Alaska more or less parallel to its arcuate shoreline. There, an older Jurassic collision between the Peninsular and Wrangellian terranes formed what is commonly referred to as the Wrangellia composite terrane (Trop and Ridgway 2007). The area is transected and complicated by younger strike slip faults, such as the Border Ranges and Denali faults, both of which have poorly constrained displacement. The Wrangellia composite terrane is interpreted to have collided with the more northerly Yukon–Tanana terrane during closure of the siliciclastic Kahiltna basin in a double northward-dipping tectonic scheme (Pavlis et al. 2019, 2020) or in a two-phase hypothesis involving southward subduction during the Jurassic within the basin, followed by northward-directed subduction along the south side of the Wrangellia composite terrane after the collision (Sigloch and Mihalynuk 2020).

The relatively low grade Wrangellian terrane (Fig. 10) comprises carbonates and a 1–3.5 km thickness of dominantly subaerial high-Ti pillow and subaerial basalts (Lassiter et al. 1995; Greene et al. 2008). They overlie volcanic and sedimentary rocks of the Permian Skolai Group (Trop et al. 2002). Following the

Fig. 11. Geologic sketch map showing distribution and ages of 133–100 Ma Chisana arc rocks sitting unconformably upon deformed rocks of the Nutzotin basin and Wrangellia (modified from Richter et al. 2006 and Manselle 2019 with additional data from Snyder and Hart 2007). Note that both volcanic (green) and plutonic (brick red) rocks plot in the arc field on a Nb vs. Y discrimination diagram of Hildebrand and Whalen (2017). See Fig. 10 for location. bio, biotite; hbd, hornblende; WMGS, White Mountain granitoid suite. [Colour online.]



collision of Wrangellia with the Peninsular terrane, the amalgamated terrane contained a variety of Jurassic rocks, including the volcano-plutonic 202–169 Ma Talkeetna (DeBari and Coleman 1989; Clift et al. 2005; Rioux et al. 2007) and Chinitna arcs (Plafker et al. 1989), as well as mid-Jurassic sedimentary rocks of the Tuxedni Group (Trop et al. 2005; Amato et al. 2007). We plotted unpublished geochemistry from plutonic rocks associated with the Talkeetna arc (M. Rioux, personal communication, 2019) on our discrimination diagrams and found that most have arc geochemistry, but one 152.7 ± 1.3 Ma pluton, mapped as transecting the Peninsular–Wrangellia suture (Rioux et al. 2007), displays typical post-collisional slab break-off characteristics with Sm/Y > 2.5, Sr/Y > 20, and Nb/Y > 0.4. Thus, map relations, timing, and geochemistry provide a minimum age for the Penisular–Wrangellian collision. The youngest arc-like pluton dated by Rioux et al. (2007) is 168.8 Ma and provides a maximum age for collision. Following the Jurassic collision and the Early Cretaceous opening of the Kahiltna basin, rocks of the Wrangellia composite terrane lay to the south of the Kahiltna basin, whereas rocks of the Yukon composite terrane lay to the north, although the precise extent of the terrane is uncertain due to unresolved displacement on the Denali fault. We start by examining the Cretaceous rocks and their interactions, then examine the Jurassic collision as it is similar in many respects to the Peninsular Ranges orogeny.

Chisana arc

Wrangellia contains the only evidence within the region for an arc complex (Chisana arc) during the Early Cretaceous. This 133–98 Ma arc, is best exposed in the Nutzotin Mountains of south-central Alaska, where andesitic lavas, along with subordinate volcaniclastic rocks, unconformably overlie deformed sedimentary rocks of the Nutzotin basin (Fig. 11), and are intruded by calc-alkaline plutons of the White Mountain granitoid suite (Manselle 2019; Snyder and Hart 2007; Trop et al. 2020). These relations, plus detrital zircon ages from sedimentary rocks of the dismembered Dezadeaash–Nutzotin basin (Lowey 1998, 2019; Fasulo 2019) — all older than about 133 Ma — demonstrate that the Nutzotzin and Kahiltna basins were not fault-bounded segments of the same basin, or even the Gravina basin to the south, as commonly claimed (Sigloch and Mihalynuk 2017, 2020; Pavlis et al. 2019, 2020). However, correlation of the Nutzotin basin with the coeval Peñasquitos, Cucurpe, and Mariposa formations farther south is possible. These successions all carry the Tithonian bivalve, *Buchia piochii*, and were deformed by around 130 Ma, prior to the deposition of overlying rocks, which were basinal and (or) arc rocks of the Peninsular Ranges orogen (Mauel et al. 2011; Peryam et al. 2012; Kimbrough et al. 2014; Manuszak et al. 2007; Manselle 2019). Rocks of the Gravina basin show a clear gap in detrital zircon ages between 147 and 122 Ma (Yokelson et al. 2015), but there is no obvious break in the stratigraphic section (Supplementary Fig. S8[2]).

