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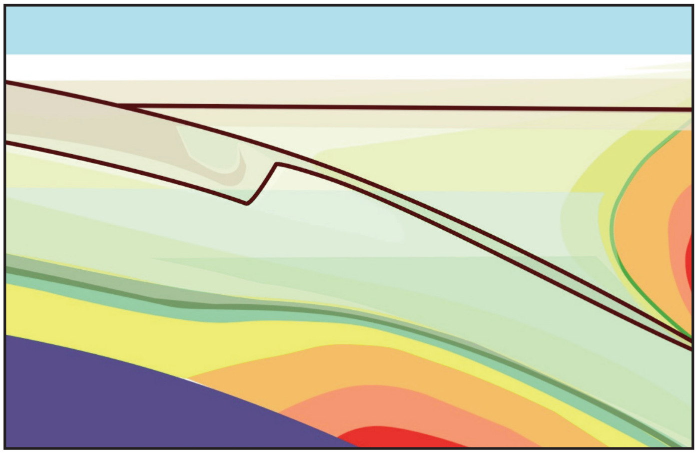
Résumé de l'article

Les reconstitutions de plaques montrent que si le panache de Yellowstone avait existé avant 50 Ma, il aurait été recouvert par la lithosphère océanique située à l'ouest de la plaque nord-américaine (PNA). Dans le contexte de modèles de subduction de longue durée vers l'est de la lithosphère océanique sous la PNA, avec le temps, la marge continentale de la PNA aurait progressivement neutralisé le panache de Yellowstone, et on devrait en voir les effets dans le registre géologique. Le rôle de ce panache de Yellowstone « ancestral » et de son renflement de surface régional associé sur l'évolution tectonique du Sud-ouest des États-Unis au Mésozoïque–Cénozoïque tardif est reconsidéré ici à la lumière de données récentes, de terrain, analytiques et géophysiques, de contraintes découlant de constructions paléogéographiques affinées, et d'idées nouvelles découlant d'une modélisation géodynamique récente de l'interaction d'un panache et d'une zone de subduction. Les modèles géodynamiques suggérant que l'ascension des panaches soient bloquée ou détruite dans les zones de subduction ont attiré l'attention sur le rôle d'hiatus ou de déchirures dans la plaque subduite qui permettent le passage du matériau du panache de la plaque inférieure à la plaque supérieure pendant la subduction. Ces modèles impliquent que le flux ascendant des panaches peut être sensiblement dévié alors que le matériau du panache migre de la plaque inférieure à la plaque supérieure, de sorte que la connexion entre la trace du point chaud calculée à partir des reconstructions de la plaque et les manifestations de l'activité du panache dans la plaque supérieure peut être bien plus diffuse que sa contrepartie du domaine océanique. D'autres modèles géodynamiques appuient l'hypothèse selon laquelle la subduction du matériau de plateau océanique sous la PNA correspond à la génération d'une plaque plate, particularité qui a longtemps été considérée comme caractéristique déterminante de l'orogenèse de Laramide dans l'ouest des États-Unis, épisode orogénique dominante de la fin du Mésozoïque au début du Cénozoïque affectant la PAN. Au cours des 20 dernières années, un nombre croissant d'éléments de preuve provenant d'une variété d'approches suggèrent qu'un panache existait bien entre 70 et 50 Ma dans le domaine océanique près de la marge la PNA, en un endroit et avec une vigueur similaires au point chaud de Yellowstone moderne. Le cas échéant, l'interaction de ce panache avec la marge aurait été précédée de celle de son renflement de surface et du plateau océanique connexe, scénario qui aurait pu générer la subduction de la plaque plate qui caractérise l'orogenèse Laramide. À moins que ce panache n'ait été détruit par subduction, il serait entré dans une période d'incubation lorsqu'il a été recouvert par la marge nord-américaine. Au cours de cette période d'incubation, le matériau du panache aurait pu migrer dans la plaque supérieure par des fenêtres ou déchirures de la plaque ou autour des marges latérales de la plaque, conformément aux modèles récents de laboratoire. La trace de l'activité magmatique résultante pourrait se trouver alors à une distance considérable de la trace du point chaud calculée. La distribution actuelle des panaches et de leurs renflements de surface suggère que leur interaction avec les zones de subduction devrait être un phénomène courant dans le registre géologique. Si tel est le cas, l'évolution du Mésozoïque–Cénozoïque tardif de l'Amérique du Nord occidentale peut représenter un analogue relativement moderne pour de tels processus.

ANDREW HYNES SERIES: TECTONIC PROCESSES



The Role of the Ancestral Yellowstone Plume in the Tectonic Evolution of the Western United States

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SUMMARY

Plate reconstructions indicate that if the Yellowstone plume existed prior to 50 Ma, then it would have been overlain by oceanic lithosphere located to the west of the North American plate (NAP). In the context of models supporting long-lived easterly directed subduction of oceanic lithosphere beneath the NAP, the Yellowstone plume would have been progressively overridden by the NAP continental margin since that time, the effects of which should be apparent in the geological record. The role of this 'ancestral' Yellowstone plume and its related buoyant swell in influencing the Late Mesozoic–Cenozoic tectonic evolution of the southwestern United States is reviewed in the light of recent field, analytical and geophysical data, constraints provided by more refined paleogeographic constructions, and by insights derived from recent geodynamic modelling of the interaction of a plume and a subduction zone.

Geodynamic models suggesting that the ascent of plumes is either stalled or destroyed at subduction zones have focused attention on the role of gaps or tears in the subducted slab that permit the flow of plume material from the lower to the upper plate during subduction. These models imply that the ascent of plumes may be significantly deflected as plume material migrates from the lower to the upper plate, so that the connection between the hot spot track calculated from plate reconstructions and the manifestations of plume activity in the upper plate may be far more diffuse compared to the more precise relationships in the oceanic domain. Other geodynamic models support the hypothesis that subduction of oceanic plateau material beneath the NAP correlates with the generation of a flat slab, which has long been held to have been a defining characteristic of the Laramide orogeny in the western United States, the dominant Late Mesozoic–Early Cenozoic orogenic episode affecting the NAP.

Over the last 20 years, a growing body of evidence from a variety of approaches suggests that a plume existed between 70 and 50 Ma within the oceanic realm close to the NAP margin in a similar location and with similar vigour to the modern Yellowstone hot spot. If so, interaction of this plume with the margin would have been preceded by that of its buoyant swell and related oceanic plateau, a scenario which could have generated the flat slab subduction that characterizes the Laramide orogeny.

Unless this plume was destroyed by subduction, it would have gone into an incubation period when it was overridden by the North American margin. During this incubation period, plume material could have migrated into the upper plate via slab windows or tears or around the lateral margins of the slab, in a manner consistent with recent laboratory models. The resulting magmatic activity may be located at considerable distance from the calculated hot spot track.

The current distribution of plumes and their buoyant swells suggests that their interaction with subduction zones should be common in the geological record. If so, the Late Mesozoic–Cenozoic evolution of western North America may represent a relatively modern analogue for such processes.

RÉSUMÉ

Les reconstitutions de plaques montrent que si le panache de Yellowstone avait existé avant 50 Ma, il aurait été recouvert par la lithosphère océanique située à l'ouest de la plaque nord-américaine (PNA). Dans le contexte de modèles de subduction de longue durée vers l'est de la lithosphère océanique sous la PNA, avec le temps, la marge continentale de la PNA aurait

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progressivement neutralisé le panache de Yellowstone, et on devrait en voir les effets dans le registre géologique. Le rôle de ce panache de Yellowstone « ancestral » et de son renflement de surface régional associé sur l'évolution tectonique du Sud-ouest des États-Unis au Mésozoïque–Cénozoïque tardif est reconsidéré ici à la lumière de données récentes, de terrain, analytiques et géophysiques, de contraintes découlant de constructions paléogéographiques affinées, et d'idées nouvelles découlant d'une modélisation géodynamique récente de l'interaction d'un panache et d'une zone de subduction.

Les modèles géodynamiques suggérant que l'ascension des panaches soient bloquée ou détruite dans les zones de subduction ont attiré l'attention sur le rôle d'hiatus ou de déchirures dans la plaque subduite qui permettent le passage du matériau du panache de la plaque inférieure à la plaque supérieure pendant la subduction. Ces modèles impliquent que le flux ascendant des panaches peut être sensiblement dévié alors que le matériau du panache migre de la plaque inférieure à la plaque supérieure, de sorte que la connexion entre la trace du point chaud calculée à partir des reconstructions de la plaque et les manifestations de l'activité du panache dans la plaque supérieure peut être bien plus diffuse que sa contrepartie du domaine océanique. D'autres modèles géodynamiques appuient l'hypothèse selon laquelle la subduction du matériau de plateau océanique sous la PNA correspond à la génération d'une plaque plate, particularité qui a longtemps été considérée comme caractéristique déterminante de l'orogenèse de Laramide dans l'ouest des États-Unis, épisode orogénique dominante de la fin du Mésozoïque au début du Cénozoïque affectant la PAN.

Au cours des 20 dernières années, un nombre croissant d'éléments de preuve provenant d'une variété d'approches suggèrent qu'un panache existait bien entre 70 et 50 Ma dans le domaine océanique près de la marge la PNA, en un endroit et avec une vigueur similaires au point chaud de Yellowstone moderne. Le cas échéant, l'interaction de ce panache avec la marge aurait été précédée de celle de son renflement de surface et du plateau océanique connexe, scénario qui aurait pu générer la subduction de la plaque plate qui caractérise l'orogenèse Laramide.

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La distribution actuelle des panaches et de leurs renflements de surface suggère que leur interaction avec les zones de subduction devrait être un phénomène courant dans le registre géologique. Si tel est le cas, l'évolution du Mésozoïque–Cénozoïque tardif de l'Amérique du Nord occidentale peut représenter un analogue relativement moderne pour de tels processus.

Traduit par le Traducteur

INTRODUCTION

Although not universally accepted (e.g. Anderson 1994; Foulger and Natland 2003; King 2007), hot spots are considered to reflect upwelling of sub-lithospheric mantle plumes (Morgan 1971, 1972). In a classical sense, a sub-oceanic mantle plume is envisaged to have a central conduit which can underplate a 400–1000 km diameter area of the lithosphere, creating an oceanic plateau above a buoyant swell (Richards et al. 1988; Sleep 1990). This buoyant swell develops a pronounced asymmetry as it becomes elongated 'downstream,' in some instances by as much as 2500 km, by the motion of the overriding plate (McKenzie 1983; Sleep 1990; Geist and Richards 1993; Ribe and Christiansen 1994). Estimates for the area of oceanic plateaus that might have been overridden by the North American plate (NAP) range up to 0.48 million km^2 (comparable in size to the state of California; Liu et al. 2010), and imply that the plateaus are significantly smaller in areal extent than the swell.

Although plumes may rise from different boundary layers in the mantle (Courtillot et al. 2003), recent tomographic images and geodynamic models suggest that many emanate from the edges of regions known as Large Low Shear Velocity Provinces (LLSVPs) which are located near the core–mantle boundary (e.g. Williams et al. 1996; Torsvik et al. 2006; Burke et al. 2008; Tan et al. 2011; Hassan et al. 2015). Instabilities within the mantle result in plumes that can be entrained, deformed and displaced by large-scale subduction-induced mantle flow (Steinberger and O'Connell 1998, 2000; Davaille et al. 2003; Steinberger et al. 2004; Davaille and Vatteville 2005). As a result, hot spots above the plumes move relative to one another, although inter-hot spot motion is much less than the motion of the lithospheric plates (O'Neill et al. 2005).

