Late Triassic–Early Jurassic Stikine–Yukon-Tanana terrane collision and the onset of accretion in the Canadian Cordillera: Insights from Hazelton Group detrital zircon provenance and arc–back-arc configuration

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ABSTRACT

The Hazelton Group is a Rhaetian–Bajocian (uppermost Triassic–Middle Jurassic) volcano-sedimentary sequence that represents both the last pre-accretionary arc volcanic cycle of Stikinia and its early synaccretionary aftermath. Hazelton magmatism of central Stikinia succeeded the Late Triassic (mainly Carnian–Norian) Stuhini arc, which ceased activity as a result of end-on collision with the pericratonic Yukon-Tanana terrane. The Hazelton volcanic belt lies to the south along strike with the coeval Whitehorse trough, the synorogenic clastic basin that developed on top of the Stikinia–Yukon-Tanana collision zone. Whereas the sources of voluminous clastic sediments in the Whitehorse trough were its rapidly exhuming shoulders, the thin clastic intervals in the Hazelton Group in northwestern British Columbia were derived from local to subregional block uplifts that supplied mainly ca. 230–215 Ma zircons eroded from the plutonic roots of the Stuhini arc. Lesser components include late Paleozoic (ca. 350–330 Ma) zircons from Stikinia’s basement and penecontemporaneous (ca. 205–172 Ma) zircons from Hazelton volcanic/subvolcanic sources. Reexamination of the four main volcanic fields that make up the lower Hazelton Group suggests that the main Hazelton volcanic belt formed a southward-convex magmatic arc from eastern Stikinia across the Skeena arch, including the Toodoggone and Telkwa belts, with the Spatsizi and Stewart–Iskut regions of northwestern British Columbia in its back-arc. The Whitehorse trough and Hazelton belt represent a collision zone to active arc pair. Southward advance of the arc and counterclockwise rotation of the Stikinia microplate contributed to closure against the Quesnellia arc and assembly of the inner Canadian Cordilleran terrane collage.

INTRODUCTION

Arc-continent collisional processes are a key driver in the formation of orogenic belts (Brown and Ryan, 2011), particularly during their incipient stages. Because no two natural examples are alike, study of a variety of currently active and well-preserved ancient systems is necessary to identify the factors that account for their diversity, from preexisting crustal architectures to angles and velocities of convergence (Brown et al., 2011). Development of the North American Cordillera as a continent-scale orogenic belt began in latest Triassic to Early Jurassic time as a localized collision between a complex of intra-oceanic arc and pericratonic terranes and the continental margin, affecting parts of present-day British Columbia, Yukon, and eastern Alaska (Nelson et al., 2013; Colpron et al., 2015). The main players in this event were the northern parts of Stikinia and Quesnellia (late Paleozoic–early Mesozoic arc terranes), the pericratonic, mid-Paleozoic arc-related Yukon-Tanana terrane, and the Cache Creek terrane, a subduction complex located between Stikinia and Quesnellia (Fig. 1). Original modeling of the collision called for counterclockwise oroclinal rotation of the outer, Stikinia limb around a hinge located within the Yukon-Tanana terrane, which led to enclosure of the Cache Creek subduction complex between the two arc terranes (Mihalynuk et al., 1994). The orocline model was an attempt to account for precollisional common geological units that link northern Quesnellia and northern Stikinia to the Yukon-Tanana terrane. Identical late Paleozoic arc sequences constitute an overlap assemblage that links northern Quesnellia and the southeastern Yukon-Tanana terrane (Nelson and Friedman, 2004). Similarly, a Paleozoic metamorphic assemblage in far northwestern British Columbia was interpreted as transitional between Stikinia and the southwestern Yukon-Tanana terrane (Mihalynuk, 1999). Late Triassic and Early Jurassic arc-related intrusive suites of Quesnellia and Stikinia provide strong evidence of interterrane contiguity. As shown in Figure 1, early Mesozoic plutonic trends of Quesnellia and Stikinia merge northward into an apparent hairpin bend within the Yukon-Tanana terrane in central Yukon and eastern...
Alaska. The plutonic belts were considered to mark the axial regions of the Quesnellia and Stikinia arcs. Prior to collision, they were modeled as adjacent intra-oceanic arc festoons that joined at a cusp on older pericratonic crust of the Yukon-Tanana terrane, similar to the way in which the present-day Aleutian and Kurile volcanic island-arc chains attach to the Kamchatka Peninsula, a rifted fragment of Eurasia (Mihalynuk et al., 1994). Oroclinal closure would have required on the order of 80° interlimb rotation to the near-parallel configuration of Stikinia and Quesnellia today. This process, however, is geodynamically improbable. Recent numerical modeling experiments involving compression of linear beams in layered media have shown that crustal ribbons do not bend around vertical axes to form oroclines; instead, they thicken and form gentle upright folds (Smith et al., 2021).

Such challenges to the inferred mechanism of accretion invite a thorough reexamination of the geological relationships that guide our understanding of the Stikinia–Yukon-Tanana–Quesnellia–Cache Creek–North America collision process. Stikinia has been described as a Philippine-style microplate that, in Early Jurassic time, during initial stages of the arc-continent collision, was the site of two outward-facing magmatic arcs separated by a central back-arc basin, extending for up to 2000 km along the axis of the terrane (Marsden and Thorkelson, 1992). It was this elongate entity that was thought to have undergone rigid-body rotation during oroclinal closure. Extensive exposures of the volcano-sedimentary Hazelton Group and associated plutons (uppermost Triassic–Middle Jurassic) in central Stikinia between latitudes 54°N and 58°N (Fig. 1) offer opportunities to address key questions about the role that this assemblage played in the collision. What was the configuration of the Hazelton arc or arcs and back-arc? Do Hazelton arc belts merge into latest Triassic–Early Jurassic plutonic suites in the Yukon-Tanana terrane, or are they separate, unrelated entities? Are there sedimentological ties between central Stikinia and the Whitehorse trough, the coeval synorogenic basin that onlapped the core of the growing orogen (Colpron et al., 2015)? How did development of the Hazelton arc or arcs influence the development and propagation of the collision zone?

Figure 1. Terrane map of the Canadian Cordillera with emphasis on Triassic–Jurassic assemblages of Stikinia and other Intermontane terranes (Quesnellia, Yukon-Tanana, Cache Creek terrane). BC—British Columbia.
This contribution presents new detrital zircon data from Hazelton Group sites along a north-south transect from Dease Lake to Stewart, British Columbia, Canada (Fig. 1), which were analyzed to identify sedimentary source units and regions and as a means of establishing the paleogeographic setting of the northern Hazelton arc–back-arc system. We then reviewed numerous recent geological studies of the Hazelton Group to construct a revised model of arc configuration and its relationship to northern Stikinia and the Whitehorse trough, which provides new insights into the evolution of this complex arc-continent collision zone.

**TECTONIC SETTING**

The focus of this study, Stikinia, is a multistage, late Paleozoic to early Mesozoic island-arc terrane located in the core of the Canadian Cordillera. It is part of the Intermontane terrane collage, located between the Yukon-Tanana terrane to the north and west and the Cache Creek terrane to the east, which in turn is tectonically juxtaposed with Quesnellia and the Slide Mountain terrane farther east (Fig. 1; Nelson et al., 2013).

**Circum-Stikinia Intermontane Terranes**

To the north, west, and east of Stikinia, the Yukon-Tanana terrane (Fig. 1) comprises a Neo-protoreozoic to Lower Devonian continental margin sequence (Snowcap assemblage) that is intruded and overlain by mid- to late Paleozoic magmatic arcs and associated basinal strata (Finlayson, Klinkit, and Klondike assemblages; Colpron et al., 2006; Piercey and Colpron, 2009; Nelson et al., 2013). Voluminous magmatism occurred at 365–330 Ma and 264–252 Ma (Nelson et al., 2006a). A belt of Late Triassic–Early Jurassic plutons in central Yukon-Tanana terrane apparently branches and continues southward into the Stikinia and Quesnellia arc systems (Fig. 1; Colpron et al., 2015). Radiometric and geochemical studies, as well as late Paleozoic macrofaunal provenance studies, have demonstrated peri-Laurentian origins for this part of Yukon-Tanana terrane (Nelson et al., 2006a). The predominance of Paleoproterozoic (ca. 2.1–1.8 Ga) and Archean (2.7–2.4 Ga) zircons in the Snowcap assemblage reflects cratonal sources of northwestern Laurentia (Piercey and Colpron, 2009). It is interpreted as a fragment of the continental margin that detached during Late Devonian–early Mississippian back-arc rifting and opening of the Slide Mountain ocean (Nelson et al., 2006a). The Slide Mountain ocean closed in Permian to Triassic time, resulting in juxtaposition of the Yukon-Tanana terrane with the outer Laurentian margin (Beranek and Mortensen, 2011).

Interpreted lateral gradations of Paleozoic Stikine assemblage and Yukon-Tanana strata suggested early linkages between the two terranes (McClelland, 1992; Mihalynuk et al., 1994; Mihalynuk, 1999). However, the mapped contact between Stikinia and the Yukon-Tanana terrane in northwestern British Columbia is not depositional but tectonic, represented by the ductile, sinistral Wann River shear zone, part of the Llewellyn fault zone of far northwestern British Columbia (Fig. 1; Mihalynuk, 1999), which is cut by a ca. 184 Ma synkinematic pluton with ca. 178 Ma Ar-Ar cooling ages (Currie and Parrish, 1993). A recent study of detrital zircons in central Stikinia showed that its pre-Devonian basement is probably very unlike that of the Yukon-Tanana terrane (George et al., 2021). These observations conflict with a key premise of the orocline model—that Stikinia and the Yukon-Tanana terrane were adjacent parts of a single, continuous crustal block prior to the onset of collision (Mihalynuk et al., 1994). We return to this point in the Discussion section.

Quesnellia, like Stikinia, is a long-lived, multi-episodic late Paleozoic to early Mesozoic arc terrane (Fig. 1; Beaty et al., 2006), linked to the Yukon-Tanana terrane at its northern end (Colpron et al., 2007). The Upper Paleozoic Lay Range volcano-sedimentary assemblage of northern Quesnellia has been correlated with the Klinkit Group of the southeast Yukon-Tanana terrane (Fig. 1; Mihalynuk et al., 1994). Early Paleontological studies of Permian limestones in seamount successions identified coralline and fusulinid faunas with strong affinities to those of mainland China and Japan. These Tethyan faunas indicate a western Panthalassan origin exotic to western Laurentia (Monger and Ross, 1971; Ross and Ross, 1983). Later revisions accept mid-Panthalassic, but nevertheless a far-traveled history (Orchard et al., 2001); in paleogeographic reconstructions, they are placed in a mid-oceanic setting, well outboard of the Stikinia and Quesnellia arcs (Belasky and Stevens, 2006). Until recently, the Cache Creek terrane has been regarded as an accretionary complex that represents the collapse of an extensive ocean basin, “the Cache Creek ocean” (Mihalynuk et al., 1994; Nelson et al., 2013). However, the northern Cache Creek terrane is now known to contain extensive remnants of late Permian–Middle Triassic suprasubduction-zone ophiolites and mafic to bimodal primitive volcanic-arc rocks (Childe et al., 1998; English et al., 2010; Schiarizza, 2012; Zagorevski et al., 2016; McGoldrick et al., 2017, 2018; Bickerton et al., 2020).
Stikinia

Stikinia is the largest accreted terrane (2000 x 300 km) within the Canadian Cordillera (Fig. 1). Described as a Philippine-style microplate (Marsden and Thorkelson, 1992), it developed as a mobile intra-oceanic arc terrain within the peri-Laurentian realm over a 200 m.y. interval from the Late Devonian through Early Jurassic. Three identified unconformity-bounded island-arc volcano-sedimentary successions are the Upper Paleozoic Stikine assemblage (Logan et al., 2000) and its correlatives; the Middle to Upper Triassic Stuhini, Tahkla, and Lewes River groups (Fig. 2; Logan et al., 2000; Monger and Church, 1977; Hart, 1997); and the uppermost Triassic to Middle Jurassic Hazelton Group (Fig. 3; Nelson et al., 2018). Mesozoic intrusive suites include the Late Triassic Stikine and Galore plutonic suites (coeval and comagmatic with the Stuhini Group; Fig. 2) and small plutons coeval and comagmatic with the Hazelton Group, such as the latest Triassic Tatogga suite near the village of Iskut and the Early Jurassic Texas Creek suite between Iskut and Kitsault (Fig. 3; Nelson et al., 2018; Nelson and van Straaten, 2021).

