Mid-Paleozoic to early Mesozoic tectonostratigraphic evolution of Yukon-Tanana and Slide Mountain terranes and affiliated overlap assemblages, Finlayson Lake massive sulphide district, southeastern Yukon

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Abstract

The Finlayson Lake massive sulphide district of southeastern Yukon is underlain by variably deformed, metamorphosed and imbricated mid- to late Paleozoic rocks of Yukon-Tanana and Slide Mountain terranes and affiliated overlap assemblages. Yukon-Tanana terrane comprises three fault-bounded successions of Upper Devonian-Lower Mississippian metavolcanic and metaplutonic rocks that were deposited on a pre-Late Devonian ensialic basement in west- or south-

1Data Repository items Murphy_Appendix1.pdf (Appendix 1) and Murphy_Appendix2.pdf (Appendix 2), are available on the CD-ROM in pocket.
west-facing forearc, arc and back-arc geodynamic settings. Upper Mississippian and Upper Pennsylvanian to Lower
Permian limestones are locally prominent, with a rarely preserved intervening Pennsylvanian succession of chert,
intermediate and mafic volcanic rocks, and coarse-grained clastic rocks. The terrane was imbricated by north- to
northeast-vergent thrust faults in the Early Permian, an event accompanied by widespread flysch sedimentation. Slide
Mountain terrane occurs primarily north of the Jules Creek fault, the main boundary between the two terranes. It
comprises Carboniferous to Lower Permian basin clastic rocks, chert, basalt and limestone that are inferred to have
been deposited on oceanic crust in the extending back-arc region behind the coeval west-facing Yukon-Tanana terrane
arc. Basalt and lesser chert of the Lower Permian Campbell Range formation were deposited on both sides of the Jules
Creek fault, implying juxtaposition of Yukon-Tanana and Slide Mountain terranes along the fault by, or during, the
Early Permian. The narrow distribution of the Campbell Range formation and related plutons around the Jules Creek
fault suggests that it was a 'leaky' transform fault. The Campbell Range formation is coeval with volcanic arc rocks to
the west, indicating that its magmatism and associated strike-slip faulting occurred in a back-arc setting.

Various geological considerations indicate a Middle Permian shift to subduction under the northern or northeastern
margin of the amalgamated terranes. This Middle Permian shift marked the onset of the closure of the Slide Mountain
ocean basin, which likely ended prior to the deposition of Upper Triassic sedimentary rocks on Yukon-Tanana terrane,
the North American margin and the remnants of the late Paleozoic Slide Mountain ocean that once separated them.

Résumé

Le district de sulfures massifs de Finlayson Lake dans le sud-est du Yukon surplombe des roches du Paléozoïque
moyen à supérieur, diversément déformées, métamorphisées et imbriquées, des terranes de Yukon-Tanana et de Slide
Mountain et des assemblages de chevauchement qui y sont affiliés. Le terrane de Yukon-Tanana comprend trois suc-
cessions limitées par failles, de roches métavolcaniques et métaplutoniques du Dévonien supérieur au Mississipien
inférieur et qui ont été mises en place sur un socle ensialique pré-Dévonien supérieur dans un contexte géodynamique
d’avant-arc, d’arc et d’arrière-arc, orienté vers l’ouest ou le sud-ouest. Par endroits, les calcaires du Mississipien
supérieur et du Pennsylvanien supérieur au Permien inférieur prédominent, avec une succession de cherts
Pennsylvanienne rarement préservée, et des horizons de roches de composition intermédiaire à mafique et des roches
clastiques grossières. Le terrane a été imbriqué au début du Permien par des failles de chevauchement de vergence
nord à nord-est, un événement accompagné par une sédimentation généralisée de flysch. Le gros du terrane de Slide
Mountain est situé au nord de la faille de Jules Creek, la principale ligne de partage entre les deux terranes. Il est
constitué de roches clastiques de bassin sédimentaire, de cherts, de basaltes et de calcaires, du Carbonifère au Permien
supérieur, vraisemblablement mises en place sur une croûte océanique dans une région d’arrière-arc en extension,
derrière l’arc du terrane de Yukon-Tanana orienté vers l’ouest. Le basalte et le chert de la formation de Campbell Range
du Permien inférieur ont été déposés sur les deux côtés de la faille de Jules Creek, ce qui implique que les terranes de
Yukon-Tanana et de Slide Mountain étaient juxtaposées de part et d’autre de la faille, au tout début ou durant le Permien
inférieur. L’étroitesse de la distribution des roches de la formation de Campbell Range et des plutons qui y sont
associés, le long de la faille de Jules Creek permet de penser qu’il s’agissait d’une faille de transformation qui « fuyait ».
La formation de Campbell Range est contemporaine des roches d’arc volcanique à l’ouest, ce qui est l’indication que
son magmatisme et ses failles de décrochement se sont produits dans un contexte d’arrière-arc.

Des considérations géologiques diverses indiquent un changement vers un régime de subduction au Permien moyen
sous la marge nord ou nord-est des terranes amalgamés. Ce changement au Permien moyen a marqué le début de la
fermeture du bassin océanique de Slide Mountain, laquelle a vraisemblablement pris fin avant le dépôt des sédiments
du Trias supérieur sur le terrane du Yukon-Tanana, sur la marge nord-américaine et dans les vestiges de l’océan Slide
Mountain du Paléozoïque supérieur, qui les séparait jadis.
INTRODUCTION

The Finlayson Lake massive sulphide district of southeastern Yukon occurs in the central part of the outlier of Yukon-Tanana and Slide Mountain terranes and affiliated overlap assemblages that lies between the Tintina fault, an Eocene dextral strike-slip fault (Roddick, 1967; Murphy and Mortensen, 2003; Gabrielse et al., in press), and the Inconnu thrust fault, the probably Jurassic fault separating the terranes from the North American continental margin sequence (Figs. 1, 2; Murphy et al., 2002). Unlike much of the remainder of Yukon-Tanana and Slide Mountain terranes in the northern Cordillera of Yukon and Alaska, the Finlayson Lake district lies in the region affected by Quaternary glaciations (Duk-Rodkin, 1999), therefore boasting some of the best exposures of these generally poorly exposed terranes and presenting an excellent opportunity to gain insight into their nature and evolution. Indeed, some of the early competing models of the origin and evolution of Yukon-Tanana terrane (cf. Tempelman-Kluit, 1979; Mortensen and Jilson, 1985; Mortensen, 1992a, see also discussion by Colpron et al., this volume), which informed subsequent debates about the terrane (Mortensen, 1992a; Hansen and Dusel-Bacon, 1998; Mihalyik et al., 1999; Hansen and Oliver, 1999), were based largely on interpretations of reconnaissance mapping and geochronological data from this better exposed part of the terrane.

The Finlayson Lake district encompasses the region underlain by the stratigraphic successions that host the five volcanogenic massive sulphide (VMS) deposits discovered, or re-discovered, beginning in 1994 (Figs. 3, 4). The deposits include the new discoveries at Kudz Ze Kayah, GP4F and Ice, as well as definition of reserves at the previously known Fyre Lake and Wolverine Lake occurrences (see Hunt, 2002 for a recent summary). The district is notable for the age range and diversity of its deposits. Fyre Lake is Late Devonian in age and associated with chloritic phylite and greenstone of boninitic composition; Kudz Ze Kayah, GP4F and Wolverine Lake are Late Devonian and Early Mississippian in age and associated with alkalic (A-type) felsic metavolcanic host rocks; and Ice occurs in Lower Permian basalt of mid-ocean ridge basalt (MORB) composition (Piercey, 2001; Piercey et al., 1999, 2001a, b, 2002a, b). The deposits occur in both Yukon-Tanana and Slide Mountain terranes, with Fyre Lake, Kudz Ze Kayah, GP4F and Wolverine Lake occurring in the former and Ice occurring in the latter.

At the time of the discoveries, the area had been mapped twice at 1:250,000-scale (Wheeler et al., 1960; Tempelman-Kluit, 1977) and has been the focus of a doctoral thesis that included 1:125,000-scale regional mapping and U-Pb geochronology (Mortensen, 1983). These data formed the bases for two different models of the tectonic evolution of the region. Tempelman-Kluit (1979) proposed that the rocks in the district represented the highly deformed - to the extent that original stratigraphy was inferred to be unrecognizable - and imbricated remnants of an early Mesozoic ensialic arc, subduction complex and marginal ocean basin that lay outboard of the North American continental margin in the late Paleozoic and early Mesozoic; and accreted to the margin during and after the early Mesozoic closure of the marginal basin by subduction beneath the arc. While acknowledging the deformed and imbricated character

Figure 1. Late Paleozoic to early Mesozoic tectonic assemblages of the northwestern North American Cordillera, (A) currently and (B) before displacement on the Tintina fault. The Finlayson Lake district and other mineral districts are indicated with black stars. Occurrences of high-pressure/low-temperature metamorphic rocks are indicated by letters: E, Permian eclogite; e, Mississippian eclogite; b, blueschist, age unknown. Communities, indicated by black dots: Wh, Whitehorse, Yukon; WL, Watson Lake, Yukon; D, Dawson, Yukon; Fb, Fairbanks, Alaska.
of the rocks of the region, Mortensen and Jilson (1985) and Mortensen (1992a) proposed that much of the region was stratigraphically intact and comprised primarily volcanic and intrusive rocks of a southwest-facing mid-Paleozoic ensialic arc, its pre-Late Devonian sialic basement, Pennsylvanian to Permian limestone and quartzite and mid-Permian volcanic and plutonic arc rocks produced by subduction under the northeastern margin of the terrane. These rocks were deformed prior to the Late Triassic and imbricated with mid-Paleozoic ophiolite (Slide Mountain terrane) after the Late Triassic, in part along a transpressive suture called the Finlayson Lake fault zone. The resultant structural stack was subsequently thrust onto the North American continental margin before the mid-Cretaceous.

The geoscience data and interpretations that were available at the time of the discoveries of the VMS deposits were insufficient to account for the diversity of the deposits in the area or to relate them in such a way as to be useful for further mineral exploration. To address this situation, a new generation of 1:50,000-scale geological mapping and integrated geochemical, metallogenic, geochronological and paleontological research was initiated in 1996. This paper summarizes what has been learned about the geological setting and evolution of the Finlayson Lake district and environs in this latest generation of research. It describes a new provisional lithostratigraphy of both Yukon-Tanana and Slide Mountain terranes and affiliated overlap assemblages and infers environments of deposition. It proposes a model of the tectonic evolution of the region that has elements in common with both previous models but differs from them in several important ways. In particular, this new work documents a metallogenically important mid- to late Paleozoic back-arc realm that includes parts of both Yukon-Tanana and Slide Mountain terranes. It also documents relationships between the two terranes that challenge how terranes are defined.

**STRUCTURAL AND STRATIGRAPHIC FRAMEWORK OF THE FINLAYSON LAKE DISTRICT**

The Finlayson Lake district comprises variably deformed and metamorphosed lower greenschist to amphibolite facies metasedimentary and metavolcanic rocks and affiliated metaplutonic suites. The district...
district is characterized by a central core of higher grade metamorphic rocks (to lower amphibolite facies) surrounded by lower grade, low greenschist facies rocks. This relatively simple metamorphic distribution is a consequence of Cretaceous dynothermal events (Murphy, 2004), not described here, that overprint and partly obscure a complex Paleozoic and early Mesozoic history, which is the topic of this paper.

In spite of locally intense deformation, regionally extensive stratigraphic units have been defined by mapping, and protoliths have been determined from locally well-preserved primary features and geochemical characteristics. The stratigraphic nature of many contacts has been inferred from a lack of convincing evidence for faulting, such as enhanced brittle or ductile deformation or contradictionary (older over younger) geochronological or biochronological data. Stratigraphic facing directions have been determined from a combination of preserved primary indicators, absolute ages determined by U-Pb geochronology and biochronology and relative ages determined from cross-cutting relationships.