Assuming that their present distribution is representative of the past, overriding of plumes at convergent margins should be common in the geological record. Fletcher and Wyman (2015) noted that 29% of mantle plumes have been located within 1000 km of a subduction zone over the past 60 Ma (Fig. 1), a statistic that also implies interaction between the subduction zone, buoyant swell and oceanic plateau should precede that of the plume itself. Plumes may vary in their buoyancy flux and the dimensions of their topographic swell (Sleep 1990). As the features of individual plumes and subduction zones are both highly variable, the features produced by plume–slab interactions may also be highly variable, and therefore difficult to decipher in the geologic record.

Oppliger et al. (1997) and Murphy et al. (1998, 2003) proposed that the Mesozoic–Cenozoic orogenic activity in western North America was profoundly influenced by plume–slab interactions, in which the ancestral Yellowstone plume, preceded by an oceanic plateau and its buoyant swell, were progressively overridden by the westerly migrating margin of the NAP (Fig. 2a, b). This style of 'plume-modified orogenesis' has also been used to explain the origin of the Karoo–Ferrar flood basalts (Dalziel et al. 2000), as well as Mesoproterozoic orogenesis in eastern and central Australia (Betts et al. 2007, 2009).

The model as applied to western North America has two fundamental requirements: (i) that the Yellowstone plume

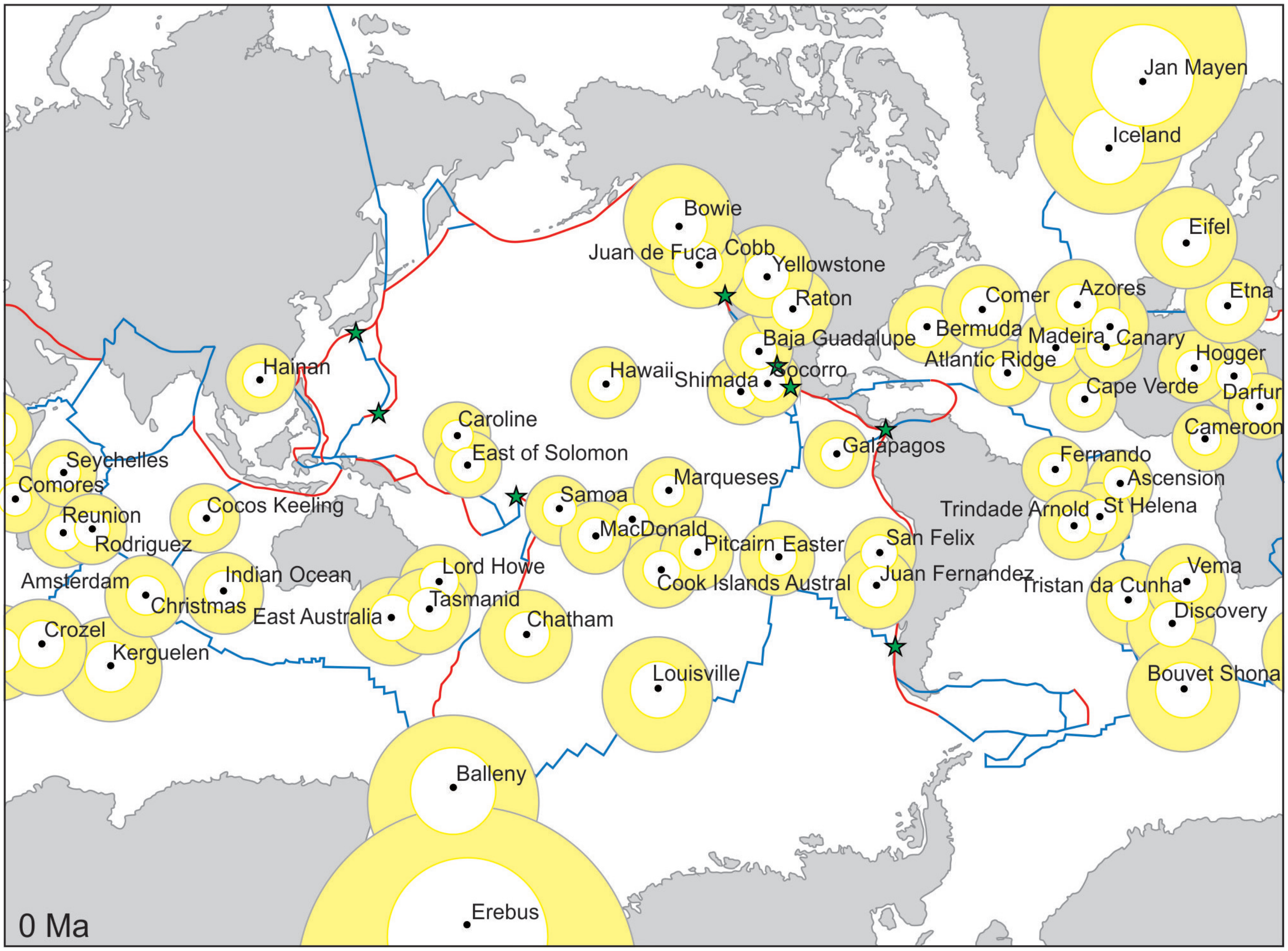

Figure 1. After Fletcher and Wyman (2015) showing zone of potential (500 km, 1000 km) for modern plumes to interact with modern subduction zones (red lines; after Seton et al. 2012). Green stars denote slab windows where a ridge intercepts a subduction zone.

existed prior to ca. 50 Ma, before which reconstructions indicate the plume would have been located beneath oceanic lithosphere (Engebretson et al. 1985; O'Neill et al. 2005; Seton et al. 2012), and (ii) that plumes maintain their integrity where they are affected by subduction processes.

If the Yellowstone plume existed prior to ca. 50 Ma, then plate reconstructions indicate that its buoyant swell, plateau, as well as hot spot oceanic islands with plume-type chemistry, would have collided with the convergent margin of western North America. Such events should be recorded in the geological evolution of western North America.

With regard to the second requirement, recent laboratory studies suggest that the integrity of plumes may be severely affected by subduction processes in that their ascent may be stalled, severely distorted or even destroyed by slab-driven subduction processes (e.g. Kincaid et al. 2013; Druken et al. 2014). On the other hand, 3D numerical models (Betts et al. 2012, 2015) identify combinations of plume and slab properties that may facilitate the transfer of plume material from the lower to the upper plate especially via gaps or tears that may develop in the slab.

The purpose of this article is to re-assess the viability of the underlying requirements for plume–slab interaction to have influenced the Mesozoic–Cenozoic evolution of the western United States in the light of recent laboratory and numerical models, tomographic data, improved plate tectonic reconstructions, as well as more recent field-based studies.

MESOZOIC–EARLY CENOZOIC OROGENESIS

It is widely recognized that episodic accretionary orogenic events have dominated the tectonic evolution of the western margin of the NAP since the Late Paleozoic, resulting in the Antler (Mississippian), Sonoma (Permian–Triassic), Nevadan (Jurassic) and Sevier (Early Cretaceous) orogenies. These events are widely interpreted to reflect predominantly eastward-dipping subduction of the Farallon oceanic plate beneath North America, which at various times in the Mesozoic also produced the voluminous ca. 130–85 Ma Cordilleran

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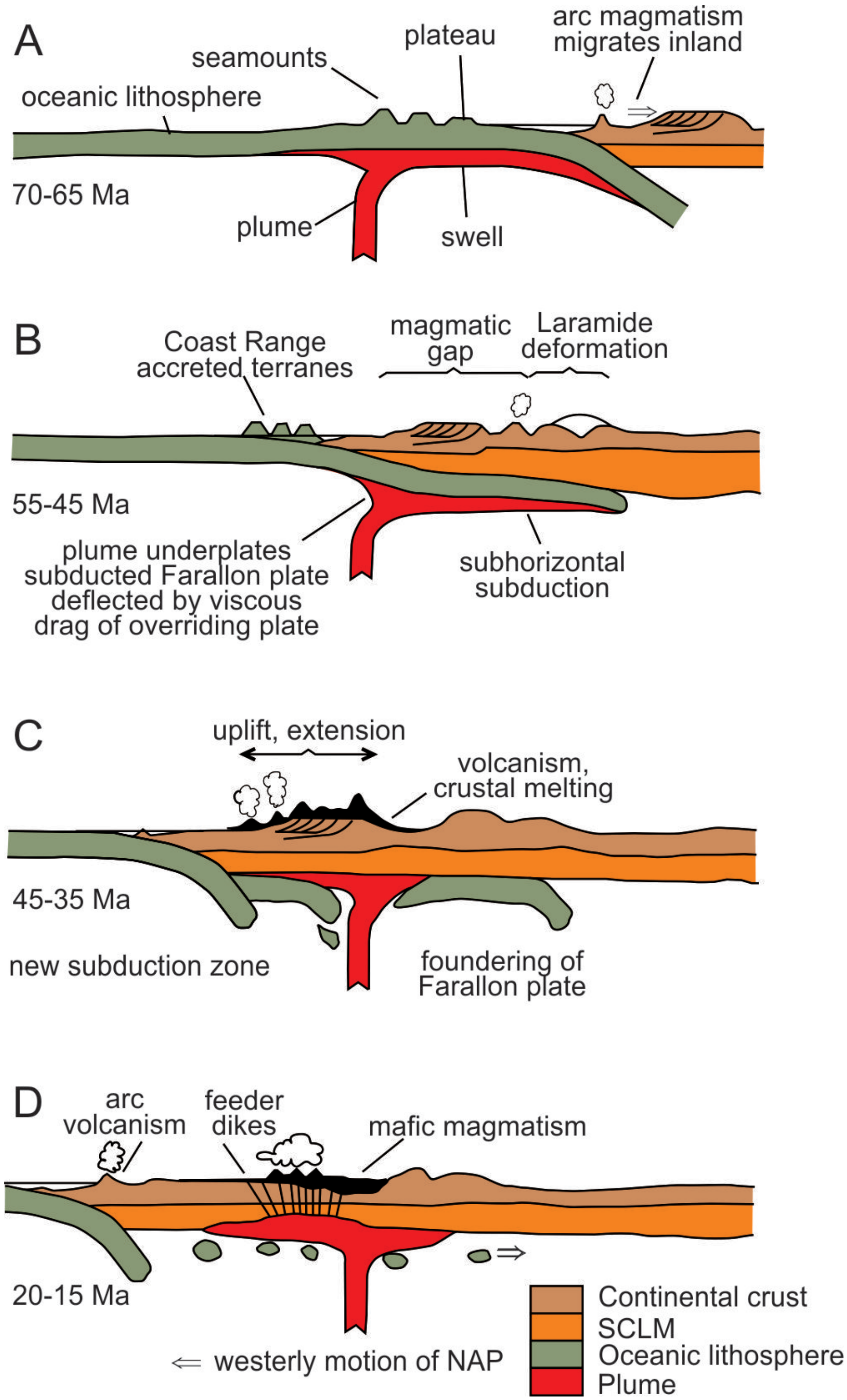

Figure 2a. Schematic diagram (after Murphy et al. 1998) showing late Mesozoic–Tertiary evolution of the southwestern United States at about 40°N (current latitude)and its proposed relationship to the ancestral Yellowstone plume. The plume is stationary, and the North American plate (NAP) moves progressively westward. (A) Plume is beneath oceanic crust where it creates oceanic terranes that subsequently accrete to NAP (Duncan 1982; Johnston et al. 1996). NAP begins to override the oceanic plateau and buoyant swell. (B) NAP continues to override the plateau and swell, resulting in flat-slab subduction, and eventually overrides the plume. (C) Assimilation of subducted portion of Farallon slab by plume, leading to generation of voluminous intracrustal melts, brittle deformation and reestablishment of dipping subduction zone at periphery of NAP. (D) 20–15 Ma breakthrough of plume-related bimodal magma, formation of dike complexes and flood basalts. SCLM: Sub-continental lithospheric mantle.