Major, long-lived structural corridors of Stikinia played an important role in controlling its internal stratigraphy and paleogeography (Febbbo et al., 2019a; Nelson and van Straaten, 2021). The terrane is transected by orthogonal, orogen-normal structural corridors: the easterly Stikine and Skeena arches, the northwesterly Nikkitkwa trough, and a through-going northerly lineament and fault corridor in the Iskut-Stewart-Kitsault region (Figs. 2 and 3). The Bowser Basin, a Middle Jurassic to middle Cretaceous clastic successor basin, lies in the center of the terrane, framed by these structural corridors. Its main fill is sediment derived from the Cache Creek tectonic belt to the northeast (Evenchick et al., 2007). For purposes of this discussion, we have subdivided Stikinia into four subregions:

(1) North-northeastern Stikinia includes the northernmost prong of Stikinia in Yukon and the Whitehorse trough (Fig. 1; Colpron et al., 2015).

(2) Northern Stikinia includes the Stikine arch region and the area to the north, excluding the Whitehorse trough (Figs. 2 and 3).

(3) Central Stikinia lies south of the Stikine arch and extends south across the Skeena arch near Terrace and Smithers. The Bowser Basin (Figs. 2 and 3) divides the older rocks of the central region into western and eastern exposures.

(4) Southern Stikinia lies south of the Skeena arch (Figs. 2 and 3). It is not discussed in this paper. Paleozoic strata are exposed in isolated block-faulted uplifts: the Stikine assemblage in northern and central-western Stikinia (Gunning et al., 2006), the Asitka Group of central-eastern Stikinia (Lord, 1948), the Zymoetz Group along the Skaena arch (Nelson et al., 2008), and the Tahkini assemblage near Whitehorse (Hart, 1997). Sparse U-Pb zircon ages from felsic volcanic rocks in these sequences cluster ca. 340–320 Ma (Childe, 1997; Gunning et al., 2006; Nelson, 2017). Late Devonian–early Mississippian (ca. 360–350 Ma) plutons are locally exposed (Ash et al., 1997a; Logan et al., 2000).

Following a late Permian–Middle Triassic hiatus, volcanism and plutonism flourished in central and northern Stikinia during Carnian–Norian time. Figure 2 shows the distribution of main Triassic supracrustal units, with overlying units of the Hazelton and Laberge Groups removed. The main magmatic axis is defined by thick accumulations of mafic to intermediate volcanic strata and accompanying ca. 229–215 Ma Stikine Plutonic Suite intrusions. It includes the Takla volcanic belt along the eastern margin of Stikinia (Takla Group of McConnell Creek area; Fig. 2; Tipper and Richards, 1976; Monger and Church, 1977), volcanic-rich facies of the Stuhini Group of the Stikine arch trend along the northern “shoulder” of the terrane (Fig. 2; Brown et al., 1996; van Straaten and Wearmouth, 2019), and the Povoas Formation (Lewes River Group) of southern Yukon (Hart, 1997; Bordet et al., 2019); together, these outline an eastward-convex, east-facing arc axis (Fig. 2; Nelson and van Straaten, 2021). Figure 2 shows the continuity of the Stuhini Group (defined in British Columbia) and the Lewes River Group (defined in Yukon) as a single volcanic-sedimentary belt. Small plutons of the Stikine (217–214 Ma) and Pyroxene Mountain (220–211 Ma) suites intrude both the Lewes River Group and adjacent Yukon-Tanana terrane, showing Late Triassic linkage between the two terranes (Sack et al., 2020).
In the McConnell Creek area of central-eastern Stikinia (Fig. 2), the volcanic-rich Takla Group is up to 4500 m thick and represents an evolution from late Carnian submarine pyroxene-phyric flows to early Norian mainly subaerial basalt and andesite (Monger and Church, 1977). This region was recognized as a northwest-trending volcanic archipelago by Tipper and Richards (1976). Along strike to the south, the Tachek pluton and Stern Creek phase of the Endako Batholith (Fig. 2) have been dated by U-Pb methods on zircon as ca. 226–219 Ma (MacIntyre et al., 2001); these Late Triassic plutons were probably emplaced into Takla volcanic centers, and thus represent a southward continuation of the arc axis.

Large, calc-alkaline Late Triassic (ca. 229–215 Ma) intrusions of the Stikline Plutonic Suite form a roughly east-west trend along the Stikine arch, which defines the axis of the Stuhini magmatic arc in that region, including the Cake Hill pluton within the Hotailuh Batholith discussed later in this paper (Fig. 2; Nelson and van Straaten, 2021). The Stuhini Group is widespread in the Stikine arch region, consisting primarily of large volcanic buildups of augite-plagioclase–phyric flows and coarse pyroclastic deposits. The Hickman Batholith is offset to the south from the main plutonic trend. Its northerly orientation is due to emplacement along preexisting lineaments or fault strands parallel to the Mess Creek fault (Fig. 2; Nelson and van Straaten, 2021). An arc volcanic–rich Stuhini section west of the Hickman Batholith is over 3500 m thick; the main components are calc-alkaline basalts, basaltic andesites, and trachyandesites derived from local eruptive centers (Brown et al., 1996).

South of the Stikine arch, between the McTagg anticlinorium and Kitsault, primary volcanic units diminish within the Stuhini Group, replaced by volcaniclastics and sedimentary units with subordinate augite-phyric basalts (Fig. 2; Massey et al., 2008). Two Late Triassic volcanogenic massive sulfide deposits occur within this area, Granduc and Rock and Roll (Fig. 2; Mihalynuk et al., 2019). Southernmost Stuhini Group equivalents near Terrace (Nelson et al., 2006b) and sparse exposures south of the Skeena arch (Diakow et al., 1997) consist of very thin (<100 m) sections of black shale, argillite,
siltstone, and chert typical of starved basin deposits. The likely tectonic setting of these relatively volcanic-poor, relatively thin marine sedimentary sequences and volcanicogenic massive sulfide occurrences (equivalent to modern seabed massive sulfides) is a back-arc basin located south and west of the main Stuhini-Takla arc axis (Fig. 2).

Stuhini–Takla–Lewes River arc activity was terminated in latest Triassic time by a regional tectonic collisional event expressed as widespread deformation of the Stuhini Group, regional uplift (Greig, 2014; Logan and Mihalynuk, 2014), and deposition of Whitehorse trough synorogenic siliciclastic strata (Lower Jurassic Laberge Group) on top of the Lewes River Group in southern Yukon (Colpron et al., 2015) and along the northeastern shoulder of Stikinia and adjacent Cache Creek terrane in northeastern British Columbia (Fig. 1). Youngest Stikine suite plutonic phases of ca. 215 Ma record the wanning stages of the Stuhini arc. Subsequently, between 212 and 205 Ma, a distinct magmatic belt, the Galore trend, developed in central-western Stikinia (Fig. 2). It is oriented north-south, orthogonal to the Stikine arch plutonic trend but parallel to prominent regional lineaments. The Galore plutonic suite that defines this trend is typified by small plutons, stocks, and rare volcanic rocks of predominantly alkaline affinity, including monzonites, syenites, ultramafic-syenite complexes, and leucite-bearing volcanic and intrusive rocks (Fig. 2; Brown et al., 1996; Logan and Koyanagi, 1994; Enns et al., 1995; Coulson et al., 1999, 2007; Romios Gold Resources, 2008; Mihalynuk et al., 2011, 2012). The anomalous northerly orientation, strong structural control, small magma volumes, and prevalent alkaline chemistry of the Galore trend suggest that it is a postsubduction feature generated by partial melting of previously subduction-metasomatized lithosphere (Nelson and van Straaten, 2021).

**Stikinia in Latest Triassic–Middle Jurassic Time**

Comparison of Figures 2 and 3 shows profound differences in the distribution of arc-related volcanic and intrusive suites in Stikinia of Carnian–Norian versus Rhaetian–Bajocian age. The prominent Late...
Triassic arc axial zone, which extended over 800 km from eastern-central Stikinia near Smithers to north of Whitehorse in Yukon (Fig. 2), became extinct by Rhaetian time. Rhaetian and younger arc-related volcanic and intrusive units, the Hazelton Group and its accompanying intrusions, are only present south of the Stikine arch (Fig. 3), representing a southward shift of ~400 km. North of the Stikine arch, Triassic volcanic sequences were onlapped by siliciclastic sedimentary strata of the Whitehorse trough (Laberge Group; Fig. 3).

As first articulated by Marsden and Thorkelson (1992), the Hazelton Group is exposed over a 300 km width across strike east to west (Fig. 2), which far exceeds the width of any modern volcanic arc. They therefore modeled it as two separate arcs with an intervening back-arc region: an eastern arc, composed of the eastern Telkwa Formation and Tooodogone Formation, and a western arc, composed of the western Telkwa Formation and the Hazelton Group in the Iskut region, separated by a narrow back-arc in the Nilkitkwa trough and Spatsizi volcanic field (Fig. 3; Marsden and Thorkelson, 1992).

In the 30 years since publication of the “two-arc” model (Marsden and Thorkelson, 1992), numerous broad geological studies have provided detailed, nuanced characterizations of the volcanic belts and fields that make up the Hazelton Group (Diakow et al., 1993, 2005; Diakow and Rhodes, 2006; Lewis, 2013; Barresi et al., 2015; Nelson et al., 2016). This wealth of new knowledge, viewed in comparison with well-studied modern examples, provides a basis for reconsidering the configuration of the Hazelton arc–back-arc system.

The Hazelton Group includes all uppermost Triassic to Middle Jurassic volcanic and sedimentary successions in Stikinia, except clastic strata of the Whitehorse trough in the north-northeastern part of the terrane (Fig. 1). In central Stikinia, the Hazelton Group has been subdivided into two parts: (1) Rhaetian to Pliensbachian volcanic and sedimentary strata of the lower Hazelton Group, and (2) the Pliensbachian to Callovian upper Hazelton Group (Fig. 3; Gagnon et al., 2012). The lower Hazelton Group represents arc and back-arc volcanism and sedimentation, whereas the upper Hazelton Group consists of postarc strata. Rocks of the lower Hazelton Group and accompanying latest Triassic to Early Jurassic plutons only crop out south of the Pitman fault, where they form four distinct volcanic fields (Fig. 3): (1) the Telkwa volcanic field along the Skeena arch and to the east in the McConnell Creek area (Tipper and Richards, 1976; Barresi et al., 2015), (2) the Tooodogone volcanic field east of the Bowser Basin (Diakow et al., 1993; Duuring et al., 2009), (3) the Stewart-Iskut volcanic field, which forms a northerly trend west of the Bowser Basin (Lewis, 2013; Nelson et al., 2018), and (4) the Klastline-Spatsizi field between the Pitman fault and the northern Bowser Basin, named for the Klastline Plateau near Iskut and the Spatsizi River area to the southeast (Thorkelson et al., 1995; Nelson et al., 2018).

In central Stikinia, postarc strata of the upper Hazelton Group comprise both regionally extensive, Pliensbachian to Callovian stratified sedimentary and sedimentary-volcanic units that overlie the lower Hazelton Group, and the Iskut River Formation, a bimodal volcano-sedimentary Middle Jurassic succession that constitutes the fill of the north-trending Eskay rift west of the Bowser Basin (Fig. 3; Gagnon et al., 2012). In northern Stikinia, along and north of the Stikine arch and west of the Mess Creek fault, the lower Hazelton Group is missing, and scattered outcrops of Pliensbachian–Toarcian upper Hazelton Group strata unconformably overlie deformed Stuhini Group and Stikine suite plutons (Brown et al., 1996; Greig, 2014; van Straaten and Nelson, 2016; van Straaten and Gibson, 2017; van Straaten and Bichlmaier, 2018), implying a >25 m.y. hiatus in sedimentation and magmatic activity.