Farther west, a number of Early Cretaceous tonalite to trondhjemitic plutons (123 ± 2 Ma) intrude uppermost Middle Jurassic metasedimentary rocks of Wrangellia (Labrado et al. 2015) and may also represent arc magmas. In a recent study, using sparse geochemistry and isotopes from the intrusions, Mahar et al. (2019) suggested the intrusions might have been emplaced above a slab-window as they are isotopically primitive low-K calc-alkaline rocks. Although most of their analyzed samples show elevated Sr/Y values, all of them have Sm/Yb <2.5, which suggests that they are arc magmas (Hildebrand and Whalen 2017).

Kahiltna trough

The now collapsed and inverted Kahiltna trough (Fig. 10) lies to the north (current coordinates) and sits between more outboard Wrangellia and the inboard metamorphic Yukon–Tanana and Farewell terranes (Box et al. 2019; Kalbas et al. 2007; Trop and

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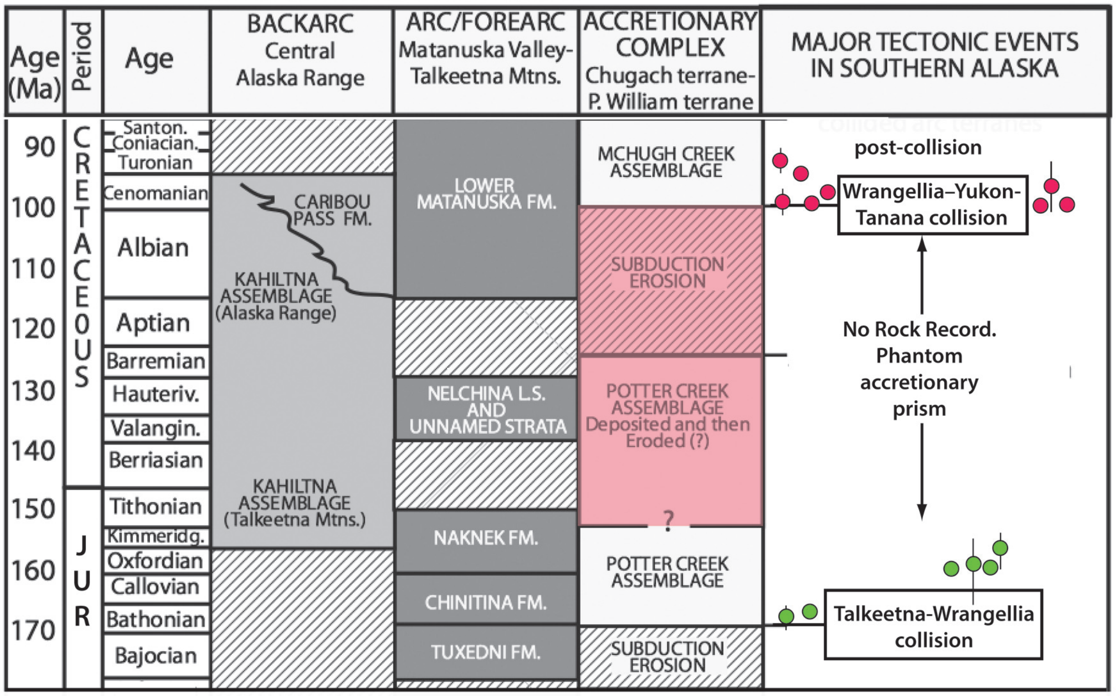
Fig. 12. Modified stratigraphic chart from Amato et al. (2013), showing their concept of the relative locations of different stratigraphic packages through time. The coloured dots represent their maximum depositional ages from detrital suites and clearly show the >50 million year gap in sedimentation, which because they held to a northerly directed subduction model, were forced to attribute to subduction erosion. Our model does not require a phantom accretionary prism because we found that subduction was southerly on the other side of Wrangellia. We attribute the sedimentation of the McHugh Creek assemblage to post-collisional exhumation on the outboard side of the collision zone. The 100–90 Ma detrital zircons were derived from post-collisional plutons emplaced following slab break-off, similar to those found in areas to the south. [Colour online.]