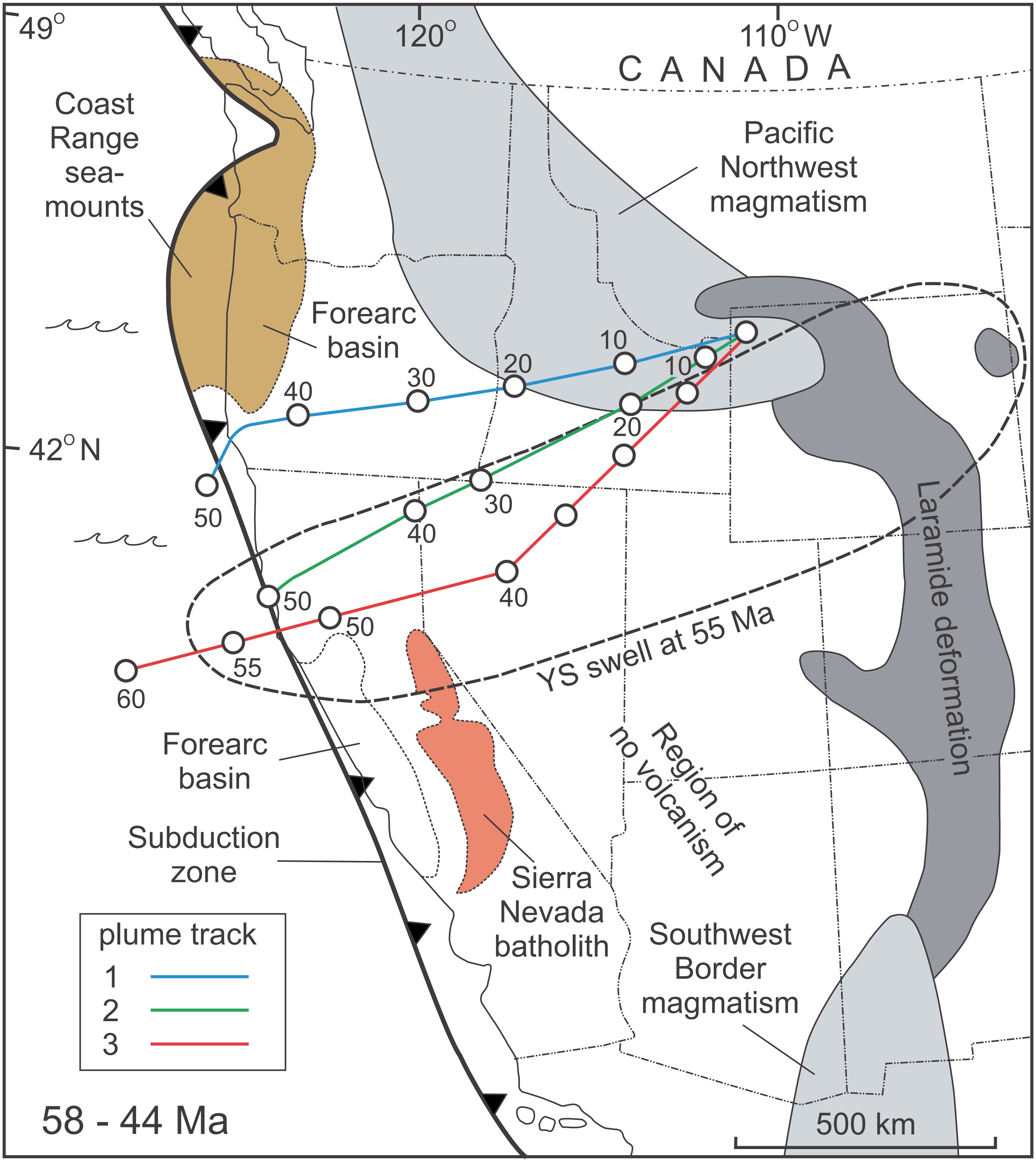

Figure 2b. Simplified mid-Eocene tectonic map of the Cordillera (modified after Dickinson 1991; Burchfiel et al. 1992; Murphy et al. 1998) showing three potential tracks of Yellowstone plume from its oceanic position before 50 Ma to its present position beneath Yellowstone (see Wells et al. 2014). Track 1 derived from O'Neill et al. 2005; Track 2 from Müller et al. 1993; Track 3 from Murphy et al. 1998. Dashed line indicates approximate region of a proposed buried plume-related swell ca. 55 Ma for Yellowstone plume following Track 3 (YS swell at 55 Ma).

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batholiths (e.g. Hamilton 1969; Dewey and Bird 1970; Lipman et al. 1972; Kistler and Peterman 1973, 1978; Dickinson and Snyder 1978; Coney et al. 1980; DePaolo 1981; Livaccari et al. 1981; Severinghaus and Atwater 1990; Burchfiel et al. 1992; Dickinson and Lawton 2001; Saleeby 2003; Busby 2004, 2012; Dickinson 2004; Monger 2014).

A rival model (see discussion in Hoffman 2013), supported by tomographic images of a vertical slab wall extending to a depth of 2000 km in the mantle (Sigloch and Mihalynuk 2013), holds that the earlier orogenic events reflect westerly dipping intra-oceanic subduction and episodes in the assembly of a superterrane (or ribbon continent) located to the west of North America (e.g. Johnston 2001, 2008; Hildebrand 2009; Hildebrand and Whalen 2014) that collided with the passive margin of North America at either ca. 150 Ma (SAYBIA; Johnston 2001, 2008; Johnston and Borel 2007) or between 125 and 100 Ma (RUBIA; Hildebrand 2009). There are variants within this model. According to Johnston (2001, 2008), the collision of SAYBIA resulted in extensive oroclinal development in the Canadian Cordillera, whereas Hildebrand and Whalen (2014) claimed that the westward subduction produced the oldest (ca. 130–100 Ma) phases of the Cordilleran batholiths as well as collision with NAP at ca. 100 Ma, which was followed by slab failure and consequent asthenospheric upwelling that resulted in the younger (100–85 Ma) Cordilleran batholiths.

In both models, there is widespread agreement that (i) most terranes had accreted to North America by ca. 80 Ma, and (ii) the Late Cretaceous to Eocene evolution of the Cordillera was influenced by interactions between the North American, Farallon, and Kula plates, and especially by the northward migration of the triple junction between them and the consequent northward propagation of subduction (Atwater 1970, 1989; Kelley and Engebretson 1994). These interactions may be even more complicated if the more recently hypothesized Resurrection plate, which would have been completely subducted by ca. 50 Ma, is verified (Haeussler et al. 2003; McCrory and Wilson 2013; Wells et al. 2014; Fig. 3). Terranes that accreted during the Tertiary include Siletzia (Fig. 4), which is exposed for 600 km along the Pacific Northwest (e.g. Snavely et al. 1968; Wells et al. 1984, 2014; Babcock et al. 1992, 1994). These terranes were translated northwards within the Kula plate and the collision of Siletzia with the NAP is interpreted to be responsible for the South Vancouver Orocline (Johnston et al. 1996; Johnston and Acton 2003). Eastward-dipping subduction then initiated along the western margin of the accreted Siletzia, producing a north–south Cascade volcanic arc, beginning at ca. 42 Ma (Fig. 3).

Between ca. 80 Ma and 45 Ma, a series of enigmatic events occurred in the southwestern United States that are traditionally assigned to the Laramide orogeny (Burchfiel et al. 1992). These events include widespread thick-skinned deformation, basement uplifts as much as 1500 km (Black Hills, South Dakota) from the continental margin, as well as simultaneous contraction and crustal thickening in both the foreland and hinterland. Voluminous magmatism ceased, and relatively minor magmatism with subduction-related geochemistry migrated inland. These features have been attributed to 'flat-slab subduction,' a region where a gently inclined subduction zone ca. 500 km wide extended about 700 km into the continental interior (Coney and Reynolds 1977; Dickinson and Snyder 1978; Livaccari et al. 1981; Severinghaus and Atwater 1990; Burchfiel et al. 1992; Saleeby 2003). According to Bird (1988), traction associated with subduction of the shallow slab could have stripped away the mantle lithosphere beneath the North American crust and transmitted the stress capable of causing the thick-skinned deformation in the foreland. In this context, the resumption of voluminous magmatism in the Eocene is attributed to the breakup and foundering of the Farallon slab (Humphreys 1995) and the re-initiation of normal-angle subduction, followed in the late Eocene by voluminous ignimbrite associated with localized extension and emplacement of metamorphic core complexes (Coney 1979; Davis and Coney 1979; Gans et al. 1989).

The flat-slab model is supported by *P–T* studies of lawsonite-bearing eclogite xenoliths in Oligocene kimberlite pipes that intrude the Colorado Plateau. These xenoliths are thought to have originated in the Farallon slab, and equilibrated at depths between 90 and 160 km and at temperatures between 500 and 700°C (Usui et al. 2003). In the Canadian and Mexican portions of the Cordillera, however, coeval development of a magmatic arc within 300 km of the trench indicates that the effects of flat-slab subduction were limited to the southwestern United States (English et al. 2003), a feature that implies segmentation of the Farallon slab into flat and steep zones (Saleeby 2003).

Rival models for the Laramide orogeny include the collision of a superterrane at ca. 125 Ma, with North America on the lower plate, followed by voluminous ca. 100–85 Ma magmatism associated with slab break-off (e.g. Hildebrand 2009, 2014; Hildebrand and Whalen 2014). As these events would have occurred before the classic Laramide events, they do not preclude the shallow-slab subduction model, which could have been initiated in the aftermath of this collision. English and Johnston (2004) pointed out, however, that fold and thrust belts in Canada and Mexico are coeval with those within the putative flat-slab region, and that a viable explanation for synchronous contraction along the entire length of the orogen remains enigmatic.