Figure 4 illustrates the temporal differences in onset of Hazelton Group sedimentation and volcanism in central versus northern Stikinia, separated by the Pitman fault. South of the fault, the base of the Hazelton Group is Rhaetian to Hettangian, and onset of volcanism varied between Rhaetian near the village of Iskut and Sinemurian on the flanks of the McTagg anticlinorium (Fig. 3, 4, columns A, B, C; Nelson et al., 2018). To the north, near Gnat Pass, Toarcian Spatsizi Formation sandstones and siltstones rest nonconformably on the ca. 216 Ma Cake Hill pluton (Fig. 4), part of the Hotailuh Batholith (Fig. 2), and in turn pass gradationally upward into the mainly volcanic Middle Jurassic Horn Mountain Formation (Fig. 4, column D; van Straaten and Nelson, 2016). This contrast suggests that northern Stikinia was an amagmatic tectonic highland throughout latest Triassic to Early Jurassic time, when the lower Hazelton Group volcanic fields were active between the Stikine and Skeena arches.

As well as the relatively longer time span across the pre-Hazelton unconformity in northern versus central Stikinia, the intensity of Triassic–Jurassic deformation and depth of erosion decrease markedly from north to south. In the western Stikinia arch region, steeply dipping, folded and faulted Norian Stuhini Group strata are overlain on a spectacular unconformity by gently to moderately dipping, Pliensbachian–Toarcian Hazelton strata (Brown et al., 1996). In the eastern Stikinia arch region, upper Hazelton Group strata lie nonconformably on the ca. 216 Ma Cake Hill pluton (van Straaten and Nelson, 2016), providing evidence for Late Triassic–Early Jurassic erosion into the plutonic roots of the Stuhini arc. In contrast, in central Stikinia, south of the Stikine arch, the basal Hazelton unconformity varies from locally angular to low-angle to paraconformable. Angular unconformities are observed east of the McTagg anticlinorium (Fig. 3; Henderson et al., 1992) and at Oweegee Dome (Greig, 1992). Further west on Snippaker Ridge, the basal Hazelton conglomerate is paraconformable above Stuhini volcanioclastic strata (Kyba and Nelson, 2015). Further south, at Red Mountain near Stewart, the contact between Stuhini black argillite and chert and Hazelton felsic tuffs and siliciclastics occurs within a marine sequence. At the southern end of the Iskut-Stewart belt near Kitsault, uppermost Triassic basal Hazelton strata, including coarse, proximal debris flows, were deposited in a series of easterly pull-apart basins (Miller et al., 2020). Near Terrace, the Hazelton Group lies with low-angle unconformity on a thin, distal Stuhini sequence and underlying Paleozoic strata (Nelson et al., 2008). In summary, while strong, regional pre-Hazelton compressional deformation and deep exhumation prevailed in the Stikine arch region of northern Stikinia, deformation in central Stikinia was less intense, with local zones of compressional deformation and extension-related
block-faulting developed in what may have been a mild transcurrent stress regime (Miller et al., 2020; Board et al., 2021), with reactivation along basement lineaments (Nelson and van Straaten, 2021).

**DETRITAL ZIRCON GEOCHRONOLOGY OF STEWART-ISKUT-STIKINE ARCH**

As part of this study, we collected detrital zircon samples along a south-to-north transect through the Stewart-Iskut and Klastline areas to the Gnat Pass area near Dease Lake (locations 1–5, Fig. 3). The transect included the northernmost extent of lower Hazelton Group volcano-sedimentary sections near Iskut village, bounded to the north by the Stikine arch. We sampled siliciclastic units to aid paleogeographic reconstructions by establishing the source units and source regions that contributed to the northern Hazelton Group, whether local or regional, within Stikinia or external to Stikinia, and how they compared with those of the Whitehorse trough.

Basal polymictic conglomerates composed mainly of well-rounded granitoid clasts in quartz-rich arkosic matrix are a unique feature of the Hazelton Group in the Stewart-Iskut region (Fig. 4, columns A–C; Figs. 5A–5D). They represent an interval of mature sedimentation that contrasts markedly with volcanic-dominated sections below and above. These conglomerates occur in the Hettangian (-Sinemurian?) Jack Formation on the periphery of the McTagg anticlinorium (Figs. 3, 5A, and 5B; Henderson et al., 1992; Lewis, 2013; Nelson and...
Kyba, 2014), the Rhaetian Snippaker unit near the lower Iskut River (Figs. 3 and 5D; Kyba and Nelson, 2015; Nelson et al., 2018), a thin lens near the Red Chris deposit (Figs. 3 and 5E; Rees et al., 2015), and possibly age-equivalent, unnamed Lower Jurassic clastic units near Mess Creek (Fig. 3; Logan et al., 1993, 2000). The Jack Formation overlies the Stuhini Group on the western flank of the McTagg anticlinalorium with angular unconformity (Fig. 5A; Lewis, 2013), whereas the Snippaker unit is paraconformable above the uppermost Stuhini volcanioclastic grit-sandstone on Snippaker Ridge (Kyba and Nelson, 2015). In measured sections, the Jack Formation is 400–900 m thick (Nelson and Kyba, 2014), and the Snippaker unit is ~90 m thick (Kyba and Nelson, 2015). Both siliciclastic units are succeeded by the volcanic Betty Creek Formation (Fig. 4, columns A and B), which is divided lithologically into the andesitic Unuk River unit and local felsic accumulations of the Mount Johnny (ca. 194 Ma) and Brucejack Lake (ca. 185–178 Ma) units (Nelson et al., 2018). Near the village of Iskut, the onset of Hazelton andesitic volcanism and accompanying intrusion of the Tatoga suite occurred some 12 m.y. earlier, ca. 207 Ma (Fig. 4, column C; Nelson et al., 2018; George et al., 2021).

The thickness of the mainly andesitic Betty Creek Formation on the flanks of the McTagg anticlinalorium varies from 0 to ~2 km, due to volcanic and basinal facies changes and subsequent pre–upper Hazelton Group erosion. The thickness in drill sections near Brucejack Lake east of the anticlinalorium is ~1700 m (Tombe et al., 2018); this is probably a maximum in the region (see Lewis, 2013). In some areas, the Betty Creek Formation is overlain by Pliensbachian–Toarcian siliciclastic strata of the Spatsizi Formation (Nelson et al., 2018), which represent volcanic quiescence and shallow-marine deposition that succeeded lower Hazelton Group volcanism. Elsewhere, the Middle Jurassic, rift-filling Iskut River Formation lies directly on Betty Creek Formation or even Jack Formation strata on the shoulders of the Eskay rift. The Iskut River Formation comprises felsic volcanic rocks of the Bruce Glacier unit, basalts of the Willow Ridge unit, fine-grained sedimentary strata of the Mount Madge unit, and fault scarp–related, coarse clastic lenses of the Kinsakan unit (Nelson et al., 2018).

Eleven samples were collected along the Stewart-Iskut-Stikine arch transect to represent lower and upper Hazelton Group units at the McTagg anticlinalorium (locations 1 and 2, Fig. 3), a Stuhini-Hazelton section on Snippaker Ridge (location 3, Fig. 3), and basal Hazelton conglomerates near the village of Iskut (location 4, Fig. 3) and at Gnat Pass (location 5, Fig. 3). They are shown in stratigraphic context in columns A–D in Figure 4. One is from the uppermost Stuhini Group, one is from the basal conglomerate of the Hazelton Group near Iskut, two are from the Snippaker unit, three are from the Jack Formation, one is from the base of the Betty Creek Formation, two are from the Spatsizi Formation, and one is from the base of the Iskut River Formation (Bruce Glacier unit). Zircon grains from conglomerate samples represent both matrix and clasts. Analytical data are available in Nelson et al. (2021).

Methods

U-Pb zircon data were obtained by laser-ablation–inductively coupled plasma–mass spectrometry at the Pacific Centre for Isotopic and Geochemical Research (PCIGR) at the University of British Columbia between 2013 and 2017, employing methods as described by Tafti et al. (2009). From each sample, 65 grains were handpicked and mounted for analysis. Instrumentation employed for LA-ICP-MS dating of zircons at PCIGR includes a New Wave UP-213 laser-ablation system and a Thermo Finnigan Element2 single-collector, double-focusing, magnetic-sector ICP-MS. Zircons greater than ~50 μm in diameter were picked from the mineral separates and were mounted in an epoxy puck along with several grains of the 337.13 ± 0.13 Ma Plešovice zircon standard (Sláma et al., 2007), together with a 416.78 ± 0.33 Ma Temora 2 reference zircon (Black et al., 2003, 2004), and brought to a very high polish. The surface of the mount was washed for 10 min with dilute nitric acid and rinsed in ultraclean water prior to analysis. The highest quality portions of each grain, free of alteration, inclusions, or possible inherited cores, were selected for analysis. Line scans rather than spot analyses were employed in order to minimize elemental fractionation during the analyses. A laser-power level of 42% and a 25 μm spot size were used. Backgrounds were measured with the laser shutter closed for 10 s, followed by data collection with the laser firing for ~35 s. The time-integrated signals were analyzed using Iolite software (Patton et al., 2011), which automatically subtracts background measurements, propagates all analytical errors, and calculates isotopic ratios and ages. Corrections for mass and elemental fractionation were made by bracketing analyses of unknown grains with replicate analyses of the Plešovice zircon standard. A typical analytical session at the PCIGR consisted of four analyses of the Plešovice standard zircon, followed by two analyses of the Temora 2 zircon standard, five analyses of unknown zircons, two standard analyses, five unknown analyses, etc., and finally two Temora 2 zircon standards and four Plešovice standard analyses. The Temora 2 zircon standard was analyzed as an unknown in order to monitor the reproducibility of the age determinations on a run-to-run basis.

Probability density and histogram plots of detrital zircon ages were created for each sample using the Isoplot 3.70 and 4.15 add-ins for Microsoft Excel (Ludwig, 2008). Analytical data and individual probability plots for all sample analyses were presented in Nelson et al. (2021). To facilitate comparison between samples, we used the Excel macro provided on the Arizona LaserChron Center’s Web site (https://sites.google.com/a/laserchron/home/) to create a series of stacked normalized probability plots (Fig. 6; Nelson et al., 2021). To identify and characterize the probable contributing sediment sources, we separated subpopulations in the samples via the following sequence of procedures (Table 1; details in individual files in Nelson et al., 2021: see sheets “Isoplot calculations”):

1. A discordance filter of <10% was applied to all samples: (1−[206Pb/238U age – 207Pb/206Pb age]/[207Pb/206Pb age]) x 100.

2. Ages of grains were tested for discontinuities via a lack of overlap of adjacent grain ages at the 2σ and 1σ levels when sorted by age, and as small breaks in plots of grain ages sorted by age (Nelson et al., 2021).
Figure 5. Photos of rocks from representative stratigraphic sections. (A) Angular unconformity at the base of the Jack Formation west of McTagg anticline. Sample 13JN01-05 just below photo. (B) Basal Jack Formation plutonic clast-rich conglomerate. Dark irregular fragments are mudstone intraclasts. Location of sample 13JN01-05. (C) Stratigraphic and structural relationships around the Iron Cap deposit at the northern end of the Kerr-Sulphurets-Mitchell porphyry trend. (D) Basal Snippaker conglomerate. Location of sample 2013JK074. (E) Basal Hazelton conglomerate at Red Chris. Location of sample 2105JK1310. (F) Basal conglomerate of the Toarcian Spatsizi Formation overlying Cake Hill pluton near Gnat Pass. Location of sample 15BVS3-14.
(3) The TuffZirc age routine in Isoplot 3.70 was applied to whole populations as well as possible subpopulations identified in steps 1–2, which identified coherent populations of 8 or more grains.

(4) The Unmix routine in Isoplot 3.70 was applied to populations identified in steps 1–2 and trial ages suggested by Isoplot. Unmix determines the Gaussian distribution that best fits two or more groups of ages for a given population of zircons.

(5) Weighted mean ages for clusters of grains identified by the TuffZirc and Unmix routines were calculated as a test for statistical probability (mean square of weighted deviates [MSWD] = 1, probability of fit > 0.05) that they represent coherent populations.

(6) Cathodoluminescence (CL) images of all grains were examined to establish growth histories and characterize features common to age groupings. Maximum depositional age (MDA) estimates were based on the youngest coherent population in a sample consisting of between 3 and 55 grains (Table 1). The sampled strata have been well dated by enclosed macrofossils and/or U-Pb geochronology on interbedded volcanic units (Table 1). Thus, MDA estimates were used, not to constrain age of deposition, but rather to show the relative contributions of penecontemporaneous (volcanic) versus older, deeply eroded sediment sources.