Rocks assigned to Yukon-Tanana terrane lie between the Tintina fault and the Jules Creek fault, the main boundary between Yukon-Tanana and Slide Mountain terranes (Fig. 2; Murphy et al., 2002). Yukon-Tanana terrane has been subdivided into a number of provisionally named fault- and unconformity-bound groups and formations (Figs. 3, 4, Murphy et al., 2001; Murphy et al., 2002 [Note that a forthcoming bulletin by the senior author will establish formal status for these units. All formation and group names in this paper are therefore provisional.]). The structurally deepest rocks which are clearly part of Yukon-Tanana terrane are pre-Late Devonian to Early Permian rocks of the Big Campbell thrust sheet, bound below by the post-Late Triassic Big Campbell thrust fault and above by the Early Permian Money Creek thrust fault. The Money Creek thrust sheet also comprises pre-Late Devonian to Early Permian rocks, but of a different depositional environment and geochemical character than the Big Campbell sheet. The Money Creek thrust sheet is bound above by the Cleaver Lake thrust fault, along which a third assemblage of Late Devonian and Early Mississippian rocks as well as retrogressed eclogite (Erdmer et al., 1998; Devine et al., this volume), were emplaced in the Early Permian.

Slide Mountain terrane comprises Mississippian (and possibly older) to Lower Permian carbonaceous metaclastic rocks and chert, Lower Permian basalt and chert, Early Permian mafic and ultramafic metaplutonic rocks, and Middle Permian limestone and quartzite. Rocks assigned to Slide Mountain terrane occur primarily between the Jules Creek fault and the Inconnu thrust fault, the fault boundary between the North American continental margin sequence and the allochthonous terranes (Fig. 2; Murphy et al., 2002). Lower and Middle Permian rocks of Slide Mountain terrane also depositionally overlie Yukon-Tanana terrane south of the Jules Creek fault, providing a firm linkage between the two terranes by the Early Permian. Clasts of Yukon-Tanana terrane occur in upper Middle Permian (-Triassic?) conglomerate deposited on Slide Mountain terrane, further affirming the linkage between the two terranes by that time. Finally, as observed in the Big Campbell window (Figs. 2, 3), rocks of Slide Mountain terrane occur in the footwall of the Big Campbell thrust fault, structurally beneath rocks of Yukon-Tanana terrane in the Big Campbell thrust sheet. In the window, meta-basalt and serpentinitized ultramafic rock of Slide Mountain terrane are overlain (depositionally?) by, and imbricated with, Upper Triassic shale and siltstone.

Slide Mountain and Yukon-Tanana terranes and overlapping rocks are juxtaposed against Triassic shale and siltstone and older rocks of the North American continental margin sequence along the Inconnu thrust fault (Figs. 2, 3). The presence of Triassic rocks beneath both the Big Campbell and Inconnu thrust faults suggests that the rocks of Slide Mountain terrane in the Big Campbell window occur in a duplex structure related to the Inconnu thrust.

The following summary starts with descriptions and paleogeographic interpretations of first Yukon-Tanana terrane, then Slide Mountain terrane and finally the affiliated overlap assemblages. It proceeds in order of age and encompasses both the rock units and structures within each interval. The discussion concludes with a summary of the geodynamic evolution of the allochthonous rocks of the Finlayson Lake district.

**YUKON-TANANA TERRANE**

**Local Basement Composition**

The mid- to late Paleozoic supracrustal rocks of Yukon-Tanana terrane in the Finlayson Lake district were deposited on and intruded into a pre-Late Devonian crustal basement, which is not exposed in the area but which can be characterized as generally sialic by isotopic and geochemical data from younger felsic igneous rocks (Mortensen, 1992b; Grant, 1997; Piercey et al., 2003). With local exceptions, εNd from felsic igneous rocks is typically highly negative and the rocks have Early to Middle Proterozoic TDM ages (Mortensen, 1992a; Grant et al., 1996; Piercey et al., 2003). Similarly, the Sr and Pb isotopic composition of felsic igneous rocks and the Pb isotopic compositions of sulphide minerals from syngeneic mineral occurrences are highly radiogenic (Mortensen, 1992a; Grant, 1997; Mortensen et al., this volume). Inherited Pb in zircons from numerous samples of felsic metavolcanic and metaplutonic rocks indicates the presence of a basement source area with an Early to Middle Proterozoic average age (Mortensen, 1992a; Mortensen et al., this volume). All of these data suggest that felsic igneous rocks in the district were either derived from older sialic crustal material or were contaminated by older sialic crustal material during passage of the melt through the crust.

In contrast to the highly evolved nature of the felsic igneous rocks in the district, the Nd isotopic signatures of mafic metavolcanic rocks are more variable, ranging from ~5 to nearly +9 (Fig. 5), suggesting the presence of small domains of more primitive character (Piercey et al., 2004). The rocks with near chondritic to slightly evolved signatures generally occur in areas where the felsic rocks are highly evolved, suggesting that the mafic rocks were contaminated during their passage through the sialic crust. The rocks with the most primitive signatures cluster in an east-west belt across the central part of the district, defining a region either underlain by more primitive lithosphere or in which magma moved through sialic crust.
Figure 3. Geological map of the Finlayson Lake district and environs. Fault abbreviations as in Figure 2.
with only limited contamination. In this region, the Fire Lake formation is made up of primitive-arc boninites, the products of melting of highly refractory mantle source regions that have been variably metamorphized by subducted slab fluids (Piercey et al., 2001a). Boninites are typically found in intra-oceanic suprasubduction zone settings (Piercey et al., 2001a and references therein), thus their presence in the Finlayson Lake district likely defines that part of Yukon-Tanana terrane that was underlain by basement of more primitive composition.

Pre-Upper Devonian North River Formation
The oldest exposed rock unit in the Finlayson Lake district, occurring in both the Big Campbell and Money Creek thrust sheets, is the pre-Upper Devonian North River formation (Figs. 3, 4). This unit consists primarily of quartzose psammite, non-carbonaceous metapelite and locally important marble, calc-schist and felsic metavolcanic members at or near its top. The age of the North River formation is constrained only to be pre-Late Devonian, older than the stratigraphically overlying Late Devonian Fire Lake formation and cross-cutting Early Mississippian intrusions.

Late Devonian and Early Mississippian Magmatism and Sedimentation
The Late Devonian to Early Mississippian period in the Finlayson Lake district is represented by three fault-bounded assemblages that formed in different depositional environments. From the structurally deepest levels in the core of the district outward these include: (1) the Grass Lakes and Wolverine Lake groups and affiliated metaplutonic rocks in the Big Campbell thrust sheet, (2) the Waters Creek and Tuchitua River formations and affiliated intrusions in the Money Creek thrust sheet, and (3) the Cleaver Lake formation and intrusions of the Cleaver Lake thrust sheet (Figs. 3, 4). Coarse-grained metamorphic rocks with Early Mississippian cooling ages and characteristics suggestive of a high pressure/low temperature metamorphic regime (Erdmer et al., 1998; Klatsa metamorphic complex of Devine et al., this volume) also occur in the Cleaver Lake thrust sheet. Structural considerations suggest that the rocks in the Money Creek thrust sheet formed at least 45 km southwest of their current location with respect to the Grass Lakes and Wolverine Lake groups (Murphy and Piercey, 2000b). The rocks of the Cleaver Lake thrust sheet in turn lay an unknown distance to the south-southwest from the rocks of the Money Creek thrust sheet. The Late Devonian and Early Mississippian period was metallogenically important as all the VMS deposits and occurrences in this part of Yukon-Tanana terrane formed during this interval (Figs. 3, 4).

Big Campbell Thrust Sheet
The Grass Lakes group (Murphy et al., 2002) comprises three map units, the Fire Lake, Kudz Ze Kayah and Wind Lake formations. The oldest unit, which directly overlies the North River formation, is the laterally extensive Upper Devonian Fire Lake formation, host of the Fyre Lake Cu-Co-Au VMS deposit (Murphy, 1998a; Foreman, 1998; Hunt, 2002). The Fire Lake formation comprises chloritic phyllite or schist (mafic volcanic and volcanioclastic rocks) and lesser carbonaceous phyllite or schist and muscovite-quartz phyllite or schist (felsic volcanic or volcanioclastic rocks). The basal contact of the Fire Lake formation is generally sharp although local, possibly structural, intercalation of mafic schist with quartz-rich schist suggests a narrowly transitional contact. Bodies of mafic and variably serpentinitized ultramafic metaplutonic rocks occur within the Fire Lake formation and in underlying rocks of the North River formation along a greater than 115 km-long, west-northwest-trending corridor (Fig. 3). These bodies occur as layer-subparallel, laterally tapering slabs within the Fire Lake formation and as discordant, dike-like bodies in the underlying North River formation. Although initially inferred to be allochthonous slices of ophiolite (Tempelman-Kluit, 1979; Mortensen and Jilson, 1985; Mortensen, 1992a), these bodies have been re-interpreted as intrusions - sills within the Fire Lake formation and dikes in the North River formation - based on their spatial association with the Fire Lake formation, their geometry and the lack of evidence for displacement along their generally sharp contacts (Murphy, 1998b).

Geochemically, the Fire Lake formation, the initial pulse of magmatism in the area, has attributes that generally indicate a rift setting but with a tendency to become more arc-like in a southerly direction. In the northern part of the Finlayson Lake district, chloritic phyllite of the Fire Lake formation is of basaltic composition with back-arc basin basalt (BABB), enriched mid-oceanic ridge basalt (E-MORB), niobium-enriched basalt (NEB), and Th-enriched NEB varieties, all typical of non-arc, rift-type settings (Fig. 6; Piercey, 2001; Piercey et al., 2004). In contrast, in the south, the majority of samples of the Fire Lake formation are boninite and island arc tholeiite (IAT) in composition. This southwestward transition coincides with a lateral transition within the rocks of the Fire Lake formation from a slightly negative εNd (-0.3) value to slightly to moderately positive εNd values (+0.5 to +3.5; Fig. 5) and by dramatic increases in thickness of the Fire Lake formation and in the volume of the intercalated mafic and ultramafic metaplutonic intrusions (Murphy, 1998b). The coincident increase in both volcanic thickness and amount of comagmatic intrusion has been attributed to down-to-the south synvolcanic normal faulting, a conclusion also supported by the presence of the Fyre Lake VMS deposit located along this trend (Murphy, 1998b; Murphy and Piercey, 2000a). These coincident transitions are attributed to rapid extension of crust with domains of more primitive and more evolved crust material and a southward change into subduction-fluid-fluxed mantle (Piercey et al., 2001a).

The age of the Fire Lake formation is constrained to be Late Devonian by two ca. 365 Ma U-Pb dates from the North Lakes metadiorite, one of the spatially associated intrusions inferred to be comagmatic with the Fire Lake formation (Table 1, Fig. 7; Table A1, Appendix 1 [see footnote 1]). Younger, stratigraphically overlying rocks are intruded by bodies of the Grass Lakes plutonic suite as old as 362.2 ± 3.3 Ma (Table 1, Fig. 8, Appendix 1).

The Fire Lake formation passes upward into a carbonaceous phyllite-dominated succession, with the felsic metavolcanic rocks that host the Kudz Ze Kayah and GP4F deposits in the lower part (Kudz Ze Kayah formation), and mafic metavolcanic rocks and quartzite in the upper part (Wind Lake formation; Murphy, 1998a).
Figure 4. Summary diagram illustrating the structural and stratigraphic relationships in the Finlayson Lake district. Numbers and numbered symbols refer to points tabulated in Tables 1-8. KMC - Klatsa metamorphic complex; WF - Whitefish limestone; WL - White Lake formation; KA - King Arctic formation; FC - Finlayson Creek limestone; NR - North River formation. Time scale of Okulitch (2002).
Figure 5. Neodymium isotopic data from mafic magmatic rocks constraining the nature of the lithosphere underlying the mid-Paleozoic rocks of Yukon-Tanana terrane in the Finlayson Lake district (from Piercey et al., 2004; S.J. Piercey and R.A. Creaser, unpublished data).