The 'hit-and-run' model of Maxson and Tikoff (1996), which attributes Laramide orogenesis to the Late Cretaceous collision followed by northward translation of the colliding terranes would refute the flat-slab model because of the timing of these hypothesized events. This model builds on the controversial Baja–BC hypothesis (Irving 1985; Irving et al. 1985, 1995, 1996), in which a wealth of paleomagnetic data implies up to 3000 km of northward (i.e. dextral) translation of terranes relative to cratonic North America since 70 Ma (Beck 1992; Wynne et al. 1995; Johnston et al. 1996; Kent and Irving 2010). Although plate reconstructions (Atwater 1970; Engebretson et al. 1985) and field evidence for dextral transpression (Oldow et al. 1989; Maxson and Tikoff 1996) are consistent with dextral translation along the North American margin of terranes embedded in the Kula plate, identifying the structures that might accommodate such a large displacement remains

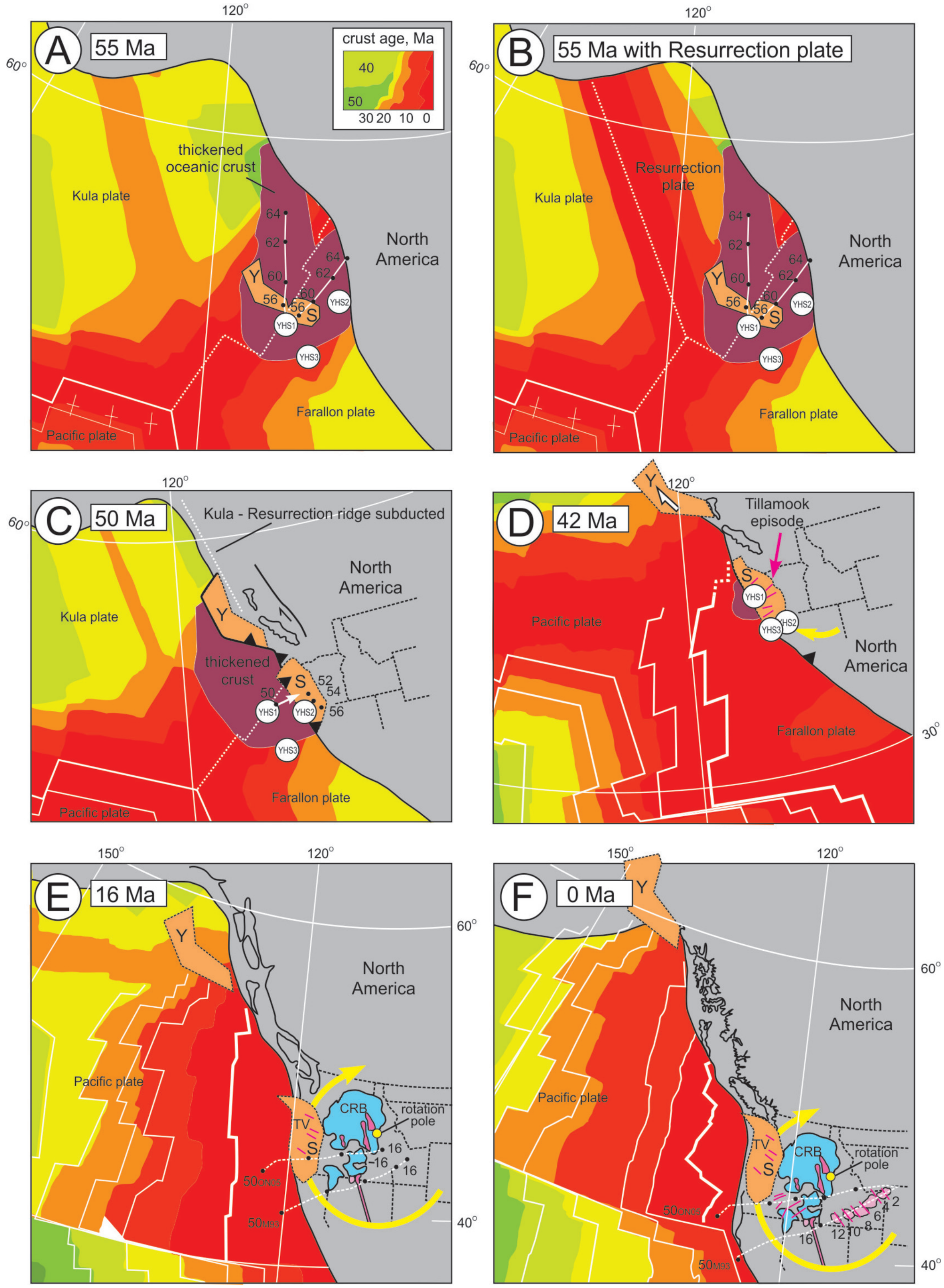

Figure 3. Schematic reconstruction of the Paleogene–Recent reconstruction of the Pacific Northwest (after Wells et al. 2014). (A) examines Pacific–Kula–Farallon–North American plate interactions, and (B) examines the additional effects of the putative Resurrection plate. Plate reconstruction model of Seton et al. (2012) showing three possible locations for the ancestral Yellowstone hot spot (YHS) as shown by Wells et al. (2014). YHS1 is the location relative to North American plate (NAP) derived from the moving hot spot reference frame (O'Neill et al. 2005; ON05); YHS2 from the plate circuit reference frame of Müller et al. (1993; M93) and YHS3 from a reference frame defined by moving hot spots in the Pacific, Atlantic, and Indian oceans (Doubrovine et al. 2012). Hot spot reference frame paths for YH1 (ON05; O'Neill et al. 2005) and YH2 (M93; Müller et al. 1993) shown in E and F by dotted lines; small dots on path show location every 10 m.y.

YHS is centered at or near the Kula–Farallon Ridge in (A) and the Resurrection–Farallon Ridge in (B). Oceanic (Siletzia, S, and the conjugate Yakutat, Y) terranes form at ridge-centred hot spot at ca. 55 Ma. (C) Accretion of oceanic terranes by 50 Ma. (D) Progressive overriding of the YHS by NAP, producing northwest-directed extension magmatism in the forearc. (E) CRB – Steens and Columbia River Basalt provinces. Yellow arrow (E and F) shows potential clockwise rotation of Coast Range which moves plume products progressively away from the hot spot track. (F) YHS under Yellowstone; age progression of Snake River Plain from 16 Ma to 0 Ma shown in pink.

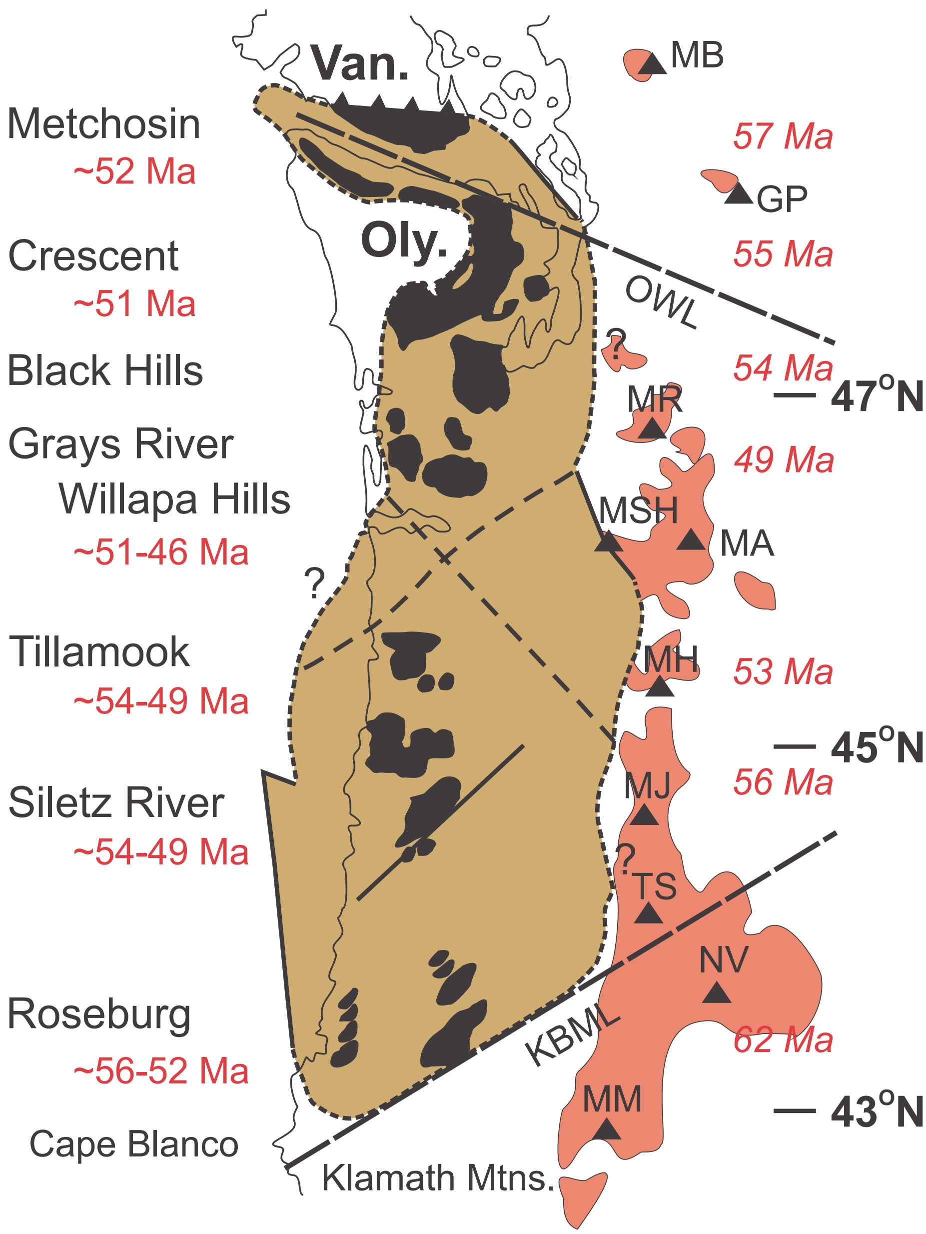

Figure 4. Outcrop (black) and inferred subsurface distribution (pink) of Siletz and Crescent mafic rocks (also known as Siletzia and as the Coast Range Basalt Province) along with age estimates for selected mafic rocks (compiled by McCrory and Wilson 2013), and for the Cascade Arc volcanoes (black triangles) from Duncan (1982). OWL – Olympic–Wallowa Lineament; KBML – Klamath–Blue Mountains Lineament. Modified from original figure by Duncan (1982).

elusive (Price and Carmichael 1986; Cowan 1994; Monger et al. 1994; Monger 2014). Recently, however, Hildebrand (2015) proposed a reconstruction in which the Texas Lineament and the Lewis and Clarke transverse zone, now 1300 km apart, were formerly contiguous, but were offset during the 80–58 Ma Laramide orogeny along N–S faults within the Cordilleran fold and thrust belt. According to Hildebrand (2015) the entire width of the Cordillera was translated northwards. By uniting two belts of plutonic rocks, the reconstruction eliminates the magmatic gap between them, and is therefore a first-order challenge to the flat-slab model. A magmatic gap could still exist between 58 and 43 Ma, but the existence of a flat slab at that time would not match the earliest stages of Laramide thick-skinned deformation.

This article assesses the potential effects of plume–slab interaction in the context of the more widely accepted model in which the Mesozoic–Cenozoic evolution of the western United States is dominated by long-lived easterly subduction. The effects of plume–slab interaction in the context of the rival models of Johnston and Hildebrand will be discussed in a separate contribution.