Results

Results are presented by region to emphasize the evolution of sediment sources through time for a given area. First, we present the McTagg anticlinorium (localities 1 and 2, Fig. 3), followed by the Snippaker Ridge section (locality 3, Fig. 3), and then individual samples from the basal Hazelton Group (Rhaetian) near the Red Chris mine (Iskut, locality 4, Fig. 3) and at Gnat Pass (Toarcian) near Dease Lake (locality 5, Fig. 3).

McTagg Anticlinorium (Localities 1 and 2, Fig. 3)

Sample 13JN1–5 is an arkosic, pebbly grit within polymictic, granodiorite clast-rich conglomerate a...
### Table 1: Summary of Detrital Zircon Age Spectra, and Provenance Interpretations for All Samples Shown on Figure 6

| Sample number | Latitude (°N) | Longitude (°E) | Description | N total/N% | N discordant grain | Probability plot peaks (Fig. 6) | Breaks in distribution (scatter plot and timeline, age in m.y.) | Tufoil* | Utnna* | Weighted averages* | Geological consequence |
|---------------|---------------|---------------|-------------|-------------|-------------------|---------------------------------|-------------------------------------------------|---------|--------|-------------------|-----------------------|
| 1SRS-03-14    | 58.237        | 129.604       | Basaltic (alveolar) pillow lava from on Cale hill pluton, Hotstool Batholith, Great Slave Lake | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 1SJK1310      | 57.703        | 129.875       | Basaltic (alveolar) pillow lava from on Cale hill pluton, Hotstool Batholith, Great Slave Lake | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 2014JK074     | 65.635        | 130.924       | Hydrothermal metamorphics from the Brucejack fault zone near Katir zone of Iron Cap alteration tuff/ignimbrite at 211.75 Ma. | 65/64       |                   | 211 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 2014JK228     | 58.350        | 130.388       | Paleozoic, granite–clast–rich intrusive from the top of Snippaker unit | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 2014JK072     | 58.660        | 131.018       | Basaltic (alveolar) pillow lava from on Cale hill pluton, Hotstool Batholith, Great Slave Lake | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 1SJK120       | 58.572        | 130.416       | Basaltic (alveolar) pillow lava from on Cale hill pluton, Hotstool Batholith, Great Slave Lake | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 1SJK107       | 58.551        | 130.521       | Spotted Sammamish Group, near Iron Cap alteration zone | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |
| 2014JK150     | 58.660        | 131.018       | Basaltic (alveolar) pillow lava from on Cale hill pluton, Hotstool Batholith, Great Slave Lake | 65/64       |                   | 215 ± 5 Ma (94 grains)          | Scatter plot shows homogenous population except for oldest grains (234, 228, 210 Ma), which are outliers. Probable xenocrysts come from older plutonic and/or volcanic source. | All zircons 215 ± 5 Ma. Single source (Cale hill pluton, Hotstool Batholith). | Not applicable, as all data from single peak. | All zircons 215 ± 5 Ma. Mean square of weighted desviations (MSWD) = 0.95, probability of Hi (prob.) = 0.97. | Uncommonly overites Hotstool Batholith (Cale hill pluton; ca. 220 Ma, Van and Nelson, 2016). Flooded locally higher in section yielded early Triassic fauna (Patterson and Perry, 1981). |

Note: Preferred fission age shown in bold.

* N total=number of grains analyzed; N%=number of grains with less than 10% discordance between measured 207Pb/206Pb and 206Pb/207Pb ages.
  * Tufoil, Utnna, and weighted average calculations are from methods of Ludwig (2002)
  * SHRIMP—thermal ionization mass spectrometry; SHuMP—sensitive high-resolution ion microprobe.
few meters above the base of the Jack Formation on the western flank of the anticlinorium (Fig. 4, column A; Figs. 5A and 5B). The probability plot shows a broad but apparently unimodal main peak at ca. 223 Ma (Fig. 6), with a weighted average of 223.0 ± 1.5 Ma. Possible subpopulations identified by small breaks at ca. 225 and 220 Ma in the scatter plot yielded weighted average ages of ca. 228, 222, and 217 Ma (Table 1). The broad peak may represent a single ca. 223 Ma source or a composite of ca. 228, 222, and 217 Ma sources; all suggest derivation from granitoid bodies of the Stikine Plutonic Suite. Notably, the provenance of this conglomerate is significantly older than its Hettangian-Sinemurian depositional age as determined with fossils (Lewis, 2013; Nelson and Kyba, 2014).

Sample 13JN11–6 is from an enclave of Jack Formation grit within the Iron Cap intrusion on the northeastern flank of the anticlinorium (Fig. 5C; Nelson and Kyba, 2014; Febbo et al., 2019a). Although highly altered, it yielded 42 concordant grains, with all except the youngest forming a broad peak at ca. 224 Ma and giving a corresponding weighted average age of 222.76 ± 1.4 Ma, similar to the basal Jack Formation (13JN1–5, Fig. 6). Based on the low probability for the weighted average of the entire Late Triassic population and a small break at ca. 230 Ma in the scatter plot, this could be a composite population with ca. 230 and 223 Ma weighted average age components (Table 1). A single youngest grain at 200.2 ± 5.2 Ma may be from the surrounding Iron Cap intrusion (part of the ca. 196–190 Ma Sulphurets intrusions; Febbo et al., 2019b; Campbell et al., 2021), or it may indicate Pb loss.

Sample 13JN17–7 was collected from a tuffaceous arkose in a slumped block within a black argillite matrix near the top of the Jack Formation near the Katir mine on the Brucejack property (Nelson and Kyba, 2014). The Jack Formation here is conformably overlain by volcanicogenic strata of the Betty Creek Formation. A block of plagioclase-phryic diorite from the basal Betty Creek volcanic conglomerate deposit yielded a 196.6 ± 0.12 Ma age (Fig. 6, Table 1). The youngest peak is based on five grains and yielded a weighted average 197.3 ± 4.5 Ma age; it probably reflects broadly penecontemporaneous volcanism and agrees within error with the age of a clast in the overlying volcanic conglomerate. A weighted average for the broad main peak yielded a 221.7 ± 1.8 Ma date with low probability (MSWD = 3.3, probability of fit (prob.) = 0.000), which suggests the presence of overlapping subpopulations. A 1σ break at ca. 202 Ma and small series breaks at ca. 228 and 213 Ma in the scatter plot suggest four possible subpopulations with Unmix peaks at ca. 234, 222, 211, and 197 Ma (the tuffaceous component).

A 25-m-thick sequence of quartz-feldspar grit, sandstone, and shale overlies deformed Stuhini Group strata north of the Johnstone fault, which marks the northern limit of the Kerr-Sulphurets-Mitchell (KSM) porphyry trend north of Iron Cap (Fig. 5C; Febbo et al., 2019a). The Johnstone fault is interpreted as a south-side-down Early Jurassic normal fault, with 1–2-km-thick sections of Jack and Betty Creek Formation strata to the south and a thinly veneered Stuhini horst block to the north (Febbo et al., 2019a). Sample 13JN10–7 from a grit in the overlying Hazelton Group unit (Spatsizi Formation) showed a main peak at ca. 186 Ma, and a subordinate older peak at ca. 224 Ma. A weighted average of all grain ages within the youngest population yielded a 186.23 ± 0.70 Ma age, and a weighted average of the oldest six grains yielded a 223.4 ± 3.1 Ma age.

Sample 13JN2–16 was collected from the base of the upper Hazelton Group (Bruce Glacier unit of the Iskut River Formation; Nelson et al., 2018) on the western flank of the McTagg anticlinorium. At this locality, the upper Hazelton Group strata overlie the lower Hazelton Group (Jack Formation) on a significant intra–Hazelton Group unconformity (Fig. 4, column A; Lewis, 2013), probably associated with a shoulder of the Eskay rift. The sample is of a welded tuff that contains accidental clasts from older units and thus offers the potential for providing recycled grains. The main peak in this sample was ca. 173 Ma, and we calculated a corresponding weighted mean age of 173.84 ± 0.80 Ma, which we interpret as the depositional age of the tuff (Fig. 6; Table 1). Older grains formed a subsidiary peak at 189 Ma, sourced from lower Hazelton Group volcanic or Texas Creek suite plutonic sources; there were also three Late Triassic grains (234–220 Ma, likely recycled from underlying Jack Formation strata) and one grain at 343 Ma, also probably recycled from lower Hazelton Group clastic units.

### Snippaker Ridge Section (Locality 3, Fig. 3)

Four samples were collected across the Stuhini-Hazelton Group unconformity and across a second unconformity between the Snippaker unit and overlying volcanic rocks of the Betty Creek Formation (column B, Fig. 4).

Sample 2014JK72 came from the uppermost stratum of the Stuhini Group, a pebbly volcanic greywacke with small hypabyssal clasts. It showed a broad main peak of ca. 212 Ma (Fig. 6; Table 1), with a weighted average age of 210.8 ± 1.4 Ma (MSWD = 2.3, probability of fit = 0.000). The low probability suggests the presence of overlapping subpopulations; utilizing small breaks at ca. 215, 209, and 201 Ma in the scatter plot yielded weighted average ages of ca. 219, 212, 206, and 199 Ma (Table 1). Based on the age scatter for the youngest four grain ages and the fact that the strata are unconformably overlain by the Snippaker unit, which contains latest Triassic ammonites (Lewis, 2013), we interpreted these youngest grains to be affected by Pb loss. We calculated a 205.8 ± 3.9 Ma weighted average age for the next youngest possible subpopulation of grain ages, which is considered to be the maximum depositional age. These ages are younger than typical Stuhini Group magmatism, but they match reported U-Pb ages from the southern Galore belt 20 km to the northwest (Romios Gold Resources, 2008; Mihalynuk et al., 2011, 2012). The oldest 11 grains, between 225 and 216 Ma, could represent xenocrysts in younger magmas or possibly older Stuhini and/or Stikine Plutonic Suite sources.

Sample 2014JK74 came from the basal conglomerate of the Snippaker unit (Fig. 5D), a few meters above the sub–Hazelton Group unconformity. The
conglomerate contains highly rounded clasts of plutonic rocks, radiolarian chert, felsic and mafic volcanic rocks, fossiliferous limestone and woody debris in a matrix of quartz-rich arkose (Kyba and Nelson, 2015). This sample had a low yield of 51 zircon grains, and 18 were rejected due to >10% discordant ages. The remaining 33 concordant grains (Table 1) showed a very broad Late Triassic peak (Fig. 6); a low probability for the weighted average age of this population (223.3 ± 4.4 Ma, MSWD = 2.1, probability of fit = 0.001) suggests the presence of overlapping subpopulations. Utilizing small breaks at ca. 230 and 218 Ma in the scatter plot yielded weighted average ages for possible subpopulations of ca. 245, 222, and 214 Ma, all with MSWD <1 and probability of fit >0.9 (Table 1). It is notable that the ca. 245 and 222 Ma peaks are significantly older than the main populations in the immediately underlying Stuhini Group. The abrupt shift from penecontemporaneous sources (ca. 212–206 Ma) in the uppermost Stuhini Group to older predominant populations in the basal Hazelton conglomerate marks the advent of tectonic unroofing. The sample also yielded three late Paleozoic grains at ca. 330, 315, and 276 Ma and two early Paleozoic grains at ca. 456 and 434 Ma. The late Paleozoic grains reflect Stikine assemblage sources. The presence of Silurian and Ordovician grains in clastic rocks of Stikinia is unusual, but not unique. A small population of 540–460 Ma detrital zircons was reported from an upper Hazelton Group sandstone at Owedge Dome (George et al., 2021), and 455–399 Ma xenocrystic zircons have been recovered from the latest Triassic Big Bulk stock near Kitsault (Miller et al., 2020).

Sample 2014JK228, collected from a polymictic, granitoid clast–rich conglomerate near the top of the Snipperk unit, yielded 64 concordant grains. The sample returned a broad ca. 221 Ma peak (Fig. 6). Excluding the youngest outlying grain age, three clusters were resolved using the Unmix routine, at ca. 229, 221 (main peak), and 210 Ma. These coincide with age clusters in 2014JK74, but they are better resolved because of the larger number of grains.

Sample 2014JK150 was collected from an immature polymictic conglomerate directly above the basal Betty Creek Formation unconformity, where it cuts through the Snipperk unit into the underlying Stuhini Group (Kyba and Nelson, 2015). It yielded 65 concordant grains. The sample showed a main peak at ca. 205 Ma (Fig. 6; Table 1), probably a recycled population from the local Stuhini Group. Lesser peaks at ca. 216 Ma and 198.5 Ma correspond to recycling of Stikine Plutonic Suite grains and initiation of Betty Creek volcanism, respectively. The sample also yielded three Pennsylvania–Permian grains.