Figure 6. Compositional variability of the Fire Lake formation in the Finlayson Lake district (from Piercey et al., 2004). Dashed line de-limits the boundary between rocks with 'arc' geochemical signatures and those with 'non-arc' signatures.
Metavolcanic rocks in the Kudz Ze Kayah and Wind Lake formations have within-plate geochemical signatures (Piercey et al., 2001b, 2002a). Felsic metavolcanic rocks of the Kudz Ze Kayah formation plot in the fields for within-plate felsic rocks on tectonic discrimination diagrams, with predominantly A-type affinities (Piercey et al., 2001b). Mafic metavolcanic rocks in the Wind Lake formation are weakly alkalic with signatures similar to ocean island basalt (OIB, Piercey et al., 2002a). These signatures are attributed to ongoing rifting, following its onset in Fire Lake formation magmatism. The evolution from felsic to mafic magmatism in the upper part of the Grass Lake group is attributed to the onset of decompression melting of the mantle following thinning, rifting and partial melting of the overlying continental crust (Piercey et al., 2002a). According to Piercey et al. (2003), a continental crustal signature is evident in the felsic rocks of the Kudz Ze Kayah formation ($\varepsilon$Nd$_t$ = -7.8) as are geochemical indications of minor crustal contamination in the overlying mafic rocks ($\varepsilon$Nd$_t$ = -2.8).

The ages of the Kudz Ze Kayah and Wind Lake formations are constrained to be younger than the ca. 365 Ma age of the Fire Lake formation and older than the 362.2 ± 3.3 Ma age of an intrusion of the Grass Lakes plutonic suite into the Wind Lake formation (Table 1, Fig. 8A, Table A1, Appendix 1).

The Grass Lakes group is intruded by texturally heterogeneous Late Devonian to Early Mississippian peraluminous granitic metaplutonic rocks of the Grass Lake plutonic suite. The most common rock type of the Grass Lakes plutonic suite, making up the batholith-sized plutons in the area, is a foliated, locally megacrystic, potassium feldspar augen granite. Foliated, medium-grained and equigranular

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Table 1. U-Pb Zircon Geochronological Constraints, Big Campbell Thrust Sheet.

<table>
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<tr>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma) ± 2$\sigma$</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
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<tr>
<td>1 HD-35</td>
<td>North Lakes metadiorite</td>
<td>365.0 ± 1.2</td>
<td>Foliated hornblende-biotite; intruded by augen-bearing phase of Grass Lakes plutonic suite</td>
<td>Mortensen (1992); this paper, Fig. 7A</td>
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<td>96DM-065</td>
<td>North Lakes metadiorite</td>
<td>366.3 ± 10.2</td>
<td>Foliated hornblende-biotite; intruded by augen-bearing phase of Grass Lakes plutonic suite</td>
<td>this paper, Fig. 7B</td>
</tr>
<tr>
<td>2 98DM-088</td>
<td>unnamed</td>
<td>362.2 ± 3.3</td>
<td>Foliated equigranular quartz monzonite; intrudes upper unit of Grass Lakes group</td>
<td>Mortensen (1992); this paper, Fig. 8A</td>
</tr>
<tr>
<td>3 GG-19</td>
<td>unnamed</td>
<td>359.9 ± 0.9</td>
<td>Foliated locally potassium feldspar megacrystic granitic to quartz monzonitic metamorphic rock; intrudes Grass Lakes group</td>
<td>Mortensen (1992); this paper, Fig. 8B</td>
</tr>
<tr>
<td>4 96DM-119</td>
<td>unnamed</td>
<td>357.3 ± 2.8</td>
<td>Slightly foliated medium-grained equigranular granite; intrudes Grass Lakes group</td>
<td>this paper, Fig. 8C</td>
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**GRASS LAKES PLUTONIC SUITE**

<table>
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<th>Sample</th>
<th>Unit</th>
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<tr>
<td>5 01DM-318</td>
<td>lower part</td>
<td>ca. 357.5</td>
<td>Youngest detrital zircons from quartzofeldspathic conglomerate unconformably overlying the Kudz Ze Kayah formation</td>
<td>this paper, Fig. 9</td>
</tr>
<tr>
<td>6 FV-21</td>
<td>lower part</td>
<td>356.2 ± 0.9</td>
<td>Quartz-feldspar-muscovite schist intercalated with carbonaceous phyllite and quartzofeldspathic sandstone and conglomerate</td>
<td>Mortensen (1992); this paper, Fig. 10A</td>
</tr>
<tr>
<td>7 P98-069A</td>
<td>Fisher porphyry-middle part</td>
<td>346.6 ± 2.2</td>
<td>Foliated quartz-feldspar porphyry intruding or extruding near top of middle part of Wolverine Lake group</td>
<td>Piercey (2001); this paper, Fig. 10B</td>
</tr>
<tr>
<td>8 GP4F-1</td>
<td>GP4F (porphyry)</td>
<td>346.9 ± 0.7</td>
<td>Foliated quartz-feldspar porphyry intruding Kudz Ze Kayah formation of the Grass Lakes group</td>
<td>this paper, Fig. 10C</td>
</tr>
<tr>
<td>9 98DM-GDR</td>
<td>unnamed</td>
<td>347.3 ± 3.7</td>
<td>Hornblende-biotite granite — relationship with nearby coeval rocks of the lower Wolverine Lake group unclear</td>
<td>this paper, Fig. 10D</td>
</tr>
</tbody>
</table>

---

Figure 7. U-Pb concordia diagrams for two samples of the North Lakes meta-diortite, inferred to be comagmatic with the Fire Lake formation. See Appendix 1 for data tables and geochronological interpretations.
The Fire Lake and Kudz Ze Kayah formations are also intruded by an undated pluton of the metaluminous Simpson Range plutonic suite near the Tintina fault (Fig. 3). The Simpson Range plutonic suite comprises hornblende-biotite granodiorite, diorite and quartz monzonite and typically occurs in the Money Creek and Cleaver Lake thrust sheets.

The geochemical signatures, spatial distribution and U-Pb age range of intrusive rocks that cut the Grass Lakes group are similar to those of the metavolcanic rocks within the group (Piercey et al., 2003). All samples of the Grass Lakes plutonic suite are geochemically and isotopically nearly identical to felsic metavolcanic rocks of the Kudz Ze Kayah formation. According to Mortensen (1992a) and Piercey et al. (2003), the plutons plot primarily as A-type within-plate granites and are isotopically evolved, with strongly negative εNd values. The single intrusion of the Simpson Range plutonic suite affinity has not been evaluated geochemically; but if it has the volcanic arc geochemical signature typical of that suite (see below), then the intrusions in the area would show the same southward transition from within-plate geochemistry to more arc-like geochemistry as is exhibited by the Fire Lake formation.

The age of the Grass Lakes plutonic suite is constrained by three U-Pb dates (Table 1, Fig. 8, Table A1, Appendix 1). A sample of foliated equigranular granite that cuts the Wind Lake formation yielded a $362.2 \pm 3.3$ Ma date (Fig. 8A, Appendix 1). A sample of foliated K feldspar-megacrystic granite yielded a concordant date of $359.9 \pm 0.9$ Ma (Fig. 8B, Table A1, Appendix 1). A weakly foliated intrusion that cuts recumbent isoclinal folds yielded a date of $357.3 \pm 2.8$ Ma (Fig. 8C, Table A1, Appendix 1), constraining the age of the first regional deformation in the area.

The Wolverine Lake group overlies different units of the Grass Lakes group in different areas, implying an angular unconformity (Figs. 3, 4). The basal unit of the succession comprises quartz- and feldspar-pebble conglomerate, grit and sandstone and carbonaceous phyllite. It grades upward into a unit of carbonaceous phyllite, quartz- and feldspar-phyric felsic metavolcanic rocks, and rare chert and locally silicified aphyric rhyolite breccia, regionally extensive silicic barite-magnetite exhalite, and at the top, massive mafic metavolcanic rocks (Bradshaw et al., 2001, 2003). Volumetrically minor subvolcanic feldspar (+/- quartz) porphyry bodies cut the middle unit of the succession and underlying rocks (Murphy and Piercey, 1998, 1999a; Bradshaw et al., 2003).

The metavolcanic rocks of the Wolverine Lake group have geochemical and isotopic signatures indicating a second cycle of within-
plate magmatism, which culminated in full sea floor spreading (Piercey, 2001; Piercey et al., 2001b, 2002a, b, 2003). Felsic volcanic and subvolcanic intrusive rocks from the lower part of the group are remarkably similar to the Kudz Ze Kayah formation with primarily within-plate, A-type signatures and evolved isotopic signatures indicative of partial melting of continental crust (Piercey et al., 2001b, 2003). Felsic volcanic rocks above the Wolverine deposit, near the top of the unit, differ in that they are subalkalic and plot in the volcanic arc field on discrimination diagrams (Piercey et al., 2001b). They pass transitionally upwards into basalt of primarily normal mid-ocean ridge basalt (N-MORB) affinity. Piercey et al. (2001b) attribute the progression from felsic A-type to felsic arc compositions to either a reduction in the temperature of melting of the continental crust or to mixing of partial melts of continental crust with high-field-strength-element-depleted mafic material; eventually pure mafic melts were erupted.

U-Pb geochronological data from igneous rocks and detrital zircons show that deposition of the Wolverine Lake group started at ca. 357 Ma and persisted to at least 347 Ma (Table 1; Figs. 9, 10; Appendices 1, 2). Detrital zircons from the basal conglomerate are as young as ca. 358 Ma (Fig. 9, Table A3, Appendix 2 [see footnote 1]). A 356.2 ± 0.9 Ma date was obtained from felsic metavolcanic rocks intercalated with the basal conglomerate (Fig. 10A, Appendix 1). Felsic metavolcanic rocks and subvolcanic intrusions in the footwall of the Wolverine deposit and intruding older stratigraphy elsewhere in the area are ca. 347 Ma (Figs. 10B, C, D; Appendix 1).

Money Creek Thrust Sheet
Stratigraphically overlying the North River formation in the Money Creek thrust sheet are felsic and minor mafic or intermediate metavolcanic rocks, carbonaceous phyllite, bedded barite, quartzite, quartz-pebble conglomerate and chert of the Waters Creek formation (Fig. 4). Felsic metavolcanic rocks of the Waters Creek formation are hornblende-phyric, and locally quartz and feldspar porphyritic.

The Waters Creek formation is intruded by sills or transposed dikes of locally megacrystic metaporphry, which are common around a large unnamed foliated pluton of the older phase of the Simpson Range plutonic suite. Composed of variably megacrystic hornblende-biotite granodiorite, the pluton and porphyry sheets resemble the hornblende-phyric rocks of the Waters Creek formation, suggesting that the pluton may be a shallow subvolcanic feeder to the volcanic rocks. The Waters Creek formation is also intruded by the Tuchitua River pluton, a weakly foliated to unfoliated body of massive, equigranular, hornblende-biotite granodiorite, which is part of a younger phase of the Simpson Range plutonic suite.

Geochemical and isotopic data from volcanic rocks in the Waters Creek formation are limited, and intrusions of the older phase of the Simpson Range plutonic suite have not been evaluated geochemically. One sample of mafic metavolcanic rock from the Waters Creek formation is Nb-enriched basalt of non-arc affinity (Piercey et al., 2001a). Grant (1997) reported highly evolved εNd, signatures...
(\(\varepsilon\)Nd, >-13.1) from siliceous carbonaceous phyllite at three locations in the Waters Creek formation.