Ridgway 2007; Hampton et al. 2007, 2010). Hampton et al. (2010) measured eight sections through the basin, collected and dated detrital zircons, and found MDAs based on the weighted mean of the three youngest zircons to range from 143 to 102 Ma. Box et al. (2019), on the basis of sandstone petrography and detrital zircon age populations, divided the rocks of the basin, which are dominantly turbiditic, into three distinct petrofacies, the lower two of which remained as separate deposystems until the Late Cretaceous: (*i*) a northwestern belt with dominantly quartzose debris with MDAs all older than 100 Ma, typically 107–103 Ma; (*ii*) a southeastern belt characterized by igneous debris with detrital zircon age peaks ranging from about 170 to 142 Ma, with MDAs ranging from 146 to 103 Ma; and (*iii*) a younger belt, also dominated by igneous debris and with Late Cretaceous detrital zircon age peaks, in places accompanied by a broad 160 Ma peak, ash beds dated at 97 and 93 Ma, and Cenomanian–Turonian MDAs. Similar to more southern sectors of the basin, such as the Gravina and Arperos sectors, the basin must have been sufficiently wide, or configured such that debris from either side did not cross the basin and intermingle in any significant abundance (Box et al. 2019).

There were two obvious episodes of deformation of the basin-fill rocks, with the oldest bracketed between 103 and 97 Ma and consisting of southeast-dipping thrust faults with tight overturned folds that verge northwest, which represent deformation during the Peninsular Ranges orogeny; whereas the youngest is bracketed to be about 80 Ma (Box et al. 2019) and probably represents Laramide deformation.

Yukon–Tanana and Farewell terranes

North of the Kahiltna basin and across the Denali fault, the metamorphic Yukon–Tanana terrane and the Farewell terrane, collectively referred to as the Yukon Composite terrane (Fig. 10), are both mainly composed of Paleozoic metasedimentary and meta-igneous rocks. The more westerly Farewell terrane comprises Cambrian to Pennsylvanian carbonates, Devonian and Triassic phosphatic black shale, barite, and sandstone, with a variety of gabbroic sills and pillowed basalts (Bundtzen and Gilbert 1983; Bradley et al. 2006). The Yukon–Tanana terrane sensu lato is highly variable with complex relations and generally poor outcrop, but it is dominated by Paleozoic greenschist to amphibolite grade metamorphic rocks, in places tectonically overlain by Mississippian to Triassic oceanic igneous and metamorphic rocks (Templeman-Kluit 1979; Dusel-Bacon et al. 2006; Colpron et al. 2006). In the extreme southeastern sector of the terrane, east of Whitehorse and crossing into the Selwyn basin, are some 115–95 Ma plutons and ignimbrites with typical slab failure geochemistry; but they are likely related to the Sevier orogenic event (Hart et al. 2004; Hildebrand 2015; Hildebrand and Whalen 2017).

Cu–Au porphyry mineralization is commonly associated with slab break-off magmatism (Solomon 1990; de Boorder et al. 1998; Cloos and Housh 2008; Hildebrand 2009; Hou et al. 2015; Hildebrand and Whalen 2017) and even though the heavily mineralized 99–88 Ma Pebble porphyry complex (Supplementary Fig. S9[2]) lies several hundred kilometres to the west, it is likely also a 100–85 Ma slab failure plutonic suite (Olson 2015) related to the Wrangellia – Yukon–Tanana – Farewell composite collision.