**Basal Hazelton Group Conglomerate near Red Chris Mine (Locality 4, Fig. 3; Column C, Fig. 4)**

This sample was located within the Klastline–Spatsizi volcanic field near Iskut village, 20 km south of the Pitman fault, which marks the southern boundary of the Stikine arch (Fig. 3). At the sample site, a thin basal Hazelton Group conglomerate (Fig. 5E) unconformably overlies Stuhini Group basalt and thin-bedded sedimentary strata (Rees et al., 2015).

Sample 15JK1310 (Figs. 5E, 6, and 7; Table 1) came from this conglomerate, which contains intrusive and chert clasts in a matrix of quartz-bearing arkose with orange-weathering carbonate cement (Fig. 5E). Detrital zircons in this sample showed a bimodal distribution (Fig. 7), with two broad peaks at ca. 334 Ma (17 grains) and 221–219 Ma (44 grains). Both peaks are composites. TuffZirc and Unmix identified separate main Triassic peaks at ca. 230 and ca. 220 Ma. Paleozoic grains ranged from 353 to 298 Ma, consistent with multiple sources spanning much of the known age range of the Stikine assemblage. This is the only sample in this study to yield a significant component of Paleozoic grains. Stikine assemblage volcanic and sedimentary strata, intruded by Mississippian plutons, are exposed in a horst within the Pitman fault system 20 km north of the sample site (Ash et al., 1997b). These rocks were probably first exhumed in the latest Triassic, providing a proximal source.

**Spatsizi Formation Conglomerate (Upper Hazelton Group) Nonconformable on Cake Hill Pluton, Gnat Pass (Locality 5, Fig. 3; Column D, Fig. 4)**

This detrital zircon sample represents the basal Hazelton Group unconformity in the Stikine arch region of northern Stikinia.

Sample 15BVS03–14 came from a conglomerate that contains abundant clasts from the immediately underlying hornblende quartz monzodiorite to quartz monzonite Cake Hill pluton, in a matrix of feldspathic arenite (Fig. 5F; van Straaten and Nelson, 2016). Seven kilometers along strike, 30 m above this basal conglomerate, there is a fossiliferous limestone that includes the early Toarcian ammonite Harpoceras (Henderson and Perry, 1981). The sample yielded a unimodal detrital zircon population with a single peak having a weighted mean age of 215.2 ± 1.5 Ma (MSWD = 0.68, probability of fit = 0.97), based on all 64 grains (Fig. 6). This age overlaps with, but is slightly younger than, the ca. 221–216 Ma U–Pb crystallization ages for the latest Triassic Big Bulk stock near Kitsault (Miller et al., 2020).
Cake Hill and related plutons (Anderson and Bevier, 1992; van Straaten et al., 2012; van Straaten and Bichlmaier, 2018). The absence of younger grains is notable and indicates that no igneous sources younger than 215 Ma existed in this part of the Stikine arch in Toarcian time.

Summary of Detrital Zircon Geochronology

Cumulative results from the 11 Hazelton Group samples in this study show a main peak at ca. 223 Ma (Fig. 8), which reflects the abundance of Stikine suite plutonic clasts in Hazelton Group basal conglomerates, and to a lesser extent recycling of Late Triassic grains into younger clastic units. In total, 13 grains have ages of 261–240 Ma, which overlap those of the primitive arc–back-arc ophiolite in the Cache Creek terrane. Most of the late Paleozoic grains in the composite results are from the Red Chris sample, with a few from other samples. They show derivation from a range of Stikine assemblage sources, probably from nearby block uplifts. Younger peaks on Figure 8 correspond to the youngest Galore belt/oldest Hazelton Group at ca. 210–205 Ma; felsic magmatism in the younger part of the Hazelton Group and associated intrusions (ca. 185 Ma); and Middle Jurassic felsic volcanism of the Iskut River Formation (ca. 173 Ma).

Cathodoluminescent images of zircons from the main age populations show distinct morphologic and growth characteristics (Fig. 9; Nelson et al., 2021), representing the outcome of magmatic histories that correspond to their inferred sources. Grains derived from Stikine suite batholiths (229–215 Ma) are stubby to equant, with subdued oscillatory zoning heavily overprinted by patchy and sector zoning: features typical of zircons in large, long-lived magma chambers (Corfu et al., 2003). In contrast, the bipyramidal prismatic shapes and bimodal, thinly laminated oscillatory zoning in the Texas Creek suite grains resulted from rapid growth in small magma chambers, such as those that incubated the Early Jurassic plutons of the Stewart-Iskut region. Zircons in the felsic pyroclastic Bruce Glacier unit (Iskut River Formation) are simply zoned, probably due to rapid magma ascent and eruption.

DISCUSSION

Local and Stikine Arch Provenance of Hazelton Group Clastic Strata in the Stewart-Iskut Region

Onset of lower Hazelton Group deposition in the Stewart-Iskut region was coeval with erosional incision into Paleozoic and Triassic stratified and plutonic rock units. The ca. 206 Ma age peak in the uppermost Stuhini Group sample at Snippaker Ridge (2014JK72) likely derives from penecontemporaneous volcanic sources in the southern Galore belt (Fig. 3). The unconformably overlying Snippaker unit (Rhaetian) as well as the slightly younger Jack Formation (Rhaetian) show significantly older peaks at ca. 225–220 Ma, which signal an abrupt change from volcanicogenic accumulation to active uplift and erosion (Fig. 6). These basal Hazelton clastic units are characterized by granitoid clast–rich conglomerates with major ca. 229–215 Ma detrital zircon age populations that match Stikine suite plutonic sources. No such plutons are exposed locally. Most likely, the Snippaker unit, Jack Formation, and possibly? unnamed Lower Jurassic strata in the Mess Creek area (Fig. 3; Logan et al., 2000) once formed an ~100-km-long continuous coarse clastic facies, derived primarily from the nearby Hickman Batholith (Fig. 2). Inferred sediment transport was north to south, from the Stikine arch toward the McTagg anticlinorium. Transport of cobble-sized clasts a hundred kilometers from their source could have been aided by high topographic gradients and by river valleys incised along prominent northerly structural lineaments, which were repeatedly active between Late Devonian and Neogene time (Nelson and van Straaten, 2021), and which control the present drainages of the Iskut River and Mess Creek. In contrast, the thin, discontinuous basal Hazelton conglomerate near Iskut village probably had local provenance. Multiple Late Triassic peaks in it derive from immediately underlying Stuhini Group and Stikine suite plutons; the significant late Paleozoic
population may be traced to the Stikine assemblage in a horst block along the Pitman fault system some 20 km to the north. Higher in Hazelton Group sections on Snippaker Ridge and in the McTagg anticlinorium, main peaks are attributable to local Early and Middle Jurassic volcanic and intrusive sources, with lesser Triassic and Paleozoic grains from continued erosional contributions from older units exposed in block uplifts and on intraformational unconformities.

Recycling of locally derived Triassic and Paleozoic grains into the lower and then upper Hazelton Group was shown by a recent study at Oweegee Dome (Fig. 3; George et al., 2021). Here, the lower Hazelton Group unconformably overlies Stikine assemblage and Stuhini Group strata (Greig, 1992). Upper Hazelton Group sandstones and siltstones (Toarcian and younger) unconformably overlie older strata (Greig, 1992; Gagnon et al., 2012, their fig. 13). Both lower and upper Hazelton Group samples contained significant Paleozoic detrital populations, probably ultimately derived from immediately underlying or nearby Stikine assemblage sources (George et al., 2021).

The basal Hazelton conglomerate at Gnat Pass north of the Stikine arch is significant in that, although Toarcian in age, it contains no Early Jurassic grains attributable to lower Hazelton Group volcanic sources or comagmatic intrusions. A sample of tuffaceous sandstone ~520 m higher in the section gave bimodal peaks at ca. 230–220 Ma and 185–170 Ma (Iverson et al., 2012), also lacking a significant 205–195 Ma component attributable to igneous sources of the lower Hazelton Group. This supports the inference that between 205 and 195 Ma, the Stikine arch was amagmatic, uplifted, and undergoing erosion, contributing sediment southward to the lower Hazelton Group in proximal parts of the Iskut region of central Stikinia.

Configuration of the Hazelton Magmatic Arc and its Back-Arc

Of the volcanic belts that make up the lower Hazelton Group, the Telkwa Formation and its related intrusions are most easily characterized...
as a segment of a typical island arc. The overall distribution of the Telkwa Formation defines a 300-km-long curvilinear volcanic arc segment, convex to the south, between Terrace and McConnell Creek (Fig. 3). Thick Telkwa Formation andesite-dacite-rhyolite accumulations are exposed in a continuous east-northeasterly trend across the Skeena arch, with mainly subaerial facies passing eastward into shallow-marine facies near Smithers (Tipper and Richards, 1976; Angen et al., 2017). In far eastern Skeena arch, a submarine facies occupies a narrow (25 km present width) north-northwest-trending belt, the Nilkitkwa trough (Fig. 3; Tipper and Richards, 1976). Telkwa sections in the Nilkitkwa trough comprise 200–1500 m of marine sedimentary strata, polymictic volcanic breccia and tuff, pillow basalt, hyaloclastite, and peperite. The north-northwest-trending Nilkitkwa trough was interpreted as the back-arc rift zone separating the eastern and western Hazleton arcs of Stikinia (Marsden and Thorkelson, 1992). Farther east, shallow-water to subaerial Telkwa volcanic sections are similar to those in the west, but they contain more abundant basalts and tend to slightly more alkaline major-element compositions (Tipper and Richards, 1976). Farthest east, in McConnell Creek area, the Telkwa Formation consists mainly of polymictic conglomerates, some of which contain abundant clasts from the underlying Asitka and Tlaka Groups; this is interpreted to represent the fill of a north-northwest-trending graben that was active during Telkwa sedimentation and volcanism (Tipper and Richards, 1976).

The western Telkwa belt near Terrace is made up of large, superimposed, emergent andesitic stratovolcanoes of latest Triassic (ca. 204–200 Ma) to Early Jurassic (ca. 194 Ma) age, with local felsic centers (Barresi et al., 2015). Measured stratigraphic thicknesses of these volcanic complexes exceed 10 km (Barresi et al., 2015) due to lateral build-out of volcanic deposits. The coeval to younger Kleanza pluton (200–180 Ma; Gareau et al., 1997; Gehrels et al., 2009), a subvolcanic body that fed the volcanic centers, follows a west-northwest trend parallel to the Skeena arch (Fig. 3). Early Jurassic meta-plutonic bodies in the Coast Mountains, which show overlapping calc-alkaline major-element geochemical trends and trace-element patterns with the Kleanza pluton, form a western extension of this trend (Fig. 3; Nelson, 2017). The west-northwesterly plutonic trend coincides with the apparent axis of Telkwa magmatism.

The Toodoggone volcanic field lies along the eastern edge of Stikinia, along strike with the eastern Telkwa Formation near McConnell Creek (Fig. 3). Although coeval with the Telkwa Formation (ca. 201–188 Ma; Diakow et al., 1993, 2005; Diakow and Rhodes, 2006; Duuring et al., 2009), the Toodoggone Formation is distinct in terms of volcanic stratigraphy and structural setting. It is exceptionally felsic compared to other lower Hazleton volcanic belts and comprises two subaerial volcanic sequences with an overall stratigraphic thickness of ~2 km (Diakow and Rhodes, 2006; Duuring et al., 2009). The lower sequence is a succession of dacitic ash-flow tuffs and high-silica andesite flows; the upper sequence comprises dacitic tuffs and basalt to basaltic-andesite flows, rhyolites, and minor volcano-sedimentary strata (Diakow et al., 2005; Diakow and Rhodes, 2006; Duuring et al., 2009). Major-element geochemistry shows that Toodoggone volcanic rocks have high-K calc-alkaline compositions (Diakow et al., 1993). The present outcrop belt is interpreted as modified from an original north-northwest-trending depositional basin controlled in part by penecontemporaneous faults (Diakow et al., 1993; Diakow and Rhodes, 2006). Easterly synvolcanic normal faults define a series of extensional subbasins (Diakow et al., 1993; Diakow and Rhodes, 2006; Duuring et al., 2009). The high percentage of dacitic ash-flow tuffs in the Toodoggone Formation and the presence of hornblende, biotite, quartz, and sanidine phenocrysts are attributed to magma ascent through a crust approaching continental thickness (Diakow et al., 1993). The felsic-dominated Toodoggone Formation contrasts markedly with variable but mainly intermediate volcanic successions of the Telkwa Formation along strike to the south. Such a change, along with the interpreted thicker crust below the Toodoggone area, might correspond to a transition from a more oceanic to more “continental” arc environment, similar to that, for instance, between the southern Kermadec intra-oceanic arc and the Taupo volcanic field of northern New Zealand (Gamble et al., 1990). The east-west normal faults are indicative of belt-parallel extension, perhaps due to southerly (present coordinates) advance of the main Telkwa arc segment.