The age of the Waters Creek formation is constrained to be Late Devonian by U-Pb dates on felsic metavolcanic rocks and cross-cutting intrusions (Table 2). A 360.8 ± 0.8 Ma zircon date has been obtained from felsic metavolcanic rocks in the formation (Fig. 11, Appendix 1). The Tuchitua River pluton, which intrudes the youngest chert member of the formation, is 356.9 ± 0.6 Ma (Fig. 12B, Appendix 1).

The Waters Creek formation is stratigraphically overlain by the Tuchitua River formation, which comprises slightly foliated, primarily intermediate and lesser felsic volcanic, volcaniclastic and epiclastic rocks. Relatively weakly deformed rocks of the Tuchitua River formation overlie different, more highly deformed members of the Waters Creek formation in different places, indicating that the base of the Tuchitua River formation is an angular unconformity.

The Tuchitua River formation and underlying rocks are intruded by weakly to unfoliated hornblende-biotite granodiorite and granite of a younger phase of Simpson Range plutonic suite (Mortensen, 1992a; Grant et al., 1996; Grant, 1997). The less highly strained nature of both the volcanic and plutonic rocks suggest that they both postdated a phase of regional deformation. As well, their compositional similarity suggests that the plutons may be the subvolcanic feeders to the volcanic rocks (Mortensen, 1992a).

Geochemical and isotopic data from the Tuchitua River formation are limited. Metavolcanic rocks have typical volcanic arc geochemical signatures (Grant et al., 1996; Grant, 1997; Piercey et al., this volume). One sample reported by Grant (1997) yielded a moderately negative \(\varepsilon\)Nd (\(-11.4\)), suggesting that the volcanic arc was ensialic.

Similarly, granitoids of the Simpson Range plutonic suite have geochemical signatures typical of a volcanic arc setting (Grant et al.,

Figure 10. U-Pb concordia diagrams for two samples of felsic metavolcanic rocks in the Wolverine Lake group and two coeval sub-volcanic intrusions. (A) sample FV-21, felsic metavolcanic rock from near base of the succession; (B) sample P98-098B (Fisher porphyry), felsic volcanic or sub-volcanic porphyry found near the middle of Wolverine Lake group; (C) GP4F-1, porphyritic felsic intrusion into the Kudz Ze Kayah formation; (D) 98DM-GDR, hornblende granite intruding the North River formation. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
Table 2. U-Pb Geochronological Constraints, Money Creek Thrust Sheet.

<table>
<thead>
<tr>
<th>#</th>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>2σ</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>FV-27</td>
<td>middle part</td>
<td>360.8</td>
<td>0.8</td>
<td>Muscovite-quartz schist</td>
<td>Mortensen (1992); this paper; Fig. 11</td>
</tr>
<tr>
<td>11</td>
<td>SL-45</td>
<td>Hasselberg Lake pluton</td>
<td>357.9</td>
<td>1.3</td>
<td>Hornblende-biotite granodiorite, may be comagmatic with part of the Tuchitua River formation</td>
<td>Mortensen (1992); this paper; Fig. 12A</td>
</tr>
<tr>
<td>12</td>
<td>TRP-22</td>
<td>Tuchitua River pluton</td>
<td>356.9</td>
<td>0.6</td>
<td>Hornblende-biotite granodiorite, intrudes Waters Creek formation, may be comagmatic with Part of the Tuchitua River formation</td>
<td>Mortensen (1992); this paper; Fig. 12B</td>
</tr>
</tbody>
</table>

TUCHITUA RIVER FORMATION

<table>
<thead>
<tr>
<th>#</th>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>2σ</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
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<tbody>
<tr>
<td>13</td>
<td>01DM-238</td>
<td></td>
<td>344.5</td>
<td>2.5</td>
<td>Muscovite-quartz phylite just below Whitefish limestone with Serpukhovian conodonts</td>
<td>this paper; Fig. 13A</td>
</tr>
<tr>
<td>14</td>
<td>PR-51</td>
<td></td>
<td>352.5</td>
<td>2.6</td>
<td>Rhyodacitic tuff breccia in upper part of formation</td>
<td>Mortensen (1992); this paper; Fig. 13B</td>
</tr>
<tr>
<td>15</td>
<td>03DM-022</td>
<td></td>
<td>346.8</td>
<td>2.0</td>
<td>Massive quartz- and feldspar-porphyritic flow(?)</td>
<td>this paper; Fig. 13C</td>
</tr>
<tr>
<td>16</td>
<td>03DM-023</td>
<td></td>
<td>347.7</td>
<td>1.4</td>
<td>Hornblende-phryic quartz-feldspar porphyry dike similar to volcanic rocks which it cuts</td>
<td>this paper; Fig. 13D</td>
</tr>
</tbody>
</table>

*Number is keyed to data points presented in Figure 4.

All studied intrusions in the Money Creek thrust sheet are hornblende-bearing, calc-alkaline, and I-type and plot in the volcanic arc field on tectonic discrimination diagram. Isotopic data from the suite yield uniformly negative εNd	extsubscript{i} values, affirming the ensialic nature of the arc basement (Grant, 1997; Piercey et al., 2003).

The age of the Tuchitua River formation is constrained by U-Pb dates on felsic volcanic rocks and possibly comagmatic rocks of the Simpson Range plutonic suite (Table 2). Muscovite-quartz phylite in the upper part of the Tuchitua River formation of inferred felsic volcanic or volcaniclastic protolith yielded a 344.5 ± 2.5 Ma zircon date (Fig. 13A, Appendix 1), Rhyodacitic tuff breccia yielded a 352.5 ± 2.6 Ma zircon date (Fig. 13B, Appendix 1). Coherent, conformable date (Fig. 13A, Appendix 1). Rhyodacitic tuff breccia yielded a 352.5 ± 2.6 Ma zircon date (Fig. 13B, Appendix 1). Coherent, conformable date (Fig. 13B, Appendix 1).

Figure 11. U-Pb concordia diagram for a sample of felsic metavolcanic rock from the Waters Creek formation. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
in the volcanic arc granite field of tectonic discrimination diagrams (Grant, 1997; Grant et al., 1996; Piercey et al., 2001b).

Neodymium isotopic data for rocks of the Cleaver Lake thrust sheet suggest derivation from an evolved sialic source or contamination during passage through an evolved sialic crust (Grant et al., 1996; Grant, 1997; Piercey et al., 2003). Samples of the Simpson Range plutonic suite are highly negative, whereas the older volcanic rocks range from moderately negative to slightly positive. These data suggest that the volcanic arc was broadly ensialic (Piercey et al., 2003).

The age of the Cleaver Lake formation is constrained to be ca. 360 Ma (and likely older) to 356 Ma (Table 3). Quartz porphyry intrusions that are inferred to be comagmatic with rhyolite flows in the middle of the Cleaver Lake formation have yielded a 360.5 ± 1.9 Ma U-Pb date (Mortensen, 1992b). Gabбро, inferred to be comagmatic with Cleaver Lake formation basalt, has yielded a 356.1 ± 0.9 Ma U-Pb date (Fig. 14, Appendix 1). Cross-cutting Simpson Range quartz monzonite intruding the Cleaver Lake formation has yielded two dates, 348.4 ± 0.8 Ma and 354.9 ± 1.8 Ma (Figs. 12C, D; Appendix 1).

The klippe of Klatsa metamorphic complex in the southern Campbell Range lies in a similar structural position with respect to footwall stratigraphy as the Money klippe, and is therefore inferred to be a part of the Cleaver Lake thrust sheet (Devine et al., 2004, this volume; Devine, 2005). Besides the mafic and ultramafic metamorphic rocks, interfoliated white mica-rich quartz schist and garnetiferous metabasite are the predominant rock types, representing clastic and basaltic protoliths, respectively. Although retrograde muscovite and chlorite are pervasive, these rocks contain relict textures and prograde minerals that indicate high pressure/low temperature metamorphic conditions (Erdmer, 1987; Erdmer et al., 1998; Devine et al., this volume; Devine, 2005). Geochemically, garnetiferous metabasite in the Simpson Range is of MORB affinity (Creaser et al., 1999).

The age of the protolith of the interfoliated schists is constrained to be Ordovician to Devonian, based on detrital zircon cores as young

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**Figure 12.** U-Pb concordia diagrams for four samples from plutons of the Simpson Range plutonic suite. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
Finlayson Lake Massive Sulphide District

as ca. 473 Ma and metamorphic zircons as old as ca. 353 Ma (Devine et al., this volume; Table 3). Igneous zircons from leucogabbro of the Klatsa metamorphic complex are ca. 368 Ma (Table 3). The Klatsa metamorphic complex was metamorphosed to eclogite facies metamorphic conditions in a subduction zone at ca. 353, and was subsequently rapidly uplifted through the \(^{40}Ar/^{39}Ar\) closure temperature for white mica by ca. 353 Ma (Table 3).

Environments of Deposition and Inferred Geodynamic Settings
The lithological, stratigraphic and geochemical characteristics of Devonian and Early Mississippian rocks in the Big Campbell thrust sheet suggest magmatism and sedimentation in an initially rapidly subsiding, restricted, rift basal environment developed on primarily sialic crust (Piercey, 2001; Piercey et al., 2001a, 2002a, 2003). Background sedimentation in the basin in which volcanism occurred

Figure 13. U-Pb concordia diagrams for samples of felsic metavolcanic rocks from the Tuchitua River formation. (A) quartz muscovite schist near top of the formation; (B) dacitic tuff breccia near base of the formation; (C) massive conformable quartz-feldspar porphyry inferred to be a flow; and (D) discordant quartz-feldspar porphyry dike. See Appendix 1 for analytical methods, data tables and geochronological interpretations.

Figure 14. U-Pb concordia diagram for gabbro inferred to be comagmatic with basalt of the Cleaver Lake formation. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
was carbonaceous, implying an anoxic water column (Bradshaw, 2003; Bradshaw et al., 2003). Most of the VMS deposits in the Finlayson Lake district occur at stratigraphic or geochemical breaks or near the top of major volcanic cycles, locally associated with exhalative rocks and in transons to carbonaceous basal sedimentation. The onset of carbonaceous background sedimentation above the non-carbonaceous North River formation was sharp, implying a rapid change. This change also coincides with the onset of within-plate magmatism throughout most of the area, implying that magmatism and subsidence were linked, a characteristic of rift basins (e.g., Storey et al., 1992). As geochemical trends in the volcanic rocks of the Fire Lake formation and lithological trends in cross-cutting plutonic rocks indicate a transition to more arc-like rocks in the south, the environment of deposition of the rocks in the Big Campbell thrust sheet is most likely that of an ensialic back-arc basin (Piercey et al., 2000a, b, 2001a, 2002a, b, this volume).

The lithological, stratigraphic and geochemical character of the Waters Creek formation in the Money Creek thrust sheet resembles that of the rocks of the Grass Lakes group, implying a similar restricted basin rift environment developed on ensialic crust. Like the Grass Lakes group, the volcanism occurred in a setting characterized by carbonaceous background sedimentation. Also, like the upward transition between the North River formation and the Grass Lakes group in the Big Campbell thrust sheet, the upward transition from the North River formation to the Waters Creek formation in the Money Creek thrust sheet is sharp and marks the rapid onset of volcanism and deposition of carbonaceous sedimentary rocks.

The main difference between the two successions lies in the hornblende-bearing nature of the volcanic rocks of the Waters Creek formation and affiliated plutonic rocks which may imply a position within the basin more proximal to a volcanic arc. Existing geochemical data on the Waters Creek formation are too sparse to further evaluate this possibility at present.

In contrast to the subsiding rift basin characteristics of the older Waters Creek formation, Grass Lakes group and the Wolverine Lake group, the lithological, stratigraphic and geochemical character of the Tuchitua River formation suggests that it was part of a volcanic and volcanioclastic arc built in an oxygenated marine setting on sialic crust. Much of the formation is composed of pale to medium green intermediate volcanic and volcanioclastic rocks. Pink flow banded rhyolite and hydacidic tuff breccias may indicate locally subaerial eruption (Mortensen, 1983, 1992a; Mortensen and Jilson, 1985). Rocks of basinal character are conspicuously lacking in the Tuchitua River formation.