MODERN YELLOWSTONE PLUME

Plume versus non-plume models for the origin of the Yellowstone hot spot have been debated ever since Morgan (1972) and Armstrong et al. (1975) related the diachronous onset of magmatism in the Snake River Plain to the migration of the NAP over a stationary plume (e.g. Humphreys et al. 2000; Pierce and Morgan 2009; Fouch 2012). Alternative 'non-plume' models include the edge effects of cratonic lithosphere (King and Anderson 1998) and 'hot-lines' (Christiansen et al. 2002). The region around the hot spot is characterized by high heat flow, pronounced hydrothermal activity, a topographic bulge 600 m high and ca. 600 km wide, and a 10–12 m positive geoid anomaly (Smith and Braile 1994).

The potential relationship between magmatism and the modern Yellowstone plume development, beginning at about 17 Ma, has been suggested on the basis of geochronological, geophysical, structural and petrological data (e.g. Armstrong et al. 1975; Hadley et al. 1976; Geist and Richards 1993; Smith and Braile 1994; Camp 1995; Glen and Ponce 2002; Hooper et al. 2007; Ito and van Keken 2007; Graham et al. 2009; Smith et al. 2009). These studies show the age progression of magmatism along the Yellowstone–Snake River Plain which matches the calculated trajectory of the hot spot track derived from plate reconstructions.

Over the past decade, the plume model has been supported by a variety of geophysical techniques. Although some studies imply a shallow source for the Yellowstone hot spot (e.g. Montelli et al. 2006), in general, the results of teleseismic tomography reveal an inclined low velocity anomaly 100 km wide within the upper mantle beneath Yellowstone that is interpreted as the thermal effects of the Yellowstone plume (Fig. 5). This anomaly extends 400 km along the length of the Yellowstone–Snake River Plain (YSRP), dips about 60 degrees WNW and can be detected to a depth of about 500 km, deflecting the 410 km discontinuity downward by as much as 12 km (Yuan and Dueker 2005; Waite et al. 2006; Smith et al. 2009). The anomaly has been modelled to reflect about 1% of partial melting at an ambient temperature of 200°C above normal (Schutt and Humphreys 2004). As a further example of the potential complexity, a *P*-wave model (Obrebski et al. 2010) shows a gap in the Juan de Fuca subducted slab in the region above the Yellowstone plume and below the YSRP.

Magnetotelluric surveys fail to detect magma directly beneath the Yellowstone hot spot and may not have the sensitivity to detect such deep structural features (Kelbert et al. 2012). These surveys do, however, detect a zone of low mantle resistivity beneath the eastern segment of the Snake River Plain, consistent with small degrees of partial melt at depths between 40 and 80 km (Zhdandov et al. 2011; Kelbert et al. 2012).

P to *S* body-wave tomography models imply an even deeper origin for the Yellowstone plume. Porritt et al. (2014) showed images of the plume to a depth of 1000 km. Other tomographic images suggest that this thermal anomaly may be connected to a much broader low-velocity anomaly in the uppermost portion of the lower mantle (e.g. Allen et al. 2008; Schmandt and Humphreys 2010; Obrebski et al. 2010; Sigloch 2011). These images also indicate that the 660 km discontinuity is deflected upwards by 12–18 km beneath Yellowstone, interpreted to reflect the presence of a plume-like upwelling (Schmandt et al. 2012). Zhao (2007) identified anomalies in both the upper and lower mantle and speculated that the complex distribution of anomalies may be due to interaction of the Yellowstone plume with the Farallon slab (see Fletcher and Wyman 2015). In addition to tomographic studies, Pierce and Morgan (2009) proposed that some of the large-scale tectonic features require a mantle plume extending to a depth of at least 1000 km.

The oldest widely accepted surface expression of the Yellowstone plume includes the main phase of the ca. 17 Ma Steen and Columbia River flood basalt provinces as well as voluminous contemporaneous silicic volcanism and is widely attributed to the initial impingement of the Yellowstone plume beneath North American lithosphere (e.g. Hooper et al. 2007; Coble and Mahood 2012; Fig. 6). The geochemistry of the basaltic rocks is complex and includes primary components typical of MORB, OIB and older mantle components (Carlson 1984; Draper 1991; Hooper and Hawkesworth 1993; Bryce and DePaolo 2004), interpreted to reflect the interaction of plume-derived material with sub-continental lithospheric mantle and overlying crust (Hooper et al. 2007; Coble and Mahood 2012). However, the volcanic centres lie on a N–S trend with the youngest centres and largest volumes occurring well to the north of the putative Yellowstone hot spot track (Barry et al. 2010). These trends have been variously explained by (i) a deflection of the Yellowstone plume by the Juan de Fuca slab (Geist and Richards 1993), (ii) exploitation by the plume of the

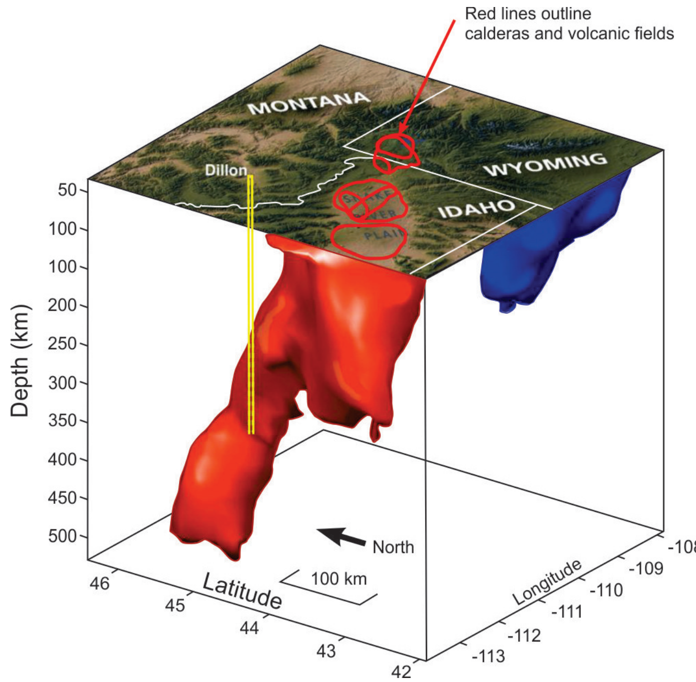

Figure 5. Seismic tomographic image (Yuan and Dueker 2005; Pierce and Morgan 2009) showing inclined conduit of warm mantle, interpreted as a plume that can be tracked northwestward from beneath Yellowstone to a depth of 500 km. Red outlines schematically show the calderas of the Snake River Plain.

old continental margin (Camp and Ross 2004), (iii) slab delamination and consequent asthenospheric upwelling that accompanied the arrival of the Yellowstone plume (Camp and Hanan 2008; Darold and Humphreys 2013), and (iv) a propagating rupture in the Farallon slab (Liu and Stegman 2012). A further explanation may be derived from numerical models which show that when the plume penetrates from the lower into the upper plate via a slab window, plume material becomes complexly distributed and does not follow a simple linear pattern expected for a hot spot track (Betts et al. 2015).

The YSRP is about 700 km in length and is classically interpreted as a northeasterly trending hot spot track (Morgan 1972; Armstrong et al. 1975) characterized by the eruption of voluminous felsic ignimbrite and overlain by a thin succession of basalt, as well as a voluminous mafic sill complex (Shervais et al. 2006). He, Pb, Sr, and Nd isotopic analyses of the mafic rocks are consistent with a mantle plume source (e.g. Craig et al. 1978; Hearn et al. 1990; Vetter and Shervais 1992; Hughes et al. 2002; Hanan et al. 2008; Graham et al. 2009). The rhyolitic volcanic rocks that characterize the YSRP system are diachronous, and their age progression of 4.5 cm/a is viewed as the composite of the migration of the NAP and Basin and Range extension over the same time interval (Shervais and Hanan 2008). The excess topography of the YSRP decreases systematically along its length away from the hot spot and has been modelled to reflect a progressively cooling and contracting lithosphere (Smith and Braile 1994).

Prior to 17 Ma, products within North America related to the Yellowstone plume are controversial. Seligman et al. (2014) provided isotopic and trace element data from 40–30 Ma vol-

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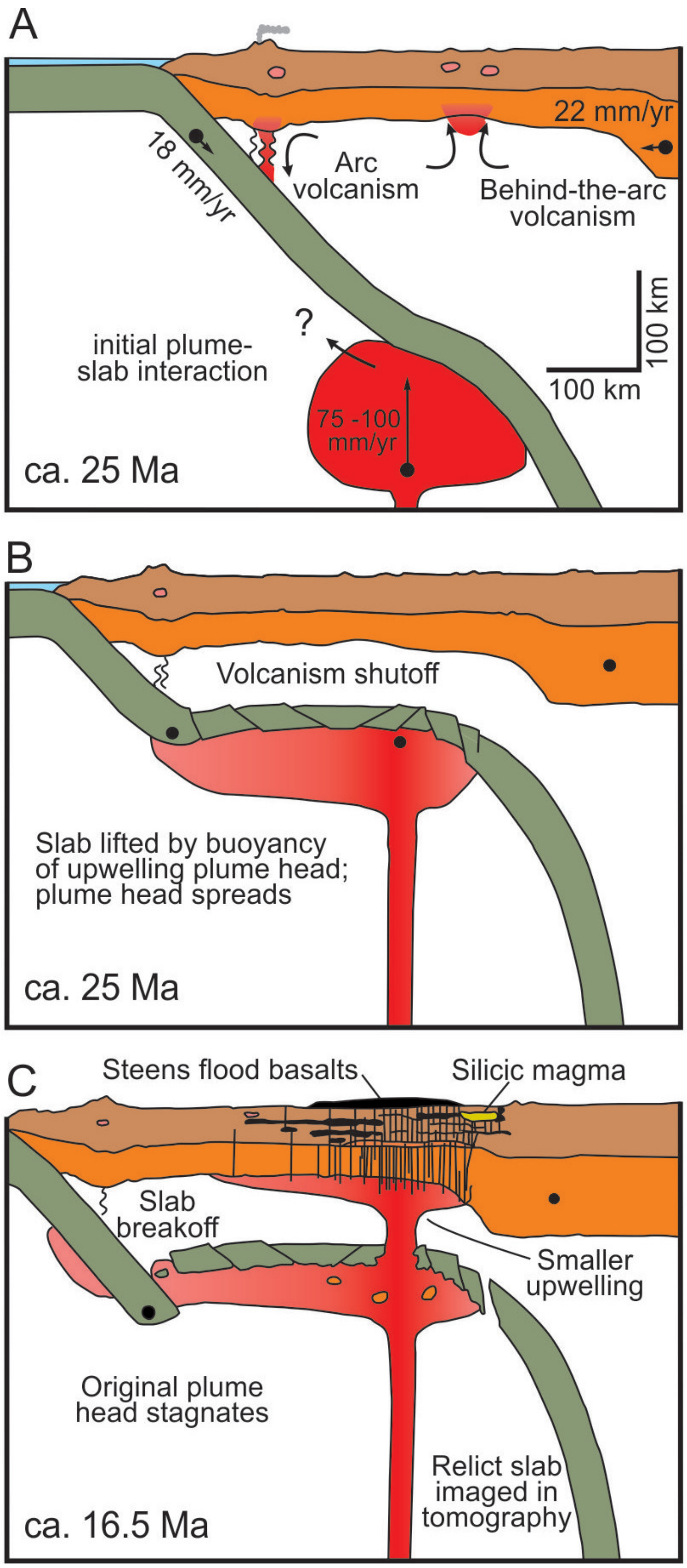

Figure 6. Model for the evolution of Yellowstone plume (after Coble and Mahood 2012) that considers the lifting of the Juan de Fuca slab at 25 Ma (A) to have resulted in volcanic quiescence (B) between ca. 23 and 17 Ma, and the distribution of voluminous bimodal basalts and coeval rhyolites (C) to represent breakthrough of plume material at ca. 17 Ma into the upper plate to form a secondary plume head.

canic centres in Oregon that are located too far east to be directly related to Farallon subduction. Basalt was derived from an enriched subcontinental lithospheric mantle, whereas some felsic complexes (e.g. Crooked River) have $\delta^{18}O$ (zircon) values that require a large heat source, extensive hydrothermal circulation and crustal recycling (Bindeman et al. 2001; Cathey et al. 2011; Watts et al. 2011; Drew et al. 2013; Seligman et al. 2014). Plate reconstructions (Seton et al. 2012; Wells et al. 2014) imply that the Yellowstone plume would have resided beneath Crooked River at that time. Seligman et al. (2014) attributed this magmatism to the earliest eruptions associated with the encroachment of the Yellowstone plume. In their model, the more widespread coeval voluminous felsic magmatism is attributed to 'plume-triggered delamination.' Although Murphy et al. (1998) attributed regional felsic magmatism to the arrival of the plume along the base of the continental lithosphere, Seligman et al. (2014) invoked a model where plume material encroaches through a tear or a gap in the subducting Farallon plate.