McHugh complex

An outboard trough, exposed along the south side of Wrangellia composite terrane as the McHugh complex, was active after 101 Ma when it received detritus, including 99–82 Ma detrital zircons (Amato and Pavlis 2010; Amato et al. 2013). These workers interpreted the McHugh complex as deposited within a fore-arc setting (Fig. 12), but temporally, spatially, and compositionally the basin is similar to the Upper Cretaceous Valle Formation of Baja California (Kimbrough et al. 2001), the Upper Cretaceous Great Valley Group of central California (Mansfield 1979), and the Upper Cretaceous Nanaimo basin of southern British Columbia (Matthews et al. 2017), all three of which developed behind the east-facing arc on the upper plate and contain Cenomanian–Turonian debris eroded from the pluton-riddled and exhuming hinterland. Additionally, plutons of the 123 ± 2 Ma tonalite–trondhjemite suite (Mahar et al. 2019), which on the basis of sparse geochemistry could be arc rocks, sit just inboard of the McHugh rocks, similar to other arc suites to the south.

As suggested earlier, these sedimentary rocks might typically be deposited in outboard troughs, or perhaps open ocean, during the later stages of arc-continent collision and mark slab failure, for it is at that time that the partially subducted lower plate is freed of its oceanic anchor, so rises rapidly (see Supplementary Fig. S10[2]). In the cases along the western margin of the Peninsular Ranges orogen, the reverse faults that separate the higher-grade hinterland from the lower-grade back-arc region probably approximate, or mark, the leading edge of the torn lower plate as it rose and exhumed the orogenic hinterland. Sedimentary debris was shed from the exhuming hinterland located to the east and deposited west of the reverse faults.

Summary of Wrangellian – Yukon–Tanana collision: implications for other models

The Cretaceous geology of south-central Alaskan parallels that seen farther south and so we include it in the Peninsular Ranges orogen. In Alaska, the Early Cretaceous seaway, recorded by strata of the Kahiltna basin, where contrasting clastic debris was shed from opposite flanks of the trough, coupled with the Early Cretaceous Chisana arc sitting on lower-grade Wrangellian basement lying to the south, and contrasting with the higher-grade metamorphic Yukon composite terrane, devoid of Early Cretaceous arc rocks, are typical relations of the Peninsular Ranges orogen farther south. The south-dipping, northward-vergent thrust faults in the basin indicate that the Yukon composite terrane was the lower plate as the basin closed, a scenario consistent with the contrasting metamorphic grades between this terrane and Wrangellia. Some possible post-collisional slab failure plutons, such as the Tok-Tetlin, the Cheslina, and the Gardiner Creek plutons have ages ranging from 94 to 88 Ma (Richter et al. 1975; Hart et al. 2004), but crop out north of the Denali fault, so may have been located elsewhere during the Upper Cretaceous.

Overall, the conjectures of Pavlis et al. (2019, 2020) fit well with the standard paradigm of eastward-dipping subduction, but are inconsistent with the geological record. For example, their model invokes two northward-dipping subduction zones, one beneath Wrangellia and another beneath Yukon–Tanana, but there is no evidence for an Early Cretaceous arc on Yukon–Tanana terrane, nor is there any evidence for an accretionary prism along the southern side of Wrangellia over the critical 150–100 Ma interval (Fig. 12). We claim that our model for west-dipping subduction beneath an Early Cretaceous arc sitting on Wrangellian basement within the Peninsular Ranges composite terrane, and attempted subduction of the higher-grade Yukon–Tanana – Farewell block during the Peninsular Ranges orogeny, is more compatible with the known geology. For example, we are not compelled to invoke any processes, such as subduction erosion, to explain the absence of an accretionary prism.

The model of Sigloch and Mihalynuk (2017, 2020) focuses largely on the Jurassic, and so does not incorporate the critical evidence for Early Cretaceous rifting and development of the arc on the western Peninsular Ranges composite terrane (Hildebrand and Whalen 2014*b*, 2017). Considerable geological evidence for the development of the Early Cretaceous seaway negates the requirement for eastward-dipping subduction beneath the amalgamated Jurassic North American block during the Cretaceous.