Like the Tuchitua River formation of the Money Creek thrust sheet, the lithological, stratigraphic and geochemical character of the Cleaver Lake formation suggests that it was a volcanic and volcanioclastic arc buildup in an oxygenated shallow marine to locally subaerial setting. Rocks of restricted basinal character are conspicuously lacking in the Cleaver Lake formation. The Nd isotopic data suggest that the crustal basement to the formation was sialic; however samples in the northern part of the Money klippe are less evolved, possibly indicating either a more primitive crustal domain or lack of interaction with sialic crust as the magma passed through it.

Data are insufficient to infer a depositional environment of the protoliths of the Klatsa metamorphic complex. However, their high-pressure/low temperature metamorphic mineral assemblage implies that part of their geological history was spent at great depth, most likely in a subduction zone (Erdmer et al., 1998; Devine et al., this volume).

### Correlations and Late Devonian-Early Mississippian Paleogeography

Although the various stratigraphic successions and plutonic suites defining Yukon-Tanana terrane in the Finlayson Lake district are fault-bounded, a Late Devonian and Early Mississippian paleogeographic model can be inferred by considering the depositional environments of coeval stratigraphic units, the spatial distribution of these units and the amount and direction of displacement on the faults that separate them. U-Pb geochronological data show that the Waters Creek and Tuchitua River formations in the Money Creek thrust sheet are respectively coeval with the Grass Lakes and Wolverine Lake groups in the footwall of the Money Creek thrust. Thus, the unconformity which separates the Grass Lakes and

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### Table 3. U-Pb Zircon Geochronological Constraints, Cleaver Lake Thrust Sheet.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>QP-30</td>
<td>unnamed</td>
<td>360.5 ± 1.9</td>
<td>Quartz porphyritic intrusion inferred to be comagmatic with rhyolite flows in Cleaver Lake formation</td>
<td>Mortensen (1992a)</td>
</tr>
<tr>
<td>ARG-52</td>
<td>unnamed</td>
<td>356.1 ± 0.9</td>
<td>Gabbro inferred to be comagmatic with basalt in Cleaver Lake formation</td>
<td>Mortensen (1992); this paper; Fig. 14</td>
</tr>
<tr>
<td>03FD231-1</td>
<td>Klatsa Metamorphic Complex</td>
<td>368.0 ± 10.0</td>
<td>Leucogabbro associated with serpentinite host of eclogitic rocks of Klatsa metamorphic complex</td>
<td>Devine et al. (this volume)</td>
</tr>
</tbody>
</table>

**SIMPSON RANGE PLUTONIC SUITE**

<table>
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<tr>
<th>Sample</th>
<th>Younger Phase</th>
<th>Age (Ma)</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
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<tr>
<td>SA-02</td>
<td>younger phase</td>
<td>348.4 ± 0.8</td>
<td>Hornblende-hornblende granodiorite</td>
<td>Mortensen (1992); this paper; Fig. 12C</td>
</tr>
<tr>
<td>3694-14</td>
<td>younger phase</td>
<td>345.9 ± 1.2</td>
<td>Hornblende-hornblende granodiorite; same body as SA-02</td>
<td>Grant (1997)</td>
</tr>
<tr>
<td>SA-28</td>
<td>younger phase</td>
<td>354.9 ± 1.8</td>
<td>Hornblende-hornblende granodiorite; same body as SA-02</td>
<td>Mortensen (1992); this paper; Fig. 12D</td>
</tr>
</tbody>
</table>

**ECLOGITE FACIES METAMORPHIC ROCKS**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>03FD039-2</td>
<td>Klatsa metamorphic complex</td>
<td>ca. 473</td>
<td>Age of youngest detrital cores in layered schist</td>
<td>Devine et al. (this volume)</td>
</tr>
<tr>
<td>03FD056-2</td>
<td>Klatsa metamorphic complex</td>
<td>353.1 ± 2.4</td>
<td>Felsic schist and metabasite; metamorphic zircons indicate age of high pressure metamorphism</td>
<td>Devine et al. (this volume)</td>
</tr>
<tr>
<td>03FD056-2</td>
<td>Klatsa metamorphic complex</td>
<td>353.0 ± 3.7</td>
<td>Felsic schist and metabasite; metamorphic zircons indicate age of high pressure metamorphism</td>
<td>Devine et al. (this volume)</td>
</tr>
</tbody>
</table>

*Number is keyed to data points presented in Figure 4.*
Late Mississippian to Early Permian Sedimentation and Magmatism

The Late Mississippian to early Permian interval is represented by four sporadically occurring stratigraphic units in the Money Creek thrust sheet: the Whitefish limestone; chert, tuff, limestone and lesser clastic rocks of the White Lake formation; clastic rocks of the King Arctic formation; and a recently recognized upper limestone, the Finlayson Creek limestone. These four units rarely occur together in a continuous complete section owing to regionally significant Pennsylvanian and Early Permian erosional unconformities. The oldest stratigraphic unit of this interval, the Whitefish limestone, is of Late Mississippian age, indicating a hiatus in deposition in the region following Early Mississippian magmatism and sedimentation. Rocks of Yukon-Tanana terrane of mid-Mississippian, essentially early Viséan, age are absent in the Finlayson Creek district and environs.

The Whitefish limestone comprises up to 200 m of recrystallized, massive to poorly bedded, locally crinoidal grey limestone. Isolated crinoid fragments are ubiquitous; the lack of crinoids in growth position suggests that the unit is bioclastic. Lenses and laterally continuous layers of whitish-grey silica are common. Four conodont collections constrain the age of the Whitefish limestone to be Upper Mississippian (late Viséan to Serpukhovian; Table 4). The basal contact of the Whitefish limestone is likely a disconformity or unconformity, based on its Late Mississippian age compared to the

**Table 4. Conodont Age Constraints on the Whitefish Limestone, White Lake Formation and the Finlayson Creek Limestone.**

<table>
<thead>
<tr>
<th>Sample#</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1 96HJ-Stop 3</td>
<td>Early Carboniferous, Serpukhovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F2 00DM-219</td>
<td>Late Early Carboniferous, late Viséan -Serpukhovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F3 00DM-324</td>
<td>Early Carboniferous, Serpukhovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F4 00DM-45</td>
<td>Early Carboniferous, late Viséan - Serpukhovian</td>
<td>Orchard (this volume)</td>
</tr>
</tbody>
</table>

**FINLAYSON CREEK LIMESTONE**

<table>
<thead>
<tr>
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<th>Age</th>
<th>Reference</th>
</tr>
</thead>
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<tr>
<td>F5 00DM-90</td>
<td>Late Carboniferous – Early Permian, Gzhelian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F6 00DM-131</td>
<td>Late Carboniferous to Early Permian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F7 00DM-131</td>
<td>Late Carboniferous, Bashkirian - Moscovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F8 00DM-329</td>
<td>Late Carboniferous, Bashkirian - Moscovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F9 00DM-9306</td>
<td>Late Carboniferous, Bashkirian - Moscovian</td>
<td>Orchard (this volume)</td>
</tr>
<tr>
<td>F10 00DM-107</td>
<td>Late Carboniferous, Bashkirian - Moscovian</td>
<td>Orchard (this volume)</td>
</tr>
</tbody>
</table>

*Number is keyed to data points presented in Figure 4.
Early Mississippian (ca. 344 Ma) youngest age of immediately underly-
ing felsic metavolcanic rocks.

Where it is not unconformably overlain by younger rocks, the
Whitefish limestone is transitionally overlain by chert, tuff, limestone
and fine clastic rocks of the White Lake formation. Pale green and
locally pink, tan and white chert and tuff are the predominant and
distinguishing rock types of the White Lake formation; crinoidal
limestone beds are common in the lower part. Where the unit is most
completely exposed, pale green siltstone and fine sandstone is com-
mon in the upper part (Devine, 2005). The lower part of the White
Lake formation is Late Viséan to Serpukhovian in age, based on
conodonts from a limestone bed near the base of the unit (Table 4).

Where preserved beneath younger unconformities, the White
Lake formation is unconformably overlain by the King Arctic forma-
tion, a succession of conglomerate, sandstone and argillite (Devine,
2005). The base of the King Arctic formation is marked by conglomer-
ate made up of rounded clasts of chert and limestone derived from
the underlying White Lake formation. Conglomerate passes upward
into about 150 m of mottled green lithic arenite and wacke. The upper
part of the formation comprises dark grey, fine grained wacke and
argillite. The age of the King Arctic formation is indirectly con-
strained to be Early Pennsylvanian, based on the age of the underlying
White Lake formation and the age of the Finlayson Creek limestone,
which is inferred to unconformably overlie the King Arctic formation
(Devine, 2005).

A sporadically preserved succession of gritty and chloritic
volcanolithic metaclastic rocks, intermediate to mafic metavolcanic
rocks, chert and minor limestone and felsic metavolcanic rock locally
lies along strike from exposures of the White Lake and King Arctic
formations, below the Finlayson Creek limestone and above a lime-
stone that is correlated with the Whitefish limestone, although no
fossils have been extracted from it. Based on this stratigraphic posi-
tion, the clastic succession is tentatively assigned a Pennsylvania
age and correlated with the White Lake and King Arctic formations.
If this is correct, then the sedimentation represented by these two
formations took place in a peri-volcanic setting. The Whitefish
limestone, the White Lake formation and correlative metavolcanic
and volcanoclastic rocks, and the Tuchitua River formation, are all
locally overlain by the Finlayson Creek limestone, the youngest rock
unit in the Late Mississippian to Early Pennsylvian interval. It is massive,
pale grey and locally crinoidal – virtually identical to the Whitefish
limestone, with which it has been confused. The base of the limestone
is locally marked by conglomerate, reflecting the unconformable
nature of its basal contact. The Finlayson Creek limestone has yielded
conodonts that range in age from Bashkirian to early Asselian
(Table 4).

**Environments of Deposition and Geodynamic Settings**
The thick-beded, bioclastic nature of both the Whitefish and Finlayson Creek limestones gives few clues to their depositional
settings. The lack of volcanic rocks in both units suggests hiatuses
in arc or back-arc volcanism in this area in the Late Mississippian
and Late Pennsylvanian. Volcanic rocks were deposited throughout
the Pennsylvanian (White Lake and King Arctic formations and
correlative volcanic rocks) although the nature and geochemical af-
finitiy of the volcanism has not been determined. Late Mississippian
to Early Permian arc volcanism has been documented in the southern
extent of Yukon-Tanana terrane (Klinkit Group; Simard et al., 2003;
and the upper division of the Lay Range assemblage; Ferri, 1997)
implying that the Finlayson Lake district may have been in a back-arc
setting during that time.

**Early Permian Synorogenic Clastic Sedimentation
and Thrust Faulting**
Geological relationships in the Finlayson Lake district suggest that
Yukon-Tanana terrane during the Early Permian was characterized
by regional shortening, uplift, erosion and synorogenic clastic sedi-
mentation. Mississippian and Pennsylvanian rocks were folded and
eroded before the unconformable deposition of carbonaceous clastic
rocks and chert of the Lower Permian Money Creek formation. The
latter formation is itself folded and overthrust by the Cleaver Lake
and Money Creek thrusts. Lower Permian rocks of the Campbell
Range formation of Slide Mountain terrane unconformably overlie
both the hanging wall and footwall of the Money Creek thrust, im-
plying that movement on the thrust occurred during the Early
Permian.

**Money Creek Formation**
The Money Creek formation (Figs. 3, 4) comprises carbonaceous
phylitic; grey, tan, pink and green chert; diamicite; and quartzofeld-
spatholithic sandstone, grit and locally conglomerate. It differs from
other older carbonaceous clastic units in the area by the immature
nature of its detritus and the occurrence of chert (Murphy, 2001).
Angular feldspar clasts are a significant component of the Money
Creek formation, and lithic clasts comprise felsic and mafic volcanic
rocks, quartzite and intraformational chert and carbonaceous phyl-
lite. Limestone-pebble conglomerate occurs where the unit overlies
the Whitefish or Finlayson Creek limestones.