ANCESTRAL YELLOWSTONE PLUME

Duncan (1982) proposed that basalt-dominated volcanic complexes accreted to the North American margin in the Eocene originated as oceanic islands above the ancestral Yellowstone hot spot which was located at that time along the Kula–Farallon ridge. This model was underpinned by plate reconstructions which implied that if the Yellowstone plume existed prior to 55–50 Ma, it would have been located beneath oceanic lithosphere.

Since that time, interpretations of the plate configuration in the oceanic realm in the Pacific Northwest have been refined. For example, Haeussler et al. (2003) attributed the migration of Eocene magmatism from Alaska to Oregon to the migration of two triple junctions, requiring the existence of a previously unrecognized oceanic plate, named the Resurrection plate. This plate, if it existed, would have been located to the east of the Kula plate, and would have been subducted, along with its bounding ridges by 50 Ma (see also Madsen et al. 2006). Despite these refinements, the basic relationship of the hypothesized ancestral Yellowstone hot spot with either the Kula–Farallon or Kula–Resurrection oceanic ridges is consistent with more recent reconstructions derived from Gplates (Seton et al. 2012) and the moving hot spot reference frames (Lonsdale 1988; Müller et al. 1993; O'Neill et al. 2005; McCrory and Wilson 2013; Wells et al. 2014).

If correct, manifestations of the Yellowstone plume, including accretion of ocean islands and interaction with oceanic plateaus related to the plume, should be recognized among the tectonic events along the convergent margin of western North America.

Possible Accreted Oceanic Complexes

Mafic complexes hypothesized to have accreted to North America include Siletzia (now fragmented into the Siletz and Crescent terranes) which is exposed in Vancouver Island (Metchosin igneous complex; Massey 1986), Oregon and Washington (Duncan 1982), and the Carmacks Group (Intermontane belt) of the Yukon Territory of Canada (Johnston et al. 1996) which oversteps previously amalgamated terranes.

Siletzia is characterized by a 56–49 Ma submarine sequence of tholeiitic–alkalic basalt and associated volcanogenic sedimentary rocks, overlain by a subaerial sequence dominated by alkalic basalt (e.g. Snavely et al. 1968; Wells et al. 1984, 2014; Babcock et al. 1992, 1994) with plume-type geochemistry (Pyle et al. 1997, 2009). The terrane is estimated to be 27 ± 5 km in thickness (Trehu et al. 1994; Graindorge et al. 2003), is exposed over 240,000 km^2, and the volume of its basaltic magmatism is estimated to exceed that of the Columbia River Basalt province by at least one order of magnitude (Duncan 1982; Wells et al. 2014). Post-accretionary magmatism continued until ca. 42 Ma with the emplacement of mafic volcanic and dike complexes in the forearc (Wells et al. 2014).

Paleomagnetic data, although complex, allow the possibility of significant post-depositional episodes of rotation (Simpson and Cox 1977; McCrory and Wilson 2013), but imply a paleolatitude similar to today. Wells et al. (2014) maintained that the oceanic composition, large volume and short duration of magmatism are characteristic of Large Igneous Provinces (LIPs) and they interpreted these rocks to collectively represent an oceanic plateau produced by the ancestral Yellowstone plume. They also showed that, in most hot spot reference frames, plate reconstructions imply these volcanic complexes would have been located close to the Yellowstone hot spot at the time, within 300 km of the continental margin in a near-ridge setting (see also McCrory and Wilson 2013). The southern Vancouver orocline is thought to have formed in response to the accretion of Siletzia to North America (Johnston and Acton 2003).

In northern Siletzia (Crescent terrane), the Metchosin Complex of Vancouver Island consists of a 60–50 Ma sequence of mafic volcanic and interbedded clastic rocks (Massey 1986). This complex is thought to comprise a part of Siletzia that was scraped from a subducting slab which underthrust previously accreted terranes (Hyndman 1995). Tomographic studies (Ramachandran 2001) suggest that the terrane beneath Vancouver Island may be as much as 25 km thick. Paleomagnetic data (Babcock et al. 1992) combined with Mesozoic–Cenozoic plate reconstructions indicate that these basaltic rocks were emplaced in a similar location to the modern Yellowstone hot spot. Murphy et al. (2003) interpreted the shallowing-upward sequence to have formed in a Loihi-type environment and estimated 4.2 km of uplift related to plume activity which yielded a plume buoyancy flux of 1.1 Mg s^{-1}, comparable in vigour to that of the modern Yellowstone hot spot.

The Carmacks Group is a ca. 70 Ma sequence of volcanic rocks that unconformably overlies the previously amalgamated northern Intermontane belt terranes, and possibly the adjacent Omineca belt terranes (e.g. Gladwin and Johnston 2006). The group comprises a thick subaerial succession dominated by alkalic basalt with shoshonitic geochemistry that is comparable with modern plume-related basalt. Paleomagnetic data combined with regional geological data constrain the eruption of the Carmacks Group to a paleolatitude similar to the modern

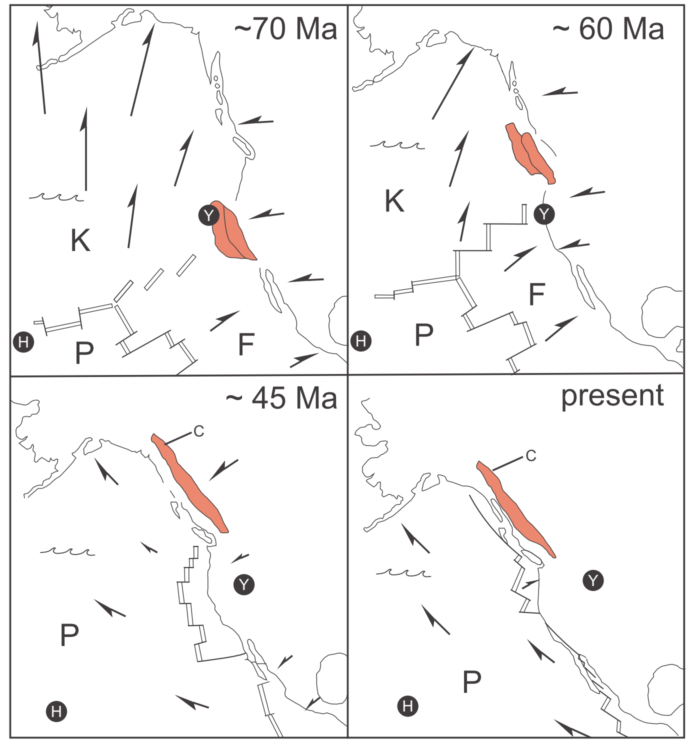

Figure 7. Yellowstone in Yukon model of Johnston et al. (1996) assuming fixed locations of Yellowstone (Y) and Hawaii (H) hot spots. In this model, the Carmacks Group (C; Intermontane belt) is generated by its passage over the Yellowstone hot spot at 70 Ma, and subsequently displaced dextrally along the margin with the North American plate. F – Farallon plate; K – Kula plate; P – Pacific plate.

Yellowstone hot spot implying a 17.2 ± 6.5° (ca. 2000 ± 600 km) northward translation since their eruption (Johnston et al. 1996; Johnston and Thorkelson 2000; Fig. 7).

The interpretation of the Carmacks Group as a representative of the Yellowstone hot spot has important implications for Cordilleran tectonics. As reconstructions place the hot spot to the west of North America at 70 Ma, this would imply the Carmacks Group, as well as the underlying Intermontane belt terranes, lay outboard of continental North America at that time. The obliteration of the oceanic tract between them could explain the lack of preservation of a hot spot track between Siletzia and Carmacks Group. The reconstruction of Hildebrand (2015) provides an explanation of the northward translation of the Carmacks Group, required to move it from an original position along the hot spot track to its modern location in the Yukon. An alternative possibility is that the matching paleolatitudes of Carmacks and Yellowstone are fortuitous, in which case these relationships place no constraints on the relationship of the Intermontane belt terranes and continental North America.

Geophysical Data and Geodynamic Models

Geodynamic models show that plumes have difficulty penetrating oceanic lithosphere (e.g. McNutt and Fischer 1987), let alone lithosphere capped by continental crust (Murphy et al. 1998). The presence of continental crust as well as remnants of the Juan de Fuca–Farallon slab detected by seismic studies

(Xue and Allen 2007; Obrebski et al. 2010) provides additional challenges to the ascent of Yellowstone plume material, suggesting the connection between the plume and corresponding hot spot is likely to be far more complex in continental environments compared to oceanic environments. For example, Geist and Richards (1993) attributed the Columbia Plateau–Snake River Plain to the deflection of the Yellowstone plume. Magmatism far removed from the YSRP track, such as the eastern Oregon Steens–Columbia River Basalt is attributed to delamination of remnant Farallon oceanic lithosphere coincident with the arrival of the Yellowstone plume (Darold and Humphreys 2013). Coble and Mahood (2012) invoked a two-stage process in which the bulk of the plume material stalled beneath the subducted Farallon slab, but some material penetrated the slab creating a secondary 'plume' which impinged on the continental lithosphere at ca. 17 Ma resulting in flood basalt volcanism and coeval silicic magmatism. This secondary plume is purported to explain why the volume of basalt produced is significantly smaller than that of a more typical Large Igneous Province.

The barriers to the ascent of plume material, such as the remnants of the Farallon slab as well as the thick lithosphere that overrides it, suggest that the Yellowstone plume must have originated significantly earlier than its first interaction with North American crust in the middle Miocene and may have been in an 'incubation' phase when it was overridden by a subduction zone (Murphy et al. 1998). According to this model, during this incubation phase plume material would have ponded beneath and progressively assimilated or thermally eroded the overlying Farallon slab following which plume material was transferred from the lower to the upper plate.