We agree with Sigloch and Mihalynuk (2017, 2020) that the Jurassic collision of the Talkeetna arc with Wrangellia was, in current coordinates, south-dipping (see Hildebrand 2013, p. 71). Along the north side of the Talkeetna arc, south-dipping thrust faults accompanied by deposition of clastic debris — sitting unconformably upon platformal siliciclastics and carbonate strata of Wrangellia — coarsens upward from marine mudstones and sandstones to conglomerate (Trop et al. 2002; Manuszak et al. 2007). These relations suggest that the leading edge of Wrangellia was pulled beneath the Talkeetna arc during the collision.

On the basis of our analysis of unpublished geochemical data (M. Rioux, personal communication, 2019) the youngest dated arc rock in the Talkeetna arc is a ~169 Ma quartz diorite pluton, which provides a maximum age for the collision of the arc with Wrangellia. And as noted earlier, the only post-collisional slab failure pluton yet recognized in the Talkeetna collision zone is 153 Ma, so there is a gap of some 16 million years between the youngest known arc rock (dated at 169 Ma) and the only documented post-collisional slab failure pluton. On the south side of the Talkeetna–Wrangellia collision zone, coarse 167–150 Ma batholithic debris of the Naknek Formation was shed southward from the hinterland belt as it was exhumed on north-dipping reverse faults (Trop et al. 2005). A more recent regional study of the formation (Herriott et al. 2017, 2019), including detrital zircon analyses, suggest that the trough represents the beginning of "ubiquitous batholithic provenance" generated by rapid exhumation of the plutonic roots of the Talkeetna arc with tectonism as the driving force. This scenario is precisely what occurs on the opposite side from the collisional suture and fore arc in the outboard back-arc setting along the Peninsular Ranges orogen from Mexico to Alaska or in younger arc-continent collisions like Papua, New Guinea (Supplementary Fig. S10[2]).

Thus, in contrast to Pavlis et al. (2019, 2020) and Sigloch and Mihalynuk (2020), we favour westerly dipping subduction for both the Jurassic and the Cretaceous collisions involving Wrangellia. We note that following collisions, coarse debris, including abundant material largely derived from post-collisional plutons and their wall rocks, was shed outboard (westward) from the hinterland belt as it was rapidly exhumed. In the Sierran paradigm of long-lived eastward subduction, these basins are considered fore-arc basins, but in our model, they form after collision, not when the arc was active. Thus, the basins should be dominated by post-collisional debris with detrital zircons largely derived from slab failure plutons rather than older arc magmatism. The spectrum of detrital zircons in the Chinitna and lower Naknek formations of the Naknek basin suggests that the Talkeetna–Wrangellian collision took place at about 167 Ma and that the stratigraphically lower Chinitna contains more arc debris (202–167 Ma) than the overlying Naknek Formation; but both formations are dominated by 161–157 Ma age peaks (Herriott et al. 2019), consistent with our back-arc model for deposition of debris from orogenic hinterlands.

Cordilleran outliers: Salinia and the Peruvian coastal batholith

Although here we progressed from south to north, we believe that another sector of the Peninsular Ranges orogen may have been dismembered by strike-slip fault(s) and is now divided between central California and coastal Peru (Hildebrand and Whalen 2014*a*). Located west of the Great Valley of California, and commonly considered by some to have formerly occupied the space between the Sierra Nevada and the Peninsular Ranges (Ducea 2001; Barbeau et al. 2005; Chapman et al. 2014), the Salinian block has long been recognized as an anomalous block (Ross 1978) and comprises 3–7.5 kbar amphibolite–granulite facies gneiss and schist lacking pure quartzite and carbonate rocks characteristic of the North American margin (Ross 1977), all cut by deformed plutons in the age range of 130–103 Ma, as well as more easterly, nondeformed 100–82 Ma plutons ranging in composition from gabbro to granodiorite (Mattinson 1978, 1990; James 1992; Kistler and Champion 2001; Kidder et al. 2003; Chapman et al. 2014). The plutons with trace element geochemistry plot as slab failure rocks (Hildebrand and Whalen 2017, p. 37), but the Salinian block does not contain low-grade, pre-100 Ma volcanics and only a few 130–103 Ma arc plutons, typical of other sectors of the Peninsular Ranges orogen. This led Hildebrand and Whalen (2014*a*) to suggest that it once belonged with the Coastal batholith of Peru, which some recognized to be another anomalous block (Loewy et al. 2004). There, 7–9 km of relatively low-grade arc rocks, the mainly latest Aptian–Albian rocks of the Casma and related groups, were deposited in the Huarmey–Cañete trough (Cobbing 1978; Atherton et al. 1985; Pitcher 1993; de Haller et al. 2006), and were deformed at 100 Ma. Although the batholith contains a few post-collisional plutons, an exhumed metamorphic hinterland and voluminous 100–85 Ma slab failure plutons are absent.