The Money Creek formation occurs in both the Big Campbell
and Money Creek thrust sheets, unconformably overlying different
rock units in different places. The formation stratigraphically overlies
the Wolverine Lake group in the Big Campbell thrust sheet and, in
different areas of the Money Creek thrust sheet, it overlies different
parts of the Tuchitua River formation, as well as the Whitefish
limestone, White Lake formation, King Arctic formation and
Finlayson Creek limestone (Fig. 3; Devine et al., 2004). The Money
Creek formation is infolded with the Whitefish limestone where it
immediately underlies the Cleaver Lake thrust. Detailed mapping
reveals that the basal contact of the Money Creek formation is folded
by longer wavelength and lower amplitude structures than the un-
derlying units, suggesting that some folding of the underlying units
had already occurred before its deposition.

The age of the Money Creek formation is constrained to be
Lower Permian by the youngest conodont collection from the under-
lying Finlayson Creek limestone (Gzhelian-Asselian; Table 4)
and the Asselian to Sakmarian age of the overlying basalt of the
Campbell Range formation (see below).
The Money Creek formation is herein interpreted as a synorogenic flysch. Deposition of the formation followed the onset of the folding of underlying rocks, and the formation is itself folded and overthrust by the Money Creek and Cleaver Lake thrust sheets. The quartzofeldspathic and volcanolithic nature of the conglomerate implies uplift and erosion of a volcanic and possibly plutonic source area. A logical source for this detritus is the volcanic arc rocks in the hinterland of the thrust belt.

**Cleaver Lake and Money Creek Thrusts**

The Cleaver Lake and Money Creek thrusts are recognized on the basis of older over younger relationships and strain zones (Murphy and Piercey, 2000a, b; Murphy, 2004). Movement on the structurally higher Cleaver Lake thrust placed the Upper Devonian Cleaver Lake formation and Mississippian intrusions above the Lower Permian Money Creek formation. Movement on the structurally deeper Money Creek thrust juxtaposed pre-Late Devonian North River formation and younger rocks above rocks as young as the Lower Permian Money Creek formation. These relationships have locally been modified by Cretaceous normal faulting. For example, the Money and North klippen of Tempelman-Kluit (1979) of the Cleaver Lake thrust sheet rest directly on rocks of the Big Campbell thrust sheet, due to displacement on the Cretaceous North River fault (Figs. 2, 3; Murphy, 2004).

In terms of fault kinematics and amounts of displacement, the Money Creek thrust cuts upsection to the north and east, as demonstrated by the occurrence of progressively younger rocks in both the immediate footwall and the immediate hanging wall in those directions (Fig. 3). This geometry implies east to northeast vergence, an interpretation supported by sparse kinematic data from the footwall strain zone (elongation lineations and various kinematic indicators such as shear bands and asymmetrical boudinage; D.C. Murphy, unpublished data, 2003). The amount of overlap of the Money Creek thrust sheet onto the footwall rocks is about 45 km, implying at least that much displacement. Hence, rocks in the Money Creek thrust sheet originally lay at least 45 km west-southwest of their current location with respect to the footwall. Similarly, the Cleaver Lake thrust overlaps a relatively thin section of Money Creek formation beneath the Money klippe, and a much thicker section of Money Creek formation under the klippe of Klatsa metamorphic complex in the east, implying it cuts upsection from approximately west to east. Sparse slickenside data from the Cleaver Lake thrust plane (Erdmer, 1985; D.C. Murphy, unpublished data, 1997, 1999) indicate northeast-directed displacement. Hence, the Cleaver Lake formation originally lay southwest of the rocks in the Money Creek thrust sheet, which in turn lay west-southwest of the rocks below the Money Creek thrust.

The age of thrusting is constrained to be Early Permian by the age of the youngest rocks offset by thrusting (Lower Permian Money Creek formation) and oldest rocks unaffected by thrusting (Lower Permian Campbell Range formation and related mafic and ultramafic intrusions).

**SLIDE MOUNTAIN TERRANE**

Rocks in the Finlayson Lake district that have been assigned to Slide Mountain terrane include the Mississippian (and older?) to Lower Permian Fortin Creek group, the Lower Permian Campbell Range formation and comagmatic to slightly younger mafic and ultramafic rocks, and the Lower to Middle Permian Gatehouse formation. While the Fortin Creek group occurs exclusively north of the northern fault boundary of Yukon-Tanana terrane, the Jules Creek fault, the Campbell Range formation and mafic and ultramafic plutonic rock occur both north and south of the Jules Creek fault, defining a narrow corridor around it. South of the Jules Creek fault, the Campbell Range and Gatehouse formations are inferred to unconformably overlie Yukon-Tanana terrane, defining the earliest stratigraphic overlap between Yukon-Tanana and Slide Mountain terranes and requiring their juxtaposition along the Jules Creek fault in the Early Permian. The weakly deformed Campbell Range formation overlies different and more highly deformed older rocks on both sides of the Jules Creek fault, implying an Early Permian unconformity and Early Permian and/or older deformation in rocks on both sides of the fault.

**Local Basement Composition**

The basement to the Fortin Creek group is not exposed in the area. However, samples of basalt of the overlying Campbell Range formation that erupted through the crust in that area are uniformly primitive (εNd, ranging from +6.4 to +8.9; S.J. Piercey and R.A. Creaser, unpublished data), implying the absence of, or lack of interaction with, sialic crust. The Fortin Creek group comprises primarily dark argillite and chert of basinal character, suggesting that it was deposited on thin crust of oceanic or transitional character.

**Mississippian to Early Permian Sedimentation and Magmatism**

In the Slide Mountain terrane of the Finlayson Lake district, the oldest, Mississippian (and older?) to Early Permian unit is the Fortin Creek group. This succession consists primarily of variably deformed dark grey, siliceous carbonaceous phyllite, and grey, green and white chert. Greyish white chert-pebble conglomerate, grey quartzite, shale-chip-bearing quartzofeldspathic grit and conglomerate, and felsic and mafic metavolcanic rock are also locally important constituents. The upper part of the group is characterized by tan, pink and green chert and argillite. Rare beds of limestone and dolomite occur throughout the succession (Murphy et al., 2002).

Preliminary geochemical data (S.J. Piercey, unpublished data) on the metavolcanic rocks of the Fortin Creek group provide evidence for a rift-type setting similar to the basinal Devonian and Mississippian rocks in the Big Campbell and Money Creek thrust sheets. A single sample of greenstone has a geochemical signature indistinguishable from that of a modern N-MORB. Six samples of felsic metavolcanic rocks at structurally high levels are weakly alkalic with trachytic compositions.

The Fortin Creek group is inferred to be Carboniferous to Early Permian in age on the basis of both local and regional constraints. Locally, variably deformed rocks of the Fortin Creek group are
overlain by weakly deformed basalt of the Lower Permian Campbell Range formation (see below) and by unfoliated dolomite, possibly correlating with the Lower to Middle Permian Gatehouse formation (see below), from which a single conodont collection of Early Permian age was obtained (Table 5). Regionally, the group is lithologically similar to and is correlated with the Mt. Aho and Rose Mountain formations, which underlie basalt of the Campbell Range formation along strike to the northwest near Faro (Pigage, 2001, 2004). Conodonts of Early Carboniferous age have been extracted from the lower part of the Mt. Aho Formation; variegated chert of the Rose Mountain Formation ranges in age from Early Carboniferous (Viséan) to Early Permian (Pigage, 2001, 2004; Table 5).

**Environments of Deposition and Inferred Geodynamic Setting**

The volcanic-poor, carbonaceous shale- and chert-dominated nature of the Fortin Creek group suggests an anoxic, deep water marine environment. The primitive nature of the Nd isotopic signature of overlying basalt suggests that the group may have been deposited on oceanic crust.

Although separated from Yukon-Tanana terrane by the Jules Creek fault, the Fortin Creek group has geological features that are compatible with having been linked geodynamically with Yukon-Tanana terrane during its evolution. With its alkaline felsic metavolcanic and carbonaceous clastic rocks, the older part of the Fortin Creek group resembles the coeval Wolverine Lake group, which records Early Mississippian rifting of Yukon-Tanana terrane ensialic crust. The occurrence of quartzofeldspathic detritus in the Fortin Creek group implies proximity to a granitic source terrane. Although the age and nature of this source terrane is not known, the Devonian and Mississippian felsic magmatic rocks of Yukon-Tanana terrane are likely candidates. The Fortin Creek group is interpreted as the part of the Slide Mountain ocean that lay adjacent to the inner rift-edge of the Yukon-Tanana terrane crustal block – on the opposite side of the ocean from the North American rifted margin. The Fortin Creek group may be a record of a ‘rift-drift’ transition, the transition to sea-floor spreading and sedimentation on new oceanic crust, in the back-arc region behind the Yukon-Tanana terrane arc.

**Early Permian Transform-Related Rifting**

The latter Early to Middle Permian interval in the Finlayson Lake district is represented by basalt and chert of the Campbell Range formation, spatially associated mafic and ultramafic plutonic rocks, and limestone and quartzite of the Gatehouse formation. These, and correlative rocks near Faro (Tempelman-Kluit, 1972; Pigage, 2001, 2004), define a 10-12 km-wide arcuate corridor that straddles the Jules Creek and equivalent Vangorda fault for over 300 km. Within this corridor, the Campbell Range formation, and locally the Gatehouse formation, overlie many of the older rock units in the Finlayson Lake district, including the Fortin Creek group north of the Jules Creek fault, and the Money Creek and Tuchitua River formations and Grass Lakes group of Yukon-Tanana terrane south of the Jules Creek fault.

Originally considered to be allochthonous with respect to underlying rocks (Tempelman-Kluit, 1979; Mortensen and Jilson, 1985; Mortensen, 1992b; Plint and Gordon, 1997), the Campbell Range formation is now considered to have been deposited unconformably on them (Murphy and Piercey, 1999a). The basal basalt contacts are sharp, and a primary depositional relationship with underlying rocks is inferred based on: (1) the observation that the formation everywhere overlies older and commonly more deformed rocks; (2) a lack of evidence for faulting (e.g., brittle or ductile deformation) at or near the basal contact; and (3) the presence of euhedral Early Mississippian detrital zircons in a greywacke near the base of the formation, where it overlies the Money Creek formation south of the Jules Creek fault (Fig. 15; Table A2, Appendix 1 [see footnote 1]). In addition, spatially associated mafic and ultramafic plutonic rocks inferred to be comagmatic with Campbell Range basalt intrude underlying strata on both sides of the Jules Creek fault, as well as the basalts themselves.

Weakly metamorphosed basalt is the predominant rock type of the Campbell Range formation (Plint and Gordon, 1997; Murphy and Piercey, 1999a). Although locally foliated south of the Jules Creek fault, basalt is generally massive, green to brown weathering, and dark green to black on fresh surfaces. Pillowed flows and reddish
and greenish mono- and heterolithic breccias are locally common. Epidote-quartz alteration is a common feature, as is pink to tan jasperoidal silica. Ribbon chert and argillite and rare limestone are intercalated with basalt. Maroon, pink and green chert and siliceous argillite are locally sufficiently widespread to form separate map units. Dark grey or carbonaceous shale occurs locally. Campbell Range basalt hosts the Ice VMS deposit and numerous other copper occurrences (Figs. 3, 4; Hunt, 2002). Diabase, leucogabbro and variably serpentinized ultramafic plutonic rocks are spatially associated with basalt and locally cut it and all underlying rocks.