However, recent geodynamic models (e.g. Kincaid et al. 2013; Druken et al. 2014) have suggested that mantle downwelling related to subduction in the vicinity of the plume may destroy the plume column (Fig. 8), implying that thermal erosion on its own is not a viable mechanism for transferring plume material from the lower to the upper plate. These results have focused attention on more specialized environments for how the plume may have penetrated to the upper plate. For example, plume material may migrate around the edges of the slab in a bifurcating fashion (Seligman et al. 2014), and/or advect into the upper plate during slab roll-back (Druken et al. 2014). In addition, numerical models (Betts et al. 2012) show that plume material may be transferred to the upper plate via a slab window which is created when the plume buoyancy stalls subduction as it interacts with the convergent margin.

In the conceptual model of Coble and Mahood (2012), plume material migrates into the upper plate via fractures or tears in the slab. A similar model was invoked by Obrebski et al. (2010) who interpreted high resolution tomographic images derived from the USArray deployment to reflect the arrival and emplacement of the ancestral Yellowstone plume beneath the Cascadia subduction zone (Fig. 9) which broke through the Juan de Fuca slab, either by exploiting pre-existing weaknesses or by promoting the tearing of the slab.

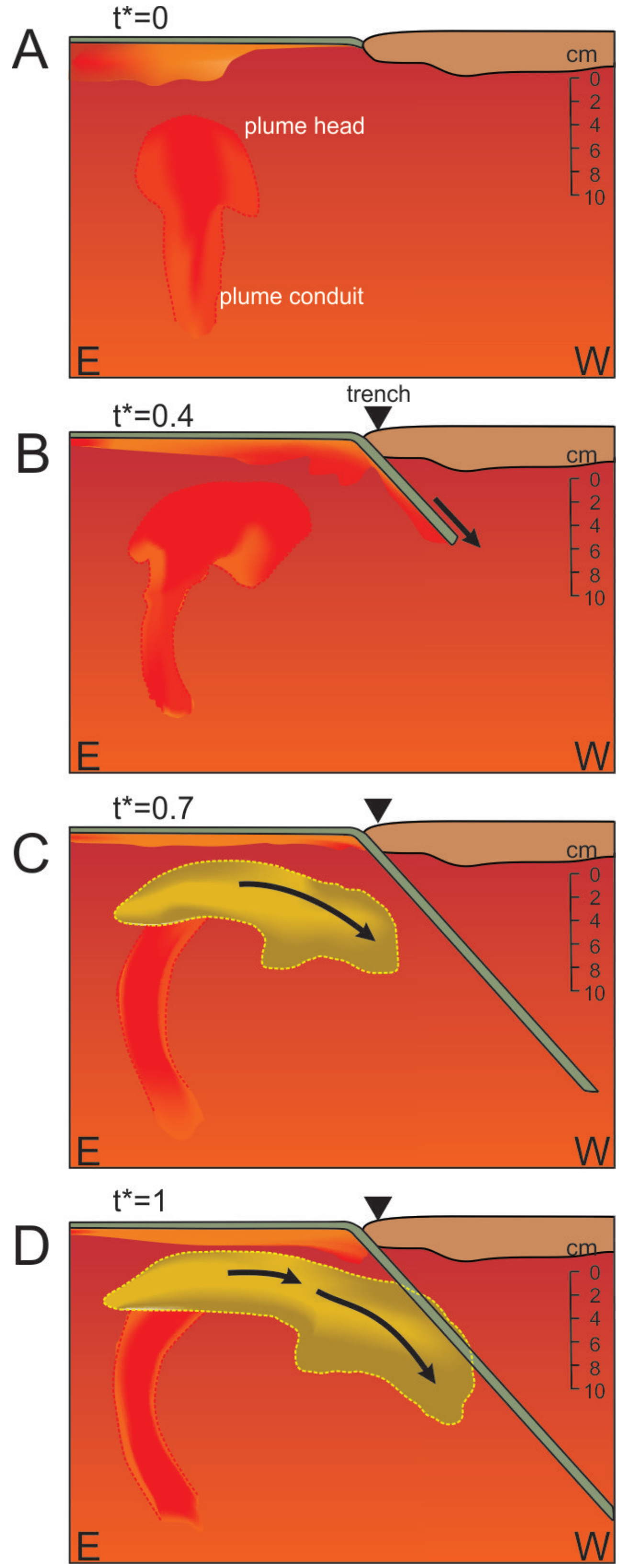

Figure 8. Example of plume-slab interaction (see Druken et al. 2014 for details). (A) Plume structure before subduction initiation. Its head rises vertically at an average rate of 2–3 cm/min. (B) and (C) Subduction induces a downward flow of plume material which is advected back into the mantle (D).

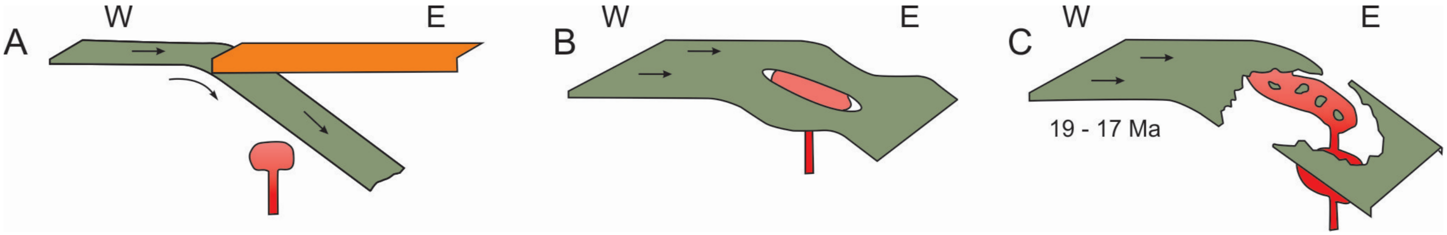

Figure 9. Model for the evolution of the Pacific Northwest after Obrebski et al. (2010). (A) The Yellowstone plume (head in pink, tail in red) penetrates the subducting slab (green) and reaches the base of the continental lithosphere (orange) resulting in slab break-up that was facilitated by pre-existing fractures in the slab (B). Slab break-off results in reduction of slab-pull and consequent decrease in rate of convergence. (C) The plume assimilated fragments of the subduction zone and generated the Columbia River flood basalt.

ANDEAN AND PACIFIC ANALOGUES

The Andean orogenic system is commonly considered to represent a modern analogue for Laramide tectonics of the southwest United States. Seismic imaging reveals that the Andean subducted slab is segmented into flat and steep domains, with the flat segments up to 500 km wide (e.g. Gutscher et al. 2000). These flat slab segments are characterized by a lack of recent magmatism, eastward migration of deformation, and by foreland thick-skinned deformation similar to the Laramide uplifts (Dalziel 1986; Allmendinger et al. 1997). These segments are spatially and temporally correlated with subducting oceanic plateaus (e.g. Pilger 1981; Gutscher et al. 2000; Gutscher 2002; Yáñez et al. 2002; Ramos and McNulty 2002; Fig. 10), implying a genetic relationship between flat slab generation and subduction of relatively thick and buoyant oceanic lithosphere.

Livacarri et al. (1981) proposed that the Laramide orogeny was due to the subduction of an oceanic plateau. According to Cloos (1993), oceanic plateaus with crustal thicknesses in excess of 30 km cause collisional orogenesis during subduction. Liu et al. (2008) reconstructed the subduction record of the Farallon oceanic lithosphere back to 100 million years ago with an inverse mantle convection model that uses stratigraphy to constrain mantle viscosity and buoyancy. Their preferred model predicts an extensive shallow-dipping slab, beginning about 90 Ma, that extended up to 1000 km to the east and north of the flat slab. They attributed the limited width of the flat slab region to the subduction of an oceanic plateau, a scenario supported by recent 3D dynamic models (Betts et al. 2015).

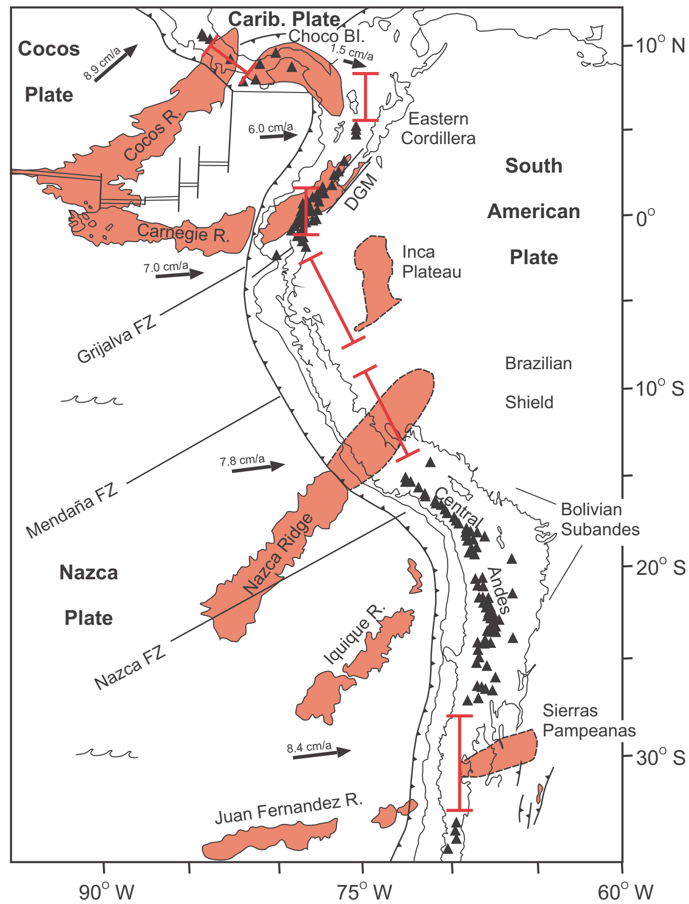

Figure 10. Relationship between flat-slab segments (thick red brackets) and subducted plateau (orange regions) projected beneath the Andean margin (from Gutscher et al. 2000).

The Hess–Shatsky large igneous province (LIP), located in the northwest Pacific (Fig. 11a), has been suggested as a conjugate for these oceanic plateaus that were subducted in the southwest Cordilleran region in Laramide time (Livacarri et al. 1981; Tarduno et al. 1985; Liu et al. 2010). If correct, these plateaus, which constitute one of Earth's largest LIPs, may provide an indication of the size and age of the oceanic plateau that was subducted in Laramide times. The Shatsky plateau alone has an area of 0.48 million km^2 (comparable in size to California) and a volume of 4.3 million km^3 (Mahoney et al. 2005; Sager 2005). The three volcanic centres have depths of 3200–2000 m, whereas the abyssal ocean floor surrounding the plateau is between 6000–5500 m below sea level (Nakanishi et al. 2015). Adjacent magnetic lineations indicate that the Shatsky plateau originated at about 130 Ma at the Pacific–Izanagi–Farallon triple junction (Tatsumi et al. 1998) and its Nd and Pb isotopic characteristics are indistinguishable from the East Pacific Rise (Mahoney et al. 2005). The age of the Hess oceanic plateau is not well constrained, but is inferred to represent renewed magmatism about 20 m.y. later (Ito and van Keken 2007). Liu et al. (2010) showed that Laramide deformation tracks the passage of the Shatsky conjugate plateau beneath North America (Fig. 11b), which was then converted to eclogite and foundered into the mantle. They interpreted Laramide uplift as isostatic rebound after this foundering.