The lower Hazelton Group in the Spatsizi area, southeast of Iskut village (Fig. 3), comprises two distinct volcanic successions with suprasubduction-zone geochemical signatures (Thorkelson et al., 1995). The older succession, the ca. 206–203 Ma Griffith Creek volcanics, comprises mafic to intermediate flows and intermediate to felsic volcaniclastic units overlying a basal polymictic conglomerate that contains clasts of Permian and Late Triassic limestone (Thorkelson et al., 1995). It is coeval with lower Hazelton Group volcanic and intrusive units near Iskut village (Fig. 4, column 3). The Red Stock, a Tatogga suite pluton that hosts the Red Chris deposit, has high-K calc-alkaline affinity (Riedell et al., 2021). The younger, ca. 195–194 Ma Cold Fish volcanics constitute an ~2-km-thick succession of mafic flows and felsic tufts and sills, including locally thick ignimbrites, which interfere upwards with Lower Pliensbachian marine sedimentary strata (Thorkelson et al., 1995). Although coeval with the Toodoggone Formation, the silica bimodality, tholeiitic to alkaline geochemistry, and partly marine setting of the Cold Fish volcanic rocks suggest that they formed in a subsiding back-arc environment located west of the Toodoggone arc axis (Thorkelson et al., 1995). Marsden and Thorkelson (1982) interpreted this as a northern extension of the Nilkitkwa trough (Fig. 3).

In the Stewart-Iskut volcanic belt, andesites, dacites, and lattes of the Lower Jurassic Betty Creek Formation and comagmatic Texas Creek plutons are coeval with the Cold Fish volcanics and in part with the Telkwa Formation. This belt was proposed as part of the west-facing arc of western Stikinia, a northern continuation of the western Telkwa Formation across a covered interval between the Skeena arch and Kitsault (Fig. 3; Marsden and Thorkelson, 1992). However, although the Betty Creek Formation and Texas Creek suite are of calc-alkaline to tholeiitic and slightly alkaline affinity and show suprasubduction-zone geochemical signatures (Alldrick, 1993; Febbo et al., 2019a), several
key aspects of this belt suggest that it may have a postcollisional origin, rather than synsubduction arc origin. Volcanism, sedimentation, and intrusion are highly structurally controlled by regional northerly and easterly lineaments (Nelson and van Straaten, 2021). Compared to the very large volcanic edifices near Terrace (Barresi et al., 2015), Betty Creek sections are generally <2 km thick. Texas Creek suite plutons are scattered, small in diameter (0.5–20 km), and commonly elongate due to local structural control. They are primarily diorites, quartz monzonites, and monzodiorites with minor monzonites and syenites, and they tend to high-K calc-alkaline compositions (Lewis, 2013; Febbo et al., 2019a). The Stewart-Iskut belt, renowned for its Cu-Au metal endowment, hosts large Cu-Au porphyry deposits such as Red Chris and Kerr-Sulphurets-Mitchell, and epithermal Au deposits such as Brucejack (Fig. 3). The combination of structural control, small magma volumes, K-enriched chemistry, and focused mineralization is recognized as characteristic of postsubduction and syncollisional magmatic-hydrothermal systems worldwide (Richards, 2009). The Stewart-Iskut belt parallels the earlier Galore belt (Figs. 2 and 3) and may represent an eastward migration of post-Stuhini magmatism in structural corridors. Southward projection of the Stewart-Iskut belt from Kitsault intersects the Telkwa belt at nearly a 90° angle (Fig. 3).

In our proposed model, the contiguous Telkwa and Toodoggone belts represent a single, curvilinear arc segment ~400 km long, which we name the Hazelton arc (Fig. 3). The Nikitkwa trough and Stewart-Iskut belt represent zones of mild extension or transtension in the back-arc region, oriented roughly normal to the arc axis defined by the Telkwa belt. The geometry of the Hazelton arc system resembles that of modern syncollisional arcs, which are characterized by short linear extents and high degrees of curvature due to rapid oceanward advance of the central segment compared to the ends (Schellart et al., 2007). Complex kinematics in syncollisional arc–back-arc systems result in peculiar extensional geometries. For instance, spreading ridges in the north Fiji Basin are oriented at 70°–90° to the New Hebrides arc, due to rotational opening around a collisional hinge zone where the West Torres oceanic plateau has entered the western part of the subduction zone (Schellart et al., 2002). In another example, short, highly curvilinear arc segments in the Mediterranean region developed during progressive collision between Africa and Europe. The Calabrian arc and its back-arc region (Balearic and Tyrrenhenian Basins) are segmented into wedges separated by strike-slip faults at high angles to the arc front, which are interpreted as the result of syncollisional extrusion of the arc into a reentrant (Mantovani et al., 2001). The New Britain arc, adjacent to the Finisterre arc-continent collision zone in eastern Papua New Guinea, provides an instructive analogy to the Hazelton (Telkwa-Toodoggone) arc (Fig. 10). Complex rift (Manus Basin) and fault geometries (Bismarck Sea seismic lineation and Weitin fault) in the New Britain back-arc are attributed to rotation of the Bismarck plate around the collision zone (Wallace et al., 2004, 2005; Brandl et al., 2020). Like the Nikitkwa trough, the Weitin fault intersects the frontal New Britain arc at a high angle (Fig. 10).

Figure 10. Current tectonics of the New Britain arc, South and North Bismarck plates, and eastern Papua New Guinea, simplified after Brandl et al. (2020). BSSL—Bismarck Sea seismic lineation and Weitin fault; Papua New Guinea, provides an instructive analogy to the Hazelton (Telkwa-Toodoggone) arc (Fig. 10). Complex rift (Manus Basin) and fault geometries (Bismarck Sea seismic lineation and Weitin fault) in the New Britain back-arc are attributed to rotation of the Bismarck plate around the collision zone (Wallace et al., 2004, 2005; Brandl et al., 2020). Like the Nikitkwa trough, the Weitin fault intersects the frontal New Britain arc at a high angle (Fig. 10).
The Tabar-Lihir-Tanga-Feni island chain in the New Ireland basin (Fig. 10) is a series of shoshonitic magmatic centers linked to recent extension and melting of previously subduction-modified lithosphere in the New Britain back-arc region (Holm et al., 2019). The Tabar-Lihir-Tanga-Feni chain is highly prospective for epithermal gold and Au-enriched porphyry deposits, most notably the giant Ladolam epithermal Au deposit on Lihir Island (Brandl et al., 2020). Magmas and hydrothermal solutions were channeled to surface along translithospheric structures, expressed at the surface as steeply dipping normal faults (Brandl et al., 2020). The back-arc, postsubduction setting of the Tabar-Lihir-Tanga-Feni trend, the high degree of structural control, and the prolific porphyry and epithermal systems are comparable to the Stewart-Iskut belt of western Stikinia. The actively uplifting Finisterre Ranges (Fig. 10), where the active arc passes into the immediate upper plate of the collision, are analogous to the Stikine arch in latest Triassic to Early Jurassic time.

**Linking the Hazelton Arc–Back-Arc, Whitehorse Trough, and Northern Stikinia–Yukon-Tanana Collision**

The Hazelton arc and back-arc system of central Stikinia existed along strike with the nascent collisional orogen of northern and north-northeastern Stikinia and the Whitehorse trough. As shown above, uppermost Triassic (Rhaetian) and Lower Jurassic (Hettangian) basal Hazelton Group conglomerates in the Iskut region contain abundant granitoid clasts and show prominent ca. 225–220 Ma detrital zircon peaks derived from Triassic plutons in the Stikine arch region of northern Stikinia. The Stikine arch was uplifted, probably amagmatic, and undergoing episodic (?) compressional deformation from Rhaetian into Pliensbachian time. The northern terminus of the Toodoggone segment of the Hazelton arc is located approximately at the Pitman fault (Fig. 3). Along strike to the north of the Pitman fault, Lower Jurassic siliciclastic strata of the Whitehorse trough overlap the northeastern margin of Stikinia and the Cache Creek terrane (Figs. 1 and 3).

Detrital zircon signatures from the Hazelton Group in central Stikinia (this study; George et al., 2021) differ in significant ways from those in coeval siliciclastic rocks of the Whitehorse trough. The predominant peak at ca. 223 Ma on the combined population density plot for all samples in this study (Fig. 9) reflects subregional (<100 km) southward transport of granitoid clasts from the Stikine arch into basal Hazelton Group conglomerates. All other peaks can be assigned to local Paleozoic basement uplifts and local episodes of Hazelton Group and Galore belt volcanism and subvolcanic intrusions. Similarly, detrital zircons from Hazelton Group rocks at Oweggee Dome and near Iskut village (locations a and b, Fig. 3) show derivation from identified nearby sources, including underlying Stikine assemblage and Stuhini Group and penecontemporaneous volcanic and/or intrusive sources between 206 and 159 Ma (George et al., 2021). U-Pb age versus εHf, plots of these data show juvenile arrays between primitive mantle values and +5 εHf, (George et al., 2021), indicative of melt sources in the subduction-modified lithosphere and noncontinental lower crust of central Stikinia.

In Yukon, the Lower Jurassic Laberge Group in the Whitehorse trough is divided into a northern, mainly shallow-marine facies and a southern, deeper-marine facies, consistent with southward deepening of the trough (Colpron et al., 2015). Detrital zircon populations from both formations yield main age peaks between 338 and 310 Ma, 210 Ma, 205 Ma, 196–190 Ma, 185 Ma, and 173 Ma, sourced from late Paleozoic and Triassic igneous units in adjacent parts of the Stikinia, Quesnellia, and Yukon-Tanana terranes, and latest Triassic–Early Jurassic synorogenic plutons (Colpron et al., 2015; van Drecht, 2019).

The ca. 223 Ma dominant peak observed in central Stikinia samples (this study) is very subdued in Whitehorse trough samples, and it is only significant in a single sample from near the northern end of the trough (16LV-001C; van Drecht, 2019). In more southerly samples, this peak is negligible (van Drecht, 2019). Its absence suggests that the large Stikine suite plutons of the Stikine arch were not a significant source for the Laberge Group. For latest Triassic–Early Jurassic Laberge populations, U-Pb age versus εHf, plots show increasing excursions into negative εHf, values between ca. 205 and ca. 185 Ma (van Drecht, 2019). These overlap εHf, values of igneous zircons from the syntectonic Minto and Long Lake plutonic suites (Fig. 1; ca. 205–194 Ma and 188–183 Ma, respectively), which intrude across the Stikinia and Yukon-Tanana terranes on the northern and western margins of the Whitehorse trough (Colpron et al., 2015; Sack et al., 2020, their appendix 10; van Drecht, 2019). Trace-element patterns and isotopic characteristics of the Minto and Long Lake suites show decreasing subduction influence and increasing crustal contamination, consistent with syncollisional as opposed to arc-related emplacement (Colpron et al., 2019; Sack et al., 2020). Negative εHf, values in Long Lake suite zircons range as low as −6, attesting to considerable contributions of evolved lower crust to the melts. The Aishihik Batholith, part of the Long Lake suite (Fig. 1), intruded the pericratonic Yukon-Tanana terrane west of the Whitehorse trough during exhumation after maximum tectonic burial ca. 200 Ma (640–650 °C, ~7 kbar), which continued through at least ca. 188 Ma (Clarke, 2017). East-directed sediment transport from proximal sources brought these grains into the trough (van Drecht, 2019).