Early geochemical work on the Campbell Range basalt documented the essential seafloor basaltic character of the unit south of the Jules Creek fault (normal and enriched mid-ocean ridge basalt types - N-MORB and E-MORB; Plint and Gordon, 1997). More recent work both north and south of the Jules Creek fault has revealed a greater degree of compositional variability and, in particular, apparent geochemical and isotopic differences across the Jules Creek fault (Piercey et al., 1999; S.J. Piercey, D.C. Murphy and R.A. Creaser, unpublished data). All samples of Campbell Range basalt north of the fault can be classified geochemically as N-MORB and BABB types; south of the fault, samples are predominantly E-MORB with lesser N-MORB and OIB (Fig. 16). Basalt north of the fault has $\varepsilon_{Nd}$, $> +6$, whereas south of the fault, $\varepsilon_{Nd}$ is between -4 and +7 (Fig. 5).

The age of the Campbell Range formation is locally constrained by one mid-Pennsylvanian to Early Permian radiolarian collection from chert intercalated with basalt (T. Harms in Plint and Gordon, 1997; Table 5) and two ca. 274 Ma U-Pb dates on cross-cutting, but likely comagmatic leucogabbro (Table 6, Fig. 17, Appendix 1). Regionally, Asselian to Sakmarian radiolaria have been extracted from pink chert immediately beneath basalt of the Campbell Range formation located on strike to the west near Faro (F. Cordey in Pigage, 2001, 2004; Table 5).

**Environments of Deposition and Inferred Geodynamic Setting**

The lithological and stratigraphic characteristics of the Campbell Range formation suggest a marine volcanic buildup in an oxygenated basinal environment. The basinal background sedimentary rocks are pink and green radiolarian chert and shale, and basalt is locally hematized. Plint and Gordon (1997) attributed the presence of breccias to synvolcanic faulting. The common occurrence of both monolithic and heterolithic breccias suggests buildup and subsequent mass failure of topographic slopes, as well as autobrecciation that is characteristic of subaqueous basaltic volcanism (e.g., McPhie et al., 1993).

Geochemical data indicate that the most likely setting for Campbell Range magmatism was a back-arc basin (Piercey et al., this volume). A coeval succession with the geological and geochemical attributes of a volcanic arc has been documented in Yukon-Tanana terrane near the British Columbia-Yukon border south of the Tintina fault (ca. 281 Ma volcanic rocks of the Klinkit Group; Roots et al., this volume; Simard et al., 2003). The geographic distribution of arc and back-arc environments suggests that, as throughout the Carboniferous, subduction in the Early Permian was east- or northeast-dipping under the western or southwestern margin of Yukon-Tanana terrane.

Three considerations suggest that the part of the Early Permian back-arc region represented by the Campbell Range formation in the Finlayson Lake district was characterized by strike-slip faulting. First, the Campbell Range formation and subvolcanic intrusions only occur within a few kilometres on either side of the Jules Creek fault. Secondly, the Jules Creek fault juxtaposes different Early Permian and older lithological successions in different areas, but does not significantly vertically offset the Campbell Range formation. Thirdly, the difference in Nd isotopic character across the fault implies that it is a significant crustal boundary. These observations are best explained if the Jules Creek fault was initially a strike-slip fault that juxtaposed Yukon-Tanana terrane against the basinal rocks of the Fortin Creek group of Slide Mountain terrane during and after deposition of the Campbell Range formation and intrusion of the comag-

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**Table 6. U-Pb Zircon Geochronological Constraints, Campbell Range Formation.**

<table>
<thead>
<tr>
<th>#</th>
<th>Sample</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>$\Delta$</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>24</td>
<td>9DM-GB</td>
<td>leucogabbro</td>
<td>273.4</td>
<td>± 1.4</td>
<td>Intrudes basalt of Campbell Range formation; spatially associated with ultramafic rocks</td>
<td>this paper, Fig. 17</td>
</tr>
<tr>
<td>25</td>
<td>GL10</td>
<td>plagiogranite</td>
<td>274.3</td>
<td>± 0.5</td>
<td>Intrudes basalt of Campbell Range formation; spatially associated with ultramafic rocks</td>
<td>Mortensen (1992a)</td>
</tr>
</tbody>
</table>

*Number is keyed to data points presented in Figure 4.*
matic mafic and ultramafic plutonic rocks. Although the Jules Creek fault likely continued to be active into the Mesozoic (see below), it, the Campbell Range formation and comagmatic intrusions are interpreted here as manifestations of a ‘leaky’ transform fault formed initially during oblique transtension in an Early Permian back-arc region.

Late Early to Middle Permian Uplift and Shallow Water Sedimentation

In the Finlayson Lake district, massive limestone or dolomite and quartzite occur in the same corridor around the Jules Creek fault as the signature basalt, chert and mafic and ultramafic metaplutonic rocks of Slide Mountain terrane. South of the Jules Creek fault near Finlayson Lake, thick-bedded, locally crinoidal limestone and overlying poorly bedded, grey orthoquartzite of the Lower to Middle(?) Permian Gatehouse formation overlie the upper Pennsylvanian to Lower Permian Finlayson Creek limestone, apparently without any intervening Lower Permian Campbell Range formation (Fig. 2). Similarly, north of the Jules Creek fault, unfoliated sandy dolomite

Figure 16. Compositional variations in the Campbell Range formation, northern Finlayson Lake district (S.J. Piercey and R.A. Creaser, unpublished data).

Figure 17. U-Pb concordia diagram for a sample of leucogabbro intruding the Campbell Range formation. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
and quartzite overlie deformed chert and carbonaceous phyllite of the Fortin Creek group.

The Gatehouse formation is included within Slide Mountain terrane because the same association of massive carbonate and locally quartzite with basalt, chert and ultramafic rocks of Slide Mountain terrane occurs in two other localities along strike to the west. Near Faro, over 150 km to the west, massive Lower to Middle Permian limestone with large fusulinids is juxtaposed with Campbell Range-like basalt (Tempelman-Kluit, 1972; Table 5), and in a similar structural position further to the west in eastern Alaska, on the southwest side of the Tintina fault, giant fusulinid-bearing Middle Permian limestone and fossiliferous quartz sandstone are spatially associated with basalt, chert and serpentinized ultramafic rocks of the Seventymile terrane, correlative with Slide Mountain terrane (Dusel-Bacon and Harris, 2003).

In the Finlayson Lake district, the age of the Gatehouse formation is locally constrained by a Late Artinskian conodont collection found near the top of the limestone member south of the Jules Creek fault (Table 5). Dolomite in the lower part of the formation north of the Jules Creek fault contains Early Permian conodonts (Table 5). Large fusulinids and conodonts from the correlative limestone to the northwest near Faro are of Early to Middle Permian age (Table 5). Massive orthoquartzite in the upper part of the formation has not been found to be fossiliferous in the Finlayson Lake district; but it may correlate with quartz sandstone in a similar setting in eastern Alaska, which contains brachiopods of Permian age and lies in the same structural plane as limestone with Guadalupian (Middle Permian) fusulinid mega fauna (Dusel-Bacon and Harris, 2003; Foster, 1976).

**Environment of Deposition**

Although seemingly incongruous with respect to the association with ocean floor basalt, limestone and quartzite of the Gatehouse formation likely represent a shallow-water depositional setting. The absence of the Campbell Range formation beneath the Gatehouse formation near Finlayson Lake suggests that its base may be an erosional unconformity.

**MIDDLE PERMIAN TO UPPER TRIASSIC SUCCESSIONS**

Middle Permian and younger assemblages in the Finlayson Lake district either are deposited on, or derived from, both Yukon-Tanana and Slide Mountain terranes, attesting to initial juxtaposition of the two terranes along the Jules Creek fault prior to this time. The Middle Permian (to Triassic?) Simpson Lake group is deposited on the Fortin Creek group and the Campbell Range and Gatehouse formations. It is made up of clastic rocks containing detritus derived from both Yukon-Tanana and Slide Mountain terranes and is intercalated with Middle Permian felsic and mafic metavolcanic rocks (Mortensen et al., 1997, 1999). With the exception of coarse-grained clastic rocks of the Faro Peak formation near Faro, Upper Triassic successions deposited on the Slide Mountain terrane resemble the Middle to Upper Triassic Jones Lake Formation of the autochthonous North American continental margin sequence, indicating that the amalgamated allochthonous terranes may have been in close proximity to the North American margin in the Early Mesozoic. However, conodont fauna with Eurasian affinity (Orchard, this volume) occur in the allochthon, suggesting that its final assembly brought together rocks that may have been deposited at great distances from one another. The final emplacement onto autochthonous North America occurred during post-Late Triassic displacement on the Inconnu thrust.

**Late Middle to Late Permian Clastic Sedimentation and Magmatism**

In the Finlayson Lake district, the Middle to Late Permian interval is represented by the Simpson Lake group, a regionally extensive but narrowly distributed, polymictic conglomerate-bearing unit. Regionally, the Simpson Lake group has a similar distribution to that of the Lower Permian Campbell Range formation and affiliated mafic and ultramafic rocks, occurring within 10 km on either side of the Jules Creek fault or equivalent structures. The group comprises variable amounts of polymictic conglomerate and breccia; locally calcareous, detrital mica-bearing siltstone, sandstone and shale; massive, locally amygdaloidal basalt; limestone and felsic metavolcanic rocks. It unconformably overlies the Fortin Creek group, the Campbell Range and Gatehouse formations and either overlies or is intercalated with late Middle to Late Permian metavolcanic or subvolcanic intrusive rocks (Figs. 3, 4); many of the conglomerate clast types reflect derivation from these units. Clasts of foliated chert and carbonaceous phyllite were likely derived from the Fortin Creek group. Clasts of basalt and serpentinized ultramafic rocks were likely derived from the Campbell Range formation and affiliated ultramafic intrusions. Clasts of quartz- and feldspar-phryic metavolcanic rock could have been derived from essentially coeval Permian metavolcanic rocks or from nearby exposures of Carboniferous felsic metavolcanic rocks. Clasts of limestone could also be locally derived, from the Whitefish or Finlayson Creek limestones and/or the Gatehouse formation. The latter formation is potentially the source for quartzite clasts.

In exposures of the Simpson Lake group north of Watson Lake (Fig. 1), clast types also include greenstone, andesite and dacite, all with volcanic arc geochemical signatures; and blueschist and eclogite (Mortensen et al., 1997, 1999; J.K. Mortensen, unpublished data). Dacite, blueschist and eclogite clasts have yielded mid-Permian to Triassic K-Ar dates. Bedrock sources for these clasts do not crop out around Watson Lake, but they are found elsewhere in either Yukon-Tanana or Slide Mountain terrane. Permian volcanic rocks occur in the Finlayson Lake district, in Division 3 of the Sylvester allochthon in northern British Columbia (Nelson, 1993), and in western Yukon and easternmost Alaska (Klondike Schist, Mortensen, 1990; Dusel-Bacon et al., this volume). Eclogite and blueschist with Middle to Late Permian 40Ar/39Ar dates occur in several places along the northeastern margin of the amalgamated Yukon-Tanana/Slide Mountain terrane (Erdmer et al., 1998).

The best constraints on the age of the Simpson Lake group suggest that it is Middle to Late Permian, possibly ranging upward
into the Triassic. A maximum age of the group is provided by the late Early Permian age of the youngest underlying rocks and the mid-Permian age of dacite clasts (Mortensen, 1997, 1999; J.K. Mortensen, unpublished data). Spatially associated felsic metaigneous rocks in Finlayson Lake map area are ca. 259 Ma (Fig. 18, Table 7, Appendix 1; late Middle to early Late Permian; Okulitch, 2002). Conodonts extracted from samples from Frances Lake and Watson Lake map areas are Permian to Early Triassic in age (Table 8).