We showed earlier that, except for a small gap in the Mojave Desert where the Peninsular Ranges orogen was transected by the Late Cretaceous – Paleogene Laramide orogen, the thrusts and post-collisional plutons of the Peninsular Ranges orogen are continuous from Mexico to Utah. Thus, instead of lying between the Sierra Nevada and the Peninsular Ranges batholiths, the Salinia block was more likely located to the south, and was possibly joined with the Arequipa block and its cover of Casma volcanics, which, on the basis of magmatic ages and deformation, collectively formed another sector of the Peninsular Ranges orogen (Hildebrand and Whalen 2014*a*, 2017). Although conflicting paleomagnetic data for the Salinian block exist (Champion et al. 1984; Whidden et al. 1998), paleontological data suggest that the faunal assemblage of Salinia is a reasonable match with those in the Peninsular Ranges of southern California (Elder and Saul 1993), and they are considered to be far-traveled rocks (Hagstrum et al. 1985).

Surface geology and mantle tomography

Three of the Jurassic and two — or possibly all — of the three Cretaceous arc-continent collisions were apparently built above westerly dipping subduction zones. The three Jurassic collisions, readily resolved in the western Sierra Nevada (Hildebrand 2013), likely occurred offshore from North America (Johnston 2008), and the initial docking of the partly assembled ribbon continent of Hildebrand with North America occurred during the Sevier orogeny at 124–120 Ma (Hildebrand 2013, 2014). This event was followed by the Peninsular Ranges orogeny at 100 Ma, and the Laramide during the Campanian, none of which produced an arc on cratonic North America.

Since the early 90s, a huge, steeply dipping fast zone in the mantle beneath eastern North America was recognized by tomography and interpreted as a fossil subducted slab (Grand 1994). As North America migrated westward during the opening of the Atlantic Ocean, it overrode the torn and sinking oceanic lithosphere previously located to the west. Sigloch and Mihalynuk (2013) showed how westerly — but not easterly — dipping slabs, could form steeply inclined to vertical slab walls in the mantle as North America migrated westward.

The immense size of the East Coast fast tomographic anomaly (Sigloch 2011), with a width of 8–10 degrees of longitude, which at 40°N are 85 km/degree, translate to a width of 680–850 km. Commonly, the torn and sinking slabs are shown as buckling concertina-fashion as they sink (Lee and King 2011; Sigloch and Mihalynuk 2017) to produce the required width; but in this case the anomaly probably represents an amalgamation of all the Jurassic and Cretaceous slabs, because the thickness of older oceanic lithosphere is about 80–100 km (Hamza and Vieira 2012). If subducted slabs are indeed amalgamated — or appear to be so at the current resolution of mantle tomography — and new ocean basins opened within the Cordillera, as documented here, then attempts to match mantle tomography with surface geology of the Cordillera cannot be quantitatively evaluated as claimed (Clennett et al. 2020).