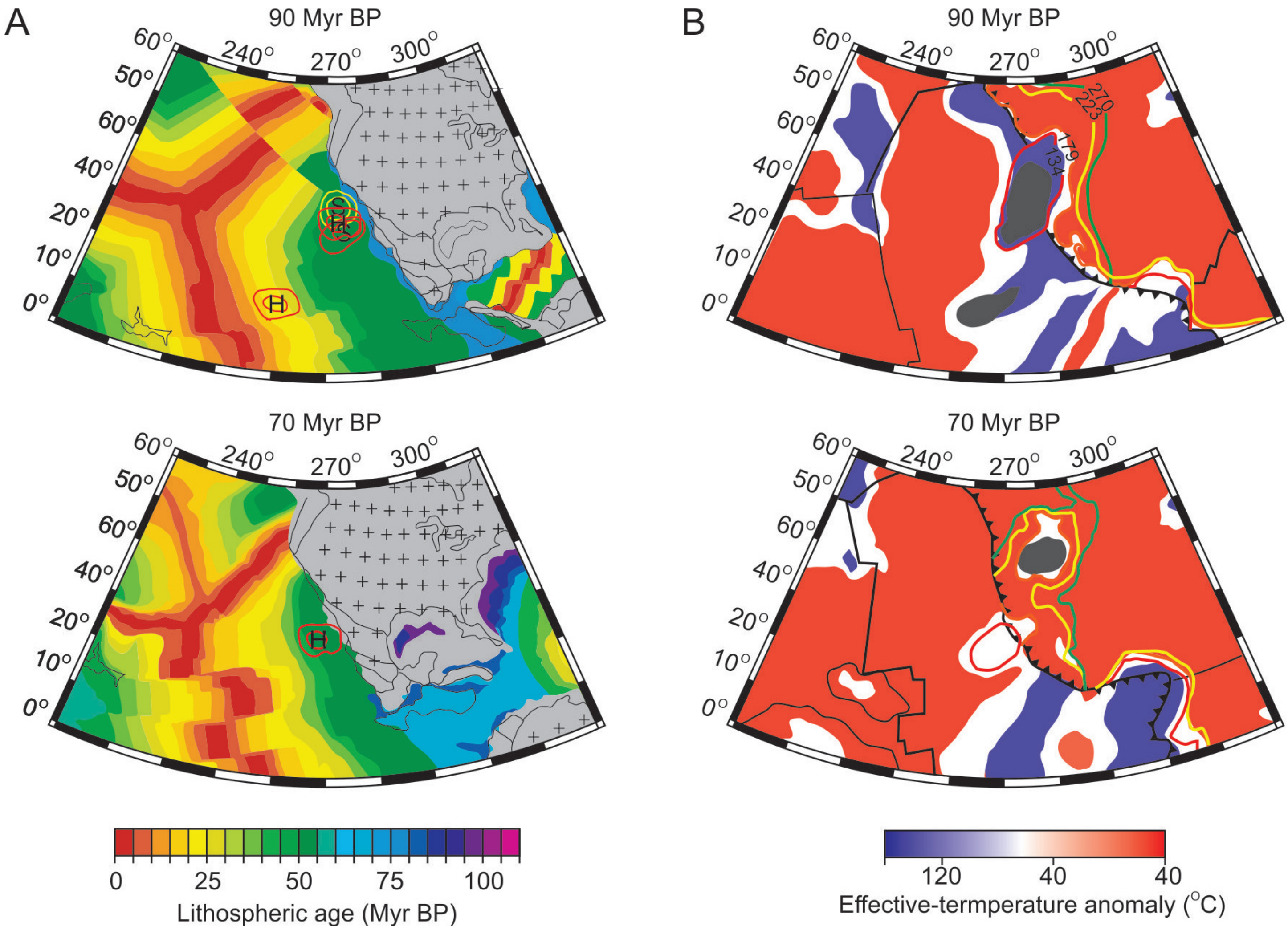

Figure 11. Potential role of the Shatsky (S) and Hess (H) conjugate plateaus in generating the flat-slab environment for the Laramide orogeny (from Liu et al. 2010). Black outlines give current locations of the main plateau. (A) Plate reconstruction of the conjugate plateaus which are inferred to have formed along the Pacific–Farallon and Farallon–Izanagi ridges, respectively. Red contours show their maximum extent, yellow contours their minimum extent. (B) Inverse convection models showing locations of the thickest part of the conjugate lithosphere above 179 km depth. Colour contours show isotherms at different depths (in km). See Liu et al. (2010) for further details.

Murphy et al. (1998, 2003) attributed flat slab subduction in the western United States to overriding of the Yellowstone plume and its associated oceanic plateau/swell by the continental margin. Using the relative motion between the North American and Farallon plates of 120 mm/a (Engebretson et al. 1985), Murphy et al. (1998) estimated that the buoyant swell about 400 km wide (Sleep 1990) would have been elongated up to 2400 km in the direction of Farallon plate motion (i.e. towards the North American trench) in the hot spot reference frame. The plateau would have interacted with the trench about 25–30 m.y. before the plume itself was overridden. Reconstructions that suggest the Yellowstone plume was overridden by the continental margin at ca. 50 Ma imply that the related oceanic plateau and underlying buoyant swell reached the trench at ca. 80–75 Ma, i.e. about the time of commencement of the Laramide orogeny. If the Carmacks Group represents a vestige of the plume, reconstructions imply the existence of an oceanic tract between the Intermontane belt terranes and continental North America at that time.

SUMMARY AND DISCUSSION

The accretion to North America at ca. 50 Ma of large oceanic terranes, such as Siletzia, with plume geochemistry and dimensions suggesting it constitutes the vestiges of a LIP (Wells et al. 2014), is strong evidence that a plume was located in the oceanic realm adjacent to the North American continent, much as envisaged by Duncan (1982). Reconstructions (e.g. McCrory and Wilson 2013; Wells et al. 2014) indicate that the ancestral Yellowstone hot spot, if it existed, would have been located close to the Kula–Farallon ridge or to the Kula–Farallon–Resurrection triple junction at that time and would be a viable source for the Siletzia ocean island basaltic rocks. In northern Siletzia, the calculated buoyancy flux (1.1 Mg s^{-1}) of the plume responsible for the Crescent volcanic rocks (Murphy et al. 2003) is similar to that of the modern Yellowstone plume (Sleep 1990). Taken together, this evidence indicates that at ca. 55–50 Ma, a plume existed in the same location as modern Yellowstone and had comparable vigour.

If so, interaction between the continental margin and the ancestral plume must have been preceded by overriding of the related buoyant swell and oceanic plateau that would have been elongate in the direction of motion of the Farallon plate. In this scenario, the swell and plateau of thick oceanic lithosphere would have been overridden by the North American margin at ca. 75 Ma, i.e. broadly coinciding with the onset of Laramide orogenesis and about 20 m.y. before the ancestral Yellowstone plume was overridden. Although back-of-the-envelope calculations suggest that the swell may supply enough positive buoyancy to counteract the negative buoyancy of 80–50 m.y. old Farallon oceanic lithosphere and hence contribute to the formation of the flat slab (Murphy et al. 1998), this contention has not been evaluated rigorously, and so the relationship between the arrival of the swell at the trench and the onset of flat slab subduction remains enigmatic. However, geodynamic models (e.g. Betts et al. 2012, 2015) show that subduction of oceanic plateaus can generate flat slabs. According to the inverse mantle convection model of Liu et al. (2008, 2010), subduction of an oceanic plateau, corresponding to the conjugate of the Shatsky plateau, beginning at about 90 Ma, could initiate the generation of the flat slab, although they acknowledged that their model yielded an initiation age which is about 10 m.y. older than geological evidence for flat slab inception. As the Shatsky rocks are about 150 Ma, its conjugate plateau could be temporally distinct from the plateau related to the ancestral Yellowstone hot spot. Alternatively, the oldest components of the plateau may not have been sufficiently buoyant to resist subduction. The space–time relationships inferred by Liu et al. (2008, 2010) make the Shatsky conjugate plateau a viable alternative to the plateau associated with the Yellowstone plume in initiating the flat slab. The location of the conjugate to the Hess plateau, for which the age is unconstrained, is intriguingly similar to that inferred for the Yellowstone hot spot. The calculated location for the Hess plateau lies adjacent to the inferred location of the Yellowstone swell. Taken together, these relationships imply the possibility of a very complicated composite plateau geometry with older and younger components being overridden by the NAP between 90 and 70 Ma, followed by the accretion of oceanic islands related to the Yellowstone plume by ca. 50 Ma.

After the plume was overridden by the NAP, its effects become more conjectural as it becomes overlain by continental lithosphere. The models of Betts et al. (2012, 2015) suggest that plume magmatism in the upper plate will be concentrated in regions where rising plume material can exploit weaknesses in the oceanic and continental lithospheres. Forearc magmatism is detected until ca. 42 Ma where the crust overlying the plume is thin and undergoing margin-parallel extension (Wells et al. 2014). More controversially, the next phase of magmatism directly or indirectly related to plume activity may be some of the 40–30 Ma magmatism in Oregon to the east of the Cascadia arc, for which isotopic data suggest a large heat source and extensive hydrothermal circulation, and plate reconstructions imply a subjacent Yellowstone plume (Seligman et al. 2014).

Plume-related magmatism between 17 and 0 Ma associated with the YSRP matches the plate motions of the NAP, much as envisaged by Morgan (1972) and Armstrong et al. (1975). Regionally extensive coeval magmatism that occurs some distance from this track may be due to secondary effects (Betts et al. 2015), such as deflection of plume material as it migrates through the upper plate and/or plume-assisted delamination (e.g. Darold and Humphreys 2013).

Laboratory studies suggesting that plumes may be severely distorted or even destroyed by slab-driven subduction processes (Kincaid et al. 2013; Druken et al. 2014) have drawn attention to the importance of tears or windows in the subducting slab where plume material can be transferred to the upper plate (e.g. Johnston and Thorkelson 1997). Betts et al. (2015) showed that the interaction between an oceanic plateau and a continental margin causes rapid trench advance, accretion of the plateau which is transferred to the upper plate, followed by re-establishment of subduction outboard of the plateau. They also showed that the additional presence of a buoyant plume beneath the plateau results in the formation of a slab window beneath the accreted plateau, where plume material can be transferred from the lower to the upper plate. Such processes provide some theoretical support for the models where plume material is deflected as it exploits weaknesses in the subducted slab (fractures, tears, windows, regions where lithosphere has been delaminated) to enter the upper plate, and explains why the products of these events may not always be located adjacent to the calculated hot spot track. Also, plume material transferred to the upper plate and deflected away from the hot spot track moves with the upper plate, and so is rapidly separated from its heat source.

The synthesis of Fletcher and Wyman (2015) that 29% of mantle plumes have been located within 1000 km of a subduction zone over the past 60 Ma implies that interaction between plumes, their plateaus and buoyant swells with subduction zones should be common in the geologic record. If so, the geologic evolution of the western United States may represent a Late Mesozoic–Cenozoic analogue for a common, but overlooked mode of orogenic activity that has occurred since plate tectonics first operated on Earth. This interaction is very complex in space and time, and a resolution of the rival models (eastward versus westward subduction in the Mesozoic) may be prerequisite to a deeper understanding of this interaction.

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