Environment of Deposition and Geodynamic Setting

The depositional environment of the Simpson Lake group is inferred to be an oxygenated marine basin into which locally derived massflows were deposited and volcanic rocks were erupted. The coarse-grained and polymictic nature of conglomerate in the Simpson Lake group suggests a structurally controlled synorogenic setting (Tempelman-Kluit, 1972, 1979), such as a foreland basin (Mortensen et al., 1997), forearc basin (Mortensen et al., 1999) or a strike-slip basin. A forearc setting is indicated by the presence of clasts of arc magmatic rocks, and blueschist and eclogite clasts from the coeval subduction zone (Mortensen et al., 1999). The narrow distribution of the conglomeratic rocks of the Simpson Lake group along the Jules Creek fault and equivalent structures would support a strike-slip setting, although under certain conditions conglomerates can be narrowly distributed near thrust faults in a foreland basin setting as well (Catuneanu et al., 1997). As the Lower Permian Campbell Range formation occurs at similar structural positions and elevations on either side of the Jules Creek fault, however, the fault is unlikely to have had much post-Early Permian vertical displacement.

Regionally, the Middle to Late Permian interval is represented by locally voluminous accumulations of volcanic arc rocks (Klondike Schist of western Yukon/easternmost Alaska, Mortensen, 1992a; Dusel-Bacon et al., this volume; and Division 3 of the Sylvester allochthon in northern British Columbia, Nelson, 1993). In addition, Middle to Late Permian ⁴⁰Ar/³⁹Ar white mica and U-Pb zircon dates have been obtained from bodies of eclogite and blueschist that are generally located on or near the northeastern edge of the amalgamated Yukon-Tanana/Slide Mountain terrane (Erdmer et al., 1998; Creaser et al., 1999). The regional distribution of Middle to Late Permian arc magmatism and high pressure/low temperature metamorphism (Fig. 1) implies subduction beneath the northeastern margin of the amalgamated Yukon-Tanana/Slide Mountain terrane (Mortensen, 1992a; Piercey et al., this volume). The locus of subduction under the terranes had therefore shifted from the southwestern margin of Yukon-Tanana terrane to the northeastern margin of the amalgamated terranes during the Middle Permian (Mortensen, 1992a).

Triassic Successions and Their Implications for the History of Amalgamation with North America

Rocks of Middle and Late Triassic age occur in four structural settings in or near the Finlayson Lake district. Weakly deformed and metamorphosed dark grey, silty, phyllitic shale; fine-grained, laminated and ripple cross-laminated, tan-brown weathering, detrital mica-bearing sandstone; and fetid, dark grey limestone of Triassic

Figure 18. U-Pb concordia diagrams for samples constraining the age of Middle to Late Permian igneous rocks in the Finlayson Lake district. (A, B) foliated quartz-feldspar porphyries of uncertain origin near the contact between the Campbell Range formation and the Simpson Lake group; (C) undeformed hornblende-phyric dike intruding the Money Creek formation. See Appendix 1 for analytical methods, data tables and geochronological interpretations.
Table 7. U-Pb Zircon Geochronological Constraints, Un-named Permian Magmatic Rocks.

<table>
<thead>
<tr>
<th>#</th>
<th>Sample#</th>
<th>Unit</th>
<th>Age (Ma)</th>
<th>±</th>
<th>Lithology / Relationship</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>26</td>
<td>01CR-140</td>
<td>ICE area volcanic or subvolcanic porphyry</td>
<td>259.8</td>
<td>± 1.2</td>
<td>Foliated quartz feldspar porphyry inferred to be depositional/intrusive on/into Campbell Range formation and overlain by Simpson Lake group</td>
<td>this paper, Fig. 18A</td>
</tr>
<tr>
<td>27</td>
<td>01DM-268</td>
<td>ICE area volcanic or subvolcanic porphyry</td>
<td>259.2</td>
<td>± 0.5</td>
<td>Foliated quartz feldspar porphyry inferred to be depositional/intrusive on/into Campbell Range formation and overlain by Simpson Lake group</td>
<td>this paper, Fig. 18B</td>
</tr>
<tr>
<td>28</td>
<td>00DM-160</td>
<td>Late dikes</td>
<td>254.7</td>
<td>± 0.6</td>
<td>Horruble discrete dike; intrudes Tuchitua River formation; similar undated dikes intrude Campbell Range formation</td>
<td>this paper, Fig. 18C</td>
</tr>
</tbody>
</table>

1 Number is keyed to data points presented in Figure 4.

age (Table 8) occur in a thrust duplex exposed in the Big Campbell window through Yukon-Tanana terrane (Figs. 2, 3). These rocks are inferred to depositional/overlie undated fragmental basalt, massive greenstone and serpentinitized ultramafic rock of Slide Mountain terrane. Secondly, lithologically similar rocks of Late Triassic age (early Norian, Table 8) overlie green-grey chert, carbonate phylite and quartzite of the Fortin Creek group immediately south of the Inconnu thrust fault. Thirdly, lithologically similar rocks of Middle to Late Triassic age overlie older rocks of the North American miogeocline north and east of the Inconnu thrust fault along much of its strike length throughout the northern Cordillera (Gordey and Makepeace, 2003; Table 8). Finally, Upper Triassic polymeric conglomerate of the Faro Peak formation (Pigage, 2001, 2004; Table 8) overlies eclogite and bluestone-bearing metamorphic rocks with Permian cooling ages south of the Vangorda fault near Faro, along strike to the northwest of the Finlayson Lake district.

With the exception of the Faro Peak formation, lithologically similar Middle and Upper Triassic clastic rocks occur on both the allochthonous rocks of the Inconnu thrust sheet and the North American autochthon. As such, these rocks may be inferred to be an overlap assemblage defining the first linkage between the autochthonous North American continental margin sequence and the allochthonous terranes (Gabrielse, 1991). In this case, the post-Late Triassic Inconnu thrust would therefore be a manifestation of shortening affecting all the assemblages following their pre-Late Triassic amalgamation. However, Orchard (this volume) reports Late Triassic conodont fauna from the allochthonous rocks in the Inconnu thrust sheet in the Finlayson Lake district that are Eurasian in affinity, unlike coeval fauna from the North American continental margin sequence and more like fauna reported from Cache Creek terrane and Wrangellia, both highly displaced tectonostratigraphic terranes in the western Cordillera. This discovery suggests that substantial displacement could have occurred within the amalgamated Slide Mountain and Yukon-Tanana terranes in the Triassic prior to their amalgamation with the continental margin in the Jurassic.

The immature, coarse-grained Faro Peak formation differs from the other Upper Triassic rocks in the region, which are typically made up of well-sorted, variably calcareous and fine-grained siliciclastic rocks. Occurring only near Faro along and south of the Vangorda Creek fault, the extension of the Jules Creek fault, the formation is isolated from all other exposures of Upper Triassic rocks and it is not possible to determine the relationship, if any, between them. It resembles the Middle to Upper Permian Simpson Lake group, both in its lithological character and its narrow distribution along the Jules Creek fault and, although there are few constraints, may be similarly interpreted as a manifestation of strike-slip faulting.

**TECTONIC EVOLUTION OF THE FINLAYSON LAKE DISTRICT**

The tectonic evolution of Yukon-Tanana terrane in the Finlayson Lake district that emerges from this and previous work is one of a
continental or transitional crustal fragment evolving through rapidly changing convergent margin geodynamic settings during the mid- and late Paleozoic. The distribution of Late Devonian and Early Mississippian arc and back-arc magmatic products is consistent with subduction beneath the western or southwestern margin of this crustal block (Mortensen, 1992a; Piercey et al., this volume). The short period of Devonian-Mississippian intra-back-arc deformation between the deposition of the Grass Lakes and Wolverine Lake groups may reflect shortening of the thermally softened overriding plate by synthetic thrusting (Colpron et al., 2000). Following the Early Mississippian cessation of arc and back-arc magmatism in this area, a mid-Mississippian hiatus, a Late Mississippian to Early Permian period of carbonate and chert deposition and volcanism, and Early Permian synorogenic flysch sedimentation, rocks from the forearc and arc were thrust over the Lower Permian flysch and underlying carbonate and inner arc and proximal back-arc rocks along the Cleaver Lake thrust. These latter rocks were in turn thrust to the east-northeast over the back-arc rocks along the Money Creek thrust. The cause for this short-lived episode of shortening is unknown; however, it and subsequent transform faulting which juxtaposed the imbricated Yukon-Tanana terrane with oceanic back-arc rocks of Slide Mountain terrane along the Jules Creek fault may be indications of an obliquely compressional environment in the Early Permian back-arc region, in contrast to the extensional Devonian-Mississippian back-arc.

At the same time as ensialic arc and back-arc rift environments existed on Yukon-Tanana terrane, Slide Mountain terrane was characterized by basinial chert and carbonaceous clastic sedimentation and rift volcanism on oceanic crust. An initial sedimentological and stratigraphic link with the Yukon-Tanana terrane back-arc region is inferred from the presence of Carboniferous quartzofeldspathic sandstone in the Fortin Creek group. A firm Early Permian link between the two terranes is indicated by the occurrence of oceanfloor basalt of the Campbell Range formation on both Yukon-Tanana and Slide Mountain terranes, deposited during juxtaposition of the two terranes by strike-slip along the Jules Creek and equivalent faults. These linkages between the terranes suggest a Carboniferous paleogeographic configuration similar to that of the Tertiary to modern Japan arc – Sea of Japan (Nelson, 1993; Creaser et al., 1999; Nelson et al., 2002, this volume). The Early Permian configuration differs somewhat, in the presence of significant strike-slip faulting in the back-arc region.

In the Middle to Late Permian, a forearc setting was established along the inboard (northeastern, present geographic coordinates) margin of the newly amalgamated Yukon-Tanana/Slide Mountain terrane. The switch to subduction along the inboard side of the terranes initiated the process of convergence between the amalgamated Yukon-Tanana/Slide Mountain terrane and North America by the near-consumption of the intervening Slide Mountain ocean basin (Nelson, 1993; Ferri, 1997; Nelson et al., this volume). Although the constraints are few, strike-slip faults may have continued to have been active in the forearc region of the allochthonous terranes, continuing possibly as late as the Late Triassic (Faro Peak Formation). Convergence with the North American continental margin culminated with the post-Late Triassic thrust displacement of the internally shuffled terranes and affiliated overlap assemblages onto the North American miogeocline along the Inconnu thrust fault.

Although the tectonic evolution of the Finlayson Lake district presented in this paper bears some resemblance to previous models, three aspects of the model are new and noteworthy. First of all, this work documents a metallogenically significant back-arc realm, elements of which occur in both Yukon-Tanana and Slide Mountain terranes. Secondly, this work identifies a type of geological relationship that is difficult to categorize using standard terrane terminology, i.e. the depositional onlap of Slide Mountain-type rocks of the Campbell Range formation onto Yukon-Tanana terrane, during their Early Permian juxtaposition along the Jules Creek fault. The Campbell Range formation is therefore both part of Slide Mountain terrane and an overlap assemblage, somewhat of a contradiction. Thirdly, this work highlights the importance of Paleozoic and early Mesozoic strike-slip faulting (e.g., Jules Creek fault) in the shuffling of original paleogeographic relationships – shuffling to the degree that two initially linked and contiguous geodynamic settings were differentiated into two “crustal block(s) or fragment(s) that preserve a distinctive geologic history that is different from the surrounding areas and that are usually bounded by faults” (modified from Coney et al., 1980), i.e. the Yukon-Tanana and Slide Mountain terranes. Furthermore, strike-slip faulting and associated magmatism provide a mechanism to explain the apparently contradictory geological relationships documented in the Finlayson Lake district, where the Campbell Range formation is both a defining part of Slide Mountain terrane, locally in fault contact with Yukon-Tanana terrane, and a stratigraphic overlap onto Yukon-Tanana terrane. Such relationships may actually be more common than has been documented in other areas where strike-slip faulting is the mechanism by which terranes were originally juxtaposed.

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FINLAYSON LAKE MASSIVE SULPHIDE DISTRICT
Finlayson Lake Massive sulphide district


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