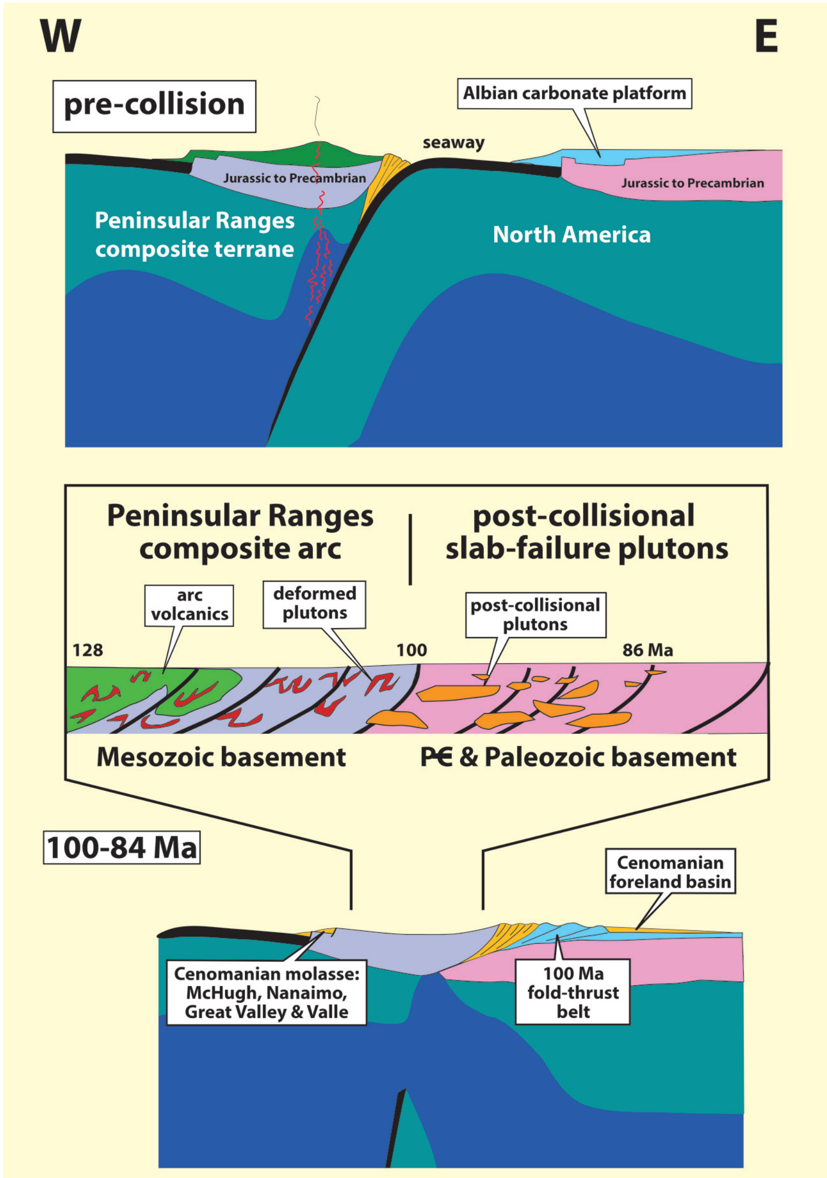
Whereas the amalgamated, or tomographically irresolvable slabs, imply a long-lived zone of westerly subduction, Hildebrand and Whalen (2014*a*) noted the two long-lived mantle upwelling zones, Jason and Tuzo (Burke and Torsvik 2004; Torsvik et al. 2008; Burke 2011; Spencer et al. 2019), and postulated that a complimentary long-lived mantle downwelling zone existed to the west of North America. They speculated that it may have formed a boundary between the Panthalassic and Pacific realms, and when the various arcs and other fragments that had accumulated above the downwelling zone collided with the Americas, Panthalassa ceased to exist. The continent then overrode the torn slabs beneath the accreted collage. We wonder whether the demise of Panthalassa marks the switchover from west-dipping subduction, which appears to have dominated Panthalassic–Cordilleran interactions, to the current eastwardly dipping Pacific plates beneath the Americas.

Conclusions

In this two-part contribution, we summarized our geological evidence over the length of the Cordillera to demonstrate that the paradigm of long-lived eastward subduction runs counter to the bulk of evidence. We conclude with the following points.