The differences between Triassic–Jurassic detrital zircons from the Whitehorse trough versus the Hazelton Group of central Stikinia result from their contrasting tectonic/paleogeographic settings. The Whitehorse trough developed as a syncollisional basin, which subsided as the basin shoulders were exhumed, providing an abundant sediment supply from upper plate (mainly adjacent north-northeastern Stikinia and northern Quesnellia) sources, along with syntectonic intrusions (Colpron et al., 2015). Sediment transport was axial (north to south), as well as inward easterly and westerly from the trough shoulders (Colpron et al., 2015; van Drecht, 2019). The lower Hazelton Group formed in central Stikinia distal to the collision zone, in an arc–backarc tectonic environment of mild local extension and transcurrent faulting. Besides locally derived sediments, basal conglomerates were fed from the uplifted Stikine arch. It is likely that a drainage divide along the Stikine arch separated central Stikinia from the Whitehorse trough. Together, these
two depocenters frame northern Stikinia as a deforming, uplifting upper plate that overrode the southwestern Yukon-Tanana terrane.

Revised Modeling for Accretion of the Intermontane Terranes

The data and interpretations presented here, along with other recent observations regarding the Late Triassic–Early Jurassic tectonic evolution of Stikinia (Colpron et al., 2015; Clarke, 2017; van Drecht, 2019; George et al., 2021), provide an increasingly nuanced understanding of its relationships to adjacent Intermontane terranes as the accretion process to the North American margin played out. These results particularly require revision of the oroclinal model of simple rotational collision between the Stikinia and Quesnellia arcs (Mihalynuk et al., 1994; Colpron et al., 2015). Figure 11 depicts inferred tectonic restorations of the Intermontane terranes in four time slices, with configurations from the previous oroclinal model (Mihalynuk et al., 1994; Colpron et al. 2015) on the left (Figs. 11A–11D) and suggested revisions on the right (Figs. 11E–11G). In its original form, the oroclinal model called for long-standing contiguity among Stikinia, the Yukon-Tanana terrane, and Quesnellia (Mihalynuk et al., 1994). As closure proceeded, the Yukon-Tanana terrane was modeled as a hinge between two oppositely rotating, rigid arc segments (Figs. 11A–11D). Although the Stikinia and Pyroxene Mountain plutonic suites (220–211 Ma) provide evidence for Late Triassic contiguity between Stikinia and the Yukon-Tanana terrane in Yukon (Sack et al., 2020), it is likely that the two terranes behaved to some extent as decoupled, independent lithospheric blocks.

In northwestern British Columbia, the boundary between them is the Llewellyn fault zone (Fig. 1), a major, long-lived shear system (Mihalynuk, 1999) that was the locus of sinistral motion in Early Jurassic time (Currie and Parrish, 1993). Pre-Devonian detrital zircon populations from central Stikinia suggest that its basement is like that of southern Wrangellia and unlike the northwestern Laurentian basement of the Yukon-Tanana terrane (George et al., 2021). We thus consider it likely that the Stikinia microplate, prior to collision, was mobile with respect to the Yukon-Tanana terrane. In this view, terrane interactions within the Intermontane belt began with a latest Triassic (212–200 Ma) end-on collision between the northern part of the Stikinia microplate (Stikine arc region) and the Yukon-Tanana terrane (Figs. 11E and 11F), similar to the ongoing impingement of the Aleutian intra-oceanic arc into adjacent pericratonic Kamchatka (Geist and Scholl, 1994; Pedoja et al., 2013). This event shut down Stuhini–Lewes River arc magmatism from the Stikine arc north, initiated deep tectonic burial of the Yukon-Tanana terrane in the lower plate, and caused widespread deformation and uplift of northern Stikinia in the upper plate (Figs. 11F and 11G).

Early Jurassic eclogite clasts (197–181 Ma cooling profiles) occur in Pliensbachian–Toarcian conglomerates of the Whitehorse trough in northwestern British Columbia (“Eclogite ridge” and Atlin Lake, Fig. 3; Kellett et al., 2018). The conglomerates also contain metamorphic and plutonic clasts (Mihalynuk, 1999; Shirvinnamad et al., 2011; Atlin Lake, Lisadale Lake areas, Fig. 3) and yield main detrital zircon ages of ca. 184 Ma (Shirvinnamad et al., 2011), derived from the exhumed Stikinia–Yukon-Tanana tectonic boundary zone and synkinematic plutons along the Llewellyn fault to the west (Fig. 3). The mixture of eclogite with Yukon-Tanana terrane–derived metamorphic clasts could be attributed to rapid exhumation along a tear in the sub-Stuhini slab below the translithospheric Llewellyn fault zone (Fig. 11G).

Subsequent to collision, the Stikinia arc axis reconfigured, moving from the abandoned Stuhini arc of the Stikine arch and Lewes River Group (Fig. 2) to the new, southeast-convex Hazelton arc of central Stikinia, facing southeastward, away from the collision zone to the north (Figs. 3, 11G, and 11H). Asymmetric advance of the Hazelton arc and counterclockwise rotation of the Stikinia microplate around the collision axis were instrumental in closure of central Stikinia against the Quesnellia arc and entrapment of the intervening Cache Creek terrane (Figs. 11G and 11H). Westward advance of the North American plate has also been recognized as an important factor contributing to the accretionary process (Colpron et al., 2015; Monger and Gibson, 2019). By ca. 189–185 Ma, the eastern edge of the Yukon-Tanana terrane and Quesnellia were accreting to the western edge of the continent, as shown by dating of syntectonic intrusions in the Turnagain area (Nixon et al., 2020) and onset of clastic deposition in the autochthonous Fernie Basin of southeastern British Columbia (Fig. 11H; Colpron et al., 2015; Panâ et al., 2019). Growth of the zone of accretion into and along the continent margin was coeval with youngest lower Hazelton Group magmatic activity (ca. 185–178 Ma), and thus was probably instrumental in the demise of the Hazelton (Telkwa-Tooedogone) arc.

Figure 12 presents a detailed view of processes and relationships between Stikinia and neighboring Intermontane terranes during main-stage Hazelton arc development ca. 195–185 Ma, on a restored paleogeographic base. Stikinia is restored 450 km to the south relative to Quesnellia along the Pinchi-Thelth fault system, such that zone boundaries between mixed Amatheit (boreal) and Tethyan Pliensbachian ammonite faunas align (Smith et al., 2001). The northern end of the terrane, including current exposures near Whitehorse and an assumed now-eroded northwestern extension, overrides the western Yukon-Tanana terrane, which in turn is carried above a salient of the North American margin. Tectonic transport during the collision is both northwesterly (orogen-parallel; Dusel-Bacon et al., 2002) and southwesterly (orogen-normal; Clarke, 2017). Early Jurassic garnet growth ages in the Yukon-Tanana terrane (Fig. 12; Clarke, 2017; Dyer, 2020; Gaidies et al., 2021) attest to Barrovian metamorphism during Early Jurassic tectonic burial. Ductile sinistral motion on the Llewellyn shear zone (Currie and Parrish, 1993) shows relative northward motion of the Stikinia block as an indenter into the western Yukon-Tanana terrane. Sediment derived from the leading end of Stikinia is transported southward to cover its northeastern edge and the adjacent Cache Creek terrane.

The Pitman fault divides the actively colliding northern part of Stikinia from a more complex arc and back-arc tectonic regime in central Stikinia. There, asymmetric advance of the Telkwa-Tooedogone (Hazelton) arc segment leads to counterclockwise rotation and accommodation by transpressional and
Figure 11. Tectonic evolution diagrams for the Late Triassic–Early Jurassic development of the northern Canadian Cordillera. Left-hand diagrams A–D are from Colpron et al. (2015); right-hand diagrams E–H are revisions based on this study. See text for details. ST—Stikinia; SM—Slide Mountain; QN—Quesnellia; YT—Yukon-Tanana terrane; CC—Cache Creek terrane; LF—Llewellyn fault.

Colpron et al. (2015)  This study

Late Sinemurian - Pliensbachian ~195-185 Ma
Late Rhaetian - Early Sinemurian ~205-195 Ma
Late Norian - Rhaetian ~212-206 Ma
Carnian-Norian ~230-215 Ma

Exhumation
active magmatism
synkinematic plutons
area of exhumation
synorogenic basin
forebulge
active subduction
thrust belt
transcurrent fault
oceanic plate vector
subduction rollback (arc advance) vector
motion of terranes with respect to North American margin
motion of North American plate
North America

outer margin  interior

0 500 km
extensional structures in the Iskut back-arc region. A sinistral transform fault is hypothesized to extend northward from the subduction zone along the eastern margin of the Stikinia block. Southeastward advance of the Hazelton arc, combined with advance of the Quesnellia arc driven by westward motion of the North American plate “continental bulldozer” (Monger and Gibson, 2019), leads to closure of the remaining ocean between the two landmasses. This reconstruction at 195–185 Ma (Fig. 11H) predicts that the eastern edge of the Cache Creek terrane will enter the active Quesnellia subduction zone, leading to Middle Jurassic southwest-vergent back-thrusting onto adjacent Stikinia, as has been documented (Evenchick et al., 2007).

### CONCLUSIONS

Detrital zircons from uppermost Triassic and Lower Jurassic clastic strata of the Hazelton Group in central Stikinia indicate both local sediment sources and semiregional (<100 km) southward transport of Late Triassic granitoid debris from the Stikine arch. Although coeval with deposition within the Whitehorse trough, they represent a distinct sedimentary system. The Whitehorse trough received voluminous sediment from uplifted Stikinia and Quesnellia adjacent to its apex, located in a collision zone between northern and north-northeastern Stikinia and the subjacent Yukon-Tanana terrane. Thin Hazelton clastic units within the predominantly volcanic succession in the Stewart-Iskut region represent internally derived sedimentation in a block-faulted back-arc region of the Hazelton
The Hazelton (Telkwa-Toodoggone) arc and Whitehorse trough were linked kinematically in an along-strike arc-continent collision zone that resembles modern arc-continent collision zones like New Britain and eastern Papua New Guinea. Counter-clockwise rotation of the Stikinia microplate and southeastward advance of the Hazelton arc contributed to the Middle Jurassic enclosure of the Cache Creek terrane, along with westward advance of the North American plate and entrained Quesnellian arc.

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REFERENCES CITED

Alldrick, J., 1993, Geology and Metallogeny of the Stewart Whitehorse trough were linked kinematically in a North American plate and entrained Quesnellian arc. The Hazelton (Telkwa-Toodoggone) arc and modern arc-continent collision zones like New

Ash, C.H., MacDonald, R.W.J., and Friedman, R., 1967a, Stratigraphy of the Tatogga Lake area, northwestern British Columbia: Implications for the origin and history of the Skeena arch, in Geological Fieldwork 2016: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2017–1, p. 1-35, https://www2.gov.bc.ca/gov/content/industry/mineral-exploration-mining/british-columbia-geological-survey/publications/paper2017

Ash, C.H., MacDonald, R.W.J., and Friedman, R., 1967a, Stratigraphy of the Tatogga Lake area, northwestern British Columbia (104H/12&13, 104G/9&16), in Geological Fieldwork 1996: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 1997–1, p. 283-290, https://www2.gov.bc.ca/gov/content/industry/mineral-exploration-mining/british-columbia-geological-survey/publications/papers/1999-1980/1999-19801997

Ash, C.H., MacDonald, R.W.J., Stinson, R.K., Fraser, T.M., Read, R.P., Pautke, J.F., Nelson, K.J., Arden, K.M., Friedman, R.M., and Lefebvre, D.V.B, 1997b, Geology and mineral occurrences of the Tatogga Lake area, NTS 104G/9NE, 16SE and 104H/12NW, 13SW: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Open-File 1997–3, scale: 1:50,000, http://cmcontent.nrs.gov.bc.ca/geoscience/PublicationCatalogue/OpenFile/BGS_OF1997-03.pdf.

Barresi, T., Nelson, J.L., and Friedman, R., 2015, Evolution of the Hazelton arc near Terrace, British Columbia: Stratigraphic, geochronological and geochemical constraints on a Late Triassic-Early Jurassic arc and Cu-Au porphyry belt: Canadian Journal of Earth Sciences, v. 52, p. 466-494, https://doi.org/10.1139/cjes-2014-0156

Beatty, T.W., Orchard, M.J., and Mustard, P.S., 2006, Geology and tectonic history of the Quesnellia terrane in the area of Kamloops, British Columbia, in Colpron, M., and Nelson, J.L., eds., Paleozoic Evolution and Metallurgy of Pericratonic Terranes in the Ancient Pacific Margin of North America, Canadian and Alaskan Cordillera: Geological Association of Canada Special Paper 45, p. 323-360.