1. The Peninsular Ranges orogen is a ∼100 Ma orogenic belt that extends from Mexico to Alaska and beyond. The orogen formed when a trough, open for about 35 million years along the western margin of North America, closed by westerly subduction, juxtaposing a Lower Cretaceous arc complex built on the ribbon-like Peninsular Ranges composite terrane along the western side of the trough, over a passive continental margin, which was locally capped by a west-facing carbonate platform developed on the eastern North American side of the trough (Fig. 13).
2. Within a million years or so following the collision, the hinterland was exhumed and intruded by gregarious tonalite–granodiorite–granite plutons, which were emplaced over a period of 10–15 million years. The timing suggests that the plutons and exhumation formed in tandem when the oceanic lithosphere broke off from the partially subducted North American plate.
3. Because the trough formed after at least one Jurassic collision and its post-collisional magmatism, many different rocks and terranes formerly attached to North America were rifted and separated at about 135 Ma, only to return at 100 Ma, likely in different places than their original locations.
4. The large mid-Cretaceous batholiths of the North American Cordillera are composed of two contrasting magmatic suites derived from distinct mantle sources and emplaced at different times. The older arc suite, which developed on the Peninsular Ranges composite terrane from Mexico to Alaska, represents a generally low-standing marine arc built on thinned lithosphere over a westward-dipping subduction zone, whereas the younger suite was post-collisional and invaded the orogenic hinterland during exhumation due to break-off and melting of the subducting slab.
5. The so-called “flare-up” events in Cordilleran arcs are the result of collision followed by slab break-off, not arc magmatism.
6. Those who utilize Andino-type, or cyclic hi-flux, models for the development of Cordilleran batholiths, fail to recognize that the transition from arc magmatism to post-collisional

Fig. 13. Our tectonic plate scale model for the Peninsular Ranges orogeny involves closure of a Lower Cretaceous seaway by west-directed subduction and arc magmatism from ~130 Ma until the collision of the arc with North America at 100 Ma. The competing buoyancies of the oceanic and cratonic lithosphere led to rapid tearing and break-off of the subducted plate and an influx of 99–84 Ma post-collisional magmatism during exhumation of the orogenic hinterland. During exhumation and plutonism, mainly 99–90 Ma debris was shed westward into the old back-arc region. These relations hold over the length of the Cordillera, from southern Mexico to Alaska. P€, Precambrian. [Colour online.]



hi-flux magmatism occurred rapidly, perhaps within a million years, so that there is simply no time to thicken the crust by underplating or for heat transfer by conduction to melt underthrust cratonic material.

7. The post-collisional magmas appear to have been derived from melting of the basaltic–gabbroic upper part of the subducted oceanic lithosphere augmented by fractional melting of the SCLM as magmas rose toward the crust. Thus, slab break-off magmas have trace element concentrations and ratios similar to slab window rocks, but where they rise through old and enriched cratonic lithosphere they have enriched radiogenic isotopes.
8. There is no compelling evidence along the entire western edge of the Peninsular Ranges composite terrane for a fore-arc basin or accretionary prism during Early Cretaceous arc magmatism. Instead, voluminous quantities of material were shed westward into the back-arc region after the 100 Ma collision and termination of arc magmatism, when abundant detrital zircons from the 100–90 Ma post-collisional plutons document rapid exhumation of the orogenic hinterland.
9. Retro-arc models for the Sevier thrust-fold belt must be reconsidered, as there was no eastward subduction beneath North America at about 120 Ma when the Sevier thrusting initiated. Instead, the 130–100 Ma Alisitos and related arc segments involved in the Peninsular Ranges orogeny were built offshore above westerly, not easterly, dipping subduction zones.

10. We suggest that the huge East Coast tomographic anomaly, which underlies eastern North America, could represent several amalgamated slabs rather than a single slab that buckled. In our conception, the three westerly subducted Jurassic slabs occurred offshore beneath the ribbon continent, but failure of the west-dipping slabs, represented by the Sevier and Peninsular Ranges orogens, occurred during and after the 120 Ma Sevier collision of the ribbon with North America. As North America migrated westward, it collided with the ribbon continent and overrode the Jurassic slabs: the two Cretaceous slabs were added to the mass of subducted slabs shortly afterwards because they could not penetrate the slab wall.

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