Belasky, P., and Stevens, C.H., 2006, Permian faunas of westernmost North America: Paleobiogeographic constraints on the Permian positions of Cordilleran terranes, in Haggart, J.W., Enkin, R.J., and Monger, J.W.H., eds., Paleogeography of the North American Cordillera: Evidence For and Against Large-Scale Displacement: Geological Association of Canada Special Paper 46, p. 71-80.

Beranek, L.P., and Mortensen, J.K., 2011, The timing and provenance record of the late Permian Klondike orogeny in northwestern Canada and arc-continent collision along western North America: Tectonic and Metallogenic Framework of Cu-Au endowment in the Tabar-to-Feni island arc chain: Ore Geology Reviews, v. 121, p. 103491, https://doi.org/10.1016/j.oregeorev.2020.103491.

Brown, D., and Ryan, P.D., 2011, Preface, in Brown, D., and Ryan, P.D., eds., Arc-Continent Collision: Berlin, Springer-Verlag, Frontiers in Earth Sciences Series, p. vi-vii.

Childe, F.C., 1992, Timing and Tectonic Setting of Volcanogenic Massive Sulfide Deposits in British Columbia: Constraints from U-Pb Geochronology, Radiogenic Isotopes and Geochemistry [Ph.D. thesis]: Vancouver, University of British Columbia, 288 p.

Childe, F.C., Thompson, J.H., Mortensen, J.K., Friedman, R.M., Schirazza, P., Belfonteinane, K., and Marr, J.M., 1998, Primitive Permo-Triassic volcanism in the Canadian Cordillera: Tectonic and metallogenic implications: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 93, p. 224-231, https://doi.org/10.1130/geo2002-0626.

Childe, F.C., Thompson, J.H., Mortensen, J.K., Friedman, R.M., Schirazza, P., Belfonteinane, K., and Marr, J.M., 1998, Primitive Permo-Triassic volcanism in the Canadian Cordillera: Tectonic and metallogenic implications: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 93, p. 224-231, https://doi.org/10.1130/geo2002-0626.

Clarke, A.D., 2017, Tectono-Metamorphic History of Mid-Crustal Rocks at Aishihik Lake, Southwest Yukon (Ph.D. thesis): Vancouver, British Columbia, Canada, Simon Fraser University, 153 p.

Cohen, K.M., Finney, S., Gibbard, P.L., and Fan, J.-X., 2013, The ICS International Chronostratigraphic Chart 2013 (Revision 3): Episodes, v. 36, no. 3, p. 198-204, https://doi.org/10.2113/gsecongeo.36.3.198

Colpron, M., Nelson, J.L., and Murphy, D.C., 2006, A tectonostatigraphic framework for the pericratonic terranes of the northern Canadian Cordillera, in Colpron, M., and Nelson, J.L., eds., Paleozoic Evolution and Metallurgy of Pericratonic Terranes at the Ancient Pacific Margin of North America, Canadian and Alaskan Cordillera: Geological Association of Canada Special Paper 46, p. 1-23.

Colpron, M., Nelson, J.L., and Murphy, D.C., 2007a, Northern Cordilleran terranes and their interactions through time: GSA Today, v. 17, p. 4-10, https://doi.org/10.1130/GSAT01704-5A.1

Colpron, M., Crowley, J.L., Gehrels, G., Long, O.G.F., Murphy, D.C., Beranek, L., and Bickerton, L., 2015, Birth of the northern Cordilleran orogen, as recorded by detrital zircons in Jurassic synorogenic strata and regional exhumation in Yukon: Lithosphere, v. 7 p. S41-562, https://doi.org/10.1130/L461.1
Coulson, I.M., Russell, J.K., and Dipple, G.M., 1999, Origins of the Triassic-Jurassic geodynamic evolution of the Cordilleran orogen, Arendale, northern British Columbia: Geological Association of Canada Special Paper 2005-01, p. 285-298.

Diakow, L.J., Panteleyev, A., and Schroeter, T.G., 1993, Geology of the northern Cache Creek terrane and implications for accretory processes in the Canadian Cordillera: Canadian Journal of Earth Sciences, v. 30, p. 217-233.

Enns, S.G., Thompson, J.H., Stanley, C.R., and Yarrow, E.W., 1995, The Galore Creek porphyry copper-gold deposits, northwestern British Columbia, in Schroeter, T.G., ed., Porphyry Deposits of the Northwestern Cordillera of North America: Canadian Institute of Mining, Metallurgy and Petroleum Special Volume 46, p. 630-643.

Evanchick, C.A., McMechan, M.E., Nicolici, V.J., and Barr, S.D., 2000, U-Pb geochronology and geochemistry of the central and southeastern Cordilleran Cordillera: Exploring links across the orogen, in Sears, J.W., Harms, T.A., and Evanchick, C.J.A., eds., Whence the Mountains?: Inquiries into the Evolution of Organic Systems: A Volume in Honor of Raymond A. Price: Geological Society of America Special Publication 433, p. 117-146, https://doi.org/10.1130/2007.2433(06).

Febbo, G.E., Friedrich, R., Nelson, J.L., Savell, M.J., Campbell, M.E., Creaser, R.A., Friedrich, R.M., van Straaten, B.I., and Stein, H.J., 2019a, The evolution and structural modification of the supergiant Mitchell Au-Cu porphyry, northwestern British Columbia: Canadian Journal of Earth Sciences and the Bulletin of the Society of Economic Geologists, v. 56, p. 303-324, https://doi.org/10.1139/eceg.2019e.4632.

Febbo, G.E., Nelson, R.K., Kennedy, L.A., Nelson, J.L., 2019b, U-Pb geochronology of the Mitchell deposit, northwestern British Columbia: Canadian Journal of Earth Sciences, v. 56, p. 659-672, https://doi.org/10.1139/e2019-039.

Gagnon, J.-F., Barresi, T., Waldon, J.W.F., Nelson, J.L., 1997, Ar/Ar data from metamorphic and plutonic rocks: Implications for the tectonic and tectono-thermal evolution of the eastern Canadian Cordillera: Tectonics, v. 16, p. 1337-1353, https://doi.org/10.1029/97TC00566.

Gareau, S.A., Friedman, R.M., Woodworth, G.J., and Childre, F., 1997, U-Pb ages from the northwestern quadrant of Terrace map area, west-central British Columbia, in Current Research 1996: Geological Survey of Canada Paper 1997-1-A/B, p. 31-40.

Gehrels, G., Rusmore, M., Woodworth, G., Crawford, M., Andronics, C., Hollister, L., Patchett, J., Ducea, M., Butler, R., Klepeis, K., Davidson, C., Friedman, R., Haggart, J., Mahoney, B., Crawford, W., Pearson, D., and Girardi, J., 2009, U-Th-Pb geochronology of the Coast Mountains batholith in north-coastal British Columbia: Constraints on age and tectonic evolution: Geological Society of America Bulletin, v. 121, p. 1341–1361, https://doi.org/10.1130/1993ER041.

Geist, E.L., and Scholl, D.W., 1984, Large-scale deformation related to the collision of the Aleutian arc with Kamchatka: Tectonics, v. 13, p. 538-560, https://doi.org/10.1029/TC010i002p00324.

George, S., Nelson, A.L., Alberts, D., Greig, C., and Gehrels, G.E., 2021, Tectonic origin and Triassic-Jurassic accretionary history of Stikinia from U-Pb geochronology and HF isotope analysis, British Columbia: Tectonics, v. 40, no. 4, p. e2020TC006505, https://doi.org/10.1002/2020TC006505.

Greig, C., 1992, Fieldwork in the Owegee and Snowsdale ranges and Kinsku Lake area, northwestern British Columbia, in Current Research, Part A: Geological Survey of Canada Paper 92-1A, p. 149-156.

Greig, C., 2014, Latest Triassic–earliest Jurassic contractional deformation, uplift and erosion in Stikinia, NW BC: Geological Society of America Abstracts with Programs, v. 46, no. 6, p. 586-587, https://doi.org/10.1130/2014BA0428.

Hart, C.J.R., 1997, A Transact Across Northern Stikinia: Geology of the Northern Whitehorse Map Area, Southern Yukon Territorial (105D13–16): Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 8, 112 p., https://data.geology.gov.yk.ca/Reference/42215/#infoTab.

Henderson, C.M., and Perry, D.G., 1981, A Lower Jurassic hettoroporid bryozoan and associated biota, Turnagain Lake, British Columbia: Canadian Journal of Earth Sciences, v. 18, p. 457-468, https://doi.org/10.1139/e81-040.

Henderson, J.R., Kirkham, R.V., Henderson, M.N., Payne, J.G., Wright, T.O., and Wright, R.L., 1992, Stratigraphy and...
structure of the Sulphurets area, British Columbia, in Cur- rent Research, Part A: Geological Survey of Canada Paper 92-1A, p. 329–332.

Holm, R.J., Tapster, S., Jelsma, H.A., Rosenbaum, G., and Mark, D.F., 2019, Tectonic evolution and copper-gold metallogene- sis of the Papua New Guinea and Solomon Islands region: Ore Geology Reviews, v. 104, p. 209–226, https://doi.org/10.1016/j.oregeorev.2018.11.007.

Iversen, O., Mahoney, J.B., and Logan, J.M., 2012, Dease Lake Geoscience Project, part IV: Tsyabalea Group: Lithological and geochemical characterization (Middle Jurassic to Cretaceous), in the Stikine arc, northwest British Columbia, in Geological Fieldwork 2011: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2012-01, p. 177–202.

In the Stikine arc, northwest British Columbia, in Geological Fieldwork 2011: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2012-01, p. 45–67, https://www2.gov.bc.ca /gov/content/industry/mineral-exploration-mining/british -columbia-geological-survey/publications/papers#2012.

Mihalynuk, M.G., Zagorevski, A., and Logan, J.M., Friedman, R.M., and Johnston, S.T., 2019, Age constraints for rocks hosting massive sulphide mineralization at Rock and Roll and Gran- duc; deposits between Iskut and Stewart, British Columbia, in Geological Fieldwork 2018: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2019-01, p. 97–111, https://www2.gov.bc.ca /gov/content/industry/mineral-exploration-mining/british -columbia-geological-survey/publications/papers#2019.

Miller, E.A., Kennedy, L.A., and van Straaten, B.L., 2020, Geology of the Kinslack Lake area and Big Bulk porphyry prospect, northwest British Columbia: Syndepositional faulting; and basin formation during the Rhaetian (latest Triassic) transition from the Stuhini to the Hazelton Group, in Geo- logical Fieldwork 2019: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2020-01, p. 77–99, https://www2.gov.bc.ca /go/industry/mineral-exploration-mining/british -columbia-geological-survey/publications/papers#2020.

Monger, J.W.H., and Church, B.N., 1973, Revised stratigraphy of the Takla Group, north-central British Columbia: Canadian Journal of Earth Sciences, v. 14, p. 318–326, https://doi.org/10.1139/e73-071.

Monger, J.W.H., and Gibson, H.D., 2019, Mesozoic-Cenozoic deformation in the Canadian Cordiller: The record of a “continental bulldozer”? in Tectonophysics, v. 757, p. 153–169, https://doi.org/10.1016/j.tecto.2018.12.023.

Monger, J.W.H., and Ross, C.A., 1971, Distribution of fusulinaceans in the western Canadian Cordiller: Canadian Journal of Earth Sciences, v. 8, p. 259–278, https://doi.org/10.1139/e71-026.

Nelson, J.L., 2013, Composite pericratonic basin of west-cen- tral Sikhonia and its influence on Jurassic magma conduits: Examples from the Terrace-Ecastll and Anyox areas, in Geological Fieldwork 2016: British Columbia Ministry of Energy, Mines and Petroleum Resources, British Columbia Geological Survey Paper 2017-1, p. 61–82, https://www2.gov.bc.ca /go/industry/mineral-exploration-mining/british -columbia-geological-survey/publications/papers#2017.

Nelson, J.L., and Friedman, R., 2004, Superimposed Quesnel (late Paleozoic–Jurassic) and Yukon-Tanana (Devonian-Mississippian) arc assemblages, Cassiar Mountains, northern British Columbia: Field, U-Pb and igneous petrochemical evidence: Canadian Journal of Earth Sciences, v. 41, p. 1201– 1235, https://doi.org/10.1139/e04-